LUNAR SCIENCE — III

REVISED ABSTRACTS OF PAPERS PRESENTED AT THE THIRD LUNAR SCIENCE CONFERENCE HOUSTON, 10-13 JANUARY 1972

EDITED BY CAROLYN WATKINS

LUNAR SCIENCE INSTITUTE CONTRIBUTION NO. 88

PREFACE

The Third Lunar Science Conference was held under the joint sponsorship of The Lunar Science Institute and NASA's Manned Spacecraft Center in Houston, 10-13 January 1972.

Prior to the meeting some 375 preliminary abstracts were printed for distribution to conference participants, with the provision that revised abstracts of up to three typed pages each could be submitted before the end of the conference. These updated "expanded abstracts" are collected here.

It is intended that these abstracts be referenced by conference participants as they prepare their manuscripts for the three-volume "Proceedings" — the collection of complete papers presented in Houston — to be issued in the late autumn of 1972.

The program was the responsibility of a committee composed of Dr. Paul W. Gast, Chairman (MSC), Dr. Peter M. Bell (Geophysical Laboratory, C.I.W.), Dr. Bruce R. Doe (U.S. Geological Survey), Professor Dieter Heymann (Rice University), and Dr. Thomas A. Mutch (U.S. Geological Survey).

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J. W. Chamberlain
Director
Lunar Science Institute

Houston
18 February 1972
Visible and near infrared diffuse reflection spectra are sensitive indicators of the average pyroxene composition of lunar rocks and soils. Wavelength positions of absorption bands arising from Fe$^{2+}$ change in a regular way from orthopyroxene to pigeonite to augite in response to changes in octahedral site symmetry. Orthopyroxenes and low-Ca pigeonites have two main bands, one near 0.90 μm and one near 1.90 μm. Augites show bands near 0.96 μm and 2.10 μm. Where several kinds of pyroxenes are present only an average spectrum is obtained because the individual bands are not resolved. Using mineral separates it has been shown that band positions in the lunar samples are governed by pyroxene only and are essentially unaffected by the presence of other minerals and of glass. Band depths, however, are strongly masked by the presence of dark glass.

Pyroxene bands in the Apollo 11 and 12 mare soils at 0.95 μm and 2.06 μm are at slightly shorter wavelengths than the bands for the mare breccias and crystalline rocks, suggesting that a foreign component of low-Ca pyroxene in the soil is shifting the average pyroxene composition. The foreign component is accounted for chemically and optically by the anorthositic and KREEP fragments, both of which contain low-Ca pyroxene.

Pyroxene bands for the Apollo 14 soils fall near 0.93 μm and 2.03 μm. These band positions indicate an average pyroxene composition in the soil that is less calcic than at the mare sites, in agreement with the chemical and modal data. In the Fra Mauro soils, however, the average pyroxene composition is significantly more calcic than the "parent" Apollo 14 breccias and basalts. It thus appears that the Fra Mauro soil has a foreign component of high-Ca pyroxene. The only known source of abundant high-Ca pyroxene is the mare regions, and it is suggested that mare material, including pyroxene, has been added as a contaminant, presumably by an impact-transport mechanism. Laboratory experiments show that addition of a few (<10) percent of mare material to the Fra Mauro breccias shifts the optical spectra to that of the Apollo 14 soil. The light-toned soil from Cone Crater has an average pyroxene composition intermediate between that of the breccias and the bulk Apollo 14 soil, suggesting that
CROSS-CONTAMINATION OF SOILS
John B. Adams

P.2

Cone Crater is young enough to be only slightly contaminated by mare material.

The increase in band depths in the reflection spectra with increasing crystal/glass ratio that was noted at the Apollo 11 and 12 sites (2,4) is further evidenced at Fra Mauro. On the basis of weak band structure in the telescopic spectrum it was predicted that the Apollo 14 soil would have a higher glass content than the Apollo 11 and 12 soils. The Fra Mauro soil samples confirm this prediction. In the laboratory, deep band structure in the spectral curve of the Cone Crater soil correlates with a very low (<10%) glass content.

The unusually low glass content of the Cone Crater soil provides further evidence that the degree of vitrification of a soil is an index of its maturity. If lunar soils are ranked in order or increasing glass content -- Cone Crater (<10%), Apollo 12 (20%), Apollo 11 (50%), and Apollo 14 (40-75%) -- it is seen that this ranking also correlates with increasing thickness of the regolith, and geologic evidence for increasing maturity of the surfaces.

Vitrification also affects the albedo of the soil, especially if opaque minerals are present. Regolith material consisting of lighter-toned silicates (plagioclase, pyroxene) and of opaques (ilmenite, metallic iron) has a "salt and pepper" appearance when crystalline, but turns uniformly dark when made into glass. A given material on the lunar surface thus becomes progressively darker with age through impact vitrification. The extent of darkening is determined primarily by the iron and titanium content of the glass and, therefore, by the amount of ilmenite, metallic iron, etc. in the parent rocks. The Fra Mauro soils, although more mature (glassy) than the mare soils, are not as dark because the parent Fra Mauro materials contain fewer opaques and have lower overall iron and titanium contents.

If the Cone Crater soil is representative of immature regolith at the Fra Mauro site it is of interest to consider whether vitrification alone would produce soil comparable in albedo to that in the surrounding area. The Cone Crater soil contains <5% opaques (including those in the lithic fragments). Laboratory experiments imply that even when completely vitrified the Cone Crater soil remains anomalously bright.

An outside source of dark glass and/or opaques appears to be required to darken the Fra Mauro regolith to the albedo of the typical surface soil. It is suggested that the source is the mare regions, and that contamination by mare material accounts for both a change in the average pyroxene composition and in the amount of dark glass in the highland soil. The presence of 11% mare glass in the Apollo 14 soil is reported by other investigators (3).

Telescopic observations imply that a glassy soil is well developed throughout the uplands, as it is at Fra Mauro, and that
CROSS-CONTAMINATION OF SOILS

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Bright craters and rays are, on average, more crystalline than the "background" areas (4). Cone Crater is too small to be measured spectrally by earth-based telescope (at present the area sensed telescopically is 18 km in diameter), but the laboratory spectral curve indicates a high crystal/glass ratio and the presence of low-Ca pyroxene. The large bright highland craters that have been observed telescopically are not as crystalline as Cone Crater, and are therefore probably older; furthermore highland craters give band positions that imply average pyroxene compositions comparable to mare material or to Fra Mauro soil. No evidence has been found for large exposures of fresh KREEP or anorthosite having dominant low-Ca pyroxene. If such rocks are exposed over large areas of the highlands their optical properties may have been obscured by a thin surface admixture of mare material.

Laboratory spectral reflectivity curves of soil samples from Apollo 11, 12, 14 and 15 sites were ratioed to each other, and the resulting curves were compared with telescopic ratio curves. McCord et al. (5) have shown that the highly accurate (~1%) telescopic ratio curves can be grouped into distinct types that are characteristic of maria and highlands and the young and old craters of each terrain. The Apollo laboratory ratio curves reproduce all of the spectral types observed telescopically, providing further evidence that the Apollo samples represent the lunar surface on a regional basis. The telescopic ratio curve for the Apollo 14 site is typical for the uplands in general. It is likely, therefore, that the contamination by mare material seen at the Fra Mauro site is regional in extent.

REFERENCES

6. We thank M. Charette for his assistance. This research was supported by NASA grants NGR 22-009-350, NGR 52-083-003. Contribution 52 of the Planetary Astronomy Laboratory, M.I.T.
PRELIMINARY RESULTS FROM THE S-161 X-RAY FLUORESCENCE EXPERIMENT


The X-Ray Fluorescence Spectrometer, carried in the SIM bay of the Command Service Module was employed principally for compositional mapping of the lunar surface while in lunar orbit, and secondarily for x-ray astronomical observations during trans earth coast. The lunar surface measurements involved observations of the intensity and characteristic energy distribution of the secondary or fluorescent x-rays produced by the interaction of solar x-rays with the lunar surface. The astronomical observations consisted of relatively long periods of measurements of x-rays from preselected galactic sources such as Cyg X-1 and Sco X-1 as well as from the galactic poles.

The x-ray experiment was turned on 84 hours into flight during the third lunar revolution. From 84 to 102 hours G.E.T. The orbit was approximately 8 X 60 nautical miles and during this period about 2 hours of undegraded x-data was obtained. After 102 hours the orbit was circularized and maintained at approximately 60 nautical miles until trans-earth coast. During the orbital period over 100 hours of surface measurements were made. A solar monitor detector was used for simultaneously monitoring the solar x-ray flux.

The results being reported here are based on quick-look data supplied in nearly real time. The data is in the form of running sums for the seven energy channels for each of the four experiment detectors (solar monitor, bare detector, detector with aluminum filter and detector with magnesium filter). The data were updated at one minute intervals. Our results as presented, are somewhat degraded in terms of spatial resolution along the projected ground track since each minute represents a three degree longitudinal displacement. Ultimately the processing of prime data, obtained for 8 second intervals should yield improved spatial information.

Examining the ground tracks in a direction from east to west the spacecraft is seen to have overflown such features as the craters Gargarin and Tsiolkovsky the backside and eastern limb highland, the mare area Smythii, Crisium, Fecunditatis, Tranquillitatis and Serenitatis, the Haemus Mts., the Apennines, Mare Imbrium and Oceanus Procellarium. Plots of Al/Si intensity ratios, determined from the quick-look data have been made along the various ground tracks.

One can make the following observations:

1. The Al/Si intensity ratios are lowest over the Mare areas and highest over the Terrae (average of 0.67 vs. 1.13). The extremes vary from 0.58 to 1.37, a factor of over 2.3.
PRELIMINARY RESULTS FROM S-161 X-RAY EXPERIMENT

I. Adler

2. The Apennine region gives a value of 0.88 and the Haemus Mts, a lower average of 0.83. On the other side of the Apennines (the Archimedes Rille area) the observed value is 0.64. On either side the values from Imbrium and Serenitatis are 0.59 and 0.58. The Apennines have values in between the mare and eastern limb highlands.

Plots were also made relating Al/Si intensity ratios to albedo values and gravitational profiles. Generally the higher Al/Si ratios correspond to higher albedo values. The deviations are due to identifiable surface features such as Copernican type craters. Thus it appears that one can use x-ray data to infer compositional information of areas with differing albedos.

With regard to the gravity profiles we observe the lowest Al/Si intensity ratios in the regions of greatest positive anomalies.

An effort has been made to arrive at the actual Al/Si concentrations using an assumed model for the solar x-ray flux and various lunar soil compositions (determined from the analysis of lunar samples) as ground truth. The results are shown in Table 1

<table>
<thead>
<tr>
<th>Table 1: Calculated Al/Si Concentration Ratios</th>
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</thead>
<tbody>
<tr>
<td>Mare Serenitatis</td>
</tr>
<tr>
<td>Mare Imbrium</td>
</tr>
<tr>
<td>Mare Crisium</td>
</tr>
<tr>
<td>Mare Tranquillitatis</td>
</tr>
<tr>
<td>Mare Fecunditatis</td>
</tr>
<tr>
<td>Mare Smythii</td>
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<tr>
<td>Archimedes Rilles Area (to Marsh of Decay)</td>
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<tr>
<td>Apennines Mts.</td>
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<tr>
<td>Haemus Mts.</td>
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<tr>
<td>Highlands E. of Serenitatis to 40 E.</td>
</tr>
<tr>
<td>Highlands W. and S. of Crisium</td>
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<tr>
<td>Highlands between Crisium and Smythii</td>
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<tr>
<td>Highlands between Smythii and Tsiolkovsky</td>
</tr>
<tr>
<td>Highlands E. of Fecunditatis</td>
</tr>
<tr>
<td>Highlands E. of Tsiolkovsky</td>
</tr>
<tr>
<td>Area N. of Schroter's Valley</td>
</tr>
<tr>
<td>Area N.E. of Schroter's Valley</td>
</tr>
</tbody>
</table>

Reference Materials
Apollo 11, Mare Tranquillitatis Bulk Soil 0.37
Surveyor, Mare Tranquillitatis Regolith 0.35
Apollo 12, Oceanus Procellarum Average of Soils 0.325
Luna 16, Mare Fecunditaitis Bulk Soil 0.416
Surveyor Vll, Rim of Tycho, Regolith 0.546
Apollo 14, Fra Mauro Soils 0.413
Apollo 15, Hadley Rille-Apennines, Soils (3 preliminary) 0.379
Some general conclusions can be drawn:

1. The Al/Si ratios confirm that the highlands and Maria do indeed have different chemical and mineralogical compositions.
2. The albedo difference between highlands and Maria are at least the expression of chemical difference.
3. The high aluminum/silicon concentration ratios of the highlands suggest they are related to the anorthositic or noritic fractions of the returned samples, while the low Al/Si concentration ratios of the Maria is consistent with the composition of the many Mare basalts analyzed.
4. The Imbrium ejecta blanket (Fra Mauro and Alpes Formations and possibly others) may be chemically different from the highlands outside such ejecta blankets. The intensity ratios suggest that they are intermediate in aluminum content between the Maria and the ejecta highlands.
THE OCCURRENCE OF GOETHITE IN A MICROBRECCIA FROM THE FRA
MAURO FORMATION

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Department of Mineralogy & Petrology, University of Cambridge

Microbreccia 14301, collected at station 61 some 150m W of the LM, is
composed of rock (dominantly micronorite) and mineral (dominantly
plagioclase) clasts, together with glassy fragments, spheres and lapilli
set in a finegrained (2-3μm) matrix of mineral fragments bonded by a film
of fawn coloured glass. Micron sized fragments in the matrix may be
devitrification products of more extensive glass. There is little sign
of the late crystallization characteristic of some microbreccias where
pyroxene, plagioclase ± sanidine develop in minute vugs, or where
pyroxene grows into the margin of plagioclase clasts, or metallic iron
has an interstitial appearance due to recrystallization of adjacent
silicate minerals.

In 14301/9, some twenty fragments of metallic iron occur in the
finegrained mineral and glass matrix. They are subrounded or angular in
shape and 25-150μm in size. Two contain visible troilite. All the
fragments are kamacite, the majority falling close to Fe 93.8%, Ni 6.0%
Co 0.2%, but a few with Fe>98%, Co<0.2%, Ni<0.2%. About half the
fragments are surrounded by a narrow halo where the matrix glass is
stained a redbrown colour and two (one with Fe>98%) show an intermittent
2-3μm wide overgrowth of, or replacement by, a grey reflecting phase
with the reddish internal reflections suggestive of goethite. The visual
relationships are not dissimilar to those between goethite and droplets
of metallic iron in some impactites from Meteor Crater, Arizona.

Stepped traverses across the boundary made with a 0.2μm beam of an
x-ray microanalysèr show an inflection at +57% Fe where "goethite" is
present, but no break where it is absent. Scanning pictures showed the
"goethite" to be rather inhomogeneous and an analysis based on data from
the electron probe and scanning electron microscope with non-dispersive
detector is quoted (Anal.3). The low total is assumed to be due to the
H₂O present. If the impurities are deducted and the analysis recalculated
to 100%, the Fe₂O₃ + NiO = 89.1% and H₂O+ = 10.9% obtained is close to
the theoretical value for goethite.

Recently, wider (+12μm) rims of goethite around a fragment of kamacite
containing troilite have been observed in 14301/19. Provisional data
confirm the initial observations and the corrected values will be presented at the meeting. It was noted that the position of the SKa peak
in the impure goethite corresponds to that of sulphur combined as
sulphate rather than sulphide. This suggests an oxidation reaction and
that a search might be made for a mineral of the Jarosite group.

The presence of minute amounts of water is suggested but not proved by two finds of amphibole previously recorded. It is further confirmed by the analyses quoted below (1,2) where $0.04\% \text{H}_2\text{O}$ is reported.

The small amount of water required to form this goethite could have come from an external or internal source. If the former, then the source is in the impacting body, a cometary or carbonaceous chondrite, a small amount of water being trapped during condensation and aggregation of the impact-produced cloud. Many authors have recognized a meteoritic component in lunar soils and microbreccias but the rarity of goethite in microbreccias and its sporadic distribution over the couple of square centimetres examined could be taken as not favouring this origin.

If the source is endogenous, several mechanisms can be proposed. The water may have been present in the original rock from which the impact ignimbrite was derived and was frozen into the glassy constituent, reacting with the iron during the subsequent cooling or thermal annealing. That the freezing in of the water is a feasible process is demonstrated by impactites derived from the Kaibab limestone at Meteor Crater, Arizona, where impactite glass has devitrified to amphibole of a similar composition (Anal.6, 7). It is also possible that the small amount of volatiles in an impact ignimbrite may be redistributed or concentrated in a manner similar to that observed in terrestrial ignimbrite sheets and that comparable recrystallization and vapour transport may have taken place in lunar analogues.

Once formed, goethite would be stable under thermal conditions at the lunar surface, provided it was sealed within the rock. If it was near enough to the surface to have access to the lunar atmosphere, the curves of Pollack et al. indicate that it should eventually break down to hematite and water vapour.

Finally, fumarolic or pneumatolytic processes may occur. Gas escape is evident in vesicular lunar basalts, so that if a basaltic intrusion were to occur within an impact ignimbrite sheet, both thermal recrystallization and vapour transport by gas escaping from the magma could affect the enclosing rocks. The metallic iron would act as a water scavenger and it is theoretically possible that the more susceptible silicates, e.g. olivine, could be involved in such reactions as the production of spinels etc or other products depending on the nature of the gas phase. However, except in the case of goethite formation, it would appear that the reactions involving pyroxenes and olivine have so far always antedated the formation of the breccias in which they are now found.

<table>
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<td>47.99</td>
<td>47.78</td>
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<td>6.71</td>
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<td>.81</td>
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<td>(1.36)</td>
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<td>P₂O₅</td>
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<td>&lt;0.1</td>
<td>0.30</td>
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<td>Cl</td>
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<td>0.65</td>
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<td>SΟ₃</td>
<td>n.d</td>
<td>n.d</td>
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<tr>
<td>S</td>
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<td>0.07</td>
<td>-</td>
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</table>

Total 100.51 100.32 89.4 92.01 100.00 100.44 100.08

1) 14003/33 Anal. J.H. Scoon
2) 14321/65 Anal. J.H. Scoon
3) Goethite from 14301/9. Low total interpreted as due to ± 10% H₂O⁺, (Probe analysis), Goethite (theoretical), Fe₂O₃ 89.86, H₂O⁺ 10.14.
4) Goethite from 14301/19 (Probe analysis) Low total interpreted as due to ± 8% H₂O⁺
5) 8.65% Jarosite, KFe₃(SO₄)₂(OH)₆ + 91.35% Goethite, FeO,OH.
6) Impactite glass, Meteor Crater, Arizona (Probe analysis)
7) Amphibole, devitrification product of Impactite glass. Meteor Crater, Arizona (Probe analysis)

* Determined by classical methods on 95% concentrate.
MINERALOGY,PETROLOGY,ANDCHEMISTRYOFLUNA16SAMPLEB-I
A. L. Albee, A. A. Chodos, A. J. Gancarz, E. L. Haines, D. A. Papanastassiou,
L. Ray, F. Tera, G. J. Wasserburg, and T. Wen, Lunatic Asylum, Division of
Geological and Planetary Sciences; California Institute of Technology,
Pasadena, California 91109. Contribution No. 2095

Luna 16 sample B-I was the largest fragment (62 mg) obtained in the
sample exchange with the USSR. Petrologic, mineralogic, and chemical
investigations have been made on this fragment in conjunction with isotopic
investigations. A Rb-Sr internal isochron and Ar$^{38}$/Ar$^{40}$ release patterns
indicate an age of about 3.45 aeons for this fragment. The exterior surface
of the equant fragment was coated with an irregular, thin, vitreous-appearing
glaze and contained numerous zap pits. The rock is a fine-grained ophitic
basalt composed predominantly of clinopyroxene (50%), plagioclase (40%),
and ilmenite (7%) with minor amounts of chromian ulvospinel, olivine cores in
pyroxene, troilite, mesostasis, apatite, and whitlockite. The texture
suggests that olivine and pyroxene crystallized first, followed by plagioclase
laths.

Plagioclase ranges from Or$_{4}$Ab$_{6}$An$_{9}$ to Or$_{2}$Ab$_{22}$An$_{76}$ and contains Fe,
Mg, K, and Ba concentrations similar to those found in other lunar samples
(Fig. 1). Two distinct compositional trends occur in pyroxene (Figs. 2, 3):
1) Ca-rich grains become less calcic with little change in Mg/Fe (e.g., Wo$_{35}$
En$_{41}$ Fs$_{24}$ grades into Wo$_{15}$ En$_{45}$ Fs$_{35}$) and 2) Ca-poor grains become more
Fe-rich (e.g. Wo$_{16}$ En$_{49}$ Fs$_{35}$ grades into Wo$_{30}$ En$_{55}$ Fs$_{75}$). No evidence exists for
the simultaneous crystallization of more than one pyroxene. The lines on
Fig. 3 are the altitudes of the triangles, which indicate the percentage of
components other than Wo, En, and Fs. The more calcic pyroxenes are highest
in Ti, Al and Cr; and the principal coupled substitution is R$^{2+}$Ti$^{4+}$ (Al$_2$O$_3$).

The bulk composition of sample B-I, as determined by defocused electron
beam analyses, is: SiO$_2$, 45.5 weight %; Al$_2$O$_3$, 13.5; TiO$_2$, 4.04; MgO, 5.95;
FeO, 17.8; MnO, 0.26; CaO, 12.0; BaO, 0.06; Na$_2$O, 0.63; K$_2$O, 0.2d; P$_2$O$_5$, 0.15;
total 100.48. Isotope dilution analyses give: Li, 10.7 ppm; K, 1396 ppm;
Rb, 1.58 ppm; Cs, 53.6 ppb; Ca, 8.25%; Sr, 437 ppm; Ba, 218 ppm. A uranium
content of 0.30 ± 0.03 ppm was determined by fission-track counting on a
1.2 mm$^2$ thin section. Compared to Apollo 11 basalts the K content is
intermediate to that of the low-K and high-K groups; its K/Rb is high, and it
contains more Si and Al and less Ti. Compared to Apollo 12 basalts sample B-I
contains similar concentrations of Ti and Si, more Al and K, and less Ca.
Compared to Apollo 14310 it contains less K, Si, and Al and more Ti and Fe.
The K/U value and the Sr content is much higher than in Apollo 11 and 12
basalts.

Isotope dilution analyses of soil sample A-2 give: Li, 8.5 ppm; K, 844
ppm; Rb, 1.79 ppm; Ca, 65.1 ppb; Ca, 8.25%; Sr, 271 ppm; Ba 168 ppm.
MINERALOGY, PETROLOGY, AND CHEMISTRY OF LUNA 16 SAMPLE B-1

A. L. Albee

Figure 1

Figure 2

Figure 3
1. \(^{40}\text{Ar-}{^{39}}\text{Ar}\) Dating: Gas retention ages have been measured using the \(^{40}\text{Ar-}{^{39}}\text{Ar}\) method for a suite of Apollo 12 samples and from an Apollo 15 sample. The ages (in units of \(10^9\) yr) are:

<table>
<thead>
<tr>
<th>Sample</th>
<th>This Work</th>
<th>Other Laboratories</th>
</tr>
</thead>
<tbody>
<tr>
<td>12002</td>
<td>3.29 ± .04 (± .06)</td>
<td>3.24 ± .05 [1]</td>
</tr>
<tr>
<td>12020</td>
<td>3.20 ± .03 (± .06)</td>
<td></td>
</tr>
<tr>
<td>12022</td>
<td>3.18 ± .04 (± .06)</td>
<td></td>
</tr>
<tr>
<td>12051</td>
<td>3.32 ± .04 (± .06)</td>
<td>3.27 ± .05 [1]</td>
</tr>
<tr>
<td>12065</td>
<td>3.23 ± .03 (± .06)</td>
<td>3.24 ± .05 [1]</td>
</tr>
<tr>
<td>15555</td>
<td>3.33 ± .05</td>
<td>3.30 ± .05 [2] 3.22 ± .03 [3]</td>
</tr>
</tbody>
</table>

Our ages are relative to an assumed monitor age of 4.56 ± .05 b.y. for the St. Severin meteorite and were calculated using a \(^{40}\text{K}\) mean life of \(1.825 \times 10^9\) yr [4]. The first error listed is the relative error for comparing one sample to another. The second error is the absolute error and is for inter-laboratory comparison. Although the ages agree well with the \(^{40}\text{Ar-}{^{39}}\text{Ar}\) ages obtained by other workers, they are not directly comparable because each laboratory is using different \(^{40}\text{K}\) decay constants. For instance, for comparison to the ages obtained by Turner [1] our results should be systematically reduced by 0.03 b.y.

The Apollo 12 results support the observation [5] that small but real age differences exist within the Apollo 12 samples. The age of 15555, a basalt from the rim of Hadley Rille, corresponds to the older ages obtained for the Apollo 12 basalts. If 15555 is representative of the basalts filling Imbrium and if the Fra Mauro ages from Apollo 14 [6] date the formation of the Imbrium basin, then the basalts filling the basin postdate its formation by at least 500 m.y. Therefore, the basalts filling Imbrium cannot be simple impact melts but must represent some lunar igneous activity which occurred long after the basin was formed.

2. Trace Element Determinations: In conjunction with our \(^{40}\text{Ar-}{^{39}}\text{Ar}\) dating program we analyze the krypton and xenon in the neutron irradiated lunar samples. Elemental abundances can be determined for those elements (Se, Br, I, Te-Ba, U) which produce isotopes of krypton and xenon in the reactor. We have previously shown that the retention of gaseous neutron capture products by the
samples varies widely for different elements. For instance, $^{128}$Xe from $^{127}$I is very loosely bound while $^{131}$Xe from $^{130}$Ba is tightly bound. The Apollo 12 and 14 samples were irradiated in evacuated quartz "break seal" capsules. We were able to measure all of the neutron produced isotopes including the low temperature fractions which had been lost in previous experiments. In addition to trace element determinations, we also were able to check for any $^{40}$Ar or $^{39}$Ar which might have been lost from the samples during the irradiation.

The following preliminary data have been obtained for Apollo 12 samples:

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>12002</td>
<td>80</td>
<td>96 ± 23</td>
<td>0.08 ± .02</td>
<td>52.0 ± 7.8</td>
<td>0.8 ± 0.3</td>
</tr>
<tr>
<td></td>
<td>350</td>
<td>20.2 ± 3.6</td>
<td>&lt;.03</td>
<td>1.9 ± 0.3</td>
<td>&lt;1.0</td>
</tr>
<tr>
<td></td>
<td>1730</td>
<td>11.7 ± 1.9</td>
<td>47.6 ± 5.7</td>
<td>0.6 ± 0.1</td>
<td>174 ± 21</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>127 ± 23</td>
<td>47.8 ± 5.7</td>
<td>54.5 ± 7.8</td>
<td>176 ± 21</td>
</tr>
<tr>
<td>12020</td>
<td>80</td>
<td>111 ± 20</td>
<td>0.18 ± .02</td>
<td>30.1 ± 4.3</td>
<td>&lt;1.6</td>
</tr>
<tr>
<td></td>
<td>350</td>
<td>5.1 ± 1.1</td>
<td>&lt;.04</td>
<td>0.8 ± 0.2</td>
<td>&lt;0.6</td>
</tr>
<tr>
<td></td>
<td>1730</td>
<td>5.7 ± 0.9</td>
<td>45.4 ± 5.4</td>
<td>0.4 ± 0.2</td>
<td>168 ± 21</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>121 ± 20</td>
<td>45.7 ± 5.4</td>
<td>31.3 ± 4.3</td>
<td>170 ± 21</td>
</tr>
<tr>
<td>12022</td>
<td>80</td>
<td>35 ± 17</td>
<td>0.17 ± .02</td>
<td>69.0 ± 9.4</td>
<td>&lt;0.9</td>
</tr>
<tr>
<td></td>
<td>350</td>
<td>6.1 ± 1.8</td>
<td>&lt;.02</td>
<td>3.2 ± 0.4</td>
<td>&lt;4.0</td>
</tr>
<tr>
<td></td>
<td>1730</td>
<td>7.6 ± 1.2</td>
<td>57.6 ± 6.9</td>
<td>0.4 ± 0.2</td>
<td>204 ± 25</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>48 ± 17</td>
<td>58.0 ± 6.9</td>
<td>72.6 ± 9.4</td>
<td>209 ± 26</td>
</tr>
<tr>
<td>12051</td>
<td>80</td>
<td>104 ± 24</td>
<td>.08 ± .02</td>
<td>14.6 ± 5.1</td>
<td>0.3 ± 0.1</td>
</tr>
<tr>
<td></td>
<td>350</td>
<td>27.6 ± 4.7</td>
<td>.03 ± .02</td>
<td>1.0 ± 0.3</td>
<td>&lt;1.5</td>
</tr>
<tr>
<td></td>
<td>1730</td>
<td>14.0 ± 2.2</td>
<td>69.5 ± 8.3</td>
<td>0.7 ± 0.1</td>
<td>245 ± 30</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>146 ± 25</td>
<td>69.6 ± 8.3</td>
<td>16.3 ± 5.1</td>
<td>247 ± 30</td>
</tr>
<tr>
<td>12065</td>
<td>80</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td></td>
<td>350</td>
<td>&lt;8.7 ± 1.8</td>
<td>&lt;.03</td>
<td>&lt;0.4 ± 0.2</td>
<td>&lt;1.2</td>
</tr>
<tr>
<td></td>
<td>1730</td>
<td>8.3 ± 1.3</td>
<td>51.1 ± 6.1</td>
<td>0.5 ± 0.1</td>
<td>201 ± 24</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>*</td>
<td>51.2 ± 6.1</td>
<td>*</td>
<td>203 ± 24</td>
</tr>
</tbody>
</table>

The 80° fraction of 12065 was lost because the "break seal" apparently was not gas tight. Some of the 350° fraction may have been lost also.

The Ba and U concentrations agree well with the values published by several other groups. The I and Br values compare well with those of Reed and Jovanović [7] for the one rock where comparison is possible. Our bromine values are systematically high than those obtained by Anders et al. [8].
Rare Gas Analyses
E. C. Alexander, Jr.

The trace elements correlate with the $^{40}$Ar-$^{39}$Ar ages. The oldest rock 12051 has the highest Br, Ba, and U contents and the lowest I content. The youngest rock, 12022, has the lowest Br content and the highest I content. The older basalts at the Apollo 12 site were enriched in the refractory elements Ba and U and depleted in the volatile element I relative to the younger basalts. Br correlates with Ba and U rather than I.

Loss of $^{40}$Ar was not significant for the Apollo 12 samples. No useful information was obtained on Se or Te concentrations.

References

The primordial isotope of Pb, $^{204}\text{Pb}$, has been measured in several Apollo 14 samples by fast neutron activation analysis. This method avoids the uncertainties caused by large blank corrections necessary in other procedures. The lability of $^{204}\text{Pb}$ was also studied by its leachability in aqueous solution. Such a measurement is of interest because lead is one of the more readily volatilized elements; it is associated with volcanic emanations as PbCl$_2$. The existence of an apparently parentless, more labile, Pb component in lunar soil and breccia has been noted by other investigators (1). A primordial labile lead possibly associated with this component was sought in this work.

In previous measurements of Pb by the activation technique, reactor irradiations were utilized; in this work, fast neutrons produced by $^9\text{Be}(d,n)^{10}\text{B}$ reactions in the ANL 60-inch cyclotron were used as well. The $^{206}\text{Pb}(n,\alpha)^{203}\text{Hg}$ reaction makes the measurement of $^{208}\text{Pb}$ feasible. Short reactor irradiations following the cyclotron irradiation permit $^{208}\text{Pb}$ measurements.

$^{204}\text{Pb}$ was measured in the "clod" 14049,35 and the soils 14163,152 and 14259,119. Sieved fractions of each of the two soils were analyzed with the assumption that vapor deposition after the formation of the soil would show a dependence upon the surface area of the grains. A 620mg piece of the fragmental rock 14321,185 broken from a larger sample was taken with care to insure that there were no previously exposed surfaces. After crushing a 470mg sample was analyzed as a "representative" sample of the mixture of medium gray matrix and the dark fragmental clasts. After irradiation all samples were leached for 10-15 minutes with a hot aqueous solution at pH 5-6. In a few cases Bi and $^{208}\text{Pb}$ were measured. Total $^{204}\text{Pb}$ concentrations (Table 1) range from 5-8 ppb in the soils and in sample 14049,35. Even under the mild leaching conditions employed 10-30% of the $^{204}\text{Pb}$ was soluble. This amount of mobile Pb is comparable to that resulting from strong acid leaching and low temperature (500-700°C) volatilization as reported by other investigators (2) despite the great difference in the leaching conditions. The sum of the sieved fractions from the soils yield nearly the same $^{204}\text{Pb}$ concentrations as the bulk
Table 1. Lead and Bismuth in Apollo 14 Samples

<table>
<thead>
<tr>
<th>Sample**</th>
<th>Residue (ppb)</th>
<th>Leach (ppb)</th>
<th>Residue (ppb)</th>
<th>Leach (ppb)</th>
</tr>
</thead>
<tbody>
<tr>
<td>14163,152</td>
<td>5.6±1.7</td>
<td>1.2±1.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14163,152&gt;150μm</td>
<td>2.3±0.1</td>
<td>0.05±0.08</td>
<td>10±1</td>
<td>2.8±0.5</td>
</tr>
<tr>
<td>14163,152&lt;150μm</td>
<td>8.2</td>
<td>2.1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14259,119</td>
<td>4.4±0.5</td>
<td>0.5±0.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14259,119</td>
<td>5.1±0.6</td>
<td>0.4±0.2</td>
<td>1.7</td>
<td>0.4</td>
</tr>
<tr>
<td>14259(119,112) &gt;150μm</td>
<td>7.0±1.4</td>
<td>4.1±1.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14259(119,112) &lt;150μm</td>
<td>5.8±0.7</td>
<td>1.1±0.2</td>
<td>6.0</td>
<td>1.6</td>
</tr>
<tr>
<td>14049,35</td>
<td>4.7±0.8</td>
<td>1.3±0.2</td>
<td>1.1</td>
<td>4.8</td>
</tr>
<tr>
<td>14321,185</td>
<td>1.1±0.2</td>
<td>0.45±0.23</td>
<td>0.9</td>
<td>0.6</td>
</tr>
</tbody>
</table>

**Sample weights were 0.3 - 1.0 gm.

samples. However the 14163,152 fine fraction contained more Pb in the residue and almost all of the leachable 204Pb whereas the 14259,119 coarse fraction contained the higher concentration in both the residue and the leach. Thus no grain size correlation can be made. This indicates that a single vapor deposition process probably cannot account for the labile 204Pb. The finer and coarser fractions appear to have had different histories, acquiring different amounts of nonleachable and leachable Pb, but ending up with nearly the same total amount of Pb.

The clod 14049,35, containing 6 ppb 204Pb of which 1.3 ppb is leachable, is indistinguishable from the soils. The rock 14231,185 has lower 204Pb, 1.55 ppb, than any other measured. Interestingly, despite the much shorter exposure of the surface of this rock, it contains about the same amount of leachable 204Pb as found in the...
other samples suggesting that the leachable $^{204}$Pb is not due to terrestrial contamination. Since this rock has not been reconstituted since formation (as soil may have been) and since vapor deposition (above) does not necessarily explain the labile Pb, it seems that this Pb may have been acquired at the time, or before, the fragments were consolidated into rock 14321,185.

The residual and leachable Bi contents in the 14321,185, 14049,35 and 14259,119 are given in the Table. The trends in $^{204}$Pb are duplicated in the Bi. Even more Bi is leachable, 4.8 vs. 1.1 ppb, in the 14049 sample. The residual and leachable $^{208}$Pb contents in 14163,152 and 14259,112 are 2 and 1 ppm and 1.3 and 0.5 ppm, respectively. The leachability of 50-40% of the $^{208}$Pb is contrasted with the unleachable nature of the U which is less than 1% leachable under the same conditions (3). The aqueous leachability of Bi as well as that of primordial and radiogenic Pb suggests that volatile transport is an important process in the distribution of some elements on the moon.

References


3. G. W. Reed, Jr., personal communication.

4. We thank the crews at the ANL 60 in. Cyclotron and the BNL High Flux Beam Reactor for irradiations. This work was supported by NASA Contract T-76356 and the USAEC.

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SECONDARY ELECTRON EMISSION CHARACTERISTICS OF LUNAR SURFACE FINES.

Calculations to date have shown that the steady-state electrostatic potential and charge distribution of the sunlit lunar surface is determined primarily by the photoemissive properties of the lunar surface material and the impacting solar wind flux. Only a small fraction of the emitted photoelectrons can escape, which implies that the sunlit lunar surface attains a small positive electrostatic potential, the magnitude of which is determined by the energy distribution of the photoemitted electrons and the solar wind particle velocity distribution and density. Secondary electron emission due to the absorption of solar wind particles has been assumed to be negligible in view of the low energies of solar wind electrons (<10 eV) and protons (~1 keV). However, recent measurements by the charged particle lunar environment experiment (CLEE) have shown that, in addition to a stable low-energy photoelectron flux, observed for energies ranging from 200 eV down to 35 eV, rapidly varying fluxes of low energy electrons of magnetospheric origin occur, with intensities greater than that of the photoelectron background. When the lunar surface is illuminated, the electron spectrum between 40 and 100 eV is dominated by the photoelectron continuum, but in the higher energy ranges, prominent peaks in the electron flux density were observed in the ranges 300 to 500 eV and at several keV respectively. These higher energy electrons will give rise to secondary electron emission at the lunar surface, which in turn, will affect the lunar surface charge and potential distribution depending on the secondary emission characteristics of the lunar material.

We have therefore measured the secondary electron yield and energy distribution of lunar dust sample 14259.116 over a primary electron energy range of 50 to 2000 eV in order to obtain a more complete description of the low energy electron layer and the electrostatic potential of the sunlit lunar surface. Detailed Auger electron analysis has, in addition, provided information concerning the elemental composition of the sample. The samples were handled in dry nitrogen atmosphere and the secondary electron yield and energy distribution measurements were made using an hemispherical collector system as described previously. Bakeout of the ultra-high vacuum system and sample was carried out at 150°C for one week in order to simulate lunar thermal conditions and to remove adsorbed nitrogen. Preliminary results for the total secondary electron yield and energy distribution curves for primary incident electron energies in the above range are shown in Fig. 1 and Fig. 2 respectively. The secondary electron yield curve, figure 1, shows a maximum of 1.4 at approximately 300-400 eV, which is somewhat lower than that observed for insulating materials such as SiO₂, Al₂O₃, CaO, FeO and MgO, the major constituents of the sample. This is expected in view of the particulate nature of the sample from which the secondaries cannot escape due to scattering within microscopic cavities. Also no elastically reflected electrons were detected,
SECONDARY ELECTRON EMISSION CHARACTERISTICS OF LUNAR SURFACE FINES.

M. Anderegg

Fig. 1 Secondary electron yield from lunar sample 14259.116 for incident primary energies of 50 to 2000 eV.

![Graph showing secondary electron yield vs. primary beam energy]

Even at primary beam energies below 50 eV which suggests that the primary electrons are either completely absorbed or, what is more likely, are only reflected back along the direction of the incident beam. This behaviour is very similar to that observed for the lunar albedo variation with phase angle being maximum at zero phase angle.10

Fig. 2 Energy distribution curves of secondary electrons emitted from lunar sample 14259.116 for various primary energies. Note the reduced size of the low energy secondary peak and the distorted spectra compared with the curve at 1750 eV primary energy due to charging of the sample.

![Graph showing energy distribution curves]
SECONDARY ELECTRON EMISSION CHARACTERISTICS OF LUNAR SURFACE FINES.
M. Anderegg

Values of the yield greater than 1 cause the dust to charge positively, i.e., for incident electron flux energies in the range 100-200 eV to 1500-2000 eV. The effect of this charging can be seen in the reduced size of the low energy secondary peak and the distortion of the energy distribution spectra for primary energies in this range, fig. 2. The 'true' secondary electron energy distribution is that shown for a primary energy of 1750 eV, for which the yield is unity. The curve is typical of that observed for insulating materials, i.e., the greater number of secondaries occur with energies less than about 4 or 5 eV. Preliminary measurements indicate the maximum positive charge on the specimen to be less than 20 eV. It is significant that the increased electron flux density observed by the CPLEE at energies of 300 to 500 eV when the moon passes through the magnetospheric tail, is also the energy at which the lunar dust sample possesses maximum secondary yield, which implies that such events will cause the lunar surface to charge even more positive and so modify both the photoelectron and secondary electron energy distributions. Incident electron energies below 100 eV, for which the secondary yield will be negligible, and above 2,000 eV, at which increased electron flux density has also been observed, will cause the surface to charge negatively.

These preliminary results therefore, indicate that the lunar surface charge and potential will depend not only on the solar wind flux and the photoemissive properties of the lunar material, but also on the energy distribution of the magnetospheric electrons and secondary electron emission of the lunar surface. Future calculations will be based on such considerations and the data presented here.

References:
3. H. Heffner, Rept. TE-7 of the 'Tycho' Study Group, University of Minnesota, Minneapolis (1965).
U, Th, Pb and REE abundances and Pb 207/206 ages of individual minerals in returned lunar material by ion microprobe mass analysis

C.A. Andersen & J.R. Hinthorne, Hasler Research Center, Applied Research Laboratories, Goleta, CA 93017

Ion microprobe analyses of Apollo 11, 12, and 14 material have shown that U, Th, Pb and REE are concentrated in accessory minerals such as apatite, whitlockite, zircon, baddelyite, zirkelite and tranquillityite. Th/U ratios are found to vary by over a factor of 40 in these minerals. K, Ba, Rb and Sr have been localized in a K rich, U and Th poor glass phase that is commonly associated with the U and Th bearing accessory minerals. Li is observed to be fairly evenly distributed between the various accessory phases. The phosphates have been found to have REE abundance patterns (normalized to the chondrite abundances) that are fairly flat, while the Zr bearing minerals have patterns that rise steeply, by factors of ten or more, from La to Gd. All the accessory minerals have large negative Eu anomalies. Radiometric age dates (Pb 207/206) of the individual U and Th bearing minerals within the fine soil material compare favorably with the Pb 207/206 age of the bulk crystalline rocks. The Th/U ratios and Pb isotopic data and the light REE abundance distributions that have been measured in the accessory mineral phases analyzed in this work are summarized in Table 1 and Figure 1, respectively. The ages given are the Pb 207/206 ages. The Pb isotopic data generally have an accuracy of about 5% and the Th/U ratios an accuracy of about 10%. The data will be discussed in detail.

* Ion Microprobe Mass Analyzer, Applied Research Laboratories, Sunland, California
### TABLE 1

Th AND U BEARING MINOR MINERAL PHASES
IN APOLLO 11, 12 AND 14 LUNAR MATERIAL

<table>
<thead>
<tr>
<th>Sample</th>
<th>Min. Phase</th>
<th>Th/U</th>
<th>Pb(^{208}/206)</th>
<th>Pb(^{207}/206)</th>
<th>Age (b.v.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>10085/7-6</td>
<td>Ti, Zr Silicate</td>
<td>2.16</td>
<td>0.59</td>
<td>0.40</td>
<td>3.97</td>
</tr>
<tr>
<td>10085/17-26</td>
<td>Zr O(_2)</td>
<td>0.49</td>
<td>0.133</td>
<td>0.455</td>
<td>4.18</td>
</tr>
<tr>
<td>12032, 46-2</td>
<td>Zircon</td>
<td>0.79</td>
<td>0.294</td>
<td>0.408</td>
<td>4.01</td>
</tr>
<tr>
<td>12013, 14</td>
<td>Zircon</td>
<td>0.48</td>
<td>0.144</td>
<td>0.483</td>
<td>4.25</td>
</tr>
<tr>
<td>12013, 14</td>
<td>Zr Phase in K Vein</td>
<td>1.22</td>
<td>0.381</td>
<td>0.414</td>
<td>4.03</td>
</tr>
<tr>
<td>12013, 14</td>
<td>Whitlockite</td>
<td>20.4</td>
<td>3.64</td>
<td>0.49</td>
<td>4.26</td>
</tr>
<tr>
<td>14257, 3-109B</td>
<td>Zirkelite</td>
<td>1.55</td>
<td>0.383</td>
<td>0.412</td>
<td>4.03</td>
</tr>
<tr>
<td>14257, 3-109B</td>
<td>Zirkelite</td>
<td>1.73</td>
<td>0.374</td>
<td>0.425</td>
<td>4.07</td>
</tr>
<tr>
<td>14163, 86-A4</td>
<td>Zr O(_2)</td>
<td>0.34</td>
<td>0.406</td>
<td></td>
<td>4.00</td>
</tr>
</tbody>
</table>
Table 1 lists the number and total areas of different types of fragments (> 0.1mm) in 14306, 14321, 14270 thin sections. Dark polymict basaltic metabreccia predominates. Among monomict fragments, light-basaltic material (i.e. noritic) predominates over anorthosite + large plagioclase crystals, with minor granitic + rhyolitic, feldspathic basalt and others. These rock types and proportions are generally similar to the 1-2mm rock fragments (Steele and Smith) from 14002, 14258.

The textures and compositions of rock and mineral fragments in 14306 imply various thermal histories with possible implications for models of thermal accretion. Breccia 14306 is polymict with a poorly-sorted, light-gray matrix that is non-porous (<5% porosity). The matrix is probably holocrystalline but distinction between fragments and matrix is arbitrary. Many tiny and some large fragments are partly glass. Fractures cut equally across fragments and matrix indicating equal coherency; however glass does not appear to pervade and bind the matrix. Hence the matrix probably cohered below the temperature of partial melting (900 ± 100°C).

Dark-gray polymict metabreccias have a non-porous matrix comprising finely-crystalline (<50μm) plagioclase and pyroxene with minor ilmenite, metal, sulfide, zircon and phosphates. Noteworthy fragments include light-gray, partly-glassy, basaltic metabreccias; pyroxenes with 5μm exsolution; partly glassy granite; deformed glassy rhyolite and irregular grains of metal-sulfide. These metabreccias cooled quickly (mins. to hrs.) from an initial temperature near 900°C as indicated by deformation of rhyolite and lack of melting in sulfide-metal grains. Marginal sodium enrichment of rhyolite suggests presence of vapor during compaction and initial cooling.

Grains with concentric shells of texturally-distinct material are rare in the darker-gray metabreccias, and contain a central kernel of rhyolite or granophyre, or rarely a light-gray metabreccia. Probably stickiness of nearly molten rhyolite and partly-liquid metabreccia resulted in attachment of solid matrix during transport or in cohesion of adhering matrix immediately upon deposition.

Round fragments of crystals and rocks occur sporadically in the darker-gray metabreccias. Evidence of melting is absent, and rounding presumably occurred by abrasion in a turbulent dust jet.

The coarser texture and mineral compositions of the light-gray, partly-glassy basaltic metabreccias suggests a higher temperature of recrystallization than attained in the matrix of darker-gray metabreccia. The Mg-distribution between ilmenite and pyroxene in the basaltic metabreccias suggests by analogy with terrestrial specimens a temperature near 1000 ± 100°C. Possibly the light-gray basaltic metabreccias were
were undergoing metamorphism near 1000°C immediately prior to disruption and incorporation into their present matrix. This is consistent with the hypothesis of a hot outer Moon at the time of formation of the Imbrium basin, as previously suggested by Opik (1969) on the basis of geometrical relations of the impact features.

Coordination of these mineralogic and petrologic data with geological maps of the Imbrium and Apollo 14 regions, together with extrapolations from terrestrial and laboratory data, suggests that these Apollo 14 breccias derived from the upper 5 to 10Kms of the pre-Imbrium "crust" during impact, and were transported in a hot, turbulent jet (~900°C) containing dust and vapor. The depth estimate assumes that the Apollo 14 material derives mainly from the initial impact with high-velocity jetting between the circumference of the incoming body and the disintegrating crust.

We attempt now to face the problem of how much material derives from the impacting body, and whether such material can be identified. We tentatively accept various arguments for the body impacting at low velocity (2-4Kms/sec) consistent with capture from Earth orbit. We expect that several percent of Apollo 14 breccias derive from the incoming body. They should result predominantly from the impacting surface and not represent the mean body if it is chemically zoned. The Apollo 14 breccias yield no positive evidence for the incoming body being an iron meteorite, and we consider possible stony meteorites. Rounded bodies are present, but are relatively rare and could be accounted for by (a) formation of droplets with later recrystallization, and (b) abrasion of existing crystals and rock fragments. We see no evidence for hydrous minerals in the breccias, or for thermal breakdown products, but recognize that both impact and thermal metamorphism might destroy any evidence. Rare fragments of dunite (olivine near Fo 90) might result from deep-seated rocks in the Moon, but could be ascribed to the incoming body. We see no evidence that conflicts seriously with existing ideas that the impacting body was a moonlet of Earth, with a composition not far removed from that of the Moon. If indeed the impacting body has a similar composition to the Moon, distinction between materials from the Moon's surface and the Imbrium moonlet may require very subtle techniques.

Turning now to the pre-Imbrium "crust", we envisage further problems. Let us assume that the dominant material in the Apollo 14 breccias is indeed from this source. From stratigraphic data we can anticipate that the pre-Imbrium surface was complex, containing debris from earlier impacts. However such material should have similar composition to pre-Imbrium "bed-rock" if the Moon is only moderately inhomogeneous laterally. Assuming that the dominant populations in the breccias actually derive from the pre-Imbrium "crust", we can use the chemical data of Steele and Smith for Im fragment since they resemble breccia populations. These suggest that the upper 5-10Kms of pre-Imbrium crust are dominated by light-colored basalts (i.e. noritic), dark-colored basalts, and feldspar-rich rocks (anorthosites and rhyolites). The chemical relations are consistent with crystal fractionation together with some liquid immiscibility, as shown or implied by Apollo 11 and 12 data of many workers. The high proportion of noritic material suggests that this is a major differentiate of the Moon implying extensive melting to a depth depending on one's estimate of the Imbrium
impact mechanics but not inconsistent with complete melting with formation of
a metal-sulfide core. Anorthosites could derive by cumulation of plagioclase
and would be expected near the surface. Cumulates of pyroxene and olivine
would be expected at greater depths, and be sampled less efficiently by the
Apollo 14 breccias: small amounts of pyroxene- and olivine-rich rocks do
indeed occur and have characteristics of deep seated crystallization.

Thus we reaffirm the viability as a working model for the Moon of the
concept of extensive early melting forming a crust composed of feldspar-rich
rocks and basalts: various data (e.g. Rb/Sr evolution) indicate that this
was an early process (~4.5 b.y.). The subsequent roles of residual liquid,
of remelting, and of impact-generated melt are unclear to us, but we
envisage extensive permeation and disruption of the early crust plus a long
history of cooling and volcanism in debris produced in mare basins.

The breccia populations are consistent with mare-type basalts forming
after the noritic material, as expected from evidence that the former enter
basins excavated in the former. However the origin and depth of formation
of the mare basalts remains controversial because of conflict in the
interpretation of crystallization sequences and chemical contents of minerals,
and of uncertainty in heat sources (radioactivity, tidal heating, impact
heating). Furthermore the mare basalts of Apollo 11 and 12 may represent
thin surface flows unrepresentative of the deeper mare fill. The extent
and nature of later melting episodes on the Moon remains equivocal, but
evidence for an early major episode seems good from the Apollo 14 breccias
and other information.

We thank Ian Steele and NASA grant 14-001-171.

**Table 1** Preliminary modal populations of breccia fragments (>1mm)

<table>
<thead>
<tr>
<th></th>
<th>dark polymict metabreccia</th>
<th>light basaltic metabreccia</th>
<th>feldspathic basalt</th>
<th>anorthositic + large plagi.</th>
<th>granitic + rhyolitic</th>
<th>other</th>
<th>Total area of thin section</th>
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<td>1/7</td>
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<td>2/2</td>
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<td>200</td>
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<td>220</td>
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<td>0</td>
<td>1/2</td>
<td>220</td>
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<tr>
<td>total 14306</td>
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<td>28/68</td>
<td>2/3</td>
<td>17/37</td>
<td>3/9</td>
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<td>1280</td>
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<td>8/12</td>
<td>0</td>
<td>8/10</td>
<td>400</td>
</tr>
</tbody>
</table>

First no. = no. of fragments; second is area in mm². *Principally norite.
SOLAR WIND INTERACTIONS WITH THE MOON: RESULTS
FROM THE APOLLO 15 SUBSATELLITE; K.A. Anderson, L. M. Chase,
R.P. Lin, Univ. of Calif., Berkeley 94720; J. E. McCoy, NASA Manned
Spacecraft Center, Houston, Texas 77058; R. E. McGuire, Univ. of Calif.,
Berkeley 94720

Information on the orbit of the Apollo 15 subsatellite is given in
Figure 1.

Properties of Solar Wind Electrons ~500 to 2200 eV. Electrons
belonging to the solar wind electron component are present in the ranges
530 to 680 eV and 1.75 to 2.25 keV. Electrons are present in the inter­
planetary medium with energies above these energies but do not exhibit
collective properties as do the 530 to 2200 eV electrons. Somewhere
between 2 keV and 6 keV electron population of a different solar or
terrestrial origin becomes dominant. An energy spectrum showing both
these interplanetary electron populations is shown in Figure 2.

The pitch angle distribution of solar wind electrons from ~500 to
2200 eV is highly anisotropic. The directional flux along the interplanete­
tary field from the sun may be 2 to 5 times the flux directed back
toward the sun. This effect can be explained by a temperature anisotropy
but it would apparently have to be somewhat larger than the 10 to 20%
reported by the VELA group. When the anisotropic pitch angle distribu­
tion is used to interpret certain features of the solar wind cavity,
"temperature" anisotropies of almost 2 to 1 would be required. How­
ever this would be required only in the highest energy portion of the
non-Maxwellian component.

The Solar Wind Cavity. There are regions behind the moon at an
altitude of ~120 km where the intensity of ~500 eV solar wind electrons
drops by at least 30 to 100 times below their intensity in front of the
moon. This is true for all directions of the interplanetary magnetic field.

When $\varphi \sim 180^\circ$ (Figure 3) the 500 eV solar wind electrons penetrate
into the cavity to distances of ~100 km which is much greater than the
gyroradius of the electrons (about 8 km), but comparable to the gyro­
radius of the most energetic ions.

For interplanetary field directions of $\varphi \sim 90^\circ$, there is also pene­
tration of the ~500 eV electrons into the night side solar wind cavity.
Electrons are completely excluded only in a sector about 25$^\circ$ wide lying
on the Eastern side of the moon's dark side. A partial cavity extends
from somewhat behind the Eastern limb all the way to within ~20$^\circ$ of the
noon meridian (Figure 4). This effect can be understood, at least in
part, in terms of a highly anisotropic pitch angle distribution with most
of the electrons moving from West to East at these times.
Upstream Interaction of the Solar Wind with the Moon. The increases in the 500 eV solar wind electron fluxes across the entire sunlit side of the moon are not associated with intrinsic variations in the solar wind. These features are highly stable in spatial location with respect to a sun-fixed reference (or the solar wind flow direction). These increases are often associated with disturbances of the magnetic field.

Although present for 2 days these features seem to have been less definite and weaker before and after the two day period.

These spatially fixed features (Figure 5) evidently represent an interaction between the solar wind and the moon involving deflection of some of the solar wind flow. For example, if only a Mach cone is present tangent to the moon slightly ahead of the solar wind terminator, there should be no effects sunward of an orbital location near or somewhat behind $\delta$ in Figure 1. The magnetic field is also considerably disturbed outside and ahead of where a Mach cone is expected to lie. We do not yet know what causes a partial deflection of the solar wind around the moon or why this should occur at some times and not others. Some form of lunar atmosphere seems to be required.
SOLAR WIND INTERACTIONS WITH THE MOON

K.A. Anderson

Figure 2

Figure 3

Figure 4

Figure 5
ELECTROSTATIC PROPERTIES OF LUNAR REGOLITH
by
G. Arrhenius, S.K. Asunmaa and R.W. Fitzgerald
University of California, San Diego
La Jolla, California 92037

Lunar regolith particles exposed to radiation on the moon attract one another electrostatically. The open cluster structure of the lunar soil results from this effect which also causes rock and grain surfaces to be coated with a tenaciously adhering layer of ultrafine particles.

Significance of study

Measurement of the forces of attraction and clarification of their nature are of importance for the following reasons:
1. to understand the cohesion and aggregation of the surface material on the Moon, and by inference of material in the space environment in general.
   These phenomena would be of particular importance in a non-gravitational environment; hence they would largely control the initial accretion of particles and the growth of small bodies in space (Alfvén and Arrhenius, 1970; Asunmaa et al., 1970).
2. to clarify the motion and redistribution of lunar surface material under varying electric field conditions as inferred by Gold (1971, 1972), and the record in lunar core samples of particle overturn (Arrhenius et al., 1971; Macdougall et al., 1972).
3. to evaluate the record of corpuscular radiation by track densities and by the distribution of implanted ions. The ultrafine particles as an adhesive coating on larger grains capture the majority of impinging low energy ions (Holland et al., 1971) and shield the coated grains from this flux, thus enhancing the apparent erosion rate indicated by the track record, and,
4. to interpret the optical properties, particularly the polarization of reflected light from the lunar surface, and other atmosphere-free bodies in the solar system.

Measurement of net charge

Samples of lunar soil particle fractions <1 mm were introduced into a horizontal electric field by vibration through a 2 mm wide polymer coated gold grid with 60-100 μm mesh openings. Deflected particles were collected on electrodes where the vertical travel distances were measured and particle dimensions esti-
ELECTROSTATIC PROPERTIES OF LUNAR REGOLITH, G. Arrhenius et al.

The electric field intensity was varied from 75 to 350 \( \text{v cm}^{-1} \). About 98 per cent of the particles were found to have a net negative charge. Assuming free fall of the particles, the net charges were found to range between \( 10^{-11} \) and \( 10^{-13} \) coul kg\(^{-1} \); in comparison triboelectric mineral particle charging typically ranges from \( 10^{-12} \) to \( 10^{-16} \) coul kg\(^{-1} \) (Zimon, 1969). These preliminary experiments also indicate that the charge density decreases with increasing particle size; charge to surface ratios around \( 10^{-6} \) coul m\(^{-2} \) were found for particles in the 10 \( \mu \text{m} \) range; the ratio average for approximately 70 \( \mu \text{m} \) particles was about three times less.

The polarity of the excess charge was verified by selective adsorption of the lunar ultrafine particles on positively charged selenium of the type used in electrophotography. It is not known at the present time if the observed charge is of lunar or terrestrial origin.

Redistribution of charge

Apollo 15 samples show more extensive surface adhesion than Apollo 11 samples stored for two years on Earth (compare also observations by Grossman et al., 1971). Grains from the bottom layers in the Apollo 12 double-core with an estimated deposition age exceeding 50 My (Arrhenius et al., 1971) on the average show lower coating density (weaker particle adhesion) than grains from the surface layer.

These qualitative observations, now being supplemented with quantitative measurements, suggest a non-linear decrease of the adhesion during \( 10^6 \) to \( 10^7 \) years.

The pronounced hydrophobic character of the lunar particle surfaces (presumably caused by formation of Si-H bonds on solar proton excited silicate surfaces (Asunmaa 1970)) apparently prevents charge leakage even at high partial pressure of water in the Earth's atmosphere.

Origin of charge asymmetry on particles

From our present results and other data we infer the main reasons for the charge asymmetry to be:

1. the damage induced by corpuscular radiation, lowering the activation energy for internal polarization. This damage is sufficiently extensive so that the most heavily irradiated surface layer, about 1000 \( \text{A} \) thick, breaks apart when polar solvents such as acetone, condensate in the surface layer.

2. the surface polarization caused by photoelectron emission from the exposed side of grains during the lunar day.
ELECTROSTATIC PROPERTIES OF LUNAR REGOLITH, G. Arrhenius et al.

3. The effect of the lunar electric field (Rhee, 1967, 1971; Grobman and Blank, 1969; Gehrels et al., 1964; Parker, 1964) for preserving polarization of the radiation damaged material and inducing persistent internal polarization.

References


Gold, T. (1971) "Erosion, Transportation and Nature of the Maria", CRSR 452, Cornell University, Center for Radiophysics and Space Research, Ithaca, N.Y.


DEFORMATION OF SILICATES IN SOME FRA MAURO BRECCIAS; Hans G. Ave'Lallemant, Dept. of Geology, Rice Univ., Houston, Texas, 77001 and Neville L. Carter, Dept. of Earth and Space Sciences, S.U.N.Y., Stony Brook, 11790.

Breccias 14068, 9, 14321, 24, 140305, 15, 14313, 42 and 14301, 83 have been examined in detail by optical methods for evidence of static and dynamic deformation of silicate crystals. There is no unequivocal evidence for static deformation in the olivines, orthopyroxenes, clinopyroxenes or plagioclases in the specimens studied but a wide range of effects due to shock deformation have been observed and the preliminary results are summarized below.

Olivine crystals are frequently fractured and one crystal contains kink bands parallel to (001) which have originated by inhomogeneous slip on \{100\}[001], the high strain rate - low temperature system observed in static experiments and in chondritic meteorites. Slip on \{hko\}[001] may also have occurred in this crystal. One olivine grain contains well-developed planar features (shock lamellae) parallel to (110). Several olivine crystals in the specimens have recrystallized to mosaics of new strain-free crystals. The relic host crystals of two such olivine grains that had not recrystallized entirely have been severely shocked as indicated by the shock mosaic structure. Recrystallization of this sort has been observed in olivine shocked experimentally to about 1 megabar and results from annealing during the period of high residual temperatures following intense shock deformation. The orientations of the recrystallized grains are closely related to that of the host with the various crystal axes being inclined at an average of 35° to similar axes of the host, a relationship that has also been observed for statically annealed silicates.

Orthopyroxenes in the specimens are commonly deformed as indicated by irregular zones of undulatory extinction and poorly defined kink bands. The external rotation axes lie near (001) and kink bands are nearly normal to c suggesting slip on \{hko\}[001]. One orthopyroxene crystal contains intergrown clinopyroxene with [010] opx ll [010] cpx; [001] opx ll [001] cpx; \gamma opx = 43°; 2\gamma = 50°. However, the expected inversion of orthopyroxene to clinopyroxene has not yet been definitely established in the specimens studied.

Clinopyroxenes have deformed by mechanical twinning on (001), fracturing, faulting and some show the shock mosaic structure. Planes of closely spaced cavities or inclusions near (001) in one crystal are interpreted to be partly annealed planar features. Similar structures in another crystal have been partly recrystallized, a feature observed previously for naturally shocked quartz. Because of the fine-grained nature of annealed clinopyroxenes, the orientations of the recrystallized
grains with respect to the host have not been determined as yet.

These breccia specimens contain many fragments of anorthosite with the plagioclases (ca. An90) showing all degrees of shock deformation from zero shock through melting. Deformation twins according to the albite, pericline and, possibly, Aka laws have been recognized and are commonly associated with faults. Planar features and shock mosaicism are also commonly associated with fault zones along which intense crushing has taken place. Regions showing the mosaic structure are commonly bounded by irregular surfaces although some may be bounded by (010) and (001) segments. The external rotation axis between host and one mosaic subgrain is near [001] and the 30° disorientation may have occurred by predominant slip on (010)[1c]. Networks of regular to irregular boundaries defined by cavities or inclusions in some of the crystals may be partially healed microfractures.

One anorthosite fragment, elliptical in shape and 1 mm in maximum dimension, contains a perfectly round hole near the center (see sketch below). Outlining the microcrater in an asymmetric fashion are prismatic crystallites with interstitial glass. The intensity of deformation diminishes away from the crater from intense shock mosaicism accompanied by fine-grained recrystallization - mosaicism - faults and cracks, some partially rehealed - sparse cracks - zero shock. Thus, in this single fragment we see all stages of shock deformation from zero shock (stresses less than ca. 10 kb) to melting (stresses of the order of 300 kb). The distance over which the shock intensity has decayed from ca. 300 kb to 10 kb is about 0.5 mm and if the crater radius (0.05 mm) is near the size of the projectile, the shock stress levels have attenuated approximately as $1/r^2$. 
TRACE ELEMENT STUDIES OF LUNAR ROCKS AND SOILS;
P.A. Baedecker, C.-L. Chou, J. Kimberlin and J.T. Wasson, University of California, Los Angeles 90024

Seven elements, Ni, Zn, Ga, Ge, Cd, In and Ir, have been determined in three soil samples, one crystalline rock, and separated clasts and matrix material from two breccias recovered from the Fra Mauro landing site. The same elements have also been measured in four soil samples obtained on the mission to Hadley-Appenines. The technique employed in the analyses was neutron activation, and the data are presented in the accompanying table. All data represent the means of replicate analyses except for 14321,184,32; 14066,21,1.01 and the Apollo 15 soil samples.

Data on Ni, Zn, Ge, Cd and Ir in samples from the Apollo 11 and 12 missions showed that these elements were strongly enriched in soil and breccia samples relative to local crystalline rocks. The enrichments were ascribed to an extralunar component which amounted to approximately 1.1% and 1.0% respectively in the "mature" soils 12070 and 10084, expressed in terms of an assumed (water-free) C1 chondrite-like composition.

The soils from Fra Mauro and microbreccia clasts from sample 14321 are characterized by significantly higher contents of these trace elements than soils from other mare locations which have been sampled. Based on the data for Ni and Ge (which are determined with the greatest precision), the extralunar component amounts to approximately 2.2% C1-like material. Correcting for mean regolith depth by means of the equation derived by Baedecker et al. (1), we calculate an integrated flux at the Apollo 14 site to be 6.7 relative to that at the Apollo 12 site taken to be 1. This is consistent with the observation that the flux of extralunar material onto the moon's surface must have been decreasing with time 3.2 to 3.9 Gyr ago (1, 2, 3), but implies an extremely rapid decay rate between 3.9 Gyr (Apollo 14) and 3.65 Gyr (Apollo 11). Alternatively, it seems possible that part of the extralunar material at the Fra Mauro site was accreted prior to the Imbrian event. For example, the mean extralunar component in the material excavated from the Imbrian basin might have been 0.5-0.9%, with the remainder of the material accreted during the past 3.9 Gyr.

Fra Mauro crystalline rock 14310 has concentrations of some siderophilic elements which are much higher than those in rocks from other lunar locations. The Ni and Ir data are consistent with a C1-like extralunar component amounting to about 1.3%, but the Ge and Zn concentrations are relatively much lower. Either
the projectile which produced 14310 was differentiated, or a fractionation of the extralunar component has occurred, perhaps during the period when the 14310 material was molten. The Ge/Ni ratio in the extralunar component in 14310 appears to be about $7 \times 10^{-4}$, which is similar to that observed in group IID iron meteorites.

It has been suggested (4) that 14310 is a representative sample of the rock type making up the bulk of the soil from the Fra Mauro site. Our data tend to refute this suggestion. The local soils (such as 14259 and 14163) have Ge/Ni ratios similar to those found in soils from other lunar locations. If 14310 material accounted for a major part of the Fra Mauro soils, then the Ge/Ni ratio in the extralunar component at this site differs significantly from those in the extralunar components at other sites. The more plausible explanation is that 14310 is not a representative sample, at least in terms of the siderophilic and volatile elements. It also has an anomalously low Ga/Al ratio of 4, whereas the Fra Mauro soils have Ga/Al ratios within the range 7.5 ± 2.5 within which 95% of the lunar ratios lie. The Ga concentration in 14310 is about half that in the Fra Mauro soils.

Major element data on 14310 indicate that it is similar in composition to the KREEP component found in the soil at the Apollo 12 site. However, Ni, Ge and Ir data on Apollo 12 soils 12033 and 12070 suggest a negligible siderophilic element content in KREEP (5). The Ga/Al ratio in KREEP was calculated to be 8.1 (5), a typical lunar value. Thus 14310 is also not a representative sample of KREEP, at least in terms of these elements.

Soil samples from two locations at the Hadley-Appenine landing site have been analyzed. Soil obtained near the St. George Crater has a relatively large extralunar component amounting to 1.3% Cl-like material, while soil from the Rille location has an extralunar component which is a factor of 2 lower. These values are, respectively, about 30% higher and 30% lower than the concentrations observed in the "mature" soil at the Apollo 12 site, where the crystalline rock ages are similar. The Ge/Ni ratio in both Apollo 15 soils is within the range observed in other lunar soils. We believe the explanation of the concentration differences to be as follows. The St. George Crater sample contains an admixture of Appenine soil, which is older, and thus richer in the extralunar component, whereas the regolith at the Rille location is younger than the underlying mare-type basalts due to a gradual loss of soil into the Rille. We do not believe that the lower extralunar component of the Rille soil results from dilution by exotic lunar materials, such as appeared to be the preferred explanation of the origin of soil 12033 (5).

Zinc and Cd are strongly correlated in the lunar material analyzed to date. All samples for which our analytical data are complete have Cd/Zn ratios in the range $(2.5-13) \times 10^{-3}$, with the exception of three Apollo 12 rocks which appear to have lower
FRA MAURO ROCKS AND SOILS
P.A. Baedecker

ratios. Zinc concentrations range over a factor of 40, Cd concentrations over a factor of about 100. The highest concentrations are always observed in the soils and breccias, indicating the presence of one or more Zn-, Cd-rich exotic components at each site. We find evidence for both extralunar and lunar Zn-, Cd-rich materials as well as for handling contamination.

The apparent contamination of most Apollo samples by the In seals utilized on the rock boxes continues to frustrate any attempt to obtain meaningful data for this cosmochemically important element.


Concentrations of seven trace elements in rocks and soils from Fra Mauro and Hadley-Appenines.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Description</th>
<th>Ni</th>
<th>Zn</th>
<th>Ga</th>
<th>Ge</th>
<th>Cd</th>
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</tr>
<tr>
<td>15531</td>
<td>fines (Rille)</td>
<td>140</td>
<td>7.7</td>
<td>4.0</td>
<td>120</td>
<td>20</td>
<td>410</td>
<td></td>
</tr>
</tbody>
</table>

We report here (1) the first magnetic study of iron-titanium-chromites (synthesized by Prof. A. Muan of Penn State U.), (2) the discovery of a characteristic magnetic signature of such compounds at their transition temperatures, $T_t$, where $80K < T_t < 140K$, thus enabling one to identify their presence in lunar samples and (3) we postulate a physical mechanism by which some lunar rocks could have acquired their stable NRM due to temperature-cycling through $T_t$ of iron-titanium-chromites. Depending on the efficiency of such a mechanism, it may become unnecessary to postulate the presence of large (~$10^3\gamma$) magnetic fields when such rocks were magnetized. In turn, this will remove the need for a once-molten lunar core and a lunar dynamo. In passing, we also note that the discovery of a low temperature magnetic transition in iron-titanium-chromites may obviate the need to postulate the presence of magnetite in lunar rock 12020 (Runcorn et al., 1971 Houston Conference), a hypothesis which is not supported by other workers who have used magnetic and non-magnetic techniques. Fig. 1, Curve A through the hollow circles, shows a sharp decrease in magnetization ($M$), measured in $116\ Oe$, at $T_t = 100K$ for synthetic iron-titanium-chromite $Fe_{1.14}Cr_{1.72}Ti_{0.14}O_4$ (mineralogical name = titanchromite). The same sharp decrease is found whether the applied field is zero or nearly 50,000 $Oe$, thus distinguishing it from magnetite where a similar sharp decrease, when measured in zero or low fields is obliterated on application of fields 10,000 $Oe$. Two other synthetic titanchromites, $Fe_{1.29}Cr_{1.42}Ti_{0.29}O_4$ and $Fe_{1.44}Cr_{1.12}Ti_{0.44}O_4$ had $T_t = 120K$ and 140K respectively while a chromian ulvospinel, $Fe_{1.88}Cr_{0.24}Ti_{0.88}O_4$ had $T_t = 120K$. Again, unlike magnetite, remanent magnetization applied at room temperature to iron-titanium-chromites does not undergo a sharp fall-off at $T_t$ when cooled but instead, shows a sharp rise as shown here for a warming experiment. On the basis of these data, we consider Curve A of Fig. 1 to be a characteristic magnetic signature of iron-titanium-chromites and thus, we can now look at data-points B (crosses) and C (solid circles) in Fig. 1, which were obtained for a sample of rock 12063, supplied by R. Hargraves of Princeton U. The points B were obtained on heating in $116\ Oe$ while the points C were obtained on cooling in $41\ Oe$. Note the different $M$-scales. In addition to a shoulder at about 60K which is characteristic of ilmenite, 12063 shows the same magnetic signature as our synthetic titanchromite (Curve A). Optical microscopy also revealed the presence of titanchromites in this sample. Using the same method we have found the presence of titanchromites in our samples of 14321 (dark clasts only) and 14310. We note that part of the magnetic evidence on the basis of which Runcorn et al. based their claim of observing magnetite in lunar rock 12020 was such a magnetic transition at 140K in a warming experiment. It would seem that titanchromite, not studied
Fig. 1. Magnetization (M) vs. Temperature (T) for titanchromite Fe$_{1.14}$Cr$_{1.72}$Ti$_{0.14}$O$_4$ (Curve A through the hollow circles) and lunar rock 12063 (Crosses B for heating in 116 Oe, Solid Circles C for cooling in 41 Oe). Note that M-scales are different for A, B and C.
Iron-Titanium-Chromite

S. K. Banerjee

P. 40

magnetically up to now, would be an equally plausible and perhaps, preferable alternative.

A third and most important result of our discovery of magnetism in iron-titanium-chromite relates to the consequences of the suggestion that this family of minerals is responsible for the stable component of Natural Remanent Magnetization (NRM) in some lunar rocks. Strangway and Fuller have suggested that the thermal cycling during lunar days and nights suffered by iron particles in lunar rocks will result in an oscillating coercive force which may result in an acquired stable NRM which can then be mis-interpreted as a Thermo-Remanent Magnetization (TRM) acquired on cooling through the Curie point of iron. Thus, a conventional paleointensity experiment in which such a cooling process is simulated in a known magnetic field cannot yield the true value of ancient lunar field. We think that such a mechanism is justified from the point of view of physics, however, the magnitude of such a remanence may be negligible for iron. This is so because this remanence, in the first order, is proportional to the ratio of the coercive force at low temperature (say, 77K) and that at high temperature (say, 550K). For iron, this ratio is about 1.5, a comparatively small value if such a remanence has to compete with conventional TRM. But, for magnetite which has a magnetic transition point at 120K (when cooling produces a sharp increase in coercive force), this ratio can be as large as 5 or 6 and a remanence by thermal cycling in magnetite could be appreciable and has already been invoked by Arrhenius and Brecher (personal communication) to explain the stable NRM in carbonaceous chondrites. In our case, we have found that in titanchromites, this ratio of coercive forces can be as high as \(10^3\) to \(10^4\), making such a remanence acquisition process very likely for titanchromites. The next problem to be solved is to show how much of such a remanence is left when heated to 300K. For \(Fe_{1.14}Cr_{1.72}Ti_{0.14}O_4\), the preliminary Curie point determined by us is about 425K and therefore, it can conceivably carry such a remanence at room temperature if the remanence is still finite at 300K. Such an experiment is important to do because if the mechanism of remanence by thermal cycling in titanchromites can prove to be more efficient than conventional TRM, the lunar ambient field need not be as large as of the order of \(10^3\) which has been postulated by other workers to explain the large stable NRM in lunar rocks.

We are reporting our results obtained on two samples returned by the Apollo 14 mission. One of these was a sample of the large breccia, 14321,221 and the other a sample of soil, 14163,159. The description of these samples has been published by L.E.P.E.T.[1].

The objective of this investigation was to obtain, on the same sample, precise and accurate isotopic ratios as well as isotope dilution assays for a variety of elements. The elements chosen were, for the most part, those for which well characterized reference standards were available.

Although the investigation for all the elements to be determined is not yet completed we report here the isotopic results for Pb, U, Th, Rb, Sr and Cu and the assay results for these elements as well as for Fe, Ti, Ni, B, Cd and Ag.


Table 1. Isotopic Compositions.

<table>
<thead>
<tr>
<th>Element</th>
<th>Sample 321</th>
<th>Sample 163</th>
<th>Terrestrial</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lead</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>208/206</td>
<td>0.98837</td>
<td>0.94909</td>
<td></td>
</tr>
<tr>
<td>207/206</td>
<td>0.54773</td>
<td>0.73775</td>
<td></td>
</tr>
<tr>
<td>204/206</td>
<td>0.001755</td>
<td>0.001945</td>
<td></td>
</tr>
<tr>
<td>Blank</td>
<td>7.72 ng</td>
<td>5.63 ng</td>
<td></td>
</tr>
<tr>
<td>Uranium</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>238/235</td>
<td>137.80</td>
<td>137.82</td>
<td>137.88 ± .20</td>
</tr>
<tr>
<td>234/235</td>
<td>0.007469</td>
<td>0.007465</td>
<td>0.007495 ± .00004</td>
</tr>
<tr>
<td>Rubidium</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>85/87</td>
<td>2.5916</td>
<td>2.5921</td>
<td>2.59265 ± .0017</td>
</tr>
<tr>
<td>Strontium</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>88/86</td>
<td>8.3752</td>
<td>8.3752</td>
<td></td>
</tr>
<tr>
<td>87/86</td>
<td>0.70799</td>
<td>0.71485</td>
<td></td>
</tr>
<tr>
<td>84/86</td>
<td>0.05655</td>
<td>0.05651</td>
<td></td>
</tr>
<tr>
<td>Copper</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>63/65</td>
<td>2.2458</td>
<td>2.2417</td>
<td>2.2440 ± .0021</td>
</tr>
</tbody>
</table>
Table 1. Isotopic Compositions (continued).

NOTES:

a NBS Certified values.
b Normal laboratory limits are 208/206 ± 0.10%, 207/206 ± 0.066%, 204/206 ± 0.10%.
c ng = 10⁻⁹ gm/gm.
d Normalized to 86/88 = 0.1194.

Table 2. Concentration Analysis.

<table>
<thead>
<tr>
<th>Element</th>
<th>Sample 321</th>
<th>Sample 163</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lead</td>
<td>3.306 ppm</td>
<td>9.757 ppm</td>
</tr>
<tr>
<td>Blank</td>
<td>0.021 ppm</td>
<td>0.014 ppm</td>
</tr>
<tr>
<td>Uranium</td>
<td>1.550 ppm</td>
<td>3.650 ppm</td>
</tr>
<tr>
<td>Blank</td>
<td>0.0007 ppm</td>
<td>0.0008 ppm</td>
</tr>
<tr>
<td>Thorium</td>
<td>5.778 ppm</td>
<td>13.21 ppm</td>
</tr>
<tr>
<td>Blank</td>
<td>0.0001 ppm</td>
<td>0.0001 ppm</td>
</tr>
<tr>
<td>Rubidium</td>
<td>5.964 ppm</td>
<td>15.41 ppm</td>
</tr>
<tr>
<td>Blank</td>
<td>0.0031 ppm</td>
<td>0.0031 ppm</td>
</tr>
<tr>
<td>Strontium</td>
<td>120.29 ppm</td>
<td>185.50 ppm</td>
</tr>
<tr>
<td>Blank</td>
<td>0.117 ppm</td>
<td>0.117 ppm</td>
</tr>
<tr>
<td>Copper</td>
<td>77.91 ppm</td>
<td>9.981 ppm</td>
</tr>
<tr>
<td>Blank</td>
<td>0.14 ppm</td>
<td>0.034 ppm</td>
</tr>
<tr>
<td>Iron (as Fe)</td>
<td>11.50%</td>
<td>7.36%</td>
</tr>
<tr>
<td>Titanium</td>
<td>1.31%</td>
<td>0.979%</td>
</tr>
<tr>
<td>Nickel</td>
<td>145 ppm</td>
<td>333 ppm</td>
</tr>
<tr>
<td>Boron</td>
<td>--</td>
<td>2.19 ppm</td>
</tr>
<tr>
<td>Cadmium</td>
<td>0.50 ppm</td>
<td>≤0.3 ppm</td>
</tr>
<tr>
<td>Silver</td>
<td>&lt;0.007 ppm</td>
<td>0.018 ppm</td>
</tr>
</tbody>
</table>

a ppm = 10⁻⁶ gm/gm.
b Procedure subsequently changed, current Sr blank 0.002.
c Determined by spark source mass spectrometry.
d Determined by differential polarography.
e Determined by nuclear track counting.

For those elements for which isotopic compositions have been measured thus far a number have compositions which are considered to be essentially constant in nature. These are uranium, thorium, rubidium and copper. In each case the measured ratio/s are, within experimental error equal to those found previously in terrestrial materials.
A small correction for the blank lead (208=52.34%, 207=22.08%, 206=24.14%, 204=1.42% and the remaining lead corrected for lead of troilite composition as listed by Oversby[4]. The resulting composition as well as the apparent lead-uranium ages are listed in Tables 3 and 4.

Table 3

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>208</th>
<th>207</th>
<th>206</th>
<th>204</th>
</tr>
</thead>
<tbody>
<tr>
<td>321</td>
<td>38.825</td>
<td>21.531</td>
<td>39.348</td>
<td>0.05771</td>
</tr>
<tr>
<td>163</td>
<td>35.265</td>
<td>27.415</td>
<td>37.177</td>
<td>0.07147</td>
</tr>
</tbody>
</table>

(after troilite correction)

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>208</th>
<th>207</th>
<th>206</th>
</tr>
</thead>
<tbody>
<tr>
<td>321</td>
<td>37.154</td>
<td>20.941</td>
<td>38.8086</td>
</tr>
<tr>
<td>163</td>
<td>33.1952</td>
<td>26.6848</td>
<td>36.5095</td>
</tr>
</tbody>
</table>

Table 4

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>207/206</th>
<th>206/238</th>
<th>207/235</th>
<th>208/232</th>
</tr>
</thead>
<tbody>
<tr>
<td>321</td>
<td>4.416</td>
<td>4.375</td>
<td>4.404</td>
<td>4.379</td>
</tr>
<tr>
<td>163</td>
<td>4.861</td>
<td>4.923</td>
<td>4.881</td>
<td>4.910</td>
</tr>
</tbody>
</table>

The rubidium and strontium data also permit the calculation of a model age with some assumptions about the $^{87}$Sr/$^{86}$Sr initial value. Papanastassiou and Wasserburg have recently published[5] data for initial values of these two samples of 0.69942 (basalt clast) and 0.69898 (BABI) for the fines. Using these values "ages" of 4.18 BY and 4.60 BY are obtained for 321 and 163 respectively. Thus the Rb-Sr ages are consistently lower in this case as has been shown generally. This is particularly conclusive since both "ages" in this case were determined on the identical sample fraction. Both samples show discordant lead-uranium-thorium ages. A line joining the two intersects the concordia line at 4.75 BY.

MOSSBAUER STUDIES OF APOLLO 14 LUNAR SAMPLES

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The University, Newcastle upon Tyne, NE1 7RU, England

The iron-bearing minerals in ten Apollo 14 lunar samples have been examined by Mössbauer spectroscopy. The results parallel earlier data on the Apollo 11 and 12 samples, but show significant differences in the mineral content.

Lunar soils

The soil samples 14003.20, 14162.48, 14163.50, and 14259.17, gave basically similar spectra at 78 K and 295 K which could be represented by three overlapping quadrupole doublets from Fe$^{2+}$ cations in ilmenite, M2 pyroxene sites/glass, M1 pyroxene sites/olivine. The ilmenite constituted 10% or less of the total iron, thereby contrasting with the ilmenite-rich Apollo 11 soils (19-27%) but being similar to those from Apollo 12. The proportion of metallic iron was also substantially less than for Apollo 11 soils, and was barely detectable as a hyperfine magnetic pattern in the bulk samples ($\lesssim$ 1%).

All four soil samples have a high glass content which results in substantial broadening of the lines; this distinguishes these soils from the soils returned by both previous missions and prevents analysis of the data in the same depth. No significant differences were seen between samples from the upper 1 cm of the regolith (14259) and from a depth of several cm (14162, 14163). The individual contributions from the orthopyroxene, clinopyroxene and olivine known to be present in sample 14259 could not be distinguished.

Our preliminary data for five Apollo 15 soil samples show a similar distribution of iron in the various minerals, but with the distinction of substantially better spectrum resolution indicative of a high degree of crystallinity. From the data for all four Apollo missions, it appears that the composition of the soil is essentially independent of the detailed sampling within each lunar region.

Rock samples

The Mössbauer spectra of six rock samples were obtained at 78 K and

295 K. The basic three-doublet pattern remained, although the lower glass content of these samples resulted in narrower resonance lines which enabled the mineral compositions to be estimated more accurately. The three fragmental rocks 14301.15, 14303.36, and 14321.17/9, were similar to each other, with \( \sim 7\% \) of the iron as ilmenite and \( \sim 24\% \) in M1 pyroxene/olivine sites, the remaining \( \sim 69\% \) being in M2 pyroxene sites. A fourth rock 14318.35 was similar in composition but showed greater linewidths and a correspondingly lower resolution at both temperatures; a greater variation in the silicate compositions or a higher glass content is indicated for this sample. Fragmental rock 14311.32 and igneous rock 14310.66 both showed a decrease in the M1 pyroxene/olivine contribution (to \( \sim 16\% \)) with substantially different parameters. The two are however distinct from each other in containing 6\% and 8\% ilmenite respectively. Any troilite or metallic iron content was below detectable limits (0.2\%) in all the rocks.

The major iron mineral in rock 14310 is clinopyroxene with a fairly limited composition range close to \( \text{Fe}_4\text{Si}_2\text{O}_6 \). With the assumption of Ca ordering onto the M2 sites, and using our measured M1/M2 population ratio, we derive a site occupancy of

\[
\begin{align*}
\text{Mg}^{0.90-0.10} & \text{M1} \\
\text{Fe}^{0.02-0.46} & \text{M2} \\
\text{Si}^{0.52-1} & \text{Si}^{2}\text{O}_6
\end{align*}
\]

The ordering of the magnesium onto the M1 site indicates equilibration at a low temperature of the order of 900 K.

Mineral separates from soil 14259.17

4.5 g of soil sample 14259.17 were subjected to several physical separation procedures in an attempt to provide specimens of the component minerals. Size fractionations from > 500 \( \mu \)m to < 75 \( \mu \)m gave no significant compositional separation. A density fractionation in Clerici solution of the > 125 \( \mu \)m particles was partially successful. The highest density fraction gave a sharp spectrum from pyroxene/olivine soil containing a small amount of ilmenite, but otherwise similar to separates from Apollo 11 rocks. Iron metal was absent. The lowest density fraction which contained mainly glass gave broadened asymmetric lines and was the only sample to show a significant resonance from iron metal. The latter is therefore mainly associated with the glassy particles. Intermediate fractions showed a graduation between the end members. A paramagnetic fractionation obtained by means of a Frantz magnetic separator from one of the intermediate density samples was largely ineffective except that a small quantity of plagioclase containing little or no iron was obtained.

By contrast, a strongly magnetic separate (7 mg) was obtained from the > 125 \( \mu \)m size fractions using a hand magnet. This sample gave a single magnetic hyperfine splitting with less than 20\% of residual silicate material. The observed magnetic field (335 kG) is 3 kG greater than in an iron foil at room temperature and the lines were considerably broader than expected for pure iron metal. These features persisted at 4.2 K, and are consistent with an average nickel content of 3\%, although the presence
of cobalt or other alloying impurities cannot be excluded. Significantly, no troilite was detected in this magnetic separate or indeed in any Apollo 14 samples examined despite its presence in Apollo 11 and 12 soils.

The presence of nickel in the iron granules has been confirmed by X-ray fluorescence measurements. The relative proportions of Ti, V, Cr, Mn, Fe, Co, and Ni were determined in the sample and compared with the analysis of the bulk soil. Considerable enrichment of nickel and cobalt was found and, after correction for the iron in the silicate residue, an average alloy composition of 4.5% Ni, 1.0% Co and 94.5% Fe was deduced. The manganese and chromium were both depleted in proportion to the residual silicate and were not present in the alloy in significant quantity. Vanadium was not detected \( \leq 0.05\% \).

Microprobe analyses on 25 Fe-Ni granules gave an average composition of 5.4 at % Ni, 0.6 % Co, 94 % Fe. Five particles featured some degree of nickel concentration zoning, and in one of these a small area of taenite (37% Ni, 0.7% Co) was found at the edge of a fragment of kamacite (5.6% Ni, 1.2% Co). The equilibration temperature was derived from the phase diagram to be \( \sim 740 \) K. Some iron-nickel was also found in rock 14321 (16% Ni, 1.8% Co).

Many of the particles had a composition outside the range normally attributed to meteorites,\(^4\) and the evidence favours an indigenous fractionation rather than an extra-lunar meteoritic origin.

Heymann and Yaniv (1971) have recently reviewed the problem of the large amounts of 40-Ar in lunar fines. The total 40-Ar concentration exceeds that which could be expected from the in situ decay of 40-K. The amount of surface correlated 40-Ar is, in relation to the solar wind 36-Ar, orders of magnitude higher than the theoretically predicted abundance. Thus, retrapping of radiogenic Ar from the lunar atmosphere seemed a process capable to account for all observations.

The investigation of the detailed release patterns of He, Ne and Ar from lunar fines during linear heating reported here is aimed at a detection of retrapped 40-argon. We expected that the retrapped 40-Ar implanted with energies in the 1 KeV range should be released more easily than the solar wind 36-Ar. We therefore subjected samples to a heating rate of 100°C/hour and observed He, Ne and Ar continuously during an uninterrupted linear heating sequence released between 100 and 900°C. The gases were analysed in two statically operated mass spectrometers connected to the diffusion oven via a high pumping speed continuously operating purification system. Two to four peaks were taken per minute in each spectrometer and properly averaged readings were taken every 5 minutes. Sieve fractions of bulk samples <25μm from the fines 10084,48; 14163,166; 15101,88 and 15231.54, were analysed, as well as separates from 14163,166 enriched in plagioclase, pyroxene and magnetic glass. These separates, sieved to 30-48 μm were chosen to distinguish the release patterns of in situ radiogenic 40-Ar and "orphan-argon". The sample designated "Plag B" -selected to reduce the solar gases- consisted of plagioclases in the size range 200-350μ which were crushed to 30-48 μm.

Table 1 lists the total amounts of gas released up to 900°C from the separates of 14163 and those for the bulk samples. The different retentivities of the separates for solar wind noble gases are clearly seen. The 40-Ar concentrations are the same within a factor of two, whereas the 36-Ar differs by a factor of about 40. The solar noble gases in the <25 μm bulk fractions show the previously noted concentration variations.
In all samples the solar wind noble gas isotopes show a clear enhancement of the release rates below 300°C. This release amounts to less than 1% of the total gas released. A striking similarity of this release with that observed for 4-He implanted with 2.5 KeV quartz leads us to attribute at least partly the low temperature release enhancement to a peculiarity of the release of gases implanted as ions.

Fig. 1 shows at the left hand side the differential release rates of 36 and 40-Ar from the four separates as a function of temperature. The solar 36-Ar release rates alone are given, on one scale, on the right hand side, together with the 40/36 ratios. We note from the graphs:

1) The low temperature release of 40-Ar is enhanced over the 36-Ar release below about 400°C; however, 40-Ar released up to 300°C is only about 1% of the total 40-Ar.

2) The 36-Ar release rates as a function of temperature are relatively similar for all separates, showing an enhancement of the release rates at temperatures below 300°C similar to that found for He and Ne in bulks and separates.

3) The maximum release rates for 40-Ar and 36-Ar are around 750°C.

4) The plagioclase release rates below 600°C are comparatively high which may be attributed to a concentration profile affected by diffusion. Plagioclase has for solar gases the lowest apparent retentivity.

5) Comparison of the release rates of 36-Ar and 40-Ar shows that the samples rich in 40-Ar release 40-Ar primarily above 400°C. Thus, we tend to assign the high temperature release partly to in situ produced radiogenic Ar.

6) All 40/36 ratios of the separates show a minimum between 750°C and 800°C, which is also true for the bulk samples. The value of
the minimum depends either on sample mineralogy or bulk composition.

We conclude that our experiments do not give evidence for substantial amounts of 40-Ar implanted at low energies. The release patterns of 36 and 40-Ar observed seem best explained by a modulation of surface correlated Ar of a 40/36 around unity, by a low temperature 40-Ar addition (retrapped Ar) and a high temperature addition of in situ produced radiogenic Ar. Such an Ar distribution would result, if the solar 40/36 ratio is higher than expected and / or if an in situ decay of 40-K contributed to the surface correlated 40-Ar, whereby at least some of that K was later redistributed. The latter possibility would be substantiated if a surface correlation of the K concentration could be established.

This work was supported by the Swiss National Foundation, grant 2.386.70.

MICROSTRUCTURE, MELTING AND CRYSTALLIZATION CHARACTERISTICS OF LUNAR, VITREOUS FINES (14163,14162,14258).

G.Bayer(1) and H.G.Wiedemann(2)

(1) Institute of Crystallography and Petrography, Eidgenöss.Techn.Hochschule, Zürich, Switzerland.
(2) Mettler Instrumente AG, Greifensee/Zürich, Switzerland.

Melting and crystallization studies were carried out on lunar soil samples (<1 mm and 1-2 mm fines) using thermoanalytical methods, x-ray and electron microscopy. Small, homogeneous fractions were obtained by sieving, by magnetic separation and by picking of individual grains. Approx. 50% of the fines were magnetic especially the black, slag-like particles but also many greyish plagioclase rock fragments. This is mainly caused by metallic, iron-rich inclusions and droplets on the surface. Most of the crystalline grains consisted of plagioclase disregarding their appearance - clear crystals, grey rock fragments, black slags. Olivine and pyroxene were very rare. The majority of the transparent vitreous material were brown glass fragments and also spheres, less glasses were green or colorless. One large, green glass sphere (diameter 1.48 mm) with black, wrinkled surface was investigated separately: scanning electron micrographs proved that the interior was homogeneous, vitreous whereas lath-shaped crystals (plagioclase?) were abundant in one part of the surface (Fig.1). The Ti-concentration was definitely lower in the surface the overall Ca-, Mg-, Fe-, and Ti-concentration of the glass sphere however was very close to that of terrestrial basalts and to that of the black, vesicular spherules (Fig.2), which were observed frequently. Microscopic investigation showed that practically all
glass particles contained gas bubbles of very different sizes. The greatest concentration of bubbles is found close to the surface.

DTA runs on non-separated lunar fines (smaller 100 \( \mu \)m) in vacuum proved that the vitreous portion of the sample recrystallizes at 780°C during heating up (4°C/min), that melting occurs in the range 1140-1170°C, that the sample still contains a small portion of crystals at 1250°C and recrystallized during cooling (6°C/min) at 1045°C to plagioclase. Recrystallization could be avoided if the samples were heated higher (1400°C) and cooled with 15°C/min. The glass obtained was colored brown. When the fines were heated in graphite crucibles in vacuum (10^{-5} torr) up to 1300°C a green-brown vesicular glass sphere was obtained which contained many metallic, magnetic spherules and white, lath-shaped crystals on the lower surface, similar as those shown in Fig.1. This reduction is obviously due to formation of CO during heating up. Analogous experiments in which the sample was heated in alumina crucibles in \( N_2-H_2 \) up to 1400°C showed a strong loss in weight (12%) above 1050°C. The glass obtained was colorless and contained white, lath-shaped crystals and tiny metallic spheres.

Combined TGA or DTA, heating x-ray and mass spectometry proved to be very useful for following the melting and crystallization of the small samples available in relation to the evolution of gases. Neon and argon loss was observed first between room temperature and 200°C with a maximum at around 100°C. The largest amount of Ne and Ar is given off above the recrystallization of the vitreous portion of the soil sample, above 900°C. Experiments in which a molecular beam of the evolved gases was directed on a balance pan showed clearly the bursting of individual bubbles in
the course of melting. The heating x-ray pattern (1°C/min) proved in accordance with DTA that the crystalline reflections (plagioclase) disappear not completely in the melting region 1140 to 1170°C. During cooling (1°C/min) recrystallization of plagioclase occurred in the same temperature region, with identical but much sharper x-ray reflections than in the original soil sample. The gas evolution (Ne, Ar) decreases strongly in the region 850 to 650°C but increases again below 600°C. Similar experiments are carried out with homogeneous, separated soil fractions.

Additional experiments in vacuum, N₂-H₂ and O₂ were carried out with terrestrial basalts. The results obtained lead to the conclusion that vacuum and low alkali concentration both favor the formation of divalent iron above the liquidus temperature. This leads to formation of glasses in contrast to oxidized melts (Fe³⁺) which could not be quenched to glasses. Reducing conditions caused the formation of metallic spheres embedded in a colorless glass matrix.
COSMIC RAY PRODUCED RADIOISOTOPES IN APOLLO XII AND APOLLO XIV SAMPLES.

F. Begemann, W. Born, H. Palme, E. Vilcsek and H. Wänke
Max-Planck-Institut für Chemie (Otto-Hahn-Institut), Mainz, Germany

A sample of lunar fines 12001 was separated into two different grain size fractions, the finest (< 45 \( \mu \)) showing a significant enrichment in feldspar and, hence, different contents of target elements relevant for the production of \(^{22}\text{Na}\), \(^{26}\text{Al}\), \(^{36}\text{Cl}\), and \(^{39}\text{Ar}\). The samples were decomposed and the activities determined using the same procedures as described for Apollo XI. Data for the relative contributions (to the total activities) by the different target elements will be presented.

Furthermore, a slice of rock 12053 was divided into three depth fractions. A top layer (0-5 mm) was taken from that segment of the slice which, judged from the presence of micro-craters, had been exposed to free space. Care was taken not to lose the very outermost layer. \(^{14}\text{C}\) and \(^{39}\text{Ar}\) were measured in this layer, in the two other ones \(^{22}\text{Na}\), \(^{26}\text{Al}\), and \(^{36}\text{Cl}\) as well. \(^{22}\text{Na}\) and \(^{26}\text{Al}\) were determined before decomposition of the samples, employing a large volume Ge(Li)-detector, as well as after a chemical separation of the pure elements. The activities of \(^{36}\text{Cl}\) and \(^{39}\text{Ar}\) are rather invariant with depth, \(^{22}\text{Na}\) and \(^{26}\text{Al}\) exhibit the by-now well established depth dependence, which is shown here to exist for \(^{14}\text{C}\) as well (Table 1).

<table>
<thead>
<tr>
<th>depth</th>
<th>dpm (^{14}\text{C}/\text{kg})</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 - 5 mm</td>
<td>72.4 ( \pm 7 )</td>
</tr>
<tr>
<td>5 - 20 mm</td>
<td>32.8 ( \pm 3 )</td>
</tr>
<tr>
<td>20 - 65 mm</td>
<td>29.7 ( \pm 3 )</td>
</tr>
</tbody>
</table>

Using the model calculations of REEDY and ARNOLD the steep slope corresponds to a value of \( R \approx 80 \text{ MV} \), but a flux of around 500 protons pro cm\(^2\) sec. As this flux value is unreasonably high as compared to the results of ARNOLD and co-workers from the analysis of radioisotopes with half-lives between 2.6 years and 4 Myr., it appears that part of the
excess $^{14}C$ in the top layer has to be accounted for in a different way. A possible explanation would be that it was contained in the solar wind, which would require a ratio $^{14}C/H \approx 6 \times 10^{-12}$. Assuming this $^{14}C$ to be produced on the sun by "thermal" neutrons via the $^{14}N (n,p)$-reaction leads to a flux of $\Phi \approx 7 \times 10^5$ neutrons/cm$^2$ sec. It is suggestive that using this flux a ratio $^{3}H/H \approx 2.5 \times 10^{-10}$ for the solar wind would be predicted from the $^{3}He (n,p) ^{3}H$-reaction, which agrees within a factor of 5 with what has been suggested by FIREMAN and co-workers in order to explain an excess of tritium in the surface of rock 12002.

Finally, in the two soil samples 14163 and 14259, which are different by about a factor of two in their $^{22}Na$ and $^{26}Al$ activities, $^{36}Cl$ and $^{39}Ar$ were measured, too. Owing to the rather high K-content of approximately 4000 ppm of these samples practically all $^{39}Ar$ is produced via the $^{39}K (n,p)$-reaction, induced by neutrons with energies above 1 MeV. Flux values for "lunar" neutrons in this energy range will be given for both samples.
INITIAL FINDINGS OF A STUDY OF CHEMICAL COMPOSITION AND CRYSTAL FIELD SPECTRA OF SELECTED GRAINS FROM APOLLO 14 AND 15 ROCKS, GLASSES AND FINE FRACTIONS (less than 1 mm), P. M. Bell and H. K. Mao, Geophysical Laboratory, Carnegie Institution of Washington, Washington, D. C. 20008

Composition

Olivine phenocrysts in a section (#A-14306,6) of breccia range in composition widely between Fo70 and Fo90 (Table 1), in contrast with those from Apollo 11 and 12 rock samples which ranged between Fo70 and Fo75. However, the iron-poor olivines of this rock contain relatively high concentrations of Cr (approximately 0.3 weight per cent Cr2O3), attesting to their having crystallized under similarly low oxygen fugacities. These data suggest the existence of precursor rocks, which were lower in iron, and perhaps were more primitive differentiates than those of Apollo 11 and 12. Further data on selected crystals will be needed to confirm this interpretation.

Large, subhedral olivine phenocrysts from Apollo 15 rock 15555 are more Fe-rich than Apollo 11, 12, and 14 phenocrysts. Previous compositional measurements on lunar olivine phenocrysts have indicated normal chemical zoning of Fe and Mg, with Mg/Fe decreasing at the grain margins or remaining constant, but several grains in the section studied (#15555,37) are reversely zoned. Relatively Fe-rich cores (Fo40) of these abnormal phenocrysts grade into Mg-rich margins (Fo60). Other smaller olivine phenocrysts in the rock are euhedral and are not zoned. It is very likely that both types of olivine phenocrystal in this rock crystallized on or immediately below the liquidus, but the compositional differences would imply that the reversely zoned crystals were inherited from an earlier stage of the process. A process of adiabatic release of pressure of 1-10 kilobars, or of a temperature rise from the enthalpy of crystallization at the lunar surface, could have caused the observed zoning. Two stages of crystallization are implied--one for the large reversely zoned olivines, the other for the smaller euhedral ones.

Spectra

The crystal field spectra of Fe2+ in crystals of olivine and pyroxene from breccias, rocks, and fine fractions (less than 1 mm) of Apollo 12, 14, and 15 samples are typical of Fe2+ bands in octahedrally coordinated sites (Figures 1, 2, 3). However, unusual bands are observed at energies between 25,000 and 15,000 wave numbers. These are thought to be caused by Ti3+ and Fe3+. Although the calculated assignments do not provide unique solutions, they support the chemical evidence for the existence of these species in lunar crystals.

The glass particles which are colored green, yellow, and red, behave differently than the crystals, in that the Fe2+ bands are nearly identical, regardless of color. The colors are caused by strong absorption in the ultraviolet, the sloping edge of which converges toward the visible range with
variable intensity. The remarkably constant infrared absorption and the
effects of the ultraviolet absorption edge of these glasses may contribute
to the uniform radiative and albedo properties of the lunar surface.

Figure 4 shows absorption spectra for a nearly colorless glass spheroid,
low in both Fe and Ti. Figure 5 shows a green glass particle with a high Fe
content, and Figure 6 a red fragment with both high Fe and Ti. Intermediate
amounts of Ti cause brown colors.

The absorption coefficient for the strong Fe$^{2+}$ band centered slightly
above 1000 nm is plotted against weight per cent FeO for several glasses in
Figure 7. The correlation is sufficient to provide a good basis for spectral
interpretation.

Plotted in Figure 8 is the wavelength ($\lambda$) of the absorption minimum
between the Ti edge and the Fe band, versus the weight per cent TiO$_2$. The
scale factors were chosen so as to spread the vertical axis. The scatter is
related to the Ti$^{3+}$/Ti$^{4+}$ ratio; a limiting value for a lunar oxygen fugacity
being given approximately by the dashed line. A family of curves, each for
a fixed oxygen fugacity, could be superimposed if they were known. A line
close to the vertical axis would correspond to all Ti as Ti$^{3+}$. The estimated
range of oxygen pressures for these glasses is $10^{-7}$-$10^{-14}$ atmospheres.

Table 1. Olivine Crystals from Breccia 14306 (Analyses by electron microprobe)

<table>
<thead>
<tr>
<th>Weight percent</th>
<th>Grain Designations</th>
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<tr>
<td></td>
<td>15</td>
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<tr>
<td>FeO</td>
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<tr>
<td>MgO</td>
<td>48.46</td>
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<tr>
<td>SiO$_2$</td>
<td>42.19</td>
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<td>Cr$_2$O$_3$</td>
<td>0.40</td>
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<tr>
<td>Total</td>
<td>101.1</td>
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</tbody>
</table>

Figures 1-8. Optical values (see text for discussion).
ON THE CONTENT OF VOLATILE ELEMENTS IN MARE FECUNDITATIS, MARE TRANQUILLITATIS AND OCEANUS PROCELLARUM

Yu. I. Belyaev and T. A. Koveshnikova
USSR Academy of Sciences, Moscow

ABSTRACT

The new atomic-fluorescence method is described and data on mercury and thallium are reported for a number of core samples of regolith returned by Luna 16, Apollo 11 and 12 missions.

Atomizer is graphite crucible (furnace) with powdered sample heated by electric current in argon atmosphere. Electrode-less high-frequency discharge lamps are used as excitation sources. Analytical signal is amplitude of fluorescence impulse of mercury and thallium vapour distilled from the sample to be analysed. The limits of detection are 0.8 ppb for mercury and 3 ppb for thallium. The precision is 15% for 10 ppb of mercury and thallium.

Consideration regarding the volatilization of mercury from lunar surface, mercury migration during lunar day-time and the condensation of mercury during lunar night-time are presented.

The results of thallium determination are discussed.
CRYSTALLIZATION HISTORIES OF PYROXENES FROM LUNAR BASALTS; A. E. Bence and J. J. Papike, State Univ. of New York, Stony Brook, N. Y. 11790

Basaltic pyroxenes collected on the Apollo 11, 12, 14, 15, and Luna 16 missions have experienced a diverse range of crystallization histories as indicated by their chemical, crystallographic, morphological and paragenetic relationships. Although the final stages of lunar basalt crystallization appear to be rapid near-surface events, the initial stages vary considerably among the different basalt types. Differences in basalt bulk rock compositions, emplacement histories, and intensive parameters (T, P, fO2) are recorded in the paragenetic sequence of the basalt crystallization, the pyroxene chemical trends in the Di- Hed- En- Fs quadrilateral and in the Ti/Al relationships in the pyroxenes. Of the basalts considered, those from Apollo 11 have the highest TiO2/Al2O3 (weight %) ratios (1.06 for sample 10057). In these basalts the paragenetic sequence was Fe-Ti opaque oxides + olivine + subcalcic augite + feldspar. Primary pigeonite is rare in the Apollo 11 rocks but has been reported by Weill et al. (1970), Kushiro et al. (1970) and Bence and Papike (1971). The extremely high TiO2/Al2O3 ratio of the basalt, the coprecipitation of augite and feldspar, and fO2 at the time of crystallization resulted in pyroxenes with a Ti/Al (atomic) ratio of ½. This is the maximum ratio to be expected in the pyroxene crystal structure if all the Ti is present as Ti4+ and the pyroxene component that reflects this ratio is R2+TI4+Al2O6. Rock 12022 has a lower TiO2/Al2O3 ratio (0.57), lower total TiO2, and a paragenetic sequence olivine + augite + augite + plagioclase + ilmenite. Consequently, in this rock the earliest pyroxenes have slightly lower Ti/Al ratios (1/2 - 1/3). Ratios less than ½ can be interpreted in terms of additional components in the pyroxene structure, namely R2+AlSiAlO6 and R2+Cr3+SiAlO6. However, in 12022 pyroxenes, R2+Cr3+SiAlO6 is very minor and octahedral Al must be present. The Luna 16 basalts have TiO2/Al2O3 ≈ 0.36 and a paragenetic sequence in which pyroxene + plagioclase coprecipitated. Our results indicate a Ti/Al ratio for these pyroxenes of approximately 1/2 - 1/3 indicating both R2+TI4+Al2O6 and R2+AlSiAlO6 components, after the subtraction of R2+Cr3+SiAlO6. In rocks 12021 and 12052, the TiO2/Al2O3 ratios are 0.34 and 0.32, respectively, and the silicate paragenetic sequence is: pigeonite + pigeonite + augite + pigeonite + augite + plagioclase + pyroxferroite. The initial Ti/Al ratios in the pigeonites and augites are 1/6 - 1/8 indicating considerable octahedral Al as the component R2+AlSiAlO6. However, when plagioclase commences to precipitate, the Ti/Al ratio of the pyroxenes changes dramatically to approximately ½ (Figure 1). We have previously interpreted (Bence et al., 1970) these relationships as indicative of a two-stage crystallization model involving a first stage in which the early clinopyroxenes formed either at depth (pressure effect) or in a melt supersaturated with respect to plagioclase in which pressure was not necessarily a contributing factor. Apollo 14 rock 14053 has a lower TiO2/Al2O3 ratio (0.19) and higher total Al2O3 (13.6%) than both 12021 and 12052, yet the Ti/Al ratios in the pyroxenes are higher (1/3 in the earliest
crystals) and the pigeonites are much less aluminous. Except in the latest stages silicate parageneses in the three rocks are very similar. Metastable pyroxferroite is a late stage mineral in both 12021 and 12052 but in 14053 the assemblage ferroaugite + fayalite + silica occurs. The latter assemblage may indicate slower late-stage crystallization. The late-stage ferroaugites have Ti/Al ratios that exceed \( \frac{1}{3} \) and we interpret this as reflecting an additional pyroxene component, \( \text{R}^2\text{Ti}^3\text{SiAlO}_6 \). Presumably, the Ti\(^3+\) results from an extremely low \( f_{O_2} \) during the final stages of crystallization of 14053. Apollo 14 rock 14310 has the lowest TiO\(_2\)/Al\(_2\)O\(_3\) ratio yet observed for lunar basalts (0.06) and a very high Al\(_2\)O\(_3\) content (20.1%). The silicate paragenetic sequence is orthopyroxene + plagioclase > pigeonite + plagioclase > augite + plagioclase. Predictably, the orthopyroxene starts to crystallize simultaneously with plagioclase) with high initial Al and a low Ti/Al ratio (\( \sim 1/6 \)) which increases to \( \frac{1}{3} \) while Ti remains constant and Al decreases. When pigeonite commences to crystallize (at higher Fe/Fe\(^{+Mg}\)), the Ti/Al ratio increases due primarily to an increase in Ti and approaches \( \frac{1}{2} \) in the late-stage augites.

Apollo 15 rocks 15499 and 15058, which have virtually identical bulk compositions, are texturally very different which indicates dissimilar crystallization histories. Both are porphyritic and have clinopyroxene phenocrysts from 2-8 mm long but 15499 has a very fine-grained variolitic groundmass whereas 15058 has a coarser, subophitic matrix. Pigeonites in 15499 crystallized, initially, with Ti/Al ratios of \( \sim 1/6 \) but with continued precipitation and the incoming of augite, Al\(_2\)O\(_3\) increased to a maximum approaching 10 wt. % in the augite. Pigeonite cores in 15058 have Ti/Al ratios of \( \sim 1/3 \) and maximum Al\(_2\)O\(_3\) of \( \sim 2 \) wt. %. Plagioclase coprecipitated with augite before Al\(_2\)O\(_3\) enrichment in the clinopyroxenes could occur and Ti/Al ratios increased discontinuously to \( \frac{1}{2} \) (Fig. 1). The foregoing observations are summarized in Table 1 and Fig. 1.

We conclude, based on these observations, that the Ti/Al ratios in lunar pyroxenes from the Luna-16 basalts and Apollo basalts 10057, 12022, 14310 can be explained by the TiO\(_2\)/Al\(_2\)O\(_3\) ratios of the rocks, near-surface one-stage crystallization, and oxygen fugacities in the range 10\(^{-13}\) bars (Brown et al., 1970). On the other hand, Apollo rock 14053 contains pyroxenes with chemical trends that suggest one-stage near-surface crystallization under lower \( f_{O_2} \) conditions. Finally, Apollo rocks 12021, 12052, and 15499 contain pyroxenes with Ti/Al ratios lower than can be reconciled with the TiO\(_2\)/Al\(_2\)O\(_3\) ratios of the rocks and one-stage near-surface crystallization. We interpret these rocks to have a two-stage, two-location history with the first crystallization taking place at depth. The first pyroxenes are thus enriched in Al\(^{IV}\) as \( \text{R}^2\text{AlSiAlO}_6 \). The second stage is interpreted as rapid near-surface crystallization and the pyroxenes thus formed contain little octahedral aluminum.
Table 1

<table>
<thead>
<tr>
<th></th>
<th>10057</th>
<th>12022, 56</th>
<th>Luna 16</th>
<th>12021, 51</th>
<th>12052</th>
<th>14053</th>
<th>14310</th>
<th>15058</th>
<th>15559</th>
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<tr>
<td>SiO₂</td>
<td>39.79</td>
<td>43.20</td>
<td>43.8</td>
<td>47.05</td>
<td>46.6</td>
<td>46.3</td>
<td>47.19</td>
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<td>CaO</td>
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<td>9.56</td>
<td>10.40</td>
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<td>11.15</td>
<td>12.29</td>
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<tr>
<td>MgO</td>
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<td>10.43</td>
<td>7.05</td>
<td>7.08</td>
<td>8.14</td>
<td>8.72</td>
<td>7.67</td>
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<tr>
<td>TiO₂</td>
<td>11.14</td>
<td>5.16</td>
<td>4.90</td>
<td>3.74</td>
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<td>2.64</td>
<td>1.24</td>
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<tr>
<td>SrO₂</td>
<td>520 ppm</td>
<td>180 ppm</td>
<td>0.04</td>
<td>1.80 ppm</td>
<td>150 ppm</td>
<td>--</td>
<td>847 ppm</td>
<td>98 ppm</td>
<td>112 ppm</td>
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<tr>
<td>Cr₂O₃</td>
<td>2400 ppm</td>
<td>3800 ppm</td>
<td>0.28</td>
<td>2400 ppm</td>
<td>0.54</td>
<td>--</td>
<td>0.18</td>
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<td>0.25</td>
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<td>0.25</td>
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<td>K₂O</td>
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<td>Fe₂O₃</td>
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<td>0.067</td>
<td>0.12</td>
<td>0.49</td>
<td>0.03</td>
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<tr>
<td>S</td>
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<td>--</td>
<td>0.17</td>
<td>--</td>
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<td>0.14</td>
<td>0.02</td>
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<tr>
<td>P₂O₅</td>
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<td>0.13</td>
<td>511 ppm</td>
<td>0.09</td>
<td>0.083</td>
<td>0.09</td>
<td>0.34</td>
<td>0.08</td>
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</tr>
<tr>
<td>TiO₂/Al₂O₃</td>
<td>1.055</td>
<td>0.571</td>
<td>0.359</td>
<td>0.341</td>
<td>0.322</td>
<td>0.193</td>
<td>0.062</td>
<td>0.1995</td>
<td>0.1953</td>
</tr>
</tbody>
</table>
Cosmic-ray Track investigation has been performed on four fragmental rocks 14 301, 14 307, 14 311, 14 321, and one igneous rock, 14 310. In the course of this study feldspar detectors have been selected with regard to their relatively good track registration and/or revelation as compared with clinopyroxenes (1), and to their low fission noise level (1, 2, 3). 14 310, 50 - Station G (?).

The igneous rock 14 310 has been a closed system for $^{40}$Ar radiogenic and Rb-Sr since $3.9 \times 10^9$ y. ago, as appears from both Rb-Sr and $^{40}$Ar/$^{39}$Ar data (4, 5). This rock was brought up on the lunar surface some $262 \pm 7$ million years ago, as indicated by heavy noble gas isotopic studies (6), probably in the same event which exposed to cosmic radiation most of the rocks of Apollo 14, if we except those picked upon or near-by the Cone crater ejecta. During the main part of this time, the average shielding of this rock was about $100 \text{ g/cm}^2$ of matter, as inferred from neutron flux measurements (6, 7). Either this rock was buried, or most probably was part of a large boulder, as a clast. Recently, it was exhumed, or broken away from the boulder by impact (?), and brought within the range of track forming cosmic-ray particles. The track-density measured in center seated location in our sample, about 2 cm away from the nearest external surface and 4-5 cm away from the nearest pitted external surface, ranges from $1.6 \times 10^6 \text{ t/cm}^2$ to $2.9 \times 10^6 \text{ t/cm}^2$. This lowest track-density corresponds to a distance of 9 cm from the external surface of our slab. Due to the present uncertainty about the irradiation geometry of the rock, the "exposure age" at the very surface may be only bracketted
between 10 and 28 million years by using our St-Severin crude normalization. Our track-density profile observed in a direction perpendicular to one of the pitted surface of the rock presents a near surface gradient, related to solar-flare particles, quite similar to that observed by D.J. Barber et al. (8) in rock 12 022. Comparison with the track-density profile observed in a piece of Surveyor 3 glass exposed 2.6 y. on the lunar surface provides information on the long-term solar energy spectrum as well as on the lunar fine scale erosion rate (a few A per year). However, such an evaluation must be considered for the best as a semi-quantitative one, taking into account: a) the possible difference in the mean flux during the exposure period to solar flare particles between Surveyor (2.6 y.) and rock 14 310; b) the difference in track registration efficiency between optical glass and feldspars; the glass register from Z>10 (9), the feldspar from Z=16 (10); c) moreover, the fact that, due to its well known frequent twinning, feldspar is rather funny as a detector; in a single crystal section, individuals of the twinning may present distinctly different characteristics of track-revelation, corresponding to different values of the ratio of the etching rate along the tracks to the etching rate normal to the etched surface.

A much more simple irradiation history is displayed by the rock 14 321. Our slab which comes from a central location (although the distance to the outside surfaces is known with a poor accuracy) shows no evidence of decreasing density with depth. The track-density in feldspar, about $4 \pm 1 \times 10^6 \text{ t per cm}^2$, is three to four times that counted in olivine. Taking into account the fact that this rock was partly buried in the soil (11), and consequently assuming a long-term $2\pi$ geometry of irradiation, we infer from this track-density an "exposure age" of $23 \pm 5 \times 10^6 \text{ y.}$, in good agreement with that derived from spallogenic noble gas analysis, $20-25 \times 10^6 \text{ y.}$ (12). This exposure age is probably related to the impact that formed the Cone crater nearby which the rock was sampled.

Our sample of 14 311 shows track-densities of $2.4 \pm 0.5 \times 10^6 \text{ t/cm}^2$ for
TRACK DATA ON APOLLO 14 ROCKS
J.L. Berdot

A first fragment, of $1.8 \pm 0.5 \times 10^6$ t/cm$^2$ for a second one. However, we have unfortunately no information on the exact location of our sample.

14 301, 55 - Station G

As for our sample of 14 301, track-densities range between $3 \times 10^6$ t/cm$^2$ and $5.4 \times 10^6$ t/cm$^2$.

14 307, 30 - Station G

Due to difficulties in track revelation probably related to the fact that the material of our sample is rather heavily shocked, we are not presently in a position to give any track data on rock 14 307. However this breccia appears exceedingly interesting in the light of preliminary noble gas data (12) which show that the dark portion is greatly enriched in solar type gases, whereas the light fragments are at least very depleted, and show predominant spallogenic rare gases. In this respect, this rock appears very similar to the so-called gas-rich dark-light meteorites (or the Kapoeta type), which does not imply at all that Kapoeta comes from the Moon!!

References
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7. G.J. Wasserburg, personal communication.
10. T. Plieninger, personal communication.
Observations of fossil charged-particle tracks in principal rock forming minerals (feldspars, pyroxenes and olivines) up to depths of 10 cm in several lunar rocks are reported. Experimental data refer to both partial and total recordable track lengths. The latter information has been obtained using the TINT method for revealing a complete track, having both the track tips within the crystals. The TINT method was used to study the charge composition of nuclei within the iron group (V, Cr, Mn, Fe, Co, Ni). Incomplete tracks, i.e., those in which only one track tip lies within the crystal were studied for measuring the abundances of nuclei of Z > 30.

The track density of V.H. nuclei at low energies, up to 50 MeV/n, are governed by erosion and is in dynamic equilibrium with the track formation rate. Figure 1 and 2 in the accompanying paper illustrate this point where the surface exposure ages for a few sun-tanned rocks are shown as a function of depth. In Fig. 2 the track production rate obtained from Surveyor glass, up to 300 microns has been used. These figures amply demonstrate the existence of an erosion equilibrium at x < 0.1 cm. The observed spread in absolute track densities at 10⁻⁵-10⁻¹ cm depth can be understood mainly as being due to variations in the rock orientation during surface exposure.

Track densities at depths greater than 1 mm do not reach an erosion equilibrium. As an illustration Fig. 1 shows the observed track profile as well as that expected on the basis of energy spectrum given earlier for the case of rock 14303. Track densities higher than about 50 x 10⁶ cm⁻² were measured by the replication technique and are appropriately normalised to conform with the optically observed track density at 300 microns. The energy spectrum up to 1 BeV/n has been obtained using the large rocks 14321 and 14310. Although most of these rocks are fragmental, no evidence has been found for their significant pre-irradiation as individual fragments. The deduced energy spectrum of VH-nuclei agrees well with that obtained from Apollo 12 rocks.

A systematic study was also undertaken to examine variations in the distribution of the recordable track lengths as a function of shielding. For this purpose we selected rock 12038, which besides leaving a simple exposure history was minerallogically suitable. Crystals taken from 0-1 mm, 1-2 mm, 2-4 mm and 1 cm have been studied by using the TINT method. The track lengths group around 5, 8, 12 and 15 microns. The largest peak appears at 11-12 microns and is probably due to iron nuclei, although a definite charge assignment cannot be made at present. The majority of 15 micron tracks in samples where track densities are less than 5 x 10⁶ cm⁻² are believed to be fissiogenic, due to spontaneous fission of U²³⁸. A definite increase in the abundances of Mn, Cr nuclei relative to Fe nuclei...
occurring in deeper samples indicating fragmentation effects.

In Figure 2, we show similar track length distribution of TINTs in pyroxenes (sp, gr. 3.3 g cm\(^{-2}\)) for two cases of track densities, for \(\rho > 10^7\) and \(< 10^7\) cm\(^{-2}\). The observed grouping in recordable lengths are similar in characteristic to those shown in Figure 2. Furthermore the peaks are at the same lengths as reported earlier. With the present considerably increased statistics, the results give creditability to the assigned peak in recordable lengths at \(\sim 12\) microns as being due to iron nuclei and about 3 micron length difference per charge in the iron group region. Increased statistics in other meteorites confirm our earlier results and strengthen the application of TRIM/Trime methods to the study of the charge composition of cosmic rays and fissionogenic tracks in pyroxenes due to U\(^{238}\) and superheavy transuranic elements.

Fig. 1 The observed track densities in a through section of rock 14305 are shown for the feldspar crystals. XZ and XYZ refers to \(\phi\) 90\(^\circ\) and random orientations of the crystals. Track densities based on replication technique may be progressively larger at depths less than 200 microns. The solid line shows the expected track profile (arbitrary units) for a vertical and radial slice based on energy spectrum deduced from Apollo 12 rocks and St. Severin meteorite.
STUDY OF HEAVY COSMIC RAYS IN LUNAR SILICATES

N. Bhandari

References

4. ibid, 2599 (1971).

Fig 2. The TINT length distribution in Patwar Pyroxenes (sp. gr 3.3 g.cm$^{-3}$) estimated to be at a depth of 2 cm ($\rho > 10^7$ cm$^{-2}$) and 6 cm ($\rho < 10^7$ cm$^{-2}$).
Craision controlled radiation history of lunar regolith

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Fossil records of nuclear tracks are studied in Apollo 14 rocks, grain aggregates and grains in a manner essentially similar to that adopted for Apollo 12 samples. In this paper, we will primarily concern ourselves with the implications of these data to the denudation processes occurring on the lunar surface.

1. Erosion due to solar wind sputtering: The presence of appreciable track densities as well as large gradients in track densities up to depths of the order of 1 mm place a limit of the order of $10^{-3}$ cm/yr. for the solar wind erosion. This conclusion is based on observation of track profiles varying as $X^{-x}$, $x = 0.7-0.9$ for the depth interval $X = 10^{-3}-0.1$ cm, indicating that the observed track densities have reached an equilibrium with production and that the production spectrum varies as $X^{-2}$ which in turn corresponds to a slope of about 3 in the power law energy spectrum of VH-nuclei (5-50 MeV/n). The calculated effective times of irradiation at equilibrium for different erosion rates are shown in Fig. 1. These should be compared with the effective exposure times based on low energy spectra deduced from Surveyor glass.

2. Surface or sun-tan ages: In the case of rocks displaying large track gradients on a given face, we have calculated their sun-tan ages based on the observed track densities at 1 mm depth and an assumed orientation guided by the absolute track densities in the erosion equilibrium region. The sun-tan ages which vary between 0-7 m.y. for Apollo 11, 12 and 14 rocks appear to indicate a positive correlation with the size of the rock.

3. Burial or the sub-decimeter ages: The fact that the track profiles show different slopes in the 10-5 cm depth interval for different rocks, the variations being outside those expected from orientation, indicate that tracks accumulate in the rock interiors in the presence of chipping (and erosion). The calculated burial ages of the rocks vary between 0 and 20 m.y., following the procedure adopted by Bhandari et al. (1971).

4. Surface irradiation ages of fines: The "surface" irradiation ages of several Apollo 14 fines have been deduced from the observed track density frequency distribution in grains of < 1, 1-2 and 2-4 mm size range, using the earlier model developed and applied extensively for Apollo 12 scoop samples and double core. The calculations yield values of 60, 80, 20 and 7 m.y. for the "surface" exposure ages in the case of the bulk (14161,40; 14162,49; 14163,121), the contingency (14001,1; 14002,1; 14003,1), the comprehensive (14257,4; 14258,23; 14259,81) and the Station C Crater (14141,42) samples respectively. The Station C Crater, South of Cone Crater dates younger than the Cone Crater for which an exposure age of 20 m.y. has
been deduced, based on spallogenic neon. The "surface" exposure ages for the two Station G trench samples (14148.37 (0-1 cm) and 14151.14; 14152.6 (36 cm deep) are found to be 20 and 15 m.y. respectively.

We present a simple model to explain the observations discussed above (1-4). It seems inescapable that meteorite impacts control the track record in the depth interval 0.1 to few centimeters. We estimate average erosion rate to be of the order of $10^{-7}$ cm/yr; this one order of magnitude higher erosion rate is not fully effective in altering the track record at $X < 0.1$ cm purely because of statistics of cratering. At larger depths, yet another fragmentation process should be taken into account - this is descapitating and chipping from larger rocks. The results are discussed in relation to data on crater counts.

The comprehensive description and analysis of the documented Apollo 14 rocks shows that the typical burial depth of rocks is about 3 cm. Some of the rocks must have been exposed only partly on the surface when they were excavated in a cratering event, e.g., 20 m.y. ago in the case of the Cone Crater throw-out. However, it seems reasonable to assume that most were buried subsequently as a result of deposition considering that even the sub-decimeter exposure ages of rocks are usually small, of the order of 15-50 m.y., and the surface exposure ages of fines exhibiting a layer thickness of 1-5 cm are typically in the range of 10-50 m.y. If we assume that most of the rocks on the surface today were either excavated from deep or later on tumbled on the surface after a fragmentation of a larger rock, 10-20 m.y. ago, then the mean burial depth of 3 cm corresponds to an "average" deposition rate of $(0.3-0.15)$ cm. m.y$^{-1}$, quite in accord with the fossil track data based age for the Apollo 12 double core, 250-500 m.y. for the 60 cm thick deposit.

The above conclusions of collision controlled radiation history of lunar rocks apply satisfactorily to meteorites; the general similarities in the two cases are discussed.

Caption to Figure: Effective irradiation times for fossil tracks are plotted as a function of depth. Fig. 1 is theoretically calculated for different erosion rates whereas Fig. 2 shows observed exposure times for certain rocks based on Surveyor production rate.
COLLISION CONTROLLED RADIATION HISTORY OF LUNAR EGGGLITH

N. Bhandari

Fig. 1

Fig. 2

References
SOLAR WIND IMPLANTATION EFFECTS IN THE LUNAR REGOLITH

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O. Eugster, Physikalishes Institut, Universitat Bern - Bern.

Ultra-thin coatings have been observed by high voltage electron microscopy, on micron-sized grains extracted from various dust samples returned to the Earth by the Apollo 11, 12, 14, 15 and Luna 16 missions. The coated grains are markedly rounded and they are generally loaded with very high densities of tracks, or they contain high concentrations of very small crystallites. Selected area electron diffraction combined to thermal annealing and chemical etching experiments, indicate that the coatings are radiation damaged layers. Artificial implantations with low energy (0.2 ≤ E ≤ 5 keV/amu) helium, neon, argon, krypton and xenon ions, strongly support the hypothesis that these layers result from an "ancient" solar wind implantation in the grains, as they reproduce the rounded habits and the superficial coatings observed in the natural dust grains.

Although the artificial coatings represent a complex dynamic balance between the lost and the formation of amorphous material, their thicknesses roughly vary as E0.5 and do not depend on the atomic number of the ions, when E is expressed in keV/amu. Therefore it is possible to study the thermal properties of the ancient solar wind by analysing the distribution of the coating thicknesses in dust samples collected at various "young" and "old" locations in the lunar regolith, particularly in well stratified core tubes. A preliminary survey of such a distribution, based on 130 different "counts", suggests that the average energy of solar wind ions stays approximatively constant over periods of about 100 to 1000 years. This distribution also reveals a surprising high frequency of periods dominated by the "apparent" emission of low energy (0.2 ≤ E ≤ 1 keV/amu) ions, and a steep decrease in the frequency of the high energy periods (E > 1 keV/amu), when the solar wind energy increases (figure 1).
The albedo of soil samples from all missions was studied as a function of several characteristics of their finest grains (400 Mesh residues), which most likely cover the coarser ones, and therefore should dominate the optical properties of the regolith. The albedo of the samples was found to be strongly correlated to their proportions in coated grains, but not to their glass contents or to their concentrations in strongly absorbing ions. Furthermore the computed reflecting power of a simple multielectric structure, constituted of a semi-infinite slab of crystalline matter with an index of refraction $n_1$, coated with a metamictized layer of index $n_2 < n_1$, and thickness $\lambda \sim 350 \text{ Å}$, qualitatively reproduces the spectral variation of lunar albedo, characterized by an increase in albedo at short and long wavelengths (figure 2). Furthermore, a very slight chemical etching of the dust grains, that only removes their ultra-thin coatings, considerably increases the albedo of sample 10084 and modify its spectral variations. Therefore the albedo of lunar soil samples certainly depends on their irradiation history in the solar wind, and is perhaps slightly "modulated" either by metamorphic events occurring subsequently in the regolith and modifying the optical properties of the coated grains, or by the mineralogical composition of the Lunar dust, as shown by the beautiful Apollo 15 X-Ray fluorescence experiment (1).

The ultra-thin coatings, considered as "markers" for the thermal properties of the ancient solar wind, have various other implications concerning for example: elemental and isotopic abundances in the solar corona, the origin of indigenous lunar carbon and methane (2), the nature and time constant of various lunar dynamic processes and ancient lunar atmospheres (3).

References
FIGURE 1

FIGURE 2
The olivine-poor basalt, 14053, has Mg/Mg + Fe ~ 0.48 (contrast ~0.40 for near cotectic Apollo 12 basalts) and represents a more primitive, less differentiated type than previous mare basalts. After opaque spinel, olivine is the liquidus silicate (1230°C), plagioclase beginning to crystallize at c. 1190°C when the charge is c. 5% crystalline, pyroxene and an opaque phase also being present 15°C lower when the charge is perhaps 20% crystalline. Using glass (1400°C) as a starting material the entries of olivine and plagioclase were c. 109°C lower, but that of pyroxene higher. M is a more olivine-rich composition (Mg/Mg + Fe ~ 0.52), and shows substantial olivine crystallization prior to plagioclase entry, which occurs at a lower temperature than in 14053. The relationship between Mg/Mg + Fe, and cotectic crystallization temperatures in Apollo 11, 12 and 14 basalts is summarised in figure 1. 14053 is not conspicuously close to a low pressure 4-phase cotectic liquid composition, and differs from previous mare surface basalts in having plagioclase as second silicate (cf. Apollo 15 samples). It may be an earlier stage liquid, suitable as a parent for a fractionation trend leading towards Apollo 12 lavas. Owing to its relatively low Mg/Mg + Fe it cannot be a primary magma derived from a mantle which had earlier yielded magmas of high Mg/Fe + Mg (e.g. majority of Apollo 14 samples inc. 14310).

Fig. 1
Interpretation of phase relations in 14321 (breccia) and 14162 (soil) can be attempted via at least three naive assumptions; that the samples represent (1) average lunar crust (2) average parental magma of some intracrustal pre-Imbrian igneous series (3) some linear and meaningful mixture of crust and uppermost mantle from the site of Mare Imbrium.

**Fig. 2**

Note expansion of divinite crystallization and suppression of plagioclase and orthopyroxene crystallization by increasing $P$-$H_2O$ (general result, observed in 14330 and many synthetic and natural systems)
14321 shows extensive olivine and spinel precipitation before plagioclase entry at 1 atmosphere. Increase of pressure suppresses the olivine liquidus, replacing it first by orthopyroxene, then by garnet (fig. 2). Under water vapour pressures of c. 1 kb, spinel replaces olivine as liquidus phase. Whichever assumption is adopted, an ultimate uppermost lunar mantle, of Mg/Mg + Fe > 0.87 would be indicated. On assumptions (1) or (2) the magma would either have to be directly derived from great depth (>160 kms if orthopyroxene is to be a residual phase in the lunar mantle) or represent an advanced partial melt of its source rock leaving olivine as the sole residuum. Such considerations are, however, academic because the 4.26 gm fragment, 14321, is probably not representative of 14321 as a whole; and because the breccias are known to contain clasts of feldspar-phyric basalts such as 14310, 14073, which in turn are known to have suffered alkali loss after their extrusion and before incorporation in the breccias. Data on 14321 cannot be meaningfully interpreted by any of the three assumptions.

14162 has a greater claim to be regarded as a fair average of Fra Mauro material, yet it must be contaminated by mare-type basalts (Reid 1971) and must contain a contribution from rocks which suffered alkali loss after reaching the lunar surface. Data for this sample are equally unsuitable for interpretation by hypotheses (1) and (3), but might lend themselves to approximate interpretation by hypothesis (2). At 1 atm., plagioclase is the liquidus phase, rapidly followed by olivine. As a near cotectic liquid in equilibrium with olivine and plagioclase at 1 atm. it could serve as a low pressure parental liquid for olivine-rich precipitates, plagioclase remaining preferentially in suspension to yield feldspar-phyric basalts. Remixing of these two components with a bias towards the olivine component could then yield a mixture with the properties of 14321. However, feldspar-phyric basalt 14310 contains much less phenocryst plagioclase than would be required to explain the plagioclase enrichment relative to the cotectic liquid, or to 14162, hence this interpretation presupposes extensive resorption of phenocrysts against which there is definite evidence in the results of Brown and Peckett (1971).

We conclude, therefore, that interpretation of the relationships between samples 14162, 14321 and the lunar crust and mantle requires more sophisticated assumptions than are entertained here.

Reexamination was aimed at resolving inter-laboratory discrepancies in data and interpretation of Apollo 12 basalts. Plagioclase remains in 12018, 12020, 12038 and 12040 at 1153°C in 3-day runs whether rock powder or glass is used as starting material. Reversal of plagioclase entry by holding 4-5 hours at upper temperature, then 5-16 hours at lower, has been accomplished on 8 samples (figure 1). Cotectic liquid compositions obtained from glass or rock powder were very similar in 12038, 12040. Temperatures are calibrated at m.p. of gold 1064°C and Li2SiO3 (12030°C).

Fig. 1. Comparative results for plagioclase entry

Calculation, from analysed glass and crystal phase composition, of percentage crystals present at plagioclase entry yields values as low as one half of those quoted by other workers relying on visual estimation. Calculation of the putative residual cotectic liquids, by extracting the visually estimated amount of crystals, yields liquids which are enriched in potential plagioclase, and depleted in potential hypersthene relative to the analysed liquids (figure 2); substantial negative MgO, and some normative nepheline (in 'presence' of pigeonite crystals) appear in some of these calculations.
Fig. 2. Diopside projection contrasting analysed and calculated cotectic liquids

The composition of the basaltic magma supplied to the surface at the Apollo 12 site is best sampled by glass and lithic fragments in the soil, or by the bulk soil; or even by averaging the total weight of rocks distributed for chemical analysis (biased towards extremes? N.B. also that glass analyses will not include iron reduced by impact heating in vacuum). These compositions resemble the analysed cotectic liquids but differ from phenocryst-enriched mixtures like 12040 (figure 3). K9, K10 are impact produced glass fragments as rich in potential olivine as the phenocryst-enriched rocks. The glassy sample 12009 shares the thermal history (rapid chilling from > 1200°C) of the impact glasses, but differs from that of most other rock samples (slower cooling from c. 1170°C-1150°C). 12009 may represent a splash of impact remelted phenocryst-enriched lava.

Fig. 3. Diopside projection comparing average rocks and cotectic liquids

The probability of the observed close approach of erupted lava compositions to low pressure cotectic liquid compositions being due to coincidence with a higher pressure primary magma composition is \( \approx 3 \) in 1000; in Apollo 11 ophitic group it is \( \approx 5 \) in 10,000, but reaches \( \approx 3 \) in 100 in the intersertal group. It is more probable that crystal-liquid fractionation at low pressures has controlled the observed lava compositions.
Prinz et al. (1971 Contrib, Mineral, Petrol. 32, 211-230) reported average composition for Apollo 11 basalts based on analyses of 21 low potassium and 18 high potassium lithic fragments in the soil. These are significantly lower in Ti, Fe, Mn and higher in Si, Al, than averages of analysed rock specimens, and accordingly much closer to the experimentally produced cotectic liquid compositions (cf. Apollo 12 lithic fragments, fig. 3).

Selected figures for comparison are:

<table>
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<tr>
<th></th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>FeO</th>
<th>MgO</th>
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<td>Cotectic Liquid (Biggar et al. 1970; average of 2)</td>
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<td>17.7</td>
<td>6.5</td>
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</tbody>
</table>

The larger rock samples may have been selectively derived from the base of a flow into which ilmenite had accumulated. The average soil is yet more feldspathic than these averages and may reflect the overlying and complementary pole of the differentiation, now fragmented to form regolith.

**Fig. 4**

SiO₂–TiO₂ plot of phase data for synthetic and natural Apollo 11 compositions.

\[ ^{87}\text{Rb} - ^{87}\text{Sr} \text{ Ages Determinations: } \]
The results for the soils returned by Apollo 14-Apollo 15 and Luna 16 missions fall on an isochron at about 4.6 b.y. (\( \alpha = 1.39 \times 10^{-11} \text{ y}^{-1} \)). The Luna 16 soil is the least radiogenic and the Apollo 14 soil the most radiogenic. These constant model age for the soils has been explained by a dominant Rb rich component. Notwithstanding its low average Rb concentration the Luna 16 soil also has a Rb-Sr model age of 4.6 b.y. We have investigated fragments from these soils. An anorthositic fragment also has a 4.6 Rb-Sr model age. The anorthositic nature of this fragment is well supported by our trace element data. A 4 mg basalt fragment was separated into three phases, one of which (\( \approx 10 \) mg) is extremely radiogenic (\( ^{87}\text{Sr}/^{86}\text{Sr} = 0.951, ^{87}\text{Rb}/^{87}\text{Rb} = 5.2 \)). The internal isochron of these basalt fragments is 3.6 \pm 0.2 b.y. We therefore believe that the 4.6 b.y. model age is not created only by the Rb rich component.

U-Th-Pb Measurements: Luna 16 and Apollo 14 measurements given for [206/204, 207/206] plot a age of 4.75 b.y. The Th-U-Pb apparent ages are discordants. The U and Th concentrations are the lowest in Luna 16 soils and highest in Apollo 14 soils in the same trend as the radiogenic character.

Lead Isotope compositions have been studied with a Cameca ionic microprobe utilizing two techniques: (1) chemical dissolution and electroplating on a gold disk and (2) directly on polished rock covered by gold grid using \( ^{2} \text{He} \)-beam.

Using mode (1) we found in a 1 mg sample of Luna 16 soils a more radiogenic lead than that in a 40 mg sample previously analyzed by conventional method. The [Pb-Pb] age is the same. A basaltic fragment (2 mg) analyzed is very radiogenic in lead with 207/206 = 0.45, 206/208 = 1 in close agreement with Rb-Sr age determinations.

Using mode (2) on a low K Apollo 11 rock we found a heterogeneity in 207/206 ratio. A Ca-U phosphate phase gives a 207/206 = 0.44 and a K rich U poor interstitial phase gives a 207/206 = 1.1, 206/208 = 0.7.

In accord with the prediction of Tatsumoto and the results of Wasserburg and all these data support a very high U/Pb ratio in the parent rocks of the mare basalts.

K-Rb-Sr-Ba and R-E Concentrations: Using isotope dilution mass spectrometry technique we have studied Apollo 14 and 15 and Luna 16 soils. Luna 16 soil is the poorest in R-E and K and has the smallest negative Eu anomaly. Apollo 14 soil (14259) is the highest in R-E and K and has the largest negative Eu anomalies. Apollo 15 soil (15021) has an intermediate concentration but a big Eu anomaly in agreement with the low Sr content. [K-Rb] and [K-Ba] plots show a good correlation between these elements for all soils.
Luna 16 microbasalt show a RE-K-Sr-Ba concentrations very close with the Apollo 11 lunar basalt. The 4.6 b.y. anorthositic fragments from the Luna 16 soil has a very big positive Eu anomaly and very low RE-Ba-Rb-K content.

Ionic probe studied were made to elucidate the trace elements distributions of K-Rb-Sr-Ba-E in low K Apollo 11 basalt. Direct picture distributions show the high concentration of K-Rb-Ba in a K rich intérstitial phase. Ba/Ce varies between Plag (6) K rich phase (2) Pyrox (0.1).

All of these data are compatible with a multicomponent mixing model for soil formation with:
1. An anorthositic rich component formed at 4.6 b.y.
2. A basaltic component of the same age remelted without Rb-Sr fractionation between 3.1-4. b.y.
3. A KREEP component formed later may be 4.2 to 4.4 b.y.

The parent rocks of those components have a structural position between 1. and 2. in the Moon and sometime contaminated the lavas flow originated from component 2.

References
AGE AND ORIGIN OF LUNAR SOILS

J. L. Birck

P. 82

**TABLE 1**

<table>
<thead>
<tr>
<th></th>
<th>Apollo 14 (20 mg)</th>
<th>Apollo 15 (30 mg)</th>
<th>Luna 16 Total Soil (20 mg)</th>
<th>Luna 16 Microbasalts (1 mg)</th>
<th>Luna 16 Anorthosite (2 mg)</th>
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**TABLE 2**

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<td></td>
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<td>0.293</td>
</tr>
<tr>
<td>Th</td>
<td></td>
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</tr>
<tr>
<td>Pb</td>
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<td>1.07 ppm</td>
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</tbody>
</table>
THERMAL RADIATION PROPERTIES OF APOLLO 14 FINES

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University of Kentucky
Lexington, Kentucky

and

James P. Dawson
Consultant
Oklahoma City

The thermal radiation properties of lunar fines sample 14163 are presented for a wavelength range from 0.36 to 14.75 μm. The spectral directional reflectance was measured between the wavelengths of 0.36 and 2.2 μm and the spectral normal emittance from 2 to 14.75 μm. These parameters were obtained for four densities from 1100 to 1600 kg/m³ and the reflectance for angles of elimination from 15 to 60 degrees.

Results

The spectral reflectance for a bulk density of 1600 kg/m³ and for a wavelength range from 0.36 to 2 μm is shown in Fig. 1. There are several prominent absorption bands in this portion of the spectrum. One band which has been found in almost every sample is that centered at approximately 0.95 μm. Other bands appear at 1.3 and 1.8 μm.

The visible appearance of this sample suggests a high reflectance. Compared with the Apollo 11 and 12 spectral reflectance results, the Apollo 14 fines reflectance is almost twice as large and this is no doubt connected with its composition. At a wavelength of 0.6 μm and a bulk density of 1600 kg/m³, the reflectance of sample 10084 was 0.082, for sample 12070, 0.10 while for 14163, 0.18. For a wavelength of 2 μm the reflectances were 0.186, 0.23 and 0.355, respectively.

The spectral emittance as a function of wavelength and bulk density of 1600 kg/m³ is shown in Fig. 2. For the wavelength range used in this study there appears to be one absorption band near 6 μm and at the Christiansen frequency (wavelength) between 8 and 9 μm; the region in the spectrum where minimum internal scattering takes place. The minimum emittance occurs near 3 μm and is approximately 0.60 and the maximum near 8.25 μm of approximately 0.97. Apollo 12 fines had values of 0.73 and 0.99 for the same wavelengths, respectively. In general, the spectral emittance of Apollo 14 fines is lower than those of Apollo 12.
From these spectral reflectance and emittance results the solar albedo and total normal emittance have been calculated. The solar albedo for an angle of illumination of 15 degrees is 0.213. Those for Apollo 11 and 12 were 0.099 and 0.119, respectively, for the same angle of illumination. The total normal emittance as a function of temperature is given by

$$e(T) = 0.9696 + 0.9664 \times 10^{-4} T - 0.31674 \times 10^{-6} T^2 - 0.50691 \times 10^{-9} T^3$$

for a temperature range from 80 to 4200 K.

![Fig. 1 Spectral Directional Reflectance of Apollo 14 Fines.](image)
Thermal Radiation Properties
Richard C. Birkebak

Fig. 2 Spectral Emittance of Apollo 14 Fines

Apollo 14163
Density = 1600 kg/m³
Angle of Viewing = 10°
○ Reflectance Measurements

Wavelength (μm)
LOW TEMPERATURE THERMOLUMINESCENCE OF APOLLO 14 LUNAR SAMPLES

I. M. Blair, A.E.R.E. Harwell, U.K.
J. A. Edgington, and R. Chen* Queen Mary College, London, U.K.
R. A. Jahn, Lanchester Polytechnic, Coventry, U.K.

We have studied the thermoluminescence (TL) of the lunar samples returned by the Apollo 14 mission as listed in Table I. The irradiations were performed at -196°C in the 160 MeV proton beam from the Harwell synchrocyclotron using a dose rate of about 140 rads per second. The heating rate was about 0.5°C per second, and TL was observed in the 435-485 nm waveband. The general feature was a broad hump extending from -50°C to +150°C, similar in nature to that observed for the Apollo 12 samples. As the magnitude was about twice as great as that of the latter, we were able to study in detail both the isothermal decay and the growth with increasing dose of the component peaks of the hump.

The isothermal decay was studied by starting the heating at different times after repeated equal irradiations. Neglecting re-trapping effects the response (R) is related to the response for zero delay (R₀) and the delay time (t) by the following expression:

\[ R = R_0 \exp(-\alpha t) \]

where \( \alpha \) is the trap leakage rate. The growth with increasing dose was studied by varying the irradiation time. Making the same assumption as above and assuming also that no new traps are produced during irradiation, the response (R) is related to its saturation value (Rₛ) and the irradiation (t) by the following expression:

\[ R = R_s [1 - \exp(-\gamma t)] \]

where \( \gamma = \alpha + \beta \), and \( \beta \) is the trap excitation rate. From our data we derive values of \( \alpha \) and \( \beta \) for each of the component peaks observed. These are listed in Table II. The values of \( \beta \) in units of rads⁻¹ (\( \beta' \)) are also given. In Figure 1 we show a typical TL curve for one of our samples, and in Figure 2 the growth and decay curves for one of the peaks observed from this sample.
Although the magnitude of the TL is about twice that observed from Apollo 12 samples it is still too small, by orders of magnitude, to provide a credible energy storage mechanism required to explain the Transient Lunar Phenomena (TLPs). Furthermore, the disturbingly large values of $\mu_G$, corresponding to mean leakage times of the order of an hour, indicate that the traps are incapable of storing energy for the required times of the order of $10^6$ years. However, it has been suggested that rather than being purely exponential, the decay also has a constant term. We require more data to check this hypothesis.

**Table I**

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<th>Harwell No</th>
<th>NASA No</th>
<th>Type</th>
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<td>Dark fraction of crushed rock</td>
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<tr>
<td>M 410a</td>
<td>14321, 147</td>
<td>Light fraction of crushed rock</td>
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<tr>
<td>M 411</td>
<td>14163, 113</td>
<td>&lt;1 mm. bulk fines</td>
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</table>

**Table II**

<table>
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<tr>
<th>Sample No.</th>
<th>Peak Temps. °C</th>
<th>$\alpha$ sec$^{-1}$ (x10$^{-4}$)</th>
<th>$\gamma$ sec$^{-1}$ (x10$^{-3}$)</th>
<th>$\beta$ sec$^{-1}$ (x10$^{-3}$)</th>
<th>$\beta'$ rad$^{-1}$ (x10$^{-5}$)</th>
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<td>0.33</td>
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<td>+ 85</td>
<td>0.33</td>
<td>0.36</td>
<td>0.33</td>
<td>0.24</td>
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<tr>
<td>M 410a</td>
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<tr>
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<td>0.37</td>
<td>0.35</td>
<td>0.25</td>
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<tr>
<td>M 411</td>
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<tr>
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<td>+ 90</td>
<td>0.67</td>
<td>0.35</td>
<td>0.28</td>
<td>0.20</td>
</tr>
</tbody>
</table>

*On attachment from the Department of Physics & Astronomy, Tel-Aviv University, Tel-Aviv, Israel.*
FIGURE 1  THERMOLUMINESCENCE OF M409a

435-485 nm WAVEBAND

FIGURE 2  GROWTH AND DECAY CURVES FOR PEAK 3, SAMPLE M409a
NOBLE GAS STUDIES ON REGOLITH MATERIALS FROM APOLLO 14 AND 15. D. D. Bogard and L. E. Nyquist, NASA Manned Spacecraft Center, Houston, TX 77058

Abundances and isotopic compositions of the five noble gases, He, Ne, Ar, Kr, and Xe, have been determined in more than twenty Apollo 14 and 15 lunar soils and breccia. In addition to analyses on bulk samples, we have analyzed portions of breccia, samples from a range of depths in a trench and a core tube, and grain size separates and stepwise temperature releases of fines 14163-14161.

Noble gas abundance patterns in Apollo 14 and 15 fines are dominated by a solar wind component and are similar to those observed for earlier lunar missions. Concentrations of Ne-20 for Apollo 14 fines range 93-121x10^-5 cc STP/g, and for Apollo 15 fines, 18-145x10^-5 cc STP/g. Other trapped solar gases show similar concentration ranges, except that He4/Ne2O abundance ratios vary up to a factor of two. Three different depths of Apollo 14 trench material possess essentially identical noble gas concentrations, which are similar to concentrations in the 60-90 micron size fraction of 14163 and several other fines with nearly equal mean grain sizes. Three different depths in core 14230 exhibit differences in gas concentrations ranging from a factor of 1.5 for He-4 to a factor of 6 for Xe-132.

Four Apollo 15 breccia exhibit solar wind noble gases in concentrations similar to lunar fines (Ne-20 = 50-120 x 10^-5 cc STP/g). Eleven Apollo 14 breccia exhibit much larger variations in noble gas contents (Ne-20 = 0.04-100 x 10^-5 cc STP/g), and the concentrations of solar wind gases in various portions of 14307 varies by two orders of magnitude. For several of these rocks cosmic ray spallation-produced noble gases are a major component. Two rocks recovered near the Lunar Module and three recovered near Cone Crater (14305, 066, 063, 068, and 321) show essentially the same Ne-21, Ar-38, and Xe-128 exposure times of about 20 x 10^6 yrs., which may reflect the age of Cone Crater. Two additional rocks (14006 and 307) show substantially larger concentrations of spallation isotopes, and presumably have had a different cratering origin. The radiogenic gas contents of these solar wind depleted breccia indicate that, if these rocks ever contained appreciable amounts of solar gases, any major degassing event must have occurred considerably prior to their explosion to the lunar surface and initiation of cosmic ray induced reactions. A definite relationship exists between the noble gas contents of these Apollo 14 breccia and a metamorphism scale based on abundance and type of glass (J. Warner, presented at 3rd Lunar Science Conference), with rocks exhibiting increasing evidence of metamorphism also possessing the lowest noble gas contents.

Concentrations of spallation produced Xe-126 and Kr-80 were determined for ten Apollo 15 fines and gas rich breccia and eleven Apollo 14 fines and gas-rich breccia. Samples from both sites exhibit variations in these components.
Noble gas studies on regolith materials....
D. D. Bogard

by factors of 3-10, which apparently cannot be attributed to variations in target element abundances, and probably represents appreciable differences in cosmic ray exposure times for these samples. Spallation Xe-126 and Kr-80 abundances in size separates of fines 14163-14161 exhibits an increase with decreasing grain size and suggest the existence of an order of magnitude difference in exposure times between 3,000 and 30 micron particles.

Noble gas analyses of grain-size separates (20 to 4,000 microns) and stepwise temperature releases were performed for fines 14163: In several important respects these two experimental methods give results which supplement each other, a test which, to our knowledge, has not been previously demonstrated. As noted by several previous investigators, gas concentrations show an inverse correlation to grain size. Smaller grain sizes exhibit somewhat lower gas concentrations per unit surface area, and a greater dependency of gas content on grain size exists for the lighter noble gases than for the heavier (i.e., He-4/Ar-36 ratio is larger for smaller grain size). Several fines and breccia show Xe-132/Ar-36 and Kr-84/Ar-36 ratios considerably lower than previously observed, even for ilmenite separates, and establish new lower limits for these elemental ratios in the solar wind of Xe-132/Ar-36 \(3.7 \times 10^{-5}\) and Kr-84/Ar-36 \(2.5 \times 10^{-5}\).

Stepwise temperature release and grain size data define mean isotopic ratios for the trapped component of Ne-20/Ne-22 = 12.65±0.10, Ne-21/Ne-22 = 0.032±0.001 and Ar-30/Ar-38 = 5.35±0.02, in good agreement with values determined for Apollo 11 and 12 fines. Trapped He-4/He-3 ratios are variable, which possibly arises from variations in the solar wind He-4/He-3. The Ar-40/Ar-30 ratios in fines and breccia are quite variable, ranging from 0.75 to 3.1 for bulk analyses, and reflect relative variations in implanted lunar atmosphere Ar-40. He, Ne, and Ar all show a minor low temperature release component which is enriched in the lighter isotope over the heavier. Higher temperature release data, the larger grain size separates, glass from 15923, and a few breccia all show the presence of a neon component which is distinct from spallation produced neon measured in meteorites and lunar basalts. This component was previously seen in high temperature release data and fines 12032 and 12033.

Low temperature krypton release of 14163 also shows characteristic enrichment of lighter isotopes observed previously by other workers, but to a much greater extent. The isotopic composition of krypton in the 30 micron fraction is, within analytical error, identical to this low temperature component. In this grain size the enrichment of Kr-84/Kr-86, Kr-83/Kr-86, and Kr-82/Kr-86 with respect to the trapped composition is 9%, 18%, and 22%, respectively, and for this component we derive Kr-82/Kr-83=1.3 and Kr-84/Kr-83=2.6. These large enrichments seem to exclude the possibility that this component results from the isotopic fractionation of solar wind krypton. Higher temperature release and coarser size separates yield data consistent with the trapped solar plus spallation Kr observed in Apollo 12 material. Xenon grain size and temperature release data is also consistent with solar plus spallation components, except that the 30 micron spallation yields are somewhat anomalous.

Isotopic data for Kr and Xe for grain size separates and temperature releases of fines 14163 permit precise determination of a spallation component.
The derived isotopic composition of this component is essentially identical for the two methods and is:

<table>
<thead>
<tr>
<th>Element</th>
<th>Isotopes</th>
<th>Abundances</th>
</tr>
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<tbody>
<tr>
<td>Kr</td>
<td>80, 81, 82</td>
<td>0.21, 0.775, 0.47</td>
</tr>
<tr>
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<td>83</td>
<td>1.00</td>
</tr>
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</table>

<table>
<thead>
<tr>
<th>Element</th>
<th>Isotopes</th>
<th>Abundances</th>
</tr>
</thead>
<tbody>
<tr>
<td>Xe</td>
<td>124, 126, 128</td>
<td>0.53, 1.00, 1.63</td>
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<tr>
<td></td>
<td>129, 130, 131</td>
<td>2.11, 1.04, 5.55</td>
</tr>
<tr>
<td></td>
<td>(132)</td>
<td>0.97, 0.13, (136)</td>
</tr>
</tbody>
</table>

This spallation component is similar to those previously derived for several lunar basalts and suggests that no major energy differences exist in the component between many bulk lunar soils and rocks. Again the fact that nearly identical results are obtained from grain size separates and temperature release data substantiates use of the latter technique.

A correlation plot of Xe-124/Xe-136 versus Xe-126/Xe-136 for a large number of Apollo 14 and 15 bulk fines and breccia, as well as grain size and temperature release data of 14163, defines a straight line of slope 0.53, and demonstrates that most of these samples possess spallation Xe-124/Xe-126 ratios of 0.52-0.65, and indicates that these particular lunar samples have not experienced major energy differences in the higher energy cosmic ray component which produces the light spallation isotopes of Xe. However, a Xe-131/Xe-136 correlation plot for these same fines and breccia shows large variations in the spallation Xe-131/Xe-126, ranging from 3 to 8. These values extend considerably higher than typical meteorite values of 2.5-4.5, but cover the same range of 131/126 spallation ratios measured by many laboratories in lunar basalts. If, as has been suggested previously, the variable excesses of Xe-131 are due to resonance capture of neutrons by Ba, the Xe-131/Xe-136 versus Xe-126/Xe-136 correlation plot allows a quantitative determination of the integrated cosmic ray exposure through the excess Xe-126 and a qualitative determination of the integrated energy of that exposure through the excess Xe-131. Thus, fines 15021 and breccia 15498 and 14307 possess excess Xe-131/Xe-126=8, and should exhibit considerable evidence of low energy neutron fluxes. Three different depths of the Apollo 14 trench sample also possess variable excesses of both Xe-126 and Xe-131. The topmost layer has received a greater integrated cosmic ray dosage and has a higher "spallation" Xe-131/Xe-126 ratio, implying a greater fraction of its exposure lifetime was spent in a more neutron rich (i.e., more shielded) environment than deeper trench samples. Core 14230 also shows small differences in this spallation ratio, along with much larger differences in spallation Xe-126 abundances.
SEARCH FOR LOW ENERGY (10 \( \leq E \leq \) 300 keV/amu) NUCLEI IN SPACE: EVIDENCE FROM TRACK AND ELECTRON DIFFRACTION STUDIES IN LUNAR DUST GRAINS AND IN SURVEYOR III MATERIAL.

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P. Peter, Laboratoire de Chimie Nucléaire, 1 NZ P3 31 Toulouse.

There is evidence based on the recent work of space "probists" for the existence of an intense flux of low energy nuclear particles in space (1), with energy smaller than 300 keV/amu. Furthermore in 1970, "trackists" at Toulouse discovered very high track densities exceeding \( \times 10^{10} \) tracks/cm\(^2\), in micron-sized lunar dust grains observed directly with a high voltage electron microscope, and they attributed these tracks to low energy solar nuclei (2). In this short paper we describe several new methods, based on track and lattice disorder observations, and developed to ascertain or disprove the existence of these nuclei in the "ancient" solar radiation.

We study tracks by using combined 1 MeV, 100 keV and scanning electron microscopies. With the 1 MeV microscope we are attempting: 1. to determine directly the track length distribution in the finest lunar dust grains, by applying a new stereoview technique; 2. to search for steep superficial track density gradients, in ultra-thin sections obtained by cutting 200 Mesh lunar grains, as well as flakes from the protective paint from the Surveyor III Spacecraft, with an ultramicrotome - such gradients constitute a characteristic "signature" for low energy ions (\( E \leq 0.5 \) MeV/amu), with short penetration in matter; 3. to establish any possible correlation between the proportion of micron-sized dust grains with track density \( \rho > 10^{11} \) tracks/cm\(^2\), and that of grains coated with very thin layers of amorphous material, produced by an "ancient" solar wind implantation (3). With the 100 keV microscope we are examining the replica of very slightly etched sections of various materials, including 200 Mesh lunar dust grains and Surveyor paint flakes, in view of also detecting very steep superficial track gradients. Finally, with the scanning electron microscope, we particularly measure the proportion of 200 Mesh grains, with \( \rho \geq 10^{9} \) tracks/cm\(^2\) at the center of the grains (this quantity essentially
depends on the irradiation of the grains in solar flare cosmic rays). The results so far obtained are: 1. the ultra-thin sections of the Surveyor paint are well adapted for a high voltage electron microscope observation of latent tracks, as the paint is made of 100% of mineral grains, with \( \approx 10\% \) of the grains giving very good dark field micrographs; 2. the observation of slightly etched external flakes from the Surveyor paint, give evidence for very high densities (\( \approx 5 \times 10^9 \) tracks/cm\(^2\)) of short tracks, registered at depths \( \approx 1 \) micron in the surface exposed to the Sun; 3. preliminary measurements - subject to considerable revision - of the track lengths in micron-sized lunar dust grains, indicate that \( \approx 50\% \) of the tracks are shorter than the thicknesses of the grains; 4. in a given dust sample, the proportion of micron-sized grains with \( \rho \approx 10^{11} \) tracks/cm\(^2\), seems to be only correlated to that of coated grains and not to the proportion of 200 Mesh grains with \( \rho \approx 10^9 \) tracks/cm\(^2\); 5. the 100 keV replica show frequently a sudden drop in the track density at depths \( \approx 2 \) microns (figure 1). Therefore these track observations, although very preliminary, still support the existence of low energy nuclei in the ancient solar radiation, but calibration experiments should now be conducted to check definitively the validity of this conclusion.

The 1 MeV electron diffraction search for high doses of low energy ions, implanted in the superficial layers of lunar dust grains, was attempted during a study of the lattice disorder observed in the finest grains from various dust samples (Luna 16-19, 10084, 12070, 14259, 15101), as well as in grains artificially irradiated with 20, 50 and 100 keV \( \alpha \)-particles. Our main results are 1. low energy \( \alpha \)-particle fluxes as low as \( 10^{16} \alpha/\text{cm}^2 \), produce a clear decrease in the total number of diffraction spots, \( \Sigma_S \), as well as a diffuse electron scattering, superimposed on the diffraction spots; 2. low values of \( \Sigma_S \) were measured in grains extracted from the Luna 16-19 and 14259 samples (figure 2); 3. a very intense diffuse electron scattering was observed in sample 15101; 4. These "disorder" features, appearing in the electron diffraction patterns of the natural dust grains, were generally strongly reduced in grains either heated 2 hours at 900°C, or slightly etched in a very dilute HF solution. From these results we conclude that both the diffuse scattering and the decrease in \( \Sigma_S \) observed for the natural dust grains, could partly be due to their irradiation in high fluxes of low energy ions, producing a shell of radiation damaged matter that do no extend completely to the center of the micron-sized lunar dust grains.

References
1. L. Frank, J. Geophys. Res., 75, 707 (1970); E. Stones, personal communication; see also the work of T. Armstrong and S. Krimigis, to be published in the Proceeding of the NASA Conference on
"Modern and ancient energetic particles from the Sun", to appear shortly in Physics to Day.


Crystal and glass phases have been probe analysed in basalts 14310, 14073, 15076, 15085 and 15555; breccias 14049, 14305, 14320; and various Apollo 14 fines. Crystal-liquid fractionation relations between picritic basalts and ferrobasalts, and between the basalts and the norites, anorthosites and plutonic fragments (peridotite and granite) in the soils and breccias, are emphasized. Feldspar-phyric basalts are brought into this scheme.

The two A14 basalts are not only feldspar-rich (53%) but carry plagioclase phenocrysts. One is more noritic (14073). Both show zoning of the phenocrysts to more sodic rims (An94 to 67) with a strong reversal to a calcic mesostasis (An94). Calculations show at least 74% loss of Na₂O between the two plagioclase generations. The pyroxene cores are strongly magnesian (Ca₃Mg₈Fe₆) and zone to augite and pyroxferroite in 14310, which is therefore not a feldspathic norite. The two rocks constitute a link between mare basalts and shocked norites and anorthosites of the regoliths. Together they contain minor whitlockite, tranquillityite, baddeleyite, Ba-sanidine, K-rhyolite glass and a third new Fe-Ti-Zr mineral (Phase X, Table 1). Iron contains up to 14% Ni. Cr-spinel is absent, and Ti is low.

The Apollo 15 lavas are mafic, siliceous basalts. 15555 is comparable with 10045 and 14053. They vary greatly in their pyroxene trends. In 15555, intergrown augite and pigeonite zone to a common focus at Ca₂₂Mg₂₂Fe₅₆. The zoning continues to an extreme ferrohedenbergite (Table 1) with fayalite and silica. In 15085, small pigeonite cores are mantled by augite zoned continuously to an extreme "subcalcic pyroxferroite" (Table 1) with no detectable fayalite. In 15076, large magnesian pigeonites have narrow rims zoned to subcalcic ferroaugite, intergrown with silica and large (0.5 mm) fayalite crystals. Varying rates of cooling at near-solidus temperatures, in three different lava flows are implied. The mesostasis includes Ba-rich rhyolite glass, tranquillityite, and a fourth new Fe-Ti-Zr mineral (Phase Y, Table 1).

Most of the A14 fragments are microbreccias often with preserved glass spheres. The brown glasses are chiefly normatively saturated or mildly oversaturated basalt. Exceptions include quartz (10%) basalt, olivine (18%) basalt, norite (68% plagioclase, 26% hypersthene), picrite (73% olivine+pyroxene) and two "sodic" aluminous basalts. The latter are normatively An₆₃ and An₇₀, with Na₂O=1.28 and 1.14%, K₂O = 1.34 and 0.89, respectively. This casts doubt on the concept that all lunar basalts are strongly depleted in Na before eruption. The two
MINERAL FRACTIONATION IN APOLLO 14 AND 15

G.M. Brown

Glasses compare closely to terrestrial olivine tholeiites, with only slightly depleted Na. The pyroxenes in the soil fragments show the most extremely magnesian orthopyroxenes so far recorded (Ca 1.5Mg 83Fe 15.5). Forsteritic olivines imply derivation of ferromagnesian phases from a lunar mantle comparable with that of the Earth in terms of Mg:Fe being at least 90:10.

The breccias contain some fragments probably derived from strongly fractionated sub-surface plutons. These include olivine-plagioclase intergrowths with crescumulate texture and granophyre with tranquillityite. A common phase is purple spinel (Al 0.61.5,Fe019.8,Mg015.0,Cr 0.23,Ti0.9). Zoning to deeper purple rims is due to rise in Fe:Mg ratio. Whitlockite occurs as an early phase within plagioclase.

The discovery of another two new Fe-Ti-Zr phases (in basalts 14310 and 15555) leads to a summary of these minerals. They can be plotted with tranquillityite, phase β (121013, Haines et al., 1971) and terrestrial zirkelite on a ZrO-FeO-TiO diagram. Contents of these and Si, Ca, Cr, Y and REE vary widely. Their affinities with rutile, baddeleyite and zirconian ilmenite in major elements, and with whitlockite in Y and REE, is evident. This brings the count to one new phase in each mission collection, from this system. Li may be an important constituent.

We now have enough data on lunar pyroxenes to re-consider the minor elements. For example, Al:Ti reaches 7:1 in the feldspar-phyric A14 basalts and some norite fragments, and is down to nearly 1:1 in the A15 basalts (ferroaugites). A-11 pyroxenes were around 2:1, and A-12 up to 5:1. This can no longer be correlated with a plagioclase control, but may be influenced more by Ti activity and the presence of trivalent Ti in late-stage pyroxenes.

X-ray fluorescence analysis of six fines (Apollo 14 regolith) show much stronger enrichment in the following sought elements relative to A-11 and A-12 fines. The maxima (ppm) are Ba(706), Nb(62), Zr(1172), Y(261), Sr(209), Rb(20), Zn(41), La(85), K(6938). This suggests a high proportion of KREEF-rich granitic material.

The evidence is now compelling that there has been strong near-surface crystal fractionation of the lunar basalts, producing contrasted varieties relative to any primary, mantle-derived parental basalt. The primitive, plagioclase-phyric basalts provide evidence for this 2-stage magmatism. Floatation of feldspar phenocrysts in a deep lava pool led to normal zoning and could have produced anorthosite and norite (with floated Mg-pyroxene attached to plagioclase grains in a 1:1 ratio) in more extreme cases. Eruption from the pool then produced Na loss, and crystallization of calcic plagioclase in the groundmass. Feldspathic basalt glasses with normative An63 (Na=K) indicate that Na-loss was a near-surface, rather than primary feature. Mafic minerals sank in these pools, to produce mafic-cumulate basalts thus not carrying plagioclase on the experimentally-derived liquidus. Plutonic cumulates in the soil fragments indicate extensive fractionation possibilities beneath the lunar surface. The crustal norites and anorthosites may be from both
local and regional fractionation. The KREEP-rich liquid was probably granitic, viscous, and of low density, and concentrated locally in the crustal layers during sub-crustal fractionation.

In terms of compositional field averages, the pyroxene data suggest a fractionation trend of A14→A12→Al5→A-11(type B) basalts with some overlaps. The most likely primitive magma is high-alumina magnesian basalt, detectable from mineral trend data. Early liquidus phases were probably plagioclase, aluminous spinel and magnesian low-aluminous orthopyroxene (olivine and iron) common in the A14 breccias and basalts. Pigeonite, chrome-spinel and augite were slightly later phases, with Ti being a late-stage enrichment. KREEP-rich rhyolite bearing Zr, U, Th-rich phases, is a product of basalt fractionation.

<table>
<thead>
<tr>
<th>TABLE 1</th>
<th>Phase X (14310)</th>
<th>Phase Y (15555)</th>
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*Includes La₂O₃.
"A DIRECT MEASUREMENT OF THE IMPACT RATE OF SECONDARY MICROMETEOROIDS ON THE LUNAR SURFACE". D.E. Brownlee, Department of Astronomy, Univ. of Washington, Seattle, Washington 98195 and P.W. Hodge, Department of Astronomy, Univ. of Washington, Seattle, Washington, 98195.

SEM analysis of the Surveyor III television camera's glass optical filters provided a unique opportunity to directly measure the impact rate of secondary micrometeoroids on the Lunar surface. The following properties of the glass filters and their lunar exposure made such a measurement possible; relatively artifact free optical quality of the surfaces, position of the filters inside the camera hood prevented surface abrasion during return to earth, orientation of the filters and shielding by the hood prevented impact of debris generated by the surveyor landing or the LM landing, the anti-reflection MgF₂ coating on the filters enables determination of azimuth for low angle impact craters, the camera hood restricted possible azimuthal incidence angles. Detailed SEM analysis of the "clear" filter revealed a large number of the micron-sized indentations which have raised rims, show evidence of plastic flow, and show associated chipping of the MgF₂ overcoating. The indentations are not hypervelocity impact sites but are identified as low velocity impact signatures of secondary lunar ejecta because of the following lines of evidence; no similar features were found on the "clear" backup filter which did not go to the moon, no similar features were found on the side-(bottom) of the "clear" flight filter which was not exposed to space, estimates of azimuthal impact angles for the features correlate with the exposure window of the filter, exact duplication of the morphological characteristics of the craters has been accomplished by firing irregular micron-sized silicate particles into MgF₂ coated glass at 0.3ms⁻¹. The measured impact rate of sub-escape velocity micron-sized particles (>1μ) averaged over the 2.6 year exposure on the lunar surface is approximately 10,000 times the estimated impact rate of primary micrometeoroids (using the Kerridge 1970 model). This measured flux of lunar ejecta particles agrees well with that predicted by Gault et. al. 1963 on the basis of artificial cratering studies.
DISTRIBUTION OF ELEMENTS BETWEEN DIFFERENT PHASES OF APOLLO 14 ROCKS AND SOILS

A. O. Brunfelt, K. S. Heier, B. Nilssen, E. Steinness and B. Sundvoll (Norway)

The following samples were received: 14163,154 fines (4.960 gr.), 14276 rock (0.236 gr.), 14303,14 breccia (32.31 gr) with two polished sections, 14310,123 rock (4.04 gr). 14276 is part of a consortium sample, and our determination of 37 elements will be reported by Professor Wasscrburg. Data on 156ol,75 (soil) are also given. All analyses were by neutron activation analysis.

Lunar regolith, 14163 See Brunfelt et al. (1) for bulk analysis. Fine fraction (< 0.12 mm) was analysed separately. It is similar to the bulk regolith composition. Seven fractions were handpicked: plagioclase, pyroxene, dark rock fragments, light rock fragments, dark irregular glass fragments, twisted glass fragments and glass spheres. Apart from the pyroxenes and plagioclase, all the fractions have similar composition, Fig 1. The REE patterns of the plagioclastes and pyroxenes are compared with the bulk fines in Fig 2. Lunar regolith 156ol is compared with 14163 and 1207o in Fig 3. REE patterns of soils from the four APOLLO missions are compared in Fig 4.

Breccia 143o3 is compared with the composition of soil 14163 in Fig 5.

Rock 14310: Ortho-pyroxene and plagioclase fractions were obtained 95 % pure. REE patterns of these and of bulk rock are shown in Fig 6.

Modal analyses (2, 3) indicate 61 % plag. vs 31 % px. and 5o % plag. vs 40 % px. respectively in this rock. The authors report clinopyroxene as the pyroxene phase while orthopyroxene dominates in 14310,123 (cpx < 1 %). Element concentrations of plag. + px. (calculated as 2/3 plag. and 1/3 px.) vs bulk rock are shown in Fig 7. (K in px and Ti in plag. taken as zero; Sr and Ga in px. as 10 and 1.1 ppm respectively). Assuming the other remaining phases to constitute 5 % of the rock, the average composition of these phases is in %: Mg (30.2), Fe (49.2), Ti (11.6), K (4.8), Mn (0.6), Cr (0.2o); in ppm: Sc (132), Co (166), Rb (216), Ba (664o), La (854), Sm (360), Tb (159), Dy (434), Yb (172), Hf (266), Ta (41), Th (148), U (48.6).
**Distribution of elements**

*O. Brunfelt*

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**Fig 1.** Distribution of elements between various fractions of 14163, 154, and bulk fines. Phases showing maximum and minimum values are shown, remaining phases plot in the indicated range (□ dark rock fragments, △ light rock fragments, * dark irregular glass fragments, V twisted glass fragments, ○ glass spheres), Na, Mg, Al, K, Ca, Ti, Cr, Mn, Fe in percent.

**Fig 2.** REE distribution in 14163, 154 pyroxene (B), plagioclase (C), bulk fines and other separated fractions (A, shaded area).

**Fig 3.** Distribution of elements between 12070 (X) and 14163 (○) vs 15601 bulk fine.

**Fig 4.** REE distribution in 10084, 12070, 14163, and 15601.

**Fig 5.** Diagram comparing the compositions of breccia 14303 with soil 14163.

**Fig 6.** REE distribution in 14310, bulk rock (A), plagioclase (B) and pyroxene (C).

**Fig 7.** Distribution of elements between plag. + px. and bulk rock in 14310, 123.

**Fig 8.** Distribution of elements between 14310 and 14163.

Note the high concentration of Ta, which is of interest in view of the Nb rich "phase β" reported earlier (4) on rock 12013 Nb was not determined by us.

The distribution of elements between 14310 and 14163 is shown on Fig 8.

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RESULTS OF THE APOLLO 14 AND 15 SOLAR WIND COMPOSITION EXPERIMENTS. F. Buehler, H. Cerutti, P. Eberhardt, and J. Geiss, Physikalisches Institut, University of Bern, Sidlerstrasse 5, 3012 Bern, Switzerland.

It is the aim of the Solar Wind Composition experiment to measure the abundances and the isotopic composition of noble gases in the solar wind. A collection technique is employed at the lunar surface by which the solar wind particles are implanted in a catcher foil and brought back to earth for laboratory determinations.

The experiment was successfully carried out during the Apollo 11, 12 (1) and also during the 14 and 15 missions (2, 3). All experiments yielded the elemental and isotopic abundances of helium and neon in the solar wind. In the Apollo 14 and 15 experiments, argon was detected for the first time. Argon data will be presented at the Conference.

In the Apollo 14 experiment the angular distribution of the arriving He\(^4\) ions was determined with higher resolution and precision than was possible in the preceding experiments.

In Table 1 the fluxes of He\(^4\) during the four exposure periods are given, and in Table 2 the relative abundances of the helium and neon isotopes. There are large differences between the helium flux measured during the four Apollo landings. In particular the Apollo 14 and 15 fluxes differ by about a factor of four, and yet the relative abundances of helium and neon, and also the isotopic abundances are very similar. Comparison with proton flux measurements obtained from unmanned spacecrafts will show whether there existed a similar difference in the fluxes of hydrogen during the Apollo 14 and 15 landings, or whether the abundances of the helium and neon isotopes changed relative to hydrogen by a large and rather uniform factor. The variations in the He\(^4\)/He\(^3\) and He\(^4\)/Ne\(^{20}\) ratios (Table 2) are small but significant. A distinct correlation between these two ratios exists. These variations are linked to the level of disturbances in the solar wind, as indicated by the observed positive correlation between He\(^4\)/He\(^3\) and geomagnetic activity (Figure 1). This correlation may actually be expected on the basis of models of solar wind ion acceleration (4). This type of correlation is useful in estimating the long-time average of the He\(^4\)/He\(^3\) ratio in the solar wind.
RESULTS OF THE APOLLO 14 AND 15 SWC EXP.
F. Buehler

The significance of the results will be discussed in relation to ques-
tions of solar evolution and abundances in the solar nebula.

The cooperation with NASA is acknowledged. This research was sup-
ported by the Swiss National Science Foundation.

References

<table>
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<th>Mission</th>
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<td>Apollo 12</td>
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<td>Apollo 14</td>
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<td>4.2 ± 0.8</td>
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<tr>
<td>Apollo 15</td>
<td>July 31, 1971</td>
<td>17.7 ± 2.5</td>
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Table 1. $^4$He flux averages during the times of foil exposure. Data for Apollo 11 and 12 from (1).
RESULTS OF THE APOLLO 14 AND 15 SWC EXP.

F. Buehler

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<tr>
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<th>He(^4)/He(^3)</th>
<th>He(^4)/Ne(^{20})</th>
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Table 2. Abundance ratios of solar wind ions determined for the foil exposure periods of the Apollo landings. Data for Apollo 11 and 12 from (1).

Figure 1. Correlation between the solar wind He\(^4\)/He\(^3\) abundance ratio and the level of disturbance in the solar wind as indicated by the geomagnetic index K\(_p\) (cf. 3). A= Apollo; S3= Surveyor 3 (5).

The record of exposure of lunar materials to galactic cosmic rays has been studied by means of Sm and Gd rare gas isotopic measurements.

In addition to the isotopic variations due to low energy neutron capture on $^{157}$Gd, we have measured Sm isotopic variations due to neutron capture on $^{149}$Sm. Enrichments in $^{150}$Sm/$^{154}$Sm and correlated depletions in $^{149}$Sm/$^{154}$Sm are in quantitative agreement with predictions for the exposure of terrestrial Sm to a low energy neutron flux. The percentage variations in $^{150}$Sm/$^{149}$Sm (up to 1.1%) are larger than corresponding variations in $^{158}$Gd/$^{157}$Gd. Neutron capture in $^{149}$Sm is enhanced in lunar materials relative to $^{157}$Gd because the lunar neutron energy spectrum has a maximum near 0.1 ev which is close to the position of a prominent resonance in $^{149}$Sm. The ratio of the average neutron capture cross section for $^{149}$Sm to that for $^{157}$Gd, obtained directly from the experimental data, is 0.83-0.89 for samples from Apollos 11, 12, and 14, 0.76 ± 0.02 for the Luna 16 G-2 soil, and 0.76 ± 0.02 for Hadley Rille soil-15601. The lower experimental cross section ratios for the Luna 16 and 15601 soils indicate that these materials were exposed to a lower energy neutron spectrum. This is qualitatively consistent with the chemical compositions in that Luna 16 G-2 and 15601 are better neutron moderators than the Apollo 11, 12 and 14 soils. The experimental cross section ratios are in reasonable agreement with those calculated theoretically by Lingenfelter, Canfield and Hampel (unpublished calculations), although the actual lunar spectrum appears to have more higher energy neutrons than the theoretical spectrum. This agreement suggests that relatively reliable neutron fluences can be calculated from either the $^{149}$Sm/$^{150}$Sm or the $^{157}$Gd ratios using average cross sections calculated by Lingenfelter, et al. From the measured $^{158}$Gd/$^{157}$Gd ratios we have calculated the fluence ($n/$cm$^2$) of low energy neutrons ($E<0.18$ ev) based on the following cross sections (in units of 10$^5$ barns) which are corrected for differences in bulk chemistry using the Lingenfelter et al. energy spectra: 89 (Apollo 11); 94 (Apollos 12 and 14); 96 (Luna 16); 98 (Apollo 15). The neutron fluences (in units of 10$^{16}$n/cm$^2$) for Apollo 14 soils 14163 (3.1) and 14259 (3.0) are slightly higher than those for Apollo 11 (2.7) and Apollo 12 (2.2-2.4). On the other hand the Luna 16G-2, 15601, 15221 and 15231 soils show
IRRADIATION HISTORY OF LUNAR SAMPLES
D. S. Burnett

distinctly higher fluences ($4-6 \times 10^{16}$ n/cm$^2$). The higher fluences for the Luna 16 and Apollo 15 samples probably reflect lower mixing rates for these sites. Assuming mixing times equal to the crystallization ages of the rocks, effective mixing depths ranging from 6 (Luna 16) to 16 (Apollo 12 soils) are obtained.

A well-defined correlation is observed between the concentrations of $^{126}$Xe produced by Ba spallation, calculated as $^{126}$Xe meas/ (Ba + 1.65 Ce), and the neutron fluences for lunar soil samples. This correlation shows that the soils represent material which have been well-mixed to depths of at least 2 meters in order to average over differences in the depth dependence of the $^{126}$Xe and $^{158}$Gd production rates. In contrast, lunar rock data show a wide scatter on this diagram, reflecting irradiations at discrete depths.

Neutron fluence measurements on grain size fractions show a tendency for the exposure to increase with decreasing grain size although the observed variations ($\leq 20\%$) are relatively small. This trend is analogous to that observed between rocks and total soil samples (rocks tend to show lower fluences). These trends are probably a reflection of long term "grinding" processes on the lunar surface.

Lunar soil accumulation and mixing models have been developed for the interpretation of neutron dosage and spallation $^{126}$Xe data from core tube and drill stem samples. For the Apollo 12 double core, the $^{126}$Xe and neutron fluences show only slight variations with depth. The data are compatible with a model in which the regolith has been accumulated at a rate greater than 3 mm/10$^6$ yr from soil which pre-irradiated to a fluence of about $2 \times 10^{16}$n/cm$^2$. On this model the deposition time of the light-colored layer (VI) is less than 60 m.y.

A sample from the bottom (2.4 meters) of the Apollo 15 deep drill stem has a neutron fluence of $3.7 \times 10^{16}$n/cm$^2$ which is very similar to that measured for Apollo 15 surface soils; however a sample from the center (1.2 m) of the drill stem shows a larger neutron fluence ($5.6 \times 10^{16}$n/cm$^2$) indicating that the drill stem is not well mixed and has preserved valuable stratigraphic information. The two measurements can be described by a model in which the soil at the drill-stem site was formed 500 m.y. ago from material having a uniform fluence of $\sim 3 \times 10^{16}$n/cm$^2$ and the deeper portions of the stem have remained unmixed.

Rock 14310 has a comparatively high neutron fluence ($3 \times 10^{16}$ n/cm$^2$), compared to the amount of Ca-produced spallation $^{36}$Ar ($410 \times 10^{-8}$ ccSTP/gm Ca). In addition, a small, but significant increase in fluence of about $2 \times 10^{15}$n/cm$^2$ is observed for an exterior piece of 14310 relative to an interior sample. These data are best explained by assuming that 14310 was irradiated in a fixed orientation in a highly shielded ($>100$ g/cm$^2$) location.

Cosmic ray exposure ages for Apollo 14 rocks, based on the amounts of $^{36}$Ar produced by Ca spallation, show a definite radial
dependence relative to Cone Crater. Samples from the flank of Cone Crater (14053, 14321, 14083) all have ages of 24 ± 2 m.y. which we ascribe to the Cone impact event. With the exception of one fragment from soil 14167 with an exposure of 29 m.y., samples from near the LM have higher ages of 110 (14073), 300 (14310), 260 (14001 fragment) and 590 (14001 fragment) million years. The low neutron fluence for 14321 (≤10^{15}n/cm^2) is consistent with a 24 m.y. surface irradiation and indicates that 14321 must have been buried at a depth greater than about 4 meters prior to the Cone Crater impact.

The ratio \( ^{126} \text{Xe}/(\text{Ba} + 1.65 \text{Ce}) \) is 0.387 x 10^{-12} ccSTP/ppm for soil 14259 and 0.498 x 10^{-12} for soil 14163 yielding model "exposure ages" of 550 and 710 m.y. which are similar to Apollo 11 and 12 soils. In contrast sized fractions of the 14141 Cone Crater fines give model \(^{126} \text{Xe}\) exposure ages of 50 (greater than 300 micron) and 99 (less than 150 micron) million years, indicating that this soil is primarily composed of previously unexposed ejecta from the Cone Crater event. The "excess exposure age" over 24 m.y. may indicate that about 4% (coarse) and 12% (fine) of the Ba in these fractions has been transported to the rim of Cone Crater from the surrounding smooth areas in the last 26 m.y. Alternatively, the excess \(^{126} \text{Xe}\) could be interpreted as residual surface soil from the pre-existing regolith at the spot of the Cone Crater event which was highly diluted with unexposed material during the impact. The amount of dilution required would correspond to mixing of material through depths of 80-250 meters during the impact.

A basaltic Luna-16 fragment (B-1) has a relatively high \(^{38} \text{Ar}\) exposure age of 475 m.y. which correlates with the high neutron fluence of the Luna 16 soil.

Apollo 15 rock 15555 has an \(^{38} \text{Ar}\) exposure age of 90 m.y. If the \(^{38} \text{Ar}\) concentration is interpreted as a steady state value, reflecting the continual erosion of the profile of the Hadley Rille, an erosion rate of 1.5 gm/cm^2/m.y. is obtained.

The 15415 anorthosite has a relatively typical \(^{38} \text{Ar}\) exposure age of 113 m.y.

References


The absorption spectra of single crystals of pyroxene and olivine minerals in rocks from the Apollo 11, 12 and 14 missions were measured in polarized light (360-2200 nm). In this preliminary study transmitted spectra were recorded for individual crystals (0.5-1.0 mm) occurring in petrographic thin sections and microprobe mounts, using polarizing microscopes equipped with a universal stage mounted in a Cary 17 spectrophotometer. As a result it was impossible to avoid working with compositionally zoned and exsolved phases, while crystal field spectral features could be detected only for the major transition metal cations Fe$^{2+}$ and, occasionally, Ti$^{3+}$.

By analogy with terrestrial silicates, the spectra of lunar pyroxenes and olivines are polarization dependent in the near infrared (800-1200 nm) where spin-allowed transitions in Fe$^{2+}$ ions occur. Absorption bands are most intense for light polarized along the b axis of pyroxenes and along a of olivines. Another very broad, spin-allowed Fe$^{2+}$ band was sometimes measured between 1700-2200 nm and beyond in lunar pyroxenes. Numerous weak, sharp peaks occur in the visible region, the majority of which represent spin-forbidden transitions in Fe$^{2+}$ ions. No evidence for Fe$^{3+}$ ions was found; in particular, the Fe$^{2+}$-Fe$^{3+}$ charge transfer band around 700 nm was absent and metal-oxygen absorption edges were displaced well into the ultraviolet. Existence of Ti$^{3+}$ ions in some pyroxenes was suggested by broad bands centered at 460-470 nm and 660-670 nm. No confirmation was obtained of the presence of Cr$^{2+}$ ions in the olivines.

In pyroxenes from rocks 10045, 10047 and 10058 light polarized along the b axis gives broad absorption maxima between 980 and 1025 nm, the variation in position and intensity of which are related to compositional zoning and proportions of exsolved augite-pigeonite lamellae. This band is assigned to a spin-allowed transition in Fe$^{2+}$ of the type $^{5}T_{2g} \rightarrow ^{5}E_g$. In pyroxenes from rocks 12021 and 12052 separate spectra were obtained from core pigeonite and rim augite. The peak maxima reflect the observed exsolution and compositional zoning and range from 900-975 nm in pigeonite and 975-1025 in augite.
These trends are similar to those found in terrestrial pigeonites and sub-calcic augites from the Mount Wellington diatase, Tasmania, in which additional features due to Fe$^{3+}$ ions are found. The pyroxenes from Apollo 14 rocks 14053 and 14310 are similar to those from rocks 12021 and 12052, but the band maxima ranges 930-980 nm indicate a lower degree of compositional zoning.

Of the numerous weak peaks observed in lunar pyroxenes at 402, 425, 460-470, 505, 550, 585 and 660-670 nm, the broad bands centered at about 465 and 665 nm were attributed to spin-allowed transitions in Ti$^{3+}$ ions originating from resolved levels of the $^2T_{2g}$ ground-state to components of the split $^2E_g$ excited state. The intensities of these Ti$^{3+}$ bands relative to the spin-forbidden peaks in Fe$^{2+}$ ions were most intense in pyroxenes from Apollo 11 rocks in which the Fe/Ti ratios may be as high as 0.1. However, the Ti$^{3+}$ bands could not be resolved by the present experimental set-up in pyroxenes from the Apollo 12 and 14 rocks, in which the Ti/Fe ratios are lower. Nevertheless, the presence of Ti$^{3+}$ ions in the Apollo 11 pyroxenes correlates with earlier evidence of low oxygen fugacities during the cooling history of the Moon, and indicates that the crystal chemistry of the lunar pyroxenes may have to be reinterpreted.
SOME SURFACE CHARACTERISTICS AND GAS INTERACTIONS OF
APOLLO 14 FINES AND ROCK FRAGMENTS, by D. A. Cadenhead (Principal Investigator), N. J. Wagner, B. R. Jones and J. R. Stetter,
Department of Chemistry, State University N.Y. at Buffalo, Buffalo, N.Y. 14214.

Over the past six months a surface study of Apollo 14 fines and rock fragments have been carried out with the following objectives: (1) A corellation of geolocial features and lunar events with surface area (2) The observation of internal pore structure (3) An evaluation of the adsorption characteristics of these materials to gases such as nitrogen, hydrogen and water vapor. Objectives 1 and 2 were realized by the observation, through both optical and electron (scanning and transmission) microscopy, of the surface features of the samples, while obtaining the surface area (and co-incidentally the helium density). Helium densities were obtained using a classical B.E.T. volumetric system. Typical density values ranged from 2.7 to 3.1 g/cc, and were used to provide accurate taring of samples for gravimetric adsorption. Surface area were determined by gravimetrically measuring nitrogen adsorption at liquid nitrogen temperatures, using an area/adsorbed nitrogen molecule of 16.2 Å². Surface areas of 0.2 to 1.0 m²/g were found for fines (<< 1 mm) and for breccia surface fragments, with considerably lower values for surface crystalline fragments. Preliminary observations indicate that geode and pressure fusion breccia formation enhance surface area, while the formation of glass lined pits through micrometeorite impact reduces surface area.

The electron microscopy in particular reveals that the measured surface areas arise in large part from small particle size (fines), from pressure fusion of small particles to form larger ones (fines and breccia) and from surface roughness. So far we have not observed internal pore structure which arises through rock crystallization accompanied by gas evolution.

Physical adsorption of molecular hydrogen on lunar fines was observed at low temperatures (-197°C) but not at higher temperatures. Possible chemisorption of molecular hydrogen at higher temperatures (up to 130°C) and atomic hydrogen at both low and high temperatures will be studied. Adsorption of water vapor on lunar fines at room temperatures takes place with typical capillary condensation hysteresis at high relative pressures but with complete reversability at low pressures. This behavior as well as good reproducibility is considerably dif-
Surface Characteristics and Gas Interactions
D. A. Cadenhead

Different from previously reported behavior on lunar fines. These differences could arise through the different nature of the samples, their different pretreatment prior to adsorption studies, difficulties in the calibration of the gravimetric system, or a combination of all three.

Our observations, so far, are that those factors which most increase surface area or roughness are the fracturing of rocks to fine particles as well as the accompanying light pressure fusion processes. The formation of geodes and glass lined micro-meteorite pits in crystalline rock materials appear to produce only small increases or decreases in surface area. Molecular hydrogen will have no effect on the chemical composition of exposed materials at low temperatures (-130 to 0°C) but may do so at higher temperatures (0° to 130°C). A more likely candidate is atomic hydrogen. Exposure of lunar materials to water vapor has not shown any rapid weathering effects, hence brief accidental exposures to an earth-like atmosphere is not likely to be harmful. Under lunar conditions, any water vapor to which these materials might be exposed would rapidly be lost.

References


Water Adsorption $T = 37.5^\circ$C

- Adsorption
- Desorption
SURVEY OF LUNAR CARBON COMPOUNDS II: THE CARBON CHEMISTRY OF APOLLO 11, 12, 14 and 15 SAMPLES. P.H. Cadogan, G. Eglinton, J.N.M. Firth, J.R. Maxwell, B.J. Mays and C.T. Pillinger, Organic Geochemistry Unit, School of Chemistry, Bristol University, U.K.

In our previous studies hydrocarbons released from lunar samples by acid dissolution were resolved into indigenous species and reaction products (deuterocarbons) by the use of deuterium-labelled reagents. The quantities of CH$_4$ and CD$_4$ released by DF etch of Apollo 11 and a variety of Apollo 12 fines were correlated with parameters relating to surface exposure. These data, taken with preliminary results from other experiments (shallow etching and analysis of an igneous rock and of size-differentiated fines) indicated that solar wind implantation in the fines plays a major role in the synthesis of endogenous hydrocarbons. Also, a dual extralunar origin (meteorite impact and solar wind implantation) appeared likely for the 'carbides', the presumed source of the deuterocarbons. Further analyses by mass spectrometry and improved gas chromatographic techniques have now provided additional evidence for the origins of these carbon species:

1. Sources of possible contamination and of artefacts. It has been suggested that methane in lunar fines arises by adsorption from the terrestrial environment or by hydrolysis of reactive lunar 'carbides' by adsorbed water. Three experiments taken together indicate to the contrary: (i) Exposure of fines (10086) to CD$_4$ for 24 hours and subsequent dissolution in HF showed that $< 0.2 \mu g/g$ of CD$_4$ was not desorbed by standard degassing procedures preceding each analysis. Further, adsorbed terrestrial CH$_4$ has never been observed in standard procedural blanks (activated Al$_2$O$_3$ exposed during sample handling), (ii) immersion of fines (10086) in D$_2$O for 41 days did not generate CD$_4$. HF dissolution after 3 months in D$_2$O also failed to reveal trapped CD$_4$. (iii) DCI dissolution of freshly-exposed interior chips of breccia 10059 afforded quantities of CH$_4$ and CD$_4$ similar to those observed for fines (10086).

In addition, vigorous heating of DCI (200°C, 2h.) with fragments of laboratory glass released CO, indicating that the single high value for CO previously reported was artefactual, as suspected. No CO was generated in the normal reaction procedure blank (100°C, 2h.) and quantities of CO released from a variety of Apollo 11, 12 and 14 samples certainly do not exceed 3 $\mu g/g$.

2. Location of carbon compounds in lunar fines. Evidence concerning the location of hydrocarbons and 'carbides' has been obtained by analysis of size-differentiated fines and mineral separates. Our data for the Apollo 11 fines, which suggested that CH$_4$ was concentrated in the finest grains, have been extended to show that both the CH$_4$ and CD$_4$ released by DCI are directly proportional (surface correlated) to the reciprocal of the mean grain radius for particles 48 to 152 $\mu$ diameter. However, for particles 152 $\mu$ to 2mm diameter the surface relationship is no longer valid; there is an additional CH$_4$ and 'carbide' component which may be associated with the microbreccias and glassy aggregates abundant in these coarse fractions. Initial studies of density-
separated mineral fractions have not revealed major differences in \( CH_4 \) and \( CD_4 \) contents, but more information is expected with sieved fractions.

3. Survey of lunar samples. We have shown that quantitation of 'carbide' based on \( CD_4 \) released from All fines, is dependent on the nature and strength of the acid; our standard method employs 38% DCl at 100°C. Hydrolysis of coarsely powdered meteoritic cohenite under standard conditions afforded ca 6% \( CD_4 \); even this improved dissolution does not permit an accurate determination of lunar 'carbide', for on this basis the All fines should contain 300 \( \mu g/g \) total carbon (measured, ca 180 \( \mu g/g \)). Our survey of the \( CH_4 \) and 'carbide' concentrations has been extended to 8 samples from depths between 9.0 cm and 39.8 cm of the A12 double core (12025, 12028), A14 fines 141, 148, 149, 156, 163, 240 and 298, A14 breccias 267, 311, 313 and 321, and A15 fines 031, 041, 231, 261, 431, 471 and 501. \( CH_4 \) and \( CD_4 \) values range from 0.5 and 1.1 (Cape Crater 14141) for light-coloured fines, to 5.1 (15231) and 23.5 (15041) \( \mu g/g \), respectively, for dark-coloured fines. The \( CD_4/NH_4 \) ratios fall in the range 2:1 - 12:1, averaging 4:1. There is no systematic variation for \( CH_4 \) and \( CD_4 \) with depth for the core and correlations with other parameters are difficult to obtain with the existing sample allocations. However, the \( CH_4 \) and \( CD_4 \) data roughly parallel the available exposure parameters. The A15 fines all give high \( CH_4 \) and \( CD_4 \) values with no apparent relationship to 'ray' or 'front' material but the most depleted sample (15431) was derived from the pedestal of 'genesis' rock. Three A14 breccias, 14267, 14321 and 14311, gave extremely low values for \( CD_4 \) and \( CH_4 \) in accord with the high temperature origin indicated petrologically for these samples. Figures for total carbon are awaited.

4. Correlation of carbon chemistry and lunar surface exposure history. We have extended our correlations for \( CH_4 \) and 'carbide' concentrations with parameters indicative of extended exposure of the fines at the lunar surface.

(i) Extent of solar wind implantation (as \( ^{36}Ar \) concentration) correlates with amounts of \( CH_4 \) with a \( ^{36}Ar/CH_4 \) ratio of 1:18. There is a similar, though less marked relationship with 'carbide', the ratio of \( ^{36}Ar/CD_4 \) being 1:81. Relative diffusion losses will need investigation. (ii) \( CH_4 \) concentrations correlate with the proportions of grains exhibiting radiation damage (as estimated by Maurette et al from the number of amorphous-coated 1-2\( \mu \)m grains and of feldspar grains (<200 mesh) having >10\(^9 \) tracks cm\(^{-2} \) (A14), and by Arrhenius et al from the pyroxene and feldspar grains (coarse fraction) having >10\(^8 \) tracks cm\(^{-2} \) (All and 12)). Other possible parameters (e.g. \( ^{13}C \) values, modal analyses of grains) for exposure and reworking will be reintroduced as data become available from other investigators.

5. Solar wind-carbon chemistry simulation studies. The physical location and correlation studies are in agreement with an extralunar origin for the hydrocarbons and 'carbides' in the fines. We have previously stated the arguments in favour of a hydrocarbon component of solar wind origin and 'carbides' of a solar wind and meteoritic origin. The possible role of solar wind hydrogen and carbon in the synthesis of lunar hydrocarbons and 'carbides' is now demonstrated by laboratory simulations of irradiation of a number of targets, including lunar fines, with \( ^{12}C \) and \( D_2 \) at energies similar to those in the solar wind. LRMS analysis of labelled \( C_1 \) compounds released by acid
dissolution or pyrolysis showed the formation from the irradiating species of hydrocarbons and material reacting as carbide. The simulation only approximates to the effects of the solar wind. For example, (i) higher dose rates were used, (ii) when both $^{13}\rm{C}$ and $D^+$ were used, irradiation was consecutive and not simultaneous, and (iii) $D_2^+$ was used instead of $D^+$ to obtain a sufficiently high flux. The data indicate that carbon and hydrogen in the solar wind could implant hydrocarbons and 'carbides' into lunar fines. No $^{13}\rm{C}O$ or $^{13}\rm{CO}_2$ were observed when irradiated lunar fines or simulated lunar fines were pyrolysed to 800°C. This suggests that solar wind carbon is not released as CO and CO$_2$ when lunar fines are pyrolysed to this temperature. Indeed, others have shown that acid dissolution and stepwise pyrolysis together with $\delta^{13}\rm{C}$ measurements indicate that the carbon is present in more than one form. Treatment of the vacuum-deposited Fe film with deuterated acid afforded CD$_4$, which was also released by dissolution of the unirradiated film. Fe$_3$C originally present in the iron filament evidently deposits 'carbide' with the iron, a possible analogue for the deposition of meteoritic carbide on the lunar surface under impact conditions. Furthermore, for the fines, a trend is evident between concentrations of 'carbide' and the volatile element bismuth considered by Anders et al as indicative of the micrometeorite contribution. The fewer published values for selenium follow the same trend. The addition of elemental material by meteoritic impacts may be partially compensated by redistribution and loss of volatile elements by the resulting melting and vaporisation of lunar material. Thus, metallic iron may be generated and vapor transported to the surface of the fines possibly generating 'carbide' by interaction with solar-wind implanted carbon. Models of lunar surface exposure history for carbon chemistry will need to take into account the processes mentioned above.

References

Six regolith samples from the Apollo 14 landing site were examined. The purpose was to characterize the regolith in terms of its components then compare the results with similar work on the local rocks so that a model for the evolution of the regolith could be devised. Three of the samples examined (259, 163, 003) were from separate locations close to the LEM, the other three were from different levels in the core tube from Station C. The samples, which were of the <1 mm size fraction, were first sieved then made into polished thin sections. From thin section examination approximately thirty different particle types were recognized and the relative proportions in each sample determined by counting approximately 500 particles in each sample. The counting was supplemented by several hundred microprobe analyses, enabling the range of chemical and mineralogical variability within each particle type to be assessed.

The particle classification scheme followed was patterned after that used by Wilshire and Jackson (1971) in classifying clasts in the Apollo 14 rocks. The classification is based entirely on optical characteristics as seen in thin section. Their scheme had to be modified slightly to account for the peculiarities of the regolith, particularly as regards glassy fragments, but in general the clast types observed in the rocks could easily be recognized in the regolith fragments. In the six samples that we examined there was very little variation in the relative proportions of the different particle types (fig. 1), although the assemblages appeared to be quite different from those described from the Apollo 11 and 12 sites (Wood, et al, 1970). Approximately 50-60% of the constituent particles were glass fragments of various kinds, 15-20% were lithic fragments with metamorphic textures, 4-15% were mineral grains and 6% or less were igneous fragments.

Three main types of glasses are distinguished, homogeneous glass, dark fragmental glass and splash glass. The three types grade into one another and all occur in various stages of devitrification. The homogeneous glass ranges in color from colorless through yellow to light green, most being light yellow. Vesicles are sometimes present but not common. Small spherical or elliptical opaque particles may be enclosed but mineral fragments of any kind are rare and generally small (<20 μm). A flow structure or banding occurs in some particles, the bands being formed by variation in glass color or strings of small inclusions. The glass most commonly forms angular fragments but most spheres are of this type. Some homogeneous glass fragments occur in the dark fragmental glass and splash glass, but fragments of the latter have not been observed in the homogeneous glass. The main characteristics of splash glass are a high degree of vesicularity at all scales, a fluid form, and a large number of enclosed fragments. The glass is commonly dark yellow but may be so full of
inclusions and microlites to appear opaque or cloudy. Many particles have a marked flow structure. Dark fragmental glass particles resemble the splash glass except that they lack vesicles and have angular rather than smooth, fluid outlines. Both the splash glass and the fragmental glass enclose all other particle types.

The next most abundant types of fragments are the hornfels breccias of which there are two types, light and dark. Light hornfels breccias are the most abundant (6-20%). They consist largely of plagioclase, orthopyroxene and ilmenite, which form an interlocking mosaic of equidimensional grains, generally in the 10-50 μm size range. The clastic nature of the fragments is apparent only in the larger grains which commonly contain mineral fragments. The dark hornfels breccias (1-9%) have a microcrystalline matrix which characteristically has fine, evenly disseminated ilmenite. Mineral fragments are common; lithic fragments, mainly dark fine grained breccias, are less common.

Mineral fragments constitute 4-15% of the particles in the samples examined. Feldspar grains are the most common (3-8%). They differ from the feldspar of the hornfels fragments in that they are larger, and subhedral rather than anhedral. They show a wide variety of strain and recrystallization textures and have compositions in the range An90 to An95. Pyroxene fragments are the next most common (3-6%). Orthopyroxenes and clinopyroxenes are approximately equal in abundance in contrast to the pyroxenes in the hornfelses which are predominantly ortho (fig. 3). Less than 1% of the regolith material consists of olivine fragments; they range in composition from Fo55 - Fo70.

Igneous fragments are the last major category and their proportion varied widely from sample 003, which contained less than 0.5%, to sample 230.103, which contained almost 6% igneous fragments. Interstertal and subophitic basalts are the most common but various kinds of vitrophyric, variolitic and cumulus textures were also observed. The normative composition of the hornfelses and glasses are compared in fig. 2. The minerals listed generally constitute more than 98% of the norm, with olivine or quartz less than 5% of the total. Several high potassium, high silica fragments are not included in the diagram. Most of the glasses and the hornfelses have a similar range of composition being essentially basaltic (or noritic). The homogeneous glasses have a wider range of composition than the other fragments. They are generally richer in titanium and have compositions representative of more mafic types of rocks. Some of the homogeneous glasses are vitrified feldspars.

The proportion of different types of fragments in the regolith is consistent with the local stratigraphy inferred by Wilshire and Jackson (1971) from the distribution of different rock types on the surface. They suggest that at the surface is a coherent, light clast dominant, rocks overlying a dark clast dominant rock which was excavated during the formation of cone crater. Our light and dark hornfels breccias correspond closely to their light and dark clast dominant rocks respectively. The preponderance of light hornfelses over dark among the regolith fragments therefore supports their model of light clast dominant rocks closest to the surface. Most of the glasses appear to be produced locally by repeated impact of the regolith; their compositions correspond closely to those of the rock fragments. The homogeneous glass may however be derived from a more diverse source since they have a wider compositional range.
PETROLOGIC AND CHEMICAL CHARACTERIZATION OF SOILS

Michael R. Carr

Figure 1. Relative proportions of components in six regolith samples.

Figure 2. Normative compositions of regolith particles.

Figure 3. Compositional range of pyroxenes in the regolith.

SUMMARY - Two oedometer and three direct shear tests have been performed on a 200-g sample of lunar soil from Apollo 12 (12001,119). This sample has not been exposed to a pressure above 2 x 10^{-6} torr since it was returned to the LRL. In an effort to maintain the soil particle surfaces as free of terrestrial contamination as possible, the test specimens were prepared in the F-401 glove chamber in the LRL at a pressure lower than 2 x 10^{-6} torr. The samples were placed in a small ultra-high vacuum chamber, which was sealed before removing it from F-401. The UHV shear chamber was then transferred to an external test stand and pumping station. The compression and shear tests were then performed after the pressure in the shear chamber had been reduced to less than 5 x 10^{-8} torr. Important data on the mechanical behavior of lunar soil has been obtained.

COMPRESSIBILITY - The data from the two vacuum oedometer tests are presented in Table 1. The void ratio has been calculated assuming the specific gravity for this sample is 3.1, as was measured for another Apollo 12 soil sample. The first sample was placed in a medium-dense condition (\( \rho = 1.84 \text{ g/cm}^3; \ e = 0.684 \)), compressed in 4 increments, and then sheared (Test 1). After shearing, the sample was then re-compressed in 4 increments to a normal stress more than twice the previous maximum load, and then sheared again (Test 1B). The void ratio during compression could not be accurately determined for this test and has not been included in Table 1. The soil was then re-placed in the shear cup as loosely as could be attained within the constraints of the F-401 vacuum system (\( \rho = 1.67 \text{ g/cm}^3; \ e = 0.854 \)). After that, it was compressed in a series of increments and then sheared (Test 2).

The compressibility of the lunar soil has been compared with that of a basalt sand simulant in air. It was found that at the medium-dense initial density, the compressibility of the lunar soil is similar to that of the basalt sand; at the lower density, the lunar soil is slightly less compressible than the basalt. The implication is that lunar soil in 1/6-g can exist at significantly different densities even at fairly great depths (>25m).

By assuming that the soil densifies with depth only due to self-weight in lunar gravity, the data in Table 1 can be used for preliminary calculations of density vs. depth, settlements due to imposed loads, and shear wave velocity vs. depth. For a given surface density, this assumption yields a lower bound estimate of the density vs. depth; that is, the actual density at a certain depth on the lunar surface can be higher than the calculated value, but it cannot be lower.

For the case of a surface density of 1.67 g/cm^3 (i.e., the same as the initial density for Test 2), it can be shown that the density would increase
very rapidly to a value of 1.72 g/cm³ at a depth of 30 cm and very slowly thereafter to a value of 1.77 g/cm³ at 8 m. It can also be shown that if a load of 10 kN/m² were imposed on the surface of this soil over a wide area (such as a lunar observatory), the settlement would be greater than 9 cm. This is a large settlement and careful foundation design would be required to maintain differential settlements within acceptable limits.

Shear wave velocity vs. depth has been calculated by means of a relationship developed by Hardin and Black (1966) for terrestrial sands in air. Assuming this relationship is valid for the lunar soil, it was found that for the Test 2 data, the shear wave velocity increases very rapidly from 0 m/sec at the surface to 61 m/sec at a depth of 30 cm, and increases very slowly with depth thereafter, the velocity being approximately proportional to the fourth-root of the depth; at 8 m, the velocity equals 140 m/sec.

SHEAR STRENGTH - The strength of the lunar sample, as measured in the three vacuum direct tests is shown in Table 2. The data are not sufficient to determine the relationship among $\phi$ (friction angle), $c$ (cohesion) and $\rho$ (density) for the lunar soil. However, both the test data and visual observation of the soil during sample preparation indicate that the value of $c$ was probably between 0 and 0.7 kN/m². For $c$ values in this range, the friction angle of the soil was $28^\circ$ for the loose sample (Test 2) and $34^\circ$ to $36^\circ$ for the dense sample (Tests 1 and 1B). As with terrestrial sands, the lunar soil strength increases with increasing density. A plot of $\tau/\sigma_0$ vs. $1/\varepsilon_0$ for the lunar soil yields a straight line passing through the origin, as is typically the case for terrestrial sands.

A comparison has been made between the strength of the lunar soil sample and that of the basalt sand simulant (measured in plane strain tests in air) at the same void ratios and vertical stresses. The lunar soil shear strength is only about 65% of the simulant strength.

GAS COMPOSITION - During consolidation, gas was released from the lunar soil each time the applied load exceeded the maximum load previously applied. These short duration pressure bursts ranged from $\Delta P = 0.6$ to $3.2 \times 10^{-8}$ torr, with the larger $\Delta P$'s occurring at the higher total loads. From a comparison of the soil pore volume to test chamber volume, it is estimated that the pressure in the soil pores was 250 times larger than the observed pressure in the chamber at the time of the bursts.

Because of the short duration of the pressure bursts, it was possible to measure only one AMU during any single burst. The following gases contributed to the bursts: $AMU = 2$ (H₂), $AMU = 4$ (He), $AMU = 28$ (unidentified). Water ($AMU = 18$), which was a major component of the background spectra, did not contribute to the observed pressure bursts. Other major gas components in the background were hydrogen and $AMU = 28$. Helium was not detected in the background spectra. Assuming that all pressure bursts were similar in composition, the gas released from the lunar soil was 60% hydrogen, 35% helium and 4.6% $AMU = 28$.

During shear, the total pressure in the test chamber rapidly rose to $4\%$ above the base pressure prior to shear. During Test 2, helium, which was monitored throughout shear, increased rapidly to $50\%$ above helium base prior to shear. Both total pressure and helium partial pressure showed periodic
very small bursts until near the maximum shear stress, at which time there was a large burst in both total pressure and helium partial pressure. The total pressure during this burst went from 8.2 to greater than $16 \times 10^{-8}$ torr, and the helium partial pressure increased from 0.6 to greater than $2.6 \times 10^{-9}$ torr. Simultaneously, a large drop in the shear force occurred.

**TABLE 1**

LUNAR SOIL VACUUM OEDOMETER DATA

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<th>Vertical Stress, $\sigma_v$ (kN/m²)</th>
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**TABLE 2**

LUNAR SOIL SHEAR TEST RESULTS

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*At end of consolidation, but prior to shear.


SUMMARY - The Apollo 14 and 15 core tubes and the Apollo 15 drill core have been tested in layered lunar soil simulants to assess the influence of core geometry and of method of advance on the quantity and quality of lunar sample recovery. The depth in the returned core samples has been related to the original depth in the lunar surface.

INTRODUCTION - The Apollo 11, 12, 14 and 15 missions have returned to earth eleven core tube samples and one drill sample representing a total of 5578 grams of regolith with well-preserved stratigraphy. The core tubes were driven into the lunar surface to various depths ranging from 25 cm for an Apollo 11 single core tube to 70 cm for an Apollo 15 double core tube. The drill core was first used on Apollo 15 and penetrated to a depth of approximately 236 cm; the six sections returned to earth contained 1344 g of lunar sample. Information concerning the true depth from which a core sample originated is of particular importance to investigators studying gradients of such properties as solar wind composition, cosmic-ray track densities, and thermoluminescence. Depth relationships for the Apollo 11 and 12 core tube samples were presented previously (Carrier, et al., 1971); the present paper reports additional results for the Apollo 14 and 15 core tubes and for the Apollo 15 drill core. The simulant and procedures used in obtaining these results are in general as described in our earlier paper; exceptions are pointed out in applicable sections of the paper.

The shapes and dimensions of the bit ends of the Apollo 14 and 15 core tubes and the Apollo 15 drill core are presented in Fig. 1. The Apollo 15 core tube has a much greater diameter and a thinner wall than did the Apollo 14 core tube. As a result, depth relationships established earlier for the Apollo 12 and 14 core tubes do not hold for the Apollo 15 core tube. The drill core is emplaced by a rotary-percussion action. There has previously been some speculation that this motion would cause considerable disturbance of the sample gathered with the drill. One purpose of our investigation has been to evaluate the degree of disturbance.

RESULTS - 1. The depth relationships for 2 of the 3 Apollo 14 core tube samples have been determined and are shown in Fig. 2. The third sample is too disturbed for analysis. In single core sample #2022 material from a depth-in-core of 16 cm came from a depth of nearly 29 cm beneath the lunar surface. In the double core (#2045 and #2044), the bit sample was 10 cm deep as measured on the recovered core but represented material from 58 cm beneath the surface. The penetration depth for the single core indicated in Fig. 2 is within ±1.5 cm and that of the double within ±2 cm because no photographs were taken of the tubes after they had been driven into the lunar surface. In addition,
There is an experimental error of ± 5% of the depth in the core sample. That is, at a depth of 10 cm in the sample, the depth in the undisturbed soil is accurate to ± 0.5 cm; and at 40 cm, ± 2 cm. Individual layers in the Apollo 14 cores are quite well preserved but the sampling process is such that there is some compaction and some material is pushed ahead of the driven core tube so that the recovery ratio is between 49% and 63%.

2. The three core tube samples of Apollo 15, which comprised five core tube sections and returned to earth 3.3 kilograms of material, provide a sample of regolith with well-preserved stratigraphy. The depth-in-core-tube versus true depth in the regolith for these three core tubes is practically one-to-one, as shown in Fig. 3. The sample lengths in the upper halves of the two doubles cannot be determined accurately from the X-radiographs, due to image distortion caused by parallax. This introduces an uncertainty in the sample length of ± 1 cm in the case of the first double and ± 0.5 cm in the case of the second double. When the core tubes are eventually opened, this uncertainty will be eliminated. There is, however, an experimental error of ± 3% for the depth in the undisturbed soil for all of the Apollo 15 core tube samples. Sample recovery with this new core tube approaches 100%.

3. Contrary to some predictions, the agitation of the Apollo 15 drill core did not greatly disturb the in situ stratigraphy either in our pre-mission simulations or in actual use on the moon. The depth relationship for the drill core sample is probably very close to one-to-one and the densities in the six stems are close to the in situ values, due to the low porosity of the soil encountered at the drill site. Additional tests are planned.

4. Bulk density calculations for Apollo 15 core tube and drill stem samples reveal a range of values from 1.36 grams/cm³ for the top half of a double core tube to 2.15 grams/cm³ for the bottom section of the six-section drill stem. These values for the returned samples are probably close to in situ densities. Bulk density ranges for the Apollo 12 and 14 core tube samples were 1.74 to 1.98 and 1.60 to 1.75, respectively. More investigative effort is needed to ascertain the reason for the bulk density variations with depth and between sites. Variations in both inter-granular and intra-granular porosity as well as compositional variations could produce the variations in density.

5. The approach applied to the Apollo 14 and 15 core tubes and drill samples to determine depth relationships should be extended to samples from subsequent missions as they become available.

REFERENCE:
CORE SAMPLE DEPTH RELATIONSHIPS

W. David Carrier, III

Figure 1. - Comparison of Apollo Core Bits. Shown are cut-away views giving diameters at various sections. The Apollo 12–14 and 15 core tubes are advanced into the lunar soil by manually pushing and, if necessary, hammering. The Apollo 15 drill core is battery-powered and is advanced by a rotary-percussion action.

Figure 2. - Depth-Relationships for the Apollo 14 Core Tube Samples. These curves were prepared by driving core tubes into a lunar soil simulant containing colored layers. The bit sample for the double was returned to earth in Bag 2N and has been designated LRL Sample No. 14411; the bit sample for the single was returned in Bag 18N and is LRL No. 14414.

Figure 3. - Depth-Relationships for the Apollo 15 Core Tube Samples. The new core tubes induce very little sample disturbance and curves are nearly 1:1.
MORPHOLOGY AND CHEMICAL COMPOSITION OF METALLIC MOUNDS PRODUCED BY \( \text{H}_2 \) AND C REDUCTION OF MATERIAL OF SIMULATED LUNAR COMPOSITION. James L. Carter, University of Texas at Dallas, Geosciences Division, Dallas, Texas 75230.

A joint project was initiated with David S. McKay at NASA/MSC, Houston to test the possibility of producing by reduction processes complex metallic iron and iron sulfide mounds which are similar to those observed by numerous workers on the surface of lunar glass particles (e.g., Agrell, et al. 1970; Carter and MacGregor, 1970; McKay, et al. 1970; Carter, 1971). A glass of composition similar to dark brown lunar glass was made from reagent grade chemicals.

The preliminary experiment involved placing approximately one gram of ground glass with various amounts of sulfur in weight percent (0.0, 0.24, 0.49, 1.0) in carbon crucibles and placing them in a glow bar furnace with an argon atmosphere at 1450°C for five minutes, quenching in air and storing the resulting glass sphere in a plastic vial. Other samples were placed in alumina boats in a glow bar furnace for three minutes at 1450°C and flushed with argon, then a gas consisting of 5% \( \text{H}_2 \), 95% argon was flowed over the samples for two minutes. One sample was reduced with the hydrogen mixture for fifteen minutes.

Preliminary results of the scanning electron microscope and electron microprobe examinations are shown in Table 1. The complex iron sulfide and metallic iron mounds formed by reduction with hydrogen are generally layered (cross section shown schematically in Fig. 1). The outer layer is iron sulfide or a mixture of iron sulfide and metallic iron, the interior is metallic iron and the mound material next to the silicate host is iron sulfide. Sometimes the mounds are multilayered. Sometimes the complex mounds have a thin waist of iron sulfide. Similar waists of iron sulfide around metallic iron mounds have been seen on lunar glasses (Agrell, et al. 1970; Carter, 1971). Dimples are sometimes present. The silicate surface of the dimple is covered by dendritic sheafs of iron sulfide and isolated metallic iron mounds. In some complex mounds spherules of silicate material are present. In one example, one to five micron in diameter spherules occur on the surface of an ameboid-shaped group of metallic iron and iron sulfide mounds. The spherules consist of a particle of what appears to be aluminum oxide which is surrounded by silicate material and in turn the margin of the spherule is surrounded by iron sulfide. No metallic mounds with inclusions of silicate spherules have been recognized on lunar glasses.

The complex iron sulfide and metallic iron mounds formed by reduction with carbon are generally layered (cross section shown schematically in Fig. 2). The interior of a mound is a mixture of iron sulfide and metallic iron. The outer layer of pure metallic iron is generally discontinuous. The surface of the mound next to the silicate host is iron sulfide. The mounds commonly have a waist of metallic iron. No complex iron sulfide and metallic iron mounds on lunar silicate spherules have been recognized with waists of metallic iron. There is a void beneath a mound. The silicate surface of the void is
Morphology of Metallic Mounds
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covered with mounds or stringers of iron sulfide.

These data suggest that 1) the ratio of sulfur to iron is of fundamental importance to the morphological nature of metallic mounds, 2) the growth time is important to the morphological nature of metallic mounds, and 3) most metallic mounds on lunar glass spheres did not form by reducing processes in situ. However, during the melting of lunar soil such as during a major meteoritic event hydrogen, as a result of trapped solar winds in the lunar soil, may play at least a secondary role in the formation of metallic iron and may be responsible in part for the formation of metallic iron mounds with waists of iron sulfide.

Table 1. Description of mounds.

<table>
<thead>
<tr>
<th>Reducing Wt. %</th>
<th>Description of mounds</th>
</tr>
</thead>
<tbody>
<tr>
<td>H₂ 0.0</td>
<td>Mounds occur as irregular stringers or web-like metallic iron objects with crinkled surfaces. Others are flat, porous, fan-shaped metallic iron objects up to 20 microns in longest dimension. A fifteen minute run resulted in a massive network of connected circular metallic iron mounds with crinkled surfaces.</td>
</tr>
<tr>
<td>C 0.0</td>
<td>Numerous individual metallic iron mounds occur up to 50 microns in diameter which are surrounded by smaller metallic iron mounds.</td>
</tr>
<tr>
<td>H₂ 0.24</td>
<td>Mounds occur as trains of connected metallic iron octahedra. Also irregularly shaped ameboid-like complex metallic iron and iron sulfide mounds occur.</td>
</tr>
<tr>
<td>C 0.24</td>
<td>Numerous individual mounds occur up to 100 microns in diameter. Some of the larger mounds are surrounded by masses of coalesced small metallic iron mounds. Some mounds are complex mixtures of iron sulfide and metallic iron; others are porous, dendritic, iron sulfide. Dimples with an inner dimple are common. The silicate surface of the inner dimple is covered with iron sulfide mounds.</td>
</tr>
</tbody>
</table>
| H₂ 0.49        | Individual complex iron sulfide and metallic iron mounds occur up to 200 microns in diameter. Trains of connecting circular metallic iron mounds (2-15 microns in diameter) with rough surfaces are present also. The larger mounds in the trains have six-sided flat-topped metallic iron objects on their surface. The metallic iron trains grade into areas of dendritic metallic iron mounds. Occasionally ameboid-shape complex iron sulfide and metallic iron mounds up to 300 microns in longest dimension are seen. Shrinkage cracks around their margins are poorly developed. Dimples are common. Some dimples have isolated circular metallic iron mounds and irregular dendritic areas of iron sulfide on their surface. One to five microns in diameter spherules occur on the surface of some of the ameboid-shape mounds. The spherules consist of a particle of what appears to be aluminum oxide which is surrounded by silicate material and in turn the margin of the spherule is sur-
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Numerous mounds occur up to 150 microns in diameter. Some of the larger mounds are surrounded by masses of coalesced small metallic iron mounds. The mounds consist of metallic iron that is surrounded by a mixture of iron sulfide and metallic iron. Metallic iron margins are common. Dimples with an inner dimple are common. The silicate surface of the inner dimple is covered with mounds and stringers of iron sulfide.

Individual mounds occur up to 300 microns in diameter and are complex mixtures of dendritic iron sulfide and metallic iron. Some mounds 10-50 microns in diameter have a patchy layer of metallic iron over a core consisting of a mixture of iron sulfide and metallic iron. Some mounds have a thin waist of iron sulfide. The larger mounds have well developed cooling cracks around their margins. Some larger mounds contain spherules of silicate material. Dimples are common. The upper margin of the dimple is textured and has dendritic areas of iron sulfide on its surface. Individual octahedra of metallic iron approximately five microns in diameter occur.

Numerous complex individual mounds of metallic iron and iron sulfide occur up to 200 microns in diameter and are sometimes surrounded by masses of coalesced small metallic iron mounds. Individual mounds have metallic iron margins. Dimples with an inner dimple are common. The silicate surface of the inner dimple is covered with iron sulfide mounds.

**Fig. 1.** Schematic cross section of complex mound formed by reduction with hydrogen.

**Fig. 2.** Schematic cross section of complex mound formed by reduction with carbon.

Morphology, chemistry and mineralogy of some glassy particles from Apollo 14 soil 14163,88 have been investigated. In the size fraction from 0.06 to 1 mm, 30% of the particles show the same typical features of the glasses already described as "ropy strands glasses" (MEYER et al., 1971). The shape is elongated and truncated, and the surface is never smooth and bright like in the glassy spheroids. The color is light yellow to yellow brown. They appear fluted and wrinkled, with the surface entirely coated by fines and welded dust generally showing a good size-sorting. Polished sections show a typical flow structure and cavities probably related to gas bubbles. Very frequently small inclusions of Fe-Ti, Fe-S and Fe-Ni-S are evidenced by SEM nondispersive detector analyses.

An impact melting origin for such forms can be justified on the ground of the dynamic conditions of the melted material during an impact phenomenon (CARUSI et al., 1971). In case of melted material ejected in form of filaments, the momentum transferred by the rarefaction wave is not balanced by the surface tension of the fluid. The cooling process of the radiating filaments lasts enough to let the body in the fluid state, capturing material from the impact plume.

In Fig. 1, the frequency distribution versus refractive indices are shown for Apollo 12 and 14 glassy spheroids. Besides a few monomineralic glasses, two classes appear: the first one (av. nD = 1.62) is the most populated in the Apollo 14 materials, whereas it hardly appears in the Apollo 12 glasses; the second one (av. nD = 1.66) is the most abundant in the Apollo 12 samples, with a noticeable dispersion, whereas it appears as a strictly distinctive family in the Apollo 14 spheroids.

The results of the electron probe analyses of fourteen glasses from sample 14163,88 are shown (squares) in ACF triangle (Fig. 2) where some Apollo 12 glasses compositions (stars) (FULCHIGNONI...
et al., 1971) (Trigila, 1971) have been also reported. A trend is evident from the gabbro-peridotite composition, near the F corner to the gabbro and norite compositions towards the A-C side. This trend fits the one already recognized for the Apollo 12 glasses from the gabbro-peridotites to the anorthosite compositions (Chao et al., 1970). A group of Apollo 14 glasses, richer in silica, alumina, alkalies and phosphorus, are slightly shifted towards the C corner, and lie very close to the "KREEP" glasses from Apollo 12 described by Meyer et al. (1971).

The distinction of the glasses in two groups is clearly shown in Fig. 3, where two distinctive trends are evident. For the glasses in which the mafic components are quite constant (triangles), the increasing of potassium and phosphorus is shown in Fig. 4, related with a good correlation between the two elements.

On the ground of these considerations, it appears that the glasses richer in silica, alumina, potassium and phosphorus can be related with the "KREEP" glasses analyzed by Meyer et al. (1971) in Apollo 12 soils.

We have to point out that the glasses we have analyzed, which belong to that group, appear as coated ropy glasses, or coated and opaque forms.

Such features seems to agree with a very large impact melting origin. In fact, such a catastrophic event can originate a plume in which, across a large volume, the temperature remains high enough to prevent the immediate quenching of the melted material. In this case, the particles can travel long enough before freezing to capture dust fragments on their still fluid surface.

Now, if we consider that the Fra Mauro formation represent the ejecta blanket from the Imbrian event, then it is likely that the soil 14163 results so enriched in those coated glasses.

References
GLASSY PARTICLES IN APOLLO 14 SOIL

Cavarretta G.

Fig. 1

Fig. 2

Fig. 3

Fig. 4

The Apollo 14 breccias are complex and are composed of various types of lithic, glass, and mineral fragments, including microbreccia fragments that belong to earlier generations. The principal fragment types of the breccias are: 1. Vuggy, very fine-grained to fine-grained hornfelsed microbreccia of basaltic composition, containing both ortho- and clino-pyroxenes; 2. Fine to medium-grained subophitic feldspathic pigeonite basalt; 3. Fine-grained holocrystalline basaltic anorthosite, containing orthopyroxene (some of these fragments contain well developed vesicles and lithic inclusions); 4. Devitrified round glass particles, and devitrified vesicular glass charged with abundant mineral fragments; 5. Undevitrified glass fragments, spheroids, and spherules; 6. Mineral fragments of pyroxene, plagioclase, olivine, ilmenite, spinel, etc. Fragments of fine-grained anorthosite, pyroxene-dominant microbreccia, and olivine-rich rocks are also present but they are comparatively rare.

Based on the relative abundance of types of fragments and fragment and whole-breccia textures and bulk compositions, the Apollo 14 breccias can be divided into four main categories: 1. White to grayish-white friable feldspathic microbreccia, characterized by dominant plagioclase fragments plus fragments of fine-grained basaltic anorthosite and purplish-brown dark, devitrified glass charged with angular mineral fragments; 2. Gray to dark-gray well indurated to moderately coherent feldspathic breccia, characterized by abundant fragments of black to gray, vuggy, hornfelsed basaltic microbreccia and fragments of feldspathic pigeonite basalt; 3. Shocked, non-porous regolith microbreccia, characterized by abundant devitrified glasses; and transported fragments and 4. Unshocked, porous, regolith microbreccia, characterized by undevitrified glass particles and a wide variety of lithic fragments similar to those described above. The first two types of breccias are largely free of undevitrified glass and are probably derived from similar source regions, whereas the other two types are products of mixed sources.

The distinction between discrete fragments and matrix in well indurated breccias is generally difficult because of the gradational size change between lithic and mineral fragments. This distinction is even more obscure in shocked and thermally metamorphosed breccias where the porosity disappears and grain or fragment boundaries merge. Nevertheless, it appears that many of the gray well indurated type 2 feldspathic breccias contain both discrete annealed fragments and annealed matrix. It is not known whether these
annealed and thermally metamorphosed matrices are characteristic of entire hand samples. In type 2 breccias which are moderately coherent, textures indicate that only the vuggy, hornfelsed microbreccia fragments are thoroughly annealed but most of the matrix fragments are not. Some of the fragments of vuggy, black, very fine-grained hornfelsed microbreccias may have formed by recrystallization or devitrification of a melt charged with numerous angular mineral xenocrysts. Others may have been intensely granulated prior to annealing. The fragments of hornfelsed microbreccias occur side by side with fragments of feldspathic pigeonite basalts or with other mineral fragments without developing any interlocking textures in moderately coherent type 2 breccias. Hence the hornfelsed microbreccias probably were thermally metamorphosed prior to incorporation into the complex breccia.

Weakly to moderately shocked mineral and rock fragments are present but un shocked fragments predominate in all breccias. Except for glasses of impact origin, strongly shocked fragments are very rare.

It is probable that the type 1 friable feldspathic microbreccias represent the white materials that can be seen in photographs of the large breccia blocks and boulders near Cone Crater. The type 2 well indurated and moderately consolidated feldspathic breccias, and possibly also the type 3 shocked regolith breccias, could represent the dominant gray to dark gray portions of such breccia blocks and boulders. These are probably the principal rock types of the Fra Mauro formation. The type 4 un shocked, porous regolith microbreccias consist largely of reworked material from the first three types of breccias, plus an additional small amount of particles from distant unknown sources, and they are probably widespread over the Fra Mauro site.

Photogeologic interpretation by other authors indicates that the Fra Mauro formation consists of ejecta which was derived from the Imbrium Basin and thrown out by the Imbrian event. If the basalt fragments in the type 2 breccias are part of the Fra Mauro formation, then the age of the Imbrian event would fall between 3.32 (the age of Apollo 15 basalt 15555) and 3.95 billion years (the age of a basalt fragment in a type 2 Apollo 14 breccia).
On the basis of color, refractive index (R.I.), and chemical composition, many of the Apollo 14 glasses of impact origin differ from those of Apollo 11 and 12. However, like those of Apollo 11 and 12, the Apollo 14 glasses tend to group with respect to chemical composition. We interpret these chemical groups as representing specific rock types.

There are at least six distinct chemical groups of Apollo 14 glasses of impact origin. Group 1 is the most abundant and it includes those similar in chemical composition to the glasses designated as Group 7 in the Apollo 11 fines (Chao et al., 1970).* It consists of pale yellow, greenish yellow, golden, to golden brown colored glasses with R.I. ranging from 1.5977 to 1.6185. The bulk chemical composition and normative composition (containing normative quartz) of these glasses are similar to and lie within the range of the regolith microbreccias and fines of Apollo 14. Hence they represent fused fines or regolith microbreccias. Group 2 consists of colorless to pale green glass particles with R.I. ranging from 1.5937 to 1.5947. This group has a very high normative anorthite content and is similar to the anorthositic glasses of Apollo 11 and 12. Group 3 consists of yellowish green to green glass particles with R.I. ranging from 1.6339 to 1.6378. They are characterized by low TiO$_2$ (0.56 weight %), Al$_2$O$_3$ (9.5%), and CaO (9.2%), and high FeO (17.7%), MgO (14.0%) and Cr$_2$O$_3$ (0.51%). The glasses of Group 4 also are green but have R.I. from 1.6589 to 1.6607. They are distinctly lower in Al$_2$O$_3$ (6.8 weight %) and CaO (8.6%), and higher in TiO$_2$ (0.91%), FeO (22.8%), and MgO (15.1%) compared to the green glasses of Group 3. Group 5 consists of reddish brown to brownish red glasses with R.I. ranging from 1.6649 to 1.6948 and is characterized by low silica (42.4 weight %) and high titania (7.8%). Group 6 consists of two colorless glass particles with very low R.I., 1.4961 and 1.4969. They are similar to those reported to be high in SiO$_2$ and K$_2$O. An additional group of heterogeneous greenish yellow to gray and opaque glasses is similar to Group 1 in bulk chemical composition except that they have a higher content of alkalies (Na$_2$O 0.60 to 0.96 weight %, K$_2$O 0.18 to 0.64%) and P$_2$O$_5$ (0.20 to 0.51%). The chemical variations of Apollo 14 glasses are similar to those of Apollo 11 and 12. R.I. varies

directly with the sum of FeO and TiO₂ content and inversely with the sum of CaO and Al₂O₃. FeO has negative correlation with Al₂O₃ and CaO has positive correlation with Al₂O₃.

None of the glasses we have analyzed so far resembles either the feldspathic pigeonite basalt (14310) or the ophitic basalt (14053). The only glass group we can correlate with a parent material is Group 1 which falls within the bulk chemical range of local Apollo 14 fines and regolith breccias. We have not been able to find large lithic fragments with bulk or normative composition similar to glasses of Groups 2 to 6. Glasses of Group 6 are very rare. Glasses of Group 5, although not as rare, are probably from a distant source such as a mare area where the titanium content of the source rock is high. The green glasses, particularly in Group 3 and to a slightly lesser extent in Group 4, are abundant in Apollo 14 soils. They may represent rock types that are included only as small chips in the various types of Apollo 14 microbreccias of the Fra Mauro formation. Glasses of Groups 3, 4 and 6 are unlike any Apollo 11 or 12 glasses; they are specific to Apollo 14 fines.

To be presented by title only
MINERALOGICAL ASPECTS OF APOLLO XIV SAMPLES:
LUNAR CHONDRULES, PINK SPINEL BEARING ROCKS, ILMENITES.

Paris VI, France, R. Pierrot, BRGM, 45 Orléans 02, France.

A.- Lunar chondrules.

A few "chondrules" - most likely of lunar origin - have been studied in soil 14003 (fines < 1 mm) by means of scanning electron microscope (SEM) and petrographic microscope. The most striking observations are reported below.

1.- A remarkable step growth of the silicates is found by SEM on the surface of a chondrule of ophitic basalt composition. The rock is composed of radiating feldspar and interstitial pyroxene accompanied by automorphous armalcolite.

2.- The surface of another chondrule is formed by distinctly visible squat pyramids suggesting a close packed arrangement of crystals. However the observation in thin section has shown no more than pyroxene dendrites developed in a glass. These dendritic aggregates have contours resembling those of single crystals.

3.- A few automorphous plagioclase crystals sticking out of the surface of a bean-shaped chondrule have been found. In thin section, rare blocky plagioclases in a microdoleritic matrix are found.

4.- One grain is composed of two cohesive chondrules separated by an iron-nickel rim; however the border is crossed by continuous feldspar laths. Skeletal olivine is the only mafic mineral. The overall morphology is reminiscent of that found in chondrites.

5.- An iron-nickel-troilite spherule of a type already discovered in Apollo 11 fines ("mini-moon") has been found during our investigation. However the shape is spherical and the surface is completely spongy and partly covered with solidified droplets.

Chondrules are much less abundant in the soil than glass spherules. Other examples have been studied which indicate, like the chondrules described here, a range of degrees of recrystallization. The cooling history of each object must be discussed individually.
Mineralogical aspects of Apollo XIV samples
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B.- Pink spinel-bearing rock.

A mineral not yet reported in Apollo 11 and 12 lunar samples has been observed in thin-section 1463-23. The presence of this mineral in several rock debris proves that is not fortuitous. It has been corroborated by LSPET results in several samples of the same site. It is a magnesian pink or brownish-pink spinel, slightly chromiferous; electron microprobe measurements yield \( \text{Mg/Mg+Fe} = 62 \text{ mole \%} \); \( \text{Cr/Cr+Al} = 8 \text{ to } 14 \text{ mole \%} \). Indices of refraction vary from 1.78 to 1.84. The debris of spinel-bearing rocks include the complete sequence of textural types found for lunar rocks: gabbro, shocked anorthosite, several partly devitrified glass fragments in which spinel is the first mineral to crystallize, breccia containing both large spinel debris and a glass similar to the one just described. Furthermore, the spinel is frequently found as individual grains. 61 crystals larger than 10 microns have been identified in a 0.75 cm² thin section.

The large number of spinel-bearing rocks contained in breccia 1463 is probably indicative of a sequence of hyperaluminous rocks not observed in the Mare regions. This sequence could be characteristic of the Fra Mauro substratum, brought to the surface by the Cone Crater event. Two spinel grains have been found in a section of breccia 14321. It should be noted that after careful search in several regolith samples from the vicinity of the LM (i.e. approximately 1 km from Cone Crater) one spinel grain has been discovered in the soil samples 14003 and 14163, none in 14259. Consequently it appears that if spinel does belong to the Cone Crater unit, the corresponding materials have not been moved very far by impact events, in particular the one responsible for the formation of the crater itself.

C.- Pseudo-armalcolite with high zirconium content.

This new phase found in fines 14003-47 has been provisionally named pseudo-armalcolite because its chemical composition is essentially that of an armalcolite, although the physical properties are different. Its formula is given below:

\[
\text{Ti}^{1.91}\text{Zr}^{0.08}\text{Fe}^{0.27}\text{Mg}^{0.10}\text{Cr}^{0.25}\text{Al}^{0.07}\text{Ca}^{0.12}\text{Si}^{0.02}
\]

It has been observed in a polished section as an optically isotropic surface 30 microns across surrounded by ilmenite. Its colour is grey, less brownish than ilmenite. Its reflectance estimated visually, is analogous to that of ilmenite. Pending radiocrystallographic investigation, it is impossible
to decide whether it is cubic, amorphous, or even whether its optical isotropy does result from a fortuitous orientation of an anisotropic crystal.

Nevertheless, the dispersion curve of its reflectances (from 400 to 700 nm) has shown that it is noticeably different by its shape and by absolute values of reflectances from those of armalcolite or ilmenite.

Its high chromium ($8.85\% \text{Cr}_2\text{O}_3$) and zirconium ($4.4\% \text{ZrO}_2$) contents are the main unusual chemical features of this pseudo armalcolite. Its chromium content would indicate (Akimoto, 1970) that if we were dealing with a genuine armalcolite, it would have been formed at very high temperature. The high zirconium content, an element usually found in specific minerals, could be explained by quenching.

Indeed it should be noted that the grain is included in a very frothy scoriaceous material, although it is observed that the associated ilmenite is not chemically unusual as to its Cr and Zr contents.

DIELECTRIC PROPERTIES OF APOLLO 14 LUNAR SAMPLES, D. H. Chung and W. B. Westphal, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139

Abstract - Laboratory characterization of dielectric properties of lunar samples 14301, 14310, 14318 and 14321 returned from Apollo 14 missions is made. Our measurements of dielectric constants and losses were made over a range of frequency from 100 hertz to 10 megahertz and temperature from -196°C to +200°C by two-terminal capacitance substitution methods. The representative values for apparent dielectric constants and losses are listed in the following:

<table>
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<tr>
<th>Sample Number</th>
<th>Sample Conditions</th>
<th>Density (gm/cm³)</th>
<th>High-Freq. K</th>
<th>High-Freq. tan δ</th>
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<tr>
<td>14301</td>
<td>dry</td>
<td>2.30</td>
<td>5 - 9</td>
<td>0.015 - 0.07</td>
</tr>
<tr>
<td>14310</td>
<td>dry</td>
<td>3.30</td>
<td>6 - 9</td>
<td>0.018 - 0.05</td>
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<tr>
<td>14321,228</td>
<td>dry</td>
<td>2.40</td>
<td>4 - 7</td>
<td>0.014 - 0.05</td>
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<tr>
<td>14321,163</td>
<td>wet</td>
<td>2.40</td>
<td>7 - 10</td>
<td>0.080 - 0.1</td>
</tr>
</tbody>
</table>

The general behavior of the dielectric constants and losses for dry lunar rocks is somewhat like that associated with the Maxwell-Wagner effect, and this effect seems to dominate the bulk dielectric properties of the lunar samples we studied. The Maxwell-Wagner effect probably arises as a result of impurity charges being limited in their motion by grain boundaries and intragranular defects. Values of the dielectric constant for lunar basalts like Sample 14310 seem to run a few percent higher than those for terrestrial basalts with similar chemical composition. Our lunar samples contained metallic free iron and other high-conductivity materials in minute amounts; we believe our observed values of relatively "high" apparent dielectric constants for lunar basalts appear consistent with the presence of these high-conductivity materials in the samples.

For a water-contaminated Sample 14321,163 the apparent dielectric constant varies with varying temperature and frequency. At temperatures where the viscosity of water decreases with increasing temperature, the apparent dielectric constant is increased rapidly. It appears that the temperature dependence of the dielectric constant is inversely proportional to the viscosity of water.

Combining the present observation with our earlier data obtained on Apollo 11 and 12 samples, we can make a generalization as follows: Because of the apparent absence of water on the lunar surface, the outer region of the moon can be considered as a low loss dielectric. This observation supports a long standing speculation that dry, powdered rocks in which the dielectric loss tangent is frequency-independent over a wide range of
frequency are present in the uppermost lunar surface layers. The surface layers of the moon are likely to have an extremely low electrical conductivity, i.e., in the order of $10^{-12}$ mhos per meter. Thus, future EM probing of the moon to a few hundred kilometers' depth is possible in the few kHz frequency range. The EM probing would provide further constraints to our understanding of the chemical and physical state of the interior of the moon, as we have indicated. If, on the other hand, subsurface water exists on the moon as indicated by recent analysis of the in situ Suprathermal Ion Detector Experiment, we would expect for the lunar interior a wide range of electrical properties. The low-frequency low-temperature dielectric constant could easily fall in the neighborhood of 100; but as frequency is increased, the dielectric constant would drop to 5 to 20 at megahertz frequencies.
OXYGEN ISOTOPE ABUNDANCES IN APOLLO 14 AND 15 ROCKS AND MINERALS;
R. N. Clayton, J. M. Hurd and T. K. Mayeda, Enrico Fermi Institute, University
of Chicago, Chicago, Illinois, 60637.

Oxygen isotopic compositions have been determined for separated mineral
phases of five crystalline rocks, and of two basaltic clasts from a fragmental
rock from the Apollo 14 collection. The data are reported in Table 1. The
isotopic compositions are very similar to corresponding results on Apollo 12
samples, mean values of which are shown for comparison. The petrographically
similar rocks 14053 and 14072 have indistinguishable isotopic patterns, and
yield a plagioclase-ilmenite isotopic temperature of 960°C, about 100°C lower
than the value for Apollo 12 crystalline rocks. The basaltic clasts 14321,
184-1A and 14321,184-3C are also very similar petrographically and isotopically
to 14053.

<table>
<thead>
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<td></td>
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</tr>
<tr>
<td>Anorthosite</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14321,184-7B</td>
<td>5.97</td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>

Apollo 12 mean values: 5.98 5.81 5.56 5.15 4.07

*Abbreviations: Pc-plagioclase; Pg-pigeonite; Cpx-calcic clinopyroxene;
Opx-orthopyroxene; Ol-olivine; Il-ilmenite

**δ18O (‰) relative to SMOW.

If the Apollo 14 rocks were excavated from some considerable depth by
the Imbrium impact event, the possibility must be considered that they might
have experienced a thermal metamorphism before impact, or a substantial heat-
ing in the impact itself. The oxygen isotopic data indicate that there has
been no re-equilibration among minerals with respect to oxygen at temperatures
below 960°C. Thus there is no record of a pre-Imbrium metamorphism. A rapid
impact heating to \( \geq 1000^\circ C \), followed by rapid quenching, as suggested by textural evidence in the fragmental rocks, would be compatible with the oxygen isotopic evidence. The K-Ar and Rb-Sr data\(^1\) imply that basaltic fragments from rock 14321 have had a similar thermal history to crystalline rock 14053. The oxygen isotope results support this conclusion.

The orthopyroxene-bearing rocks 14073, 14276 and 14310 all exhibit typical "igneous" values of oxygen isotopic fractionations, although the absence of measurable amounts of ilmenite prevents assignment of an isotopic temperature. Sample 14321,184-7B was a cluster of millimeter-sized plagioclase crystals within the fragmental rock 14321. This feldspar has the same isotopic composition as that in the crystalline rocks.

Oxygen isotopic compositions of Apollo 14 soils, and of microbreccia clasts taken from two fragmental rocks, are given in Table 2. The microbreccias have isotopic compositions equal to the whole-rock compositions of the basalts, as estimated from the mineral analyses. Thus, they differ from the Apollo 11 soil microbreccias, which are enriched in \( ^{18}O \) by about 0.2\(^{\circ}/oo \) relative to Apollo 11 basalts.

### TABLE 2

<table>
<thead>
<tr>
<th>Sample</th>
<th>( ^{18}O )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soils</td>
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<tr>
<td>14003,11</td>
<td>6.23</td>
</tr>
<tr>
<td>14163,62</td>
<td>6.14</td>
</tr>
<tr>
<td>14163,62 glass spheres</td>
<td>6.07</td>
</tr>
<tr>
<td>14259,60</td>
<td>6.20</td>
</tr>
<tr>
<td>Apollo 11 mean value</td>
<td>6.18</td>
</tr>
<tr>
<td>Apollo 12 mean value</td>
<td>5.81</td>
</tr>
<tr>
<td>Luna 16</td>
<td>6.20</td>
</tr>
<tr>
<td>Microbreccias</td>
<td></td>
</tr>
<tr>
<td>14306,35-10</td>
<td>5.72</td>
</tr>
<tr>
<td>14306,35-11</td>
<td>5.64</td>
</tr>
<tr>
<td>14321,184-4</td>
<td>5.77</td>
</tr>
</tbody>
</table>

Apollo 14 soils show an \( ^{18}O \) enrichment of about 0.5\(^{\circ}/oo \) relative to the crystalline rocks, as had been observed in Apollo 11 samples. A lesser enrichment of only about 0.2\(^{\circ}/oo \), was found for Apollo 12 soils.\(^2\) It was pointed out by Epstein and Taylor\(^3\) that this enrichment is correlated with the content of solar wind gases in the soils. Glass spheres separated from soil 14163 have a similar isotopic composition to the whole soil (< 1mm), and to spheres separated from an Apollo 11 soil.\(^4\)

Oxygen isotopic compositions of some Apollo 15 rocks and soils are given in Table 3. The \( ^{18}O \) values of plagioclase in the anorthosite 15415, and of pyroxene in the vitrophyre 15597 are at the low end of the ranges of previous
OXYGEN ISOTOPE ABUNDANCES IN APOLLO 14 & 15 SAMPLES

R. N. Clayton

P. 143

measurements of these minerals in lunar samples.

TABLE 3

OXYGEN ISOTOPIC COMPOSITIONS OF APOLLO 15 ROCKS AND SOILS

<table>
<thead>
<tr>
<th>Sample</th>
<th>$\delta^{18}O$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crystalline rocks</td>
<td></td>
</tr>
<tr>
<td>15415,13 anorthosite</td>
<td>5.78</td>
</tr>
<tr>
<td>15597,3 pyroxene</td>
<td>5.41</td>
</tr>
<tr>
<td>15597,3 devitrified glass</td>
<td>5.60</td>
</tr>
<tr>
<td>Glassy rock</td>
<td></td>
</tr>
<tr>
<td>15426,42 whole rock</td>
<td>5.41</td>
</tr>
<tr>
<td>15426,42 green glass spheres</td>
<td>5.27</td>
</tr>
<tr>
<td>Soils</td>
<td></td>
</tr>
<tr>
<td>15270,1 (&lt; 1 mm)</td>
<td>5.96</td>
</tr>
<tr>
<td>15421,20 (&lt; 1 mm)</td>
<td>5.80</td>
</tr>
<tr>
<td>15600,1 (&lt; 1 mm)</td>
<td>5.68</td>
</tr>
</tbody>
</table>

The glass-rich rock 15426 has a lower $\delta^{18}O$ than any lunar rock previously analyzed. Green glass spheres separated from this rock are almost 10/oo lower in $\delta^{18}O$ than glass spheres in Apollo 11 and 14 soils. Further interpretation of these isotopic compositions, and those of the Apollo 15 soils, must await the results of chemical and petrographic studies.

References:
This paper describes the continuation of observations of the interaction of far infrared electromagnetic radiation with lunar samples.

In the case of the fines an observed increase of mass electromagnetic attenuation coefficient with density indicates that part of the attenuation results from scattering: at 338 μm the pure absorption and scattering coefficients are equal at a porosity of \( \frac{\rho_0 - \rho}{\rho_0} = 0.7 \).

Within the wavelength range 300 - 3000 μm the scattering coefficient decreases with increasing wavelength although the ratio of the scattering to pure absorption terms becomes greater as the wavelength is increased. Within the same wavelength range the attenuation coefficient shows a marked temperature dependence; the relevance of this effect to absorption mechanisms centred in the infrared is discussed. Previous attenuation measurements in the range 1000 - 3000 μm have been confirmed using a new filter spectroscopic technique.
The results are discussed in connection with measurements of the thermal conductivity of lunar fines and direct thermal measurements on the lunar surface. It is concluded that there is a relatively sharp variation of conductivity with depth in the upper layers of the lunar regolith.
We have measured K, Rb, and Sr concentrations by isotope dilution and Sr isotope composition on rocks, breccias, and soils from Apollo 14 as well as a single crystalline rock from Apollo 15.

The Apollo 14 materials are distinguishable from Apollo 11 and 12 by their high K and Rb concentrations, and the lower contrast in the concentration of these elements in crystalline rock and soils.

The three soil samples analyzed (14049, 14163, and 14259) have similar K, Rb, and Sr concentrations and Sr isotopic composition. K concentrations range from 4000 to 4900 µg/g, Rb from 14.6 to 15.8 µg/g and Sr concentrations from 17.8 µg/g to 18.7 µg/g. When plotted on a Sr evolution diagram, all three samples plot near to one another, and have model ages ranging from 4.59 to 4.75 b.y. using \((\text{Sr}^{87}/\text{Sr}^{86})_0 = 0.6990\) and \(\lambda = 1.39 \times 10^{-11} \text{ yr}^{-1}\). Sample 14049 was originally described as a rock sample by LSPET. Our results are in agreement with the later description by Horz that the rock is actually a loosely aggregated dust clod.

Studies are in progress on breccia sample 14066, in collaboration with a consortium currently headed by C. Alexander at UC Berkeley. An individual clast (#17) which is itself a breccia has been separated from this sample. Fragments of crystalline rock have been hand-picked from this clast. These fragments consist of subhedral tablets of plagioclase in a pyroxene ground mass. This material is similar to light colored lithic fragments we have previously analyzed from Apollo 12 soil sample 12032. Measurements on whole rock and plagioclase separates on this crystalline material yield an age of 4.0 b.y., with an initial \(\text{Sr}^{87}/\text{Sr}^{86}\) ratio of 0.702. With the exception of fragments separated from breccia 12013, the whole-rock represents the most radiogenic lunar crystalline rock heretofore reported \((\text{Sr}^{87}/\text{Sr}^{86} = 0.7379)\). More abundant fine grained dark microbreccia clasts have also been separated from clast 17 of sample 14066,17. These are much less radiogenic.

The Rb-Sr data suggest that there has been at least some metamorphic migration of radiogenic strontium from the highly radiogenic crystalline fragments to these dark microbreccia clasts and the bulk sample of clast 17, although complete isotopic equilibrium may not have been achieved.

We have also analyzed a single basalt clast from breccia 14321,184. Data on cristobolite, pyroxene, and plagioclase de-
K, Rb, & Sr Measurements in Apollo 14 and 15 Material

R. A. Cliff

...fine an isochron corresponding to $4.01 \times 10^9$ yr.

Whole rock analyses have been made on crystalline rock 14310. These yield a model age of 4.55 b.y., which differs from the value of 4.37 b.y., reported by Papanastassiou and Wasserburg; our duplicate analyses for Rb and Sr concentration are in agreement with one another and distinctly higher. These differences require further investigation—the possible presence of xenoliths in this rock mentioned by Gancarz et al may be important in this context.

Measurements on pyroxene, plagioclase and whole rock from Apollo 15 crystalline rock 15555 indicate an age of 3.34 b.y.; however the results on a magnetic fraction deviate from this isochron.

The Apollo 14 samples have been interpreted as an ejecta blanket from the impact producing the Imbrium basin, and the age of approximately 3.9 b.y. reported by Papanastassiou and Wasserburg on these rocks, as well as our data on 14066, suggest that the Imbrium event occurred at this time, or possibly somewhat later. Mare basalt 15555 clearly postdates this event, and is very similar in age to the Apollo 12 basalts from Oceanus Procellarum. Combining these results with crater frequencies on mare ejecta blankets, and mare surfaces, suggest that an intense flux of interplanetary debris persisted as long as 0.6 b.y. after lunar formation, and during the interval 4.0 to 3.6 b.y. ago was declining with a half-life of about 100 m.y. This interpretation implies that during the first several hundred million years of lunar history, the moon was saturated with mare basins and their ejecta blankets. The original lunar crust will have been broken up and redistributed into breccia, and the only promising way to find older crystalline rocks is by study of breccia fragments from the oldest available ejecta blankets.
The Apollo 15 subsatellite was launched into a lunar orbit with an average altitude of 110 km and an inclination of 28° to the moon's equatorial plane. The subsatellite is spin stabilized and the spin axis is essentially parallel to the moon's axis. The spin rate is 12 rpm.

The subsatellite magnetometer is a two-axis fluxgate. The sensor is mounted at the end of a six foot boom. The two sensor elements are oriented parallel and transverse to the spin axis of the spacecraft. The measured variables that define the vector field are $B_p$, the parallel component, $B_T$, the absolute value of the transverse component, and $\phi$ the angle between the transverse component and the component of the sun-spacecraft vector transverse to the spin axis. Measurements of $B_p$ and $B_T$ only were available for use in the preliminary analysis reported upon here. The subsatellite carries a data storage unit that records field measurements on the far side of the moon for playback when the subsatellite is in view from the earth.

Remnant Magnetization. At times of relatively low levels of geomagnetic activity, the magnetic field in the geomagnetic tail is quite constant. Our preliminary analysis of the quick-look data recorded during the first traversal of the geomagnetic tail revealed the existence of measurable levels of remnant magnetism over much of the subsatellite orbit.

The major features of the structure of the measured remnant field are apparently associated with large craters. The most obvious feature is that apparently associated with the crater Van de Graff, which produces a $1 \gamma$ variation in the field as the satellite sweeps past it. Van de Graff is approximately 90 across and its center is located about 8° from the satellite ground track. Other prominent features of the data are tentatively associated with the craters Hertzsprung, Korolev, Gagarin, and Milne.

The magnetic field measurements used in this preliminary analysis do not include the final pre-flight calibrations. Thus
Magnetic Fields Near the Moon
Paul J. Coleman, Jr.

Although the measured variations are accurate, the absolute values of $B_p$ are not necessarily correct. Further, the data processing performed to date has been done entirely by hand and this precluded our determining the orientation of the component perpendicular to the satellite spin axis.

The measurements suggest that the remnant field is smoother and possibly weaker on the near side than on the far side and that most of the major craters produce measurable fields at 110 km altitude. This near side/far side asymmetry leads us to the speculation that the remnant field observed is due to irregularities in a magnetized crust. This crust has been disturbed over a broad region of the near side, possibly by the processes that created the ringed maria, but disturbed primarily by more localized crater formation on the back side.

Samples returned from the Apollo 11 and 12 sites show remnant magnetization as great as $10^{-2}$ emu/cm$^3$. If one takes this value as an upper limit on the magnetization of lunar material, then the minimum scale size of a spherical body magnetized at this level and producing a $1\gamma$ variation at the subsatellite orbit is approximately 10 km. The field at the surface of such a region, and therefore the maximum field that could be produced by such a region on the surface of the moon, is roughly 1000 $\gamma$. Such a volume would have a magnetic dipole moment of approximately $10^{16}$ gauss cm$^3$. For a more typical remnant magnetization of $10^{-5}$ emu/cm$^3$, the scale size would be 100 km and the surface field would be about 10 $\gamma$ for this dipole moment. The data also indicate that any lunar centered magnetic dipole must have a magnetic moment less than $4 \times 10^{19}$ gauss cm$^3$ corresponding to a surface field strength in the range 1.5 to 3 $\gamma$.

These preliminary results show that we will be able to obtain a detailed map of lunar remnant magnetization. This will provide additional information on the ancient magnetizing field and the history of the magnetized material subsequent to its magnetization.

Electrical Conductivity. Information on the electrical conductivity of the moon's interior has been obtained through an analysis of simultaneous magnetic field measurements at the Apollo 12 site and at the lunar orbiting satellite, Explorer 35. The results already obtained include a radial conductivity profile that has been interpreted in terms of models of mantle-core stratification, the mantle temperature, the near-surface
Magnetic Fields Near the Moon
Paul J. Coleman, Jr.

Data recorded at the lunar subsatellite during several successive orbits when the moon was in the solar wind show that there is greater variability of the magnetic field on the day side or upstream side of the moon. This behavior suggests that the magnetic field measured at the subsatellite when the moon is in the solar wind includes a component due to lunar induction. The presence of this component indicates that data from the subsatellite magnetometer, along with simultaneous data from the lunar surface magnetometers and Explorer 35 magnetometer, can be used to produce a detailed, three-dimensional model of the interior conductivity. Alternatively, the effect could be caused by ionized components of gases emitted from the moon.

Boundary Layer Studies. Observations of the magnetic field and plasma obtained with the lunar orbiter, Explorer 35, have revealed that a so-called diamagnetic cavity exists behind the moon, or downstream from the moon, when the moon is in the solar wind. The essential magnetic feature of this cavity is an interior magnetic field stronger by about 1.5 $\gamma$, on the average, than the exterior field. At the boundary of this cavity, there is a sharply localized decrease in the field magnitude approximately coincident with the boundary of the moon's optical shadow. The preliminary analysis of the data from the subsatellite magnetometer indicates that a diamagnetic increase also appears at the lower altitude of the subsatellite.

Data from Explorer 35 have also revealed the existence of sporadic field disturbances adjacent to the rarefaction wave at the boundary of the diamagnetic cavity. Studies of these disturbances have shown that they occur in the solar wind flow across the limbs when certain regions of the lunar surface are at the limbs. The greatest concentration of disturbance sources was found to be in a 15° square near the crater Gagarin.

Our preliminary analysis of the subsatellite magnetometer data indicates that strong disturbances, or limb effects, are present most of the time and, thus, over near-side as well as far-side regions. Thus, on the one hand the detection of relatively strong remnant fields in the vicinity of Gagarin is consistent with the earlier suggestion that the limb effects detected at Explorer 35 are caused by localized regions of enhanced magnetic fields. On the other hand, an indication that the limb disturbances are present more often than not, and that they are just as great when many other regions are at the limb, suggests that further study is required to establish their causes.
A precise mineral age of 4.00 ± .04 b.y. was found for the non-mare type basalt, 14310, using mesostasis intergrown with pyroxene in the higher density fractions to obtain samples of high Rb/Sr. Handpicked plagioclase and pyroxene gave low Rb/Sr samples. Our result is some 3.0% greater than the mineral age obtained for the same rock by the Cal. Tech. group, which we interpret as an interlaboratory bias in determining Rb/Sr.

A well-defined mineral age, not distinguishable from that of 14310, was obtained for the basalt 14072, using ilmenite and plagioclase concentrates to obtain a wide range in Rb/Sr. The major element chemistry of this rock (Table 1) clearly distinguishes it as a mare basalt.

Two fragments of basalt and one of troctolite were separated from breccia 14321, and their ages determined using mineral concentrates separated from ~100 mg samples. Both basalts appear slightly older than 14310 at 4.15 ± 0.1 b.y.; allowing for interlaboratory bias, their age is probably the same as the mineral age of 12013. The age of the troctolite is not distinguishable from 14310 and may be less than the associated basalts. The measured value depends critically upon the analysis of olivine as a high Rb/Sr phase (or olivine with K-feldspar inclusions), and the olivine result is sensitive to Rb contamination during processing which would depress the age. However, plagioclase, total-rock and K-feldspar data independently control the troctolite age as not exceeding that of the basalts.

Electron microprobe analyses of the basalt clasts, specifically the higher Cr$_2$O$_3$ of their pyroxenes and their less calcic plagioclases, indicate that they are not related to 14310. On the other hand, the troctolite (Table 1) which has a granular equilibrated mineral assemblage of olivine (Fo$_{86}$) plagioclase (An$_{95}$) and accessory K-feldspar, could be an accumulate from a magma resembling 14310. However any such relationship must
predate the mineral ages, as the initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 14310 at 4.0 b.y., 0.7004, greatly exceeds that of the troctolite, 0.6996. This conclusion holds whether 14310 formed by impact melting of regolith near the lunar surface at 4.0 b.y. or whether it was derived from the mantle at that time. Crystal fractionation of the troctolite at or earlier than 4.0 b.y. is indicated by its excessively old total-rock model age.

A 100 mg sample of the Apollo 15 basalt, 15555, has been dated at 3.53 ± 0.18 b.y., at the older limit of ages measured by us for Apollo 12 basalts. Petrographic and chemical data (Table 1) show that 15555 is a mare basalt similar to some Apollo 12 basalts.

The Apollo 14 basalts as representative total-rock samples combine with the Apollo 12 and low K Apollo 11 basalts to define an approximate pre-crystallization "isochron" of 4.5 b.y., which is also fitted by the various fines. This demonstrates that Rb-Sr model ages of igneous or glass fragments are maximum estimates for crystallization ages, and that in general terms, the moon is composed of 4.5 b.y. old material. However, no specific material which crystallized earlier than 4.15 b.y. has been identified. The preservation of the 4.5 b.y. alignment during the formation of the basaltic magmas has been interpreted in three ways: (i) by production of the magmas from a lunar mantle which is heterogeneous in Rb/Sr on a regional scale; (ii) by non-equilibrium partial melting of a mantle in which most of the Rb and radiogenic Sr are situated in accessory minerals, whereas most of the common Sr is in a low Ca pyroxene; (iii) by assuming that most of the basalts collected so far were first formed on the lunar surface at 4.5 b.y. but were later totally remelted without change in their bulk chemistry. The first interpretation requires that Rb/Sr in the melt must equal that of the source to within a few percent for small degrees of partial melting. This conflicts with expectations for a lunar mantle composed of olivine, pyroxenes and accessory phosphates. The second requires that the crystal size of the major minerals in the mantle should be very large (of the order 1 m) and the diffusion constant for Sr very low, in order that radiogenic Sr produced in accessory phases will equilibrate only very slowly with common Sr dispersed in pyroxene. Under these conditions, a non-equilibrium melt composed mainly of accessory phases and the exterior parts of major phases would incorporate the daughter $^{87}\text{Sr}$ of the source without separation from its parent Rb, and hence preserve the source's model age. The third interpretation can be excluded at
least for the mare basalts if the age of the Fra Mauro Formation can be placed at $\leq 4.0$ b.y. This will be so provided the dated basalt clasts were assembled within breccia 14321 during the Imbrium impact. If the 4.0 b.y. interpretation for the Imbrium event can be confirmed, it follows that the Apollo 11 and Apollo 12 basalts which are stratigraphically younger than the Fra Mauro Formation must have formed after this time, not close to 4.5 b.y.

Eight samples of Apollo 14 fines show some variation in Rb-Sr model age but are remarkably similar in major element chemistry. Their average composition is shown in Table 1.

<table>
<thead>
<tr>
<th>Crystalline rocks</th>
<th>Troctolite clast</th>
<th>&lt;1 mm fines</th>
</tr>
</thead>
<tbody>
<tr>
<td>14053</td>
<td>14072</td>
<td>14310</td>
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<tr>
<td>SiO$_2$</td>
<td>46.18</td>
<td>45.15</td>
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<tr>
<td>TiO$_2$</td>
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<td>2.57</td>
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<td>Al$_2$O$_3$</td>
<td>12.84</td>
<td>11.07</td>
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<tr>
<td>FeO</td>
<td>17.09</td>
<td>17.82</td>
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<tr>
<td>MnO</td>
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<td>0.27</td>
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<td>Mgo</td>
<td>8.59</td>
<td>12.16</td>
</tr>
<tr>
<td>CaO</td>
<td>11.18</td>
<td>9.84</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.44</td>
<td>0.32</td>
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<td>K$_2$O</td>
<td>0.11</td>
<td>0.08</td>
</tr>
<tr>
<td>P$_2$O$_5$</td>
<td>0.13</td>
<td>0.08</td>
</tr>
<tr>
<td>Cr$_2$O$_3$</td>
<td>0.37</td>
<td>0.51</td>
</tr>
<tr>
<td>S</td>
<td>0.10</td>
<td>0.12</td>
</tr>
<tr>
<td>Total</td>
<td>100.23</td>
<td>99.99</td>
</tr>
<tr>
<td>OxS</td>
<td>0.05</td>
<td>0.06</td>
</tr>
<tr>
<td>Total</td>
<td>100.18</td>
<td>99.93</td>
</tr>
</tbody>
</table>

$<1 \, \mu$m fines is average of analyses of samples 14141, 14148, 14149, 14156, 14161, 14162, 14163 and 14259.

Troctolite analysed by microprobe analysis of fused sample.
PARTICLE TRACK RECORD OF THE SEA OF PLENTY, G.M.
Comstock, R.L. Pleissher, and H.R. Hart, Jr., General Electric Research and Development Center, Schenectady, New York 12301.

Particle track counts in 41 crystals taken from layers 1 and 4 of the Luna 16 core sample reveal only slight differences between the two levels. The high track densities we observe indicate that virtually all of the grains have been highly irradiated, on or near the surface for a great length of time (in contrast to Apollo 11 and 12 soils which include many grains with \(<10^8\) tracks/cm\(^2\)). Grains could have acquired the observed median density of \(1-2 \times 10^9/\text{cm}^2\) in \(\sim 10^6\) years on the very surface (for \(\sim 100-\mu\) diameter grains) or in \(\sim 3 \times 10^9\) years at a depth of 1 cm. Since there are 3000 100-\(\mu\) layers in 30 cm (the depth to level 4), the total time required for each of them to have spent \(\sim 10^6\) years on the surface is also \(\sim 3 \times 10^9\) years, comparable to the age of the maria surfaces. Not only have nearly all of the Luna 16 grains been very close to the surface, but there has also been little "recent" admixture of previously shielded and therefore low track density material from below 30 cm. Further there is no time available in the "exposure budget" for material much below 30 cm at the Luna 16 site to have been exposed near the surface. We conclude that the regolith is unusually thin at
the Luna 16 site. This is consistent with the observation that
the Luna 16 core coarsens near the bottom.

The slight decrease in median track density observed in
going from layer 1 to layer 4 is consistent with the results of
Kashkarov on olivines from layers 1 through 5. Our cumulative
track density distribution is given in Table I and the detailed
distributions in Figure 1.

Table I: Cumulative Track Density Distribution

<table>
<thead>
<tr>
<th>DENSITY [cm$^{-2}$]</th>
<th>LEVEL 1</th>
<th>LEVEL 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>$&gt;4 \times 10^7$</td>
<td>100% (20)</td>
<td>95% (20)</td>
</tr>
<tr>
<td>$10^8$</td>
<td>100% (20)</td>
<td>90% (19)</td>
</tr>
<tr>
<td>$10^9$</td>
<td>&gt;35% (7)</td>
<td>&gt;33% (7)</td>
</tr>
</tbody>
</table>
Distribution of track densities from Level 1 (0-8 cm) and Level 4 (28-32 cm). All definite values were measured using a transmission electron microscope. In addition, the numbers to the left of the broken bars with arrows indicate the numbers of crystals for which lower limits have been estimated optically.
Results of optical and scanning electron microscopic examination of the Apollo 14 and 15 command module heat-shield windows emphasize: (1) the morphology of the micrometeoroid craters; (2) implications as to the dynamic properties of the meteoroid; (3) analysis of any meteoritic residue in the crater; and (4) correlation with the flux of particles deduced from satellite meteoroid detectors, the recovered Surveyor III camera housing, lunar rocks, and earlier spacecraft window examinations.

Ten possible meteoroid impacts are identified of which five were found on Apollo 7, one each on Apollos 8, 9, and 13 and two on Apollo 14. No impacts larger than 200 microns diameter were found on the Apollo 15 window in a preliminary survey.

A scanning electron microscopic examination of the two Apollo 14 impact craters, one about 25 microns diameter, and the other 85 x 130 microns, reveal low angle spallation and radial fractures, a shattered central zone and lack of a rounded pit. They are typical of low to medium velocity impacts and reduce the probability of a meteoroid origin.

The Apollo 13 impact crater is characteristic of a hypervelocity event with a roughly circular spall zone of 430 to 460 microns diameter and a molten central pit. The pit is unusual in that it is distinctly elliptical measuring 130 x 110 microns. The spall zone is formed by low angle fractures and contains about 10 major radial fractures. About 3/4 of the way in from the outer edge, the spall steepens sharply and forms an annulus surrounding the pit. It is clear that the lip of the pit was spalled away and only a portion of the original pit remains. The pit and spall zone contain the same particulate contamination from the spacecraft thrusters, sea-water and mylar heat-shield coating as found on all the window surfaces from previous missions. No trace of any projectile material was found in or around either the Apollo 13 or 14 pits.

Meteoroid mass calculations for each of the probable impacts are based on a spall diameter to meteoroid diameter ratio of 100 to 1 and a mass density of 2 gm/cm³. Flux calculations consider partial shielding of the window surfaces by the Lunar Module in cislunar space and total shielding during the lunar orbital phase. The results are compared with a near-Earth meteoroid environment model (1) shown by the solid line in Figure 1.

The flux estimates obtained from Apollos 8, 10, 12, 13, 14, and 15 are seen to be an extension of the lunar surface data points obtained from the Surveyor III camera examination (2) and the Lunar Orbiter penetration detectors (3); and the Apollo 7 result in Earth orbit is in agreement with the reference environment model. The ratio of the near-Earth meteoroid flux to the interplanetary flux is 2.04 by comparing the Apollo 7 result with the other Apollo data. This is in agreement with the expected gravitational enhancement of the meteoroid flux due to the Earth’s field.

Assuming that the Surveyor III interplanetary mass distribution for the lunar
Apollo Window Meteoroid Experiment, S-176

B. G. Cour-Palais

\[ \log_{10} N (m^{-2} s^{-1}) = -10.80 - 0.56 \log_{10} m(g) \]

is applicable to the lunar rock crater distributions of Hörz et al (4), the resulting data points for rocks 12017 II and 12073 III show very good agreement, if an average exposure age of 2800 years is used. This would indicate that the current interplanetary meteoroid mass distribution has not changed significantly over an appreciable time.

References

Fig. 1. Logarithmic cumulative mass flux plot for meteoroids less than $10^{-2}$ g mass. The Apollo 7 data point is the only one for near-Earth orbit. The other Apollo data were obtained essentially outside the Earth's gravity field and compare well with the Surveyor III shroud and rock results.
The design of lunar systems from a thermal point of view and the calculation of lunar heat flow both require an accurate knowledge of the thermophysical properties of the lunar regolith, which consists primarily of fine material. The present paper reports a set of measurements of one of these properties, the thermal conductivity, for the returned Apollo 14 fines. This property is determined over the approximate range of temperatures experienced by the lunar surface layer and for a density of 1100 kg/m$^3$ corresponding to the lowest density obtainable in the laboratory.

The thermal conductivity was measured over the temperature range 120-360K by a modified line heat-source technique described by Cremers. The test cell, which holds 5-8 grams of material, depending on the sample density, was mounted in a thermal environmental chamber which permitted ambient temperature control from about 100K to 600K. This in turn was suspended in a vacuum chamber in which the pressure was held at about $10^{-6}$ torr, well below the level at which gaseous effects are important.

The effective thermal conductivity of the Apollo 14 fines was measured for a density of 1100 kg/m$^3$ which is the same as the density of the material in the returned core-tube sample. The data are shown in Figure 1. Also shown is a least-squares curve fitted through the data according to the relation

$$k = A + BT^3$$  \hspace{1cm} (1)

This temperature dependence of the conductivity of powdered evacuated dielectrics is predicted by elementary theory and fits the data fairly well. These data also show a similar temperature dependence. The physical interpretation of Equation (1) is that the effective conductivity of a porous evacuated dielectric material is made up of an approximately constant component representing solid conduction plus a term proportional to temperature cubed representing internal radiative transfer. The coefficients determined from the data for use in Equation (1) are: $A = 8.36 \times 10^{-4}$ and $B = 2.09 \times 10^{-11}$.

It is of interest to compare the thermal conductivity of the Apollo 14
fines with that of other samples. However, it was not possible to achieve such a comparably low density with the Apollo 11 and 12 fines. In both these cases it was not possible to prepare a sample with a density less than about 1300 kg/m. This indicates a significant physical difference in the Apollo 14 fines as compared with those from the two previous flights that will probably be explained when the particle sizing analyses are completed. The conductivities of the Apollo 11 and 12 samples at greater densities are larger than those given in Figure 1 as one would expect.

Although data for other lunar samples were not available at densities close to 1100 kg/m, it was possible to obtain such data for powdered terrestrial basalt under vacuum. This is shown as a least-squares curve of the form of Equation (1) by the dashed line in Figure 1. The

![Figure 1. The thermal conductivity of Apollo 14 fines at a density of 1100 kg/m.](image-url)
The density of the sample was 1130 kg/m³ which should be sufficiently close for comparison. The agreement between the curves for the Apollo 14 samples and the powdered basalt is excellent. This has not always been the case previously with the Apollo 11 and 12 samples.

Acknowledgement

The author is indebted to D. R. Talley and M. W. Vannier for their efforts in obtaining and analyzing the data. Financial assistance from the National Aeronautics and Space Administration under Grant NGR 18 001 060 is also gratefully acknowledged.

References

High energy (500-1500 ev) photoelectrons are ejected from the directly illuminated surfaces of partially illuminated rocks which are located in the dusk lunar terminator. A monopole positive charge will develop on the partially illuminated surface and will force the return of the high energy photoelectron flux to such a rock. Accretion of a fraction of the return flux on the dark, down-sun, surface of the rock results in the creation of a stable multipole charge distribution across the light/dark boundary of the rock. The intense, multipolar electric field (>10^2 volts/centimeter) evoked across the light/dark boundary can levitate micron size soil particles which become electrically charged during the charge separation process. The monopole charge will not develop if low energy photoelectrons can flow onto the directly illuminated region of the rock from adjacent dark areas. Low energy photoelectrons are produced by ultraviolet solar photons which are scattered from the directly illuminated rock to the foreground soil and then back to the dark areas of the rock. The ratio of the solar x-ray to the solar ultraviolet flux must exceed a critical value before levitation of soil particles occurs.

Forward diffraction of sunlight by such levitated particles can explain the post-sunset glow photographed on the western lunar horizon by Surveyors 6 and 7.

The physical factors which govern levitation are: 1) the photoelectric work function of the lunar material; 2) the flux level of solar x-rays with wavelengths less than 25Å; 3) the distribution of x-ray sources over the solar disk; and 4) the attenuation of solar ultraviolet photons which are multiply scattered between lunar rocks and the soil.

A possible excess of fission xenon in rock 14301 was first reported by the preliminary examination team. In confirmation of this report we find fission xenon in 14301 which must be attributed to the decay of Pu\textsuperscript{244}.

Xenon data from step-wise heating of 14301 are shown in Figures 1 and 2. Extraction temperatures in hundreds of degrees celsius are indicated by the numbers. The heavy xenon isotopes demonstrate a linear correlation indicative of two-component mixing between solar xenon (SUCOR) and a fission component lying on a north-east extension of the line. Figure 2 shows the light (fission shielded) xenon isotopes. The straight-line correlation suggests a two component mixing between solar xenon and a single cosmic-ray spallation component. We therefore assume that for each temperature fraction, xenon from 14301 can be constructed from various mixtures of the three components; solar, spallation and fission.

A unique feature of 14301 is a clear separation in the release of the spallation and fission components, e.g. the most spallogenic xenon was extracted at 1200°C, a fraction low in fissiogenic xenon, while the most fissiogenic xenon was liberated at 500°C, the second lowest fraction in spallation. This property allows us to determine the isotopic composition of the pure spallation and pure fission components shown in Table 1. The computed fission spectrum is essentially identical with Pu\textsuperscript{244} fission (Alexander et al, Science 172, 1971) and different from the spectrum of uranium spontaneous or particle-induced fission.

Additional evidence for Pu\textsuperscript{244} comes from the quantity of fission xenon. If we attribute all the fission xenon to the decay of U\textsuperscript{238} the required uranium content would be 74 ppm, in contrast with our measured uranium content for an aliquot of this sample of 5 ± 2 ppm (PET value for 14301 is 3.6 ppm).

Xenon in rock 14306 is dominated by spallation and fission components. The rare gases in the light and dark regions of this rock are virtually identical as are the uranium contents. Both regions have similar Kr\textsuperscript{81}-Kr\textsuperscript{83} exposure ages with an average value of 24.4 ± 1.6 m.y. The exposure age of 14301 is 102 ± 30 m.y. In contrast with 14301 the total fission xenon content of 14306 is consistent with U\textsuperscript{238} spontaneous fission.

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The extensively studied meteorite Pasamonte has been found to contain \(8.5 \times 10^{-11}\) ccSTP Xe\textsuperscript{136} fission per microgram of uranium (Hohenberg et al., J. Geophys. Res. 72, 1967). The corresponding number for 14301 is \(3 \times 10^{-11}\). If we assume that the initial ratio of Pu\textsuperscript{244} to U\textsuperscript{238} was uniform for the original material of these two objects, and if we assume that no plutonium fission xenon has been added to 14301 by processes other than insitu fission, we can compute the time of xenon retention in 14301 relative to that of Pasamonte. It would then appear that rock 14301 formed no later than 120 m.y. after Pasamonte. Therefore, if 14301 represents pre-Imbrium crustal material which was ejected over the Fra Mauro region by the Imbrium event, then the crust and the meteorites are nearly contemporaneous, having begun xenon retention early in the primitive solar system. This interpretation however leaves unexplained the difference between 14301 and 14306. Additional measurements on a variety of Apollo 14 breccias are clearly desirable.

<table>
<thead>
<tr>
<th>ISOTOPE</th>
<th>14301 FISSION SPECTRUM</th>
<th>ISOTOPE</th>
<th>14301 SPALLATION SPECTRUM</th>
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<td>136</td>
<td>100.</td>
<td>124</td>
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<td>128</td>
<td>173. ± 20.</td>
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<td>29.4 ± 3.3</td>
<td>(129)</td>
<td>96. ± 7.</td>
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<tr>
<td>(129)</td>
<td>25.6 ± 3.1</td>
<td>130</td>
<td>84. ± 2.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>131</td>
<td>484. ± 9.</td>
</tr>
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</table>
EVIDENCE FOR EXTINCT $^{244}\text{Pu}$

G. Crozaz

P.164

TRACK STUDIES

We previously reported (Burnett et al., Geochim. Cosmochim. Acta, Suppl. 2, Vol. 2, 1971) that the fossil track density in whitlockite crystals in rock 12040 was greatly in excess of the number expected from slowing down heavy cosmic rays (as measured in adjacent feldspar grains) or spontaneous fission from $^{238}\text{U}$. Similar track excesses are found in whitlockites from Apollo 12 and 14 rocks. These anomalous densities, which are based on total pit counts, result from spallation recoils. Samples of 12040 were first annealed and then irradiated with $5.5 \pm 1.1 \times 10^{15}$ p/cm$^2$ of 28 Gev protons. The induced spallation recoil track density had an average value of $3.5 \times 10^{8}$/cm$^2$. Internal isochrons on two lunar rocks of known exposure age give a production rate roughly half this value. This high production value shows that major elements, probably Ca and also Fe impurities, are responsible for the spallation tracks. The spallation recoil tracks are very short and it is also possible to isolate the fission component by using plastic replicas and counting long tracks. Correction for spallation contributions remove the excess track densities and give reasonable agreement with known ages. The one exception to this are certain "hot spots" found in a whitlockite crystal of 14321. Track measurements have not yet been made in rock 14301. In rock 14306 the crystals appear to be too heavily shocked to give useful track data.

An extensive search was made for uranium rich inclusions in a 60 mg chip of rock 15415 with negative results. The uranium concentration in the plagioclase was $< 20 \pm 2$ ppb. The background fossil track density of $1.4 \times 10^6$ t/cm$^2$ would require a fission enhancement due to $^{244}\text{Pu}$ of a factor of 200 and is thus likely due instead to slowing down cosmic rays.

The rocks of Apollo 14 show a remarkable diversity in the distribution of uranium. In the breccia 14306, which has a pronounced light–dark structure, the dark areas tend to give a fairly uniform pattern of distributed uranium with a concentration of $\sim 18 \pm 5$ ppm (however some dark areas are noticeably lower in uranium). The white areas show regions that are largely devoid of distributed uranium but which contain intense fission stars. Qualitatively the pattern is similar to that in rock 12013. A section of rock 14311 on the other hand has a fairly uniform background of $\sim 5$ ppm with occasional zircon crystals up to 100µ in size with uranium concentration of $\sim 100$ ppm. The large crystalline rock 14310 is studded with high uranium stars separated by regions with virtually no uranium. In all cases where the star-producing grains can be identified they are associated with P or Zr bearing minerals as in previous lunar samples.
SOLAR FLARE AND GALACTIC COSMIC RAY STUDIES OF APOLLO 14 SAMPLES

G. Crozaz, R.-Drozd, C. M. Hohenberg, H. P. Hoyt, Jr., D. Ragan, R. M. Walker and D. Yuhas, Washington Univ., Lab. for Space Physics, St. Louis, Mo. 63130.

A step-wise thermal extraction of the rare gases in both light and dark fractions of 14306 shows a clear separation of spallation and fission components with little solar type gas.

The Kr-Kr ages are respectively $25.4 \pm 2.9$ and $23.4 \pm 1.4 \times 10^6$ yrs. In rock 14321, the maximum surface exposure age inferred from track counts is $25 \pm 3 \times 10^6$ yrs. The agreement of these ages and those measured in other Apollo 14 breccias suggests that this is the age of Cone Crater. Rock 14301, which has both solar gas and excess fission gas seems to have had a more complicated history giving a Kr-Kr age of $102 \pm 30 \times 10^6$ yrs.

In most lunar soil samples the majority of crystals in the >200 mesh fraction have track densities in excess of $10^8$ t/cm$^2$. This is also true for all six positions in the Apollo 15 long drill core. Exceptions are Layer VI of the Apollo 12 double core, 12033, 12030, and two samples from Apollo 14 - Cone Crater soil (14141) and the bottom of the trench (see Fig. 1). From thermoluminescence (TL) we estimate the average depth of sample 14141 to be $2.1 \pm .4$ cm. Taking twice this as the maximum depth and assigning the lowest track density crystal to this position we calculate an exposure age of $\sim 18 \times 10^6$ yrs for an undisturbed layer. In fact, the distribution of track densities are consistent with a $25 \times 10^6$ yr age and a maximum stirring (or sampling) depth of $\sim 5$ cm. Thus 14141 appears to be a sample of original Cone Crater ejecta that has lain relatively undisturbed since the original impact.

The trench bottom has similar track densities and could be Cone Crater ejecta that has recently been covered over by older, more irradiated material. The data also show that the comprehensive and bulk fines do not consist solely of Cone Crater ejecta.

The depth dependence of fossil track density and TL was measured in a vertical section of 14310. The TL signal reflects an equilibrium between the rate of ionization and thermal draining. Because thermal gradients are low, the signal in rocks is determined by the rate of ionization. The rapid rise of TL towards the surface (Fig. 2) is expected from solar flares. The TL at any glow temperature also has a characteristic equilibrium time determined by the activation energy for draining. The effect is thus analogous to radioactivity measurements and can be used to look for fluctuations in solar activity over an estimated time interval of $10^2$-$10^5$ yrs. No large fluctuations are seen.

In Fig. 3 we show a log-log plot of track density vs depth for selected lunar rocks from three missions. The data for 14310 lie close to those for 12063 and our previous conclusions on the long-term average solar flare...
SOLAR FLARES AND COSMIC RAYS

G. Crozaz

Energy spectrum \( (dN/dE = CE^{-2.6}) \) and rock erosion rates \( (3 \times 10^{-8} \text{ to } 10^{-7} \text{ cm/10}^6 \text{ yrs}) \) remain unchanged. We re-emphasize that the erosion rates refer to microerosion and set only lower limits for mass wastage.

Detailed studies of galactic cosmic ray tracks in the interior of 14310 will be given as part of a consortium report. From our data alone we would predict a true surface exposure age of \(<3 \times 10^6 \text{ yrs} \) leading to the prediction that Mn\(^{53}\) should be undersaturated at the surface.

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Fig. 1. Track densities (total SEM pit counts) in feldspars from Apollo 14 and 15 fines. The samples from Cone Crater and the bottom of the trench are clearly less irradiated than the others and must have been added to the regolith relatively recently.
Fig. 2. Normalized TL as a function of depth in rock 14310. The points were obtained by dividing the natural TL intensity at 350°C by the intensity found in a second heating of the sample after a β irradiation of 38 krad. The near surface samples are believed to be low because of optical bleaching.

Fig. 3. Variation of track density with depth in selected lunar rocks. The tracks in this depth range are produced predominantly by solar flare particles.
The viscous flow behaviors of lunar compositions 14259 and 14310 have been determined over a wide range of viscosity. Measurements have been carried out using a combination of rotating cylinder and bending beam viscosimeters to cover the range from 1 to $10^{14}$ poises. Both viscosimeters are provided with controlled atmosphere high-temperature furnaces to permit measurements in non-oxidizing atmospheres. In most cases, meaningful data could not be obtained over an intermediate range of temperature where crystallization takes place to a significant extent during the time required for accurate determination of the viscosity. The viscosity vs. temperature relations for these ranges are therefore estimated by interpolation between the higher temperature and lower temperature data. In all cases, the measurements were made on synthetic compositions, prepared by melting in molybdenum crucibles at high temperatures in a reducing atmosphere.

The results of the viscosity measurements are compared with predictions based on the model of Battinga and Weill for flow at relatively high temperatures. Useful agreement is found with the predictions of the model over the full temperature range where it is suggested to be applicable. The apparent activation energies for viscous flow in the various ranges of temperature are similar to those observed previously with terrestrial silicate liquids.

The viscosity data are also used to clarify the glass-forming characteristics and crystallization behavior of the two compositions. In this clarification, use is made of the growth rate data determined in a companion study. These data were obtained on synthetic materials prepared and tested under identical conditions to those used in the present investigation.
CRYSTALLOGRAPHIC STUDIES OF SOME APOLLO 14 PLAGIOCLASES


We are reporting results from preliminary optical, chemical and X-ray studies of lunar plagioclases from Apollo 14 rocks 14053 and 14310 (basaltic), from rock 14301 (fragmental) and from sample 14163 (fines). Some specimens were investigated before and after heat treatment.

It is known that calcic plagioclase single crystals may be composed of "domains". The size of these "domains" depends on sodium content and temperature and can be characterized by the intensity and diffusiveness of certain X-ray reflections (c-reflections) [Laves and Goldsmith, (1954), Laves, Czank, Schulz (1970)]. In addition, c-reflections investigated at room temperature provide information on the history of the earlier life of the crystal. The room temperature behaviour of the reflections informs on conditions which must have occurred in earlier times before the crystal was cooled down to the present state.

CHEMICAL AND OPTICAL DATA

Chemical analyses of the plagioclases have been carried out by electron microprobe (sample 14301) and by optical methods (samples 14310, 14053). The weakly zoned cores of the crystals from samples 14310 and 14053 have An$_{89+3}$. Four crystals from sample 14301 range from An$_{77}$ to An$_{94}$/Ab$_{19}$ to Ab$_{5}$/Or$_{3}$ to Or$_{0.7}$. Thin margins of lower An content (down to $\sim$An$_{40}$) were noticed. No antiperthitic exsolution in the feldspars investigated was observed.
CRYSTALLOGRAPHIC STUDIES OF SOME APOLLO 14 PLAGIOCLASES
M. Czank

Nearly all the plagioclases are intimately twinned. The twin laws observed are albite, carlsbad, albite-carlsbad, baveno and pericline. There are indications of twin lamellae smaller than 1 micron.

SINGLE-CRYSTAL X-RAY DATA

Seven plagioclase single crystals were examined by the precession method. These crystals were chosen to be optically as little twinned as possible. All crystals show very similar diffraction patterns with respect to a and b reflections \((h + k + l \text{ even})\) which are always sharp. The c reflections \((h + k \text{ even}, l \text{ odd})\) are fairly weak and are always very diffuse; they are elongated approximately parallel to \(b^*\) in the \(b^*c^*\) plane.

Two of the crystals have been heated at \((1015\pm10)\)°C for 36 hours and then re-examined at room temperature. The diffraction pattern of a crystal from sample 14163.166 (from fines) shows c-reflections which are less diffuse and slightly more intensive than before the heat treatment. The pattern of a crystal from chip 14053.45 (from a basaltic rock) has b-reflections whose intensity and size are not basically changed but c-reflections which are weaker. No basic change of the twinning was observed in these two crystals after heating.

Considering twin orientations, all diffraction patterns can generally be correlated with the optical observations. The X-ray intensity ratio of twins according to the albite and the carlsbad laws is in accordance with the estimated volume ratio. However, in one case there is a discrepancy between X-ray intensity and optical volume ratios, thus indicating that this particular crystal might be submicroscopically twinned.

CONCLUSIONS

1) The chemical composition of the plagioclases is similar to the results found for Apollo 11 and 12 samples.

2) The twin laws observed are similar to those reported for Apollo 11 and 12 plagioclases. However, it has to be further investigated if these twins should be considered as growth or transformation twins.
3) The heat treated plagioclase from the basalt (14053.45) must have been cooled down from about 1000°C in a longer time than the cooling rate in our experiment (from 1015°C to 300°C in 1 hour).

4) The thermal history of the crystal from the fines (14163.166) may be comparable with the conditions of heat treatment in our experiment, i.e. the material cooled slowly from ~1000°C, or was later annealed not higher than ~1000°C.

We thank Drs. Jagodzinski and Korekawa, Munich, Germany, for valuable discussions on the twinning, reported in this paper, that has been found by them, too.

Literature
IMPACT-GENERATED SHOCK AND THERMAL METAMORPHISM IN FRA MAURO LUNAR SAMPLES

M.R. Dence, Earth Physics Branch, Dept. of Energy, Mines & Resources, Ottawa

Large scale hypervelocity impacts in crystalline rocks produce a diverse suite of breccias, glasses and igneous rocks as the result of both primary shock and secondary thermal and dynamic events. Samples from Fra Mauro show strong textural similarities to terrestrial examples of such rocks and are interpreted from these analogs. In contrast to Apollo 11 and 12 suites, crystalline rocks of volcanic origin are a minor component of Fra Mauro samples.

Glasses observed in soil 14258, 34, both as individual fragments and as inclusions in breccias, and in breccia 14315, 11 show the common characteristics of single stage shock melted materials: variable composition, though on average close to the mean of the soils; commonly developed schlieren, inclusions (some shocked) generally concentrated in layers; complete absence of phenocrysts. Most common are pale yellow glasses of which glasses 9 and 12 (Table 1) are representative. Glasses of essentially the same colour and composition were recorded in some Apollo 12 fines and breccias (Proc. Second Lunar Sci. Conf.), but at Fra Mauro, they are similar to the average bulk composition of the soil (LSPET) and thus are representative of its major component. Glass fragments of other compositions are most commonly (eg. 20) depleted in normative ilmenite, apatite and potash feldspar relative to the average, though others are enriched (18, 5) in one or more of these components.

Direct shock melting may also be the origin of crystalline rock 14310. Our analysis of 14310, 4 shows subhedral plagioclase (An90-98) up to 1mm long intergrown with anhedral pyroxene of comparable size in clots 2 to 3mm across. These are interspersed with regions of finer grain size in which boxworks of laths of plagioclase contain interstitial pyroxene, alkali feldspar, ilmenite, iron, troilite and other minor phases, in addition to a highly siliceous potassic brown glass. Compositional zoning in the pyroxenes is simple. Thus some pyroxenes have low calcium cores rimmed by augite which grades to an iron-rich margin (eg. Wo4En75Fs21 → Wo35En38Fs27 → Wo14En22Fs64). While these and other features compare closely with some mare basalts, they are also compatible with near surface crystallization of a large sheet of impact melt as is found in large (>20km) terrestrial impact craters. This possibility is strengthened by the compositional similarity of 14310 to the Fra Mauro fines, the presence of inclusions (LSPET), and of schreibersite (El Goresy et al, 1971).
Impact - generated shock
M.R. Dence

P. 175

Table 1. Electron microprobe analyses of some glasses
from the sample of 1-2mm fines, 14258,34

<table>
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<tr>
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(18) bright yellow, homogeneous, small inclusions; (20) vv. pale yellow, vesicular, inclusions; (5) v. pale yellow, vesicular, homogeneous; (9) pale yellow, small vesicles, homogeneous; (12) pale yellow-brown, inhomogeneous, many inclusions; (8) orange-brown core and clear margin, interstitial with crystals.

Breccias, the most abundant rocks at Fra Mauro, include those with glass fragments (eg. 14315, 11) and those in which all matrix and glass components have been recrystallized to fine grained aggregates of pyroxene, plagioclase and ilmenite (14305,5, 14311,88 and 14314,135). Glass-bearing breccias in soil 14258,34 have compositions (Table 2, col. 1) similar to the average soil and predominant yellow glass (Table 1), while the thermally metamorphosed (annealed) breccias show more variation (Table 2, cols. 2-6) from the most common, moderately mafic variety (col. 2) to varieties richer in plagioclase. There is, however, a dearth of fragments comparable to the anorthositic or potassic granite fragments of previous sites, or of crystalline silica phases. Instead, some annealed breccias contain patches of interstitial clear to orange-brown glasses (Table 1, glass 8) generally containing crystallites of potash feldspar and other phases, comparable to residual, highly siliceous glasses in some mare basalts. These glasses appear to have been generated from siliceous fragments during annealing which also modifies other inclusions, some of which form reaction haloes. Most distinctly affected are strongly shocked fragments which are considerably more common in the annealed breccias than in components which post-date the thermal metamorphic events. A typical result is partial to complete recrystallization of maskelynite.
While the lightly welded breccias with glass fragments are probably, as at previous Apollo sites, the products of local impact events, the annealed breccias (and rock 14310) may be original Fra Mauro Formation components. By analogy with terrestrial craters it seems unlikely that the thermal effects they record 500 km from the edge of the Imbrium basin were generated by the Imbrium event, but rather by earlier intense, large impacts into noritic uplands terrain.
VOLATILIZED LEAD FROM APOLLO 12 AND 14 SOILS
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INTRODUCTION. Silver (5) and Tatsumoto (7) have made a case for mobile lead in the lunar soils. Studies of the isotopic composition of lead volatilized from lunar soils 14259, 12070, and 12033 as a function of temperature (in stepwise increments: 0°-600°, 600°-800°, 800°-1000°, and 1000°-1350°C) have been undertaken in an attempt to better understand this mobilized lead. All investigations of 12070 showed the whole-soil to lie below concordia (1, 6, 9, 10) which, coupled with data on whole-soil 12033 and breccia 12034, confirmed Tatsumoto's (8) findings on breccia 12013 that materials collected by Apollo 12 had been subjected to "third events" in the age range of 1000 ± 1000 m.y. In spite of the known complexity of the U-Th-Pb system in the soil, 12070, both for the whole-soil and for density fractions, the lead isotope volatilization data already given (2, 6) look no more complex than was observed on 10084, whose whole-soil U-Th-Pb system gave concordant ages.

We performed additional volatilization experiments on 12033 which has a whole-soil analysis that lies even more below concordia than does 12070, and in which all the density fractions are strongly affected by the "third events." The prime differences between 12070 and 12033 appear to be more basalt and very much more cindery glass in 12070 and more homogeneous and very much more ropy, brown glass in 12033 (3, 4). The brown ropy glass resembles the dark phase of breccia 12013 (3, 4) and therefore there is perhaps little surprise that the whole-soil sample lies on the same discordia line for "third events" as does 12013 [between about 800 and about 4400 m.y. (8)] and that 12070 whole-soil, poor in ropy brown glass, does not lie on the same "third events" discordia line. In conducting volatilization experiments on 12033, a distinctly different kind of soil is being investigated.

Lastly, volatilization studies were performed on 14259 because the whole-soil has a 207Pb/206Pb > 0.57. Like the other two samples, however, it also is "normally" discordant, i.e.--it lies below concordia.

DISCUSSION. Data from volatilization experiments are given for samples 14259, 12070, and 12033 in the table and the graph. Certain features are conspicuous:

(1) In all three samples, the 600°C value of 206Pb/204Pb is lower than the values at other temperatures, in accord with the data from similar studies on 10084 and 12070 by Huey et al. (2) and Silver (5, 6).

(2) The 206Pb/204Pb at 1000°C is greater than that at 800°C.

(3) The value of 207Pb/206Pb at 600°C is greater that ratio at 800°C which in turn is greater than that ratio at 1000°C, in accord with the trend given for 10084 and 12070 by Huey et al. (2) and Silver.
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(5, p. 1571; 6).

(4) The value of $^{207}\text{Pb}/^{206}\text{Pb}$ at 1350°C, however, is always greater than that ratio at 1000°C. Data on a 1350°C volatilization, where the sample is molten, have not previously been published.

(5) The volatilization data for all samples overlap at a ($\Delta^{207}\text{Pb}$)/($\Delta^{206}\text{Pb}$) model age of about 4500 m.y. (see figure), and the ($\Delta^{207}\text{Pb}$)/($\Delta^{206}\text{Pb}$) model age for the leads released in the 1350°C volatilization is always close to 4500 m.y. This is an observation of possible great importance and may indicate the presence of a refractory component in the three soils that is about 4500 m.y. old.

(6) The older the ($\Delta^{207}\text{Pb}$)/($\Delta^{206}\text{Pb}$) model age of the whole-soil, the greater the span of the values of $^{207}\text{Pb}/^{206}\text{Pb}$ for the volatilization extractions of each sample.

(7) Although the data from the volatilization experiments in general are similar to data obtained in the density fractions by Tatsumoto et al. (9) and acid leaches of 10084 by Silver (5), there are some important departures. For example, there is more spread in $^{207}\text{Pb}/^{206}\text{Pb}$ of the volatilization results on 12033 (Table 1), primarily owing to the 1350°C data, than was found in the density fractions; therefore, the sample is internally more heterogeneous than might have been guessed from the density separation experiments. On the other hand, the spread in $^{207}\text{Pb}/^{206}\text{Pb}$ for volatilized leads from 12070 is less than was found in the density fractions.

REFERENCES


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Table 1 Data for volatilization extractions of lead from lunar soils 12033, 12070 and 14259.

<table>
<thead>
<tr>
<th>Sample (and weight)</th>
<th>Temp (°C)</th>
<th>Total Pb recovered (g)</th>
<th>206Pb/204Pb</th>
<th>207Pb/204Pb</th>
<th>208Pb/204Pb</th>
<th>206Pb/204Pb</th>
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<td>12033 (0.889 g)</td>
<td>600</td>
<td>0.264</td>
<td>276.1</td>
<td>151.4</td>
<td>387.8</td>
<td>0.9445</td>
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<td></td>
<td>800</td>
<td>0.418</td>
<td>386.1</td>
<td>188.3</td>
<td>570.7</td>
<td>0.9169</td>
<td>1.1664</td>
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<td></td>
<td>1000</td>
<td>1.445</td>
<td>1825.7</td>
<td>870.3</td>
<td>4688.1</td>
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<td>0.8151</td>
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<td>1350</td>
<td>0.491</td>
<td>1042.9</td>
<td>460.6</td>
<td>974.0</td>
<td>0.9799</td>
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<td>12070 (0.844 g)</td>
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<td>47.9</td>
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<td>14259 (0.735 g)</td>
<td>600</td>
<td>0.356</td>
<td>242.1</td>
<td>123.4</td>
<td>262.5</td>
<td>0.9218</td>
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<td>800</td>
<td>2.318</td>
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Concentration of lead in whole soil 12033 averages 4.22 ppm; sum of volatilization extractions is 3.40 ppm.
Concentration of lead in whole soil 12070 averages 3.51 ppm; sum of volatilization extractions is 3.16 ppm.
Concentration of lead in whole soil 14259 averages 7.70 ppm; sum of volatilization extractions is 6.82 ppm.
OPTICAL POLARIMETRIC AND PHOTOMETRIC STUDIES OF LUNAR SAMPLES

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The photometric and polarisation characteristics of three Apollo 14 and one Luna 16 lunar fines samples in the visible and U.V. regions have been investigated using a photoelectric photo-polarimeter; Apollo 14 crystalline rocks and breccia surfaces are also being studied. Optical properties are compared with microscopic structures shown by a scanning electron microscope.

Aims

Study of the structure of planetary, asteroidal, and satellite surfaces has hitherto been confined to remote sensing by photometry, polarimetry and spectroscopy using Earth-based telescopes. Interpretation of the ensuing data has been effected by comparative means using terrestrial minerals, meteorites and returned lunar samples. The polarimetric properties of the lunar surface having already been well established from telescopic observations, the purpose of studying returned samples is threefold: to compare their polarimetric properties with lunar measurements collected at the telescope; by similar means, to investigate the optical properties at other lunar sites so as to be able to predict and account for compositional, microstructural and stratigraphic variations; to accumulate a more numerous and diverse body of data on minerals exposed to space environment, with a view to interpret photometric and polarimetric measurements collected remotely on other atmosphereless celestial bodies.

Special emphasis is given to the last consideration since the remote photometric and polarimetric sensing of Mercury, Mars, some asteroids and planetary satellites will shortly be carried out, not only by telescopic observations but from space-probes (namely the "Grand Tour" mission).
OPTICAL POLARIMETRIC AND PHOTOMETRIC STUDIES

A. DOLLFUS

Results

a) Fines: Photo-polarimetric analysis of lunar surface fines from Apollo 11 showed that their polarisation curves (that is, plots of degree of polarisation against phase angle) do not depart sensibly from those obtained from Earth-based telescopic observations of the Mare Tranquillitatis landing site. In contrast, the polarisation curves for Apollo 12, Apollo 14 and Luna 16 surface fines did not agree with remote measurements made on the respective landing sites, for which the telescopic measurements take an average.

The maximum degree of polarisation and the normal albedo, together with their spectral variation, are the parameters which best describe lunar samples. Results for Apollo 11, 12 and 14 fines, and the Luna 16 fines are given in Figs. 1 and 2.

The landing sites of Apollos 12 and 14 provided samples having a great variety of polarimetric properties. However, no individual sample departs from the linear relationship established from telescopic observations characterising a limited range of optical opacity everywhere on the lunar surface (Fig. 1). The range of spectral variation is also limited (Fig. 2), although 12028.98 (core-tube) and 12032 (ray material) depart slightly. Pulverised terrestrial extrusive rocks display greater variety.

b) Rocks and breccia: Measurements were made on several crystalline rocks and breccia, most notably the large breccia 14267, for which different regions of the exposed surface were analysed; these were freshly broken, dust-covered and glass-coated areas. Emphasis was given to the negative branch of polarisation occurring at small phase angles, as it corresponds to the range of phases covered by telescopic observations on planetary satellites and asteroids. Interpretation of the shape of these curves is given in terms of microstructure revealed by a scanning electron microscope.

Conclusion

The variety of lunar material already available is most valuable in improving our understanding of the relationship between optical properties and the nature and texture of material exposed to space. These results are being used to interpret telescopic observations on planetary satellites and asteroids, and will in future serve in the analysis of telemetered data from space-probes investigating many other atmosphereless planetary bodies.
Fig. 1. Normal albedo $A$ versus maximum degree of polarisation $P_m$ (logarithmic scales) for 144 lunar regions and 16 lunar fines samples; low $A$/high $P_m$ is at bottom left of diagram. New data show that lunar region points should be moved up by 0.15 in log $A$.

Fig. 2. Log $A$/log $P_m$ diagram together with spectral variation for 11 lunar samples. Wavelengths used are 3520, 3790, 4350, 5200, 5800 Å (1-5 respectively as for sample h); shorter wavelengths are at the lower left of each line joining sample measurements. Generally, the longer the line the redder is the sample.
THE HIGH RESOLUTION TRACK AND TEXTURE RECORD OF LUNAR BRECCIAS AND GAS-RICH METEORITES.

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Nuclear particle tracks and texture features stored in the constituent grains of micron-sized lunar dust aggregates as well as in lunar and meteoritic breccias, have been studied by combined high voltage and scanning electron microscopies, and compared to those registered in lunar dust grains artificially heated. The main purposes of this study were to detect track metamorphism in the constituent grains of the breccias in view of: 1. looking for possible differences in the nature or intensity of dynamic processes potentially active in the regolith of atmosphereless and magnetic field free planets, as a function of the distance to the Sun; 2. studying the characteristics of low energy solar nuclear particle fluxes at various locations in the solar system and at different times in the past, by finding breccia grains still showing the track record produced when they were "individually" exposed to solar radiation, before their compaction into the breccias; 3. understanding the nature of the first dust accretion step in the "early" solar nebula; 4. interpreting the differences in albedo between the "light" and the "dark" parts of peculiar types of gas-rich meteorites.

With lunar breccias the main results so far obtained are: 1. the distribution of etched tracks in polished sections, examined by scanning electron microscopy, show two very distinct groups of breccias characterized by low ($\sim 10^7$ tracks/cm$^2$) or high ($\sim 10^9$ tracks/cm$^2$) track densities respectively (figure 1); 2. the low "density" group (samples 14305, 14083, 14063, 14371) contains breccias characterized by the highest microhardnesses and the highest albedo; 3. high voltage electron microscope studies of micron-sized grains extracted from both groups of breccias, strongly suggest that the high "density" group (samples 10046, 10059, 14049, 14161) corresponds to breccias or dust "clods", in which the constituent grains still partially show their "dust" type track record, produced during their individual exposure in
solar cosmic rays, before their compaction into breccia; therefore the dynamic processes responsible for the formation of this group of breccias, have involved a relatively middle heat metamorphism of the grains. This metamorphism, which can be further scaled in temperature by measuring the size of the microcrystallites observed by 1 MeV transmission electron microscopy in the finest grains from the breccias, is perhaps correlated to the "base-surge" brecciation process, described by Mc Kay et al (1); 4. if this hypothesis is confirmed by further work, the low "density" group should characterize breccias in which the constituent grains have been processed by a much more extensive heat metamorphism, that has completely annealed any "dust" type track record in the grains; then the similarity in the degree of track metamorphism estimated for the Apollo 11 breccias is in striking contrast to the variability of the same parameter observed for the Apollo 14 breccias; 5. the formation of lunar dust clods (samples 14049 and 14161) involves processes producing no apparent track metamorphism in the grains; 6. dust particles returned to the Earth by the Luna 16 mission show clear evidence for track metamorphism; this observation suggests a complex irradiation history for these grains, which have perhaps been temporarily incorporated into "high track density" type breccias.

The track and texture records observed in the constituent grains of several "gas-rich" meteoritic breccias are different from those registered in lunar breccias, and their origin and significance is not very well understood at the time of this writing. The results and tentative conclusions of this preliminary survey still are not in good agreement with those obtained by other groups (2); they are: 1. the track record in pyroxene grains extracted from the dark part of Kapoyeta and Weston cannot be used to infer with certainty that these grains have been individually irradiated in low energy solar VH nuclei, because they show a very important spallation contribution due to galactic cosmic rays, producing both high track densities and inhomogeneous track gradients in the grains; 2. however the olivine grains in Weston have no spallation contribution; therefore their track gradients, which are generally inhomogeneous and observed in about 10% of the grains, could be indicative of the individual irradiation of the grains in solar flare VH nuclei, but many problems have to be solved before definitively ascertaining such a conclusion; 3. high voltage electron microscopy fails to reveal very clear evidence for an heavy irradiation of the finest dark part grains in several gas-rich meteorites; in particular by examining about 150 micron-sized grains in Weston, we only discovered 2 grains with texture features which could be optimistically interpreted as solar wind ultra-thin coatings and as high densities of thermally metamorphized tracks; 4. several high resolution features in gas-rich meteorites, such as the extensive microfracturing of the Kapoyeta grains,
have not yet been observed in lunar breccias; 5. the difference between
the albedo of the light and dark parts of gas-rich meteorites is not understood
and remains a stumbling block for any albedo theory, including that recently
proposed by the Orsay group to explain the optical properties of the lunar soil.

References:
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   posium, Stockholm (1971); D. J. Barber, R. Cowsik, I. D. Hutcheon 
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A NOTE ON THREE IMBRIUM SPINELS, AND A TWINNED PIGEONITE IN HIGH ALUMINA BASALT 14310

by


Imbrium Spinels

Pink, aluminous spinels in Fra Mauro feldspathic breccias have a compositional range (Haggerty 1971, Drever et al. 1972) quite different from that of the spinels in lunar basalts. In these basalts the trend is toward enrichment in Ti and Fe for Cr and Mg respectively and, according to experimental evidence, there is a miscibility gap between aluminous and titaniferous spinels (Kuan et al. 1971). Only rarely have aluminous spinels been found in samples from the Apollo 11 Mission (Keil et al. 1970), the Apollo 12 Mission (Wood et al. 1971; Reid 1971) and Luna 16 (Jakes et al. 1972).

A useful comparison can be made between the spinels in 14306 and 14055 (Drever et al. 1972) and the spinels, recently analysed by Henderson and Suddaby (1971), in a thin chrome-spinel seam between a feldspathic and an olivine rich layer in the ultrabasic layered intrusion on the isle of Rhum in Scotland (Fig. 1).

Fig. 1. Al and Mg enrichment of spinels (6-11 represents the successive compositions upward in a chrome-spinel seam) believed to be due to post-depositional reaction of chromite with olivine, plagioclase and interstitial liquid in the Rhum layered ultrabasic intrusion (Henderson and Suddaby 1971). Contrasted with the compositional range of these spinels are the compositions and possible fractionation trend of three lunar spinels (compositions A, B, C corresponding respectively with (a), (b), (c) in Drever et al. (1972); X represents the composition of spinel in the Sharps chondrite (Dodd 1971).
In the feldspathic breccias of the Fra Mauro Formation, the paragenetic association of the highly refractory aluminous spinels and aluminous plagioclase may have an important bearing on the origin of anorthositic rocks on the moon. Very light coloured, anorthite-rich material forms a prominent component of the blocks from Cone Crater, and there is little reason to doubt that these blocks and other associated breccias formed part of the original Imbrium ejecta blanket. It is therefore probable that the petrogenetic process responsible for anorthite enrichment operated at depth at the site now occupied by the Imbrium basin. The Fra Mauro aluminous spinels neither originated in rocks of the lunar highlands nor as fractionation products of mare basalts (Drever et al. 1972).

Fra Mauro breccias, relatively rich in Ca and Al, and with Cr prominent in spinel, are ultrabasic in affinity. So also are the associated clasts in which mafic minerals predominate. If these ultrabasic samples are deep-seated in origin they might be usefully compared with terrestrial rocks from layered ultrabasic intrusions. It is perhaps premature to expand this analogy before all the data on the clasts and spinels in Fra Mauro breccias have become available. But the comparison with recently analysed aluminous spinels from Rhum demonstrates (Fig. 1) that the origin of the lunar aluminous spinels is unlikely to be due to reactions of the type referred to as postcumulus (Cameron 1969) or post-depositional (Henderson and Suddaby 1971). On the other hand, the aluminous spinel enclosed in the olivine of a chondrite is remarkably similar in composition to the aluminous lunar spinels. According to Dodd (1971), this spinel crystallised early in a calcic olivine melt from which silica and alkalis may have been removed by volatilisation.

**Multiple-twinned Pigeonite**

Two crystals of pigeonite, in sub-ophitic relationship with plagioclase in Apollo sample 14310,110, exhibit multiple twinning. Examination of these crystals on the U-stage revealed that the twin axis in both is \( \{122\} \), a direction lying almost in the (100) plane of each unit of the twinned crystal. Pyroxenes twinned on this axis are rare in terrestrial rocks, having been found only in basalt from Bohemia (Iddings, 1911, p. 317), and to the best of the writers' knowledge have not previously been observed in lunar rocks. In crystal (A) the composition plane is \( \{122\} \) but in crystal (B) it is an irrational plane sub-parallel to the zone \( 120-121-120 \), making an angle of approximately 60° with \( \{122\} \). An irrational composition plane in a multiple-twinned pyroxene in a lunar rock has previously been reported by Gay et al. (1971), the twin axis in this case being \( [001] \). As suggested by these authors, the twinning in the pigeonite is probably a primary feature, possibly associated with rapid growth.

The optic properties determined for these pigeonites are: - \( b = \beta, 2H = 27° \) and \( c' = 35° \), but in both cases 2H could only be measured on one set of twin units. Rotation of crystal (B) on the U-stage revealed the presence of a narrow discontinuous zone confined to one margin, but 2H of this zone could not be determined.

Spot analyses of these crystals by electron probe yielded the following ranges in chemical composition when calculated in terms of atomic % Ca, Mg and Fe: Ca: 7.9-10.2; Mg: 60.4-64.7; Fe: 27.4-30.3 in
THREE IMBRIUM SPINELS, ETC.

H.I. Drever

There is no appreciable variation in composition between the two sets of twin units in either crystal, the variation in Fe/Mg ratio being areal in character. Crystal (A) shows the usual slight increase in Fe/Mg ratio outward in both directions from the centre, but the increase in this ratio in crystal (B) is apparently unidirectional inward from the margin. The compositions agree quite closely with those obtained for the cores of zoned pigeonites in Apollo 12 sample 12065 by Gay et al., (1971, Fig. 3, 1-6).

A line scan of crystal (B) by electron probe confirmed the presence of the narrow marginal zone and indicates that it is richer in Ca and Fe and poorer in Mg than the interior of the crystal. Its estimated composition (Ca17Mg52Fe30) is closely comparable to that recorded for the outermost part of a pigeonite in sample 12065 by Gay et al., (1971).

References


SOME TEXTURES IN LUNAR IGNEOUS ROCKS AND TERRESTRIAL ANALOGS

by

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One of the most distinctive textural characteristics of lunar igneous rocks of basaltic type is the crystallisation of olivine, pyroxene and plagioclase in immature forms. The immaturity extends, also characteristically but not invariably, to phenocrysts in porphyritic basalts. It is with this immature crystallisation and the textures it characterises that our investigation is principally concerned. Many igneous rock samples from the Apollo 12 and Apollo 15 Missions come into this category.

The approach is a selective and comparative one, particular attention being paid to textures in Procellarum samples 12002, 12009, 12021 and 12065, together with those textures which, falling within our direct experience, represent the closest terrestrial analogs. The validity of this approach depends on the relatively much greater degree to which the geological context of the terrestrial rocks is known and the better basis we have for petrogenetic interpretation. Some compositional differences, in both the bulk chemistry and the mineral components, must be borne in mind, although they do not invalidate this approach. The analogs selected are mainly from two remarkably similar Tertiary minor intrusions of ultrabasic affinity and rich, relatively to more common terrestrial igneous rocks, in Mg and Ca (Drever and Johnston 1967). The chemistry of the terrestrial samples with analogical textures is upgraded, and the following elements determined by thermal neutron activation: Na, Mg, Al, Ca, Sc, V, Cr, Mn, Fe, Co, Rb, Sr, La, Ce?, Sm, Eu, Yb, In?, Hf, Ta. Chemical analyses by electron probe should be indivisibly linked with textural analyses.

Analogous lunar and terrestrial textures are subjected to rigorous optical analysis and the results consolidated on equal area projections, particular attention being paid to a comparison of radiating and fascicular (or plumose) associations of plagioclase and pyroxene and to the development of pyroxene cores within the plagioclase. A new textural term, intra-fasciculate is introduced to denote a characteristic tendency of the pyroxene to develop in elongated shape within hollow plagioclase.

Of all igneous textures, porphyritic texture is the one with which the interpretation of the major differentiation of lunar basalts is mainly concerned. Many lunar rocks have this texture, particularly those from the Apollo 12 and Apollo 15 Missions, the most typical phenocrysts being lime-
TEXTURES IN LUNAR INTERGRANULAR ROCKS

H. J. DREVER

Poor pyroxene; less common are forsteritic olivine and anorthitic plagioclase still less. That the pyroxene phenocrysts have developed very commonly as immature hollow crystals has been stressed by Hollister et al. (1971). Conspicuously plagioclase-phyric rocks appear to be lacking on the moon although well represented on earth. The Fra Mauro high-alumina basalt 14310 is more seriate than porphyritic.

(a) Fig. 1 (b) Fig. 2

Fig. 1 (a). Drawing of a cross-section of an elongated skeletal olivine and its associated plagioclase and augite. (b) Crystallographic orientations of the minerals in (a) on an equal area projection. The rock sample was obtained from the top contact facies of the Gars-bheinn picritic intrusion, Skye, Scotland.

Fig. 2. Electron probe analysis of augites (slightly zoned) and olivine in the same sample (Fig. 1). The indicated fractionation trend is similar to certain other terrestrial augite trends (Evans and Moore 1968; Smith and Lindsley 1971) and to the initial fractionation trend of augites in lunar basalts.

In assessing the value of the methods adopted in this comparative approach reference is made here to one of a number of terrestrial analogs. This analog is found at the top contact of a picritic intrusion in which olivine, plagioclase and augite are intergrown in immature, elongated, plumose forms. In superficial terms this rock might be regarded as an aphyric association of these minerals in subophitic to subvariolitic textural relationship, suggesting rapid cotectic crystallisation of all three. On the basis of a more detailed analysis it can be discerned that there are: (1) an order of nucleation (olivine, plagioclase, augite), (2) a crystallographic radiate relationship between the feldspars and the olivine (Figs. la, lb), and (3) a metastable fractionation trend in the augites (Fig. 2).

Immature or skeletal crystals that attain pegmatitic dimensions cannot be regarded as the result of 'quenching'. If such crystals occur as phenocrysts they could have been formed by rapid, early-stage crystallisation from well-dispersed nuclei. The size of a crystal may be less critical than its form: even the huge olivine crystals in the Rhum harratite are of immature type (Drever and Johnston 1972). Lunar liquids appear to be reluctant to nucleate before attaining a higher degree of supersaturation than terrestrial liquids, the rapid growth rate after nucleation being due to this, to the relatively low viscosity (Weill et al. 1971) and relatively high
thermal conductivity. Volatilisation of alkalis at any stage would result in viscosity reduction to which there should be a textural response.

The textures investigated provide reliable information on the nucleation density, on a sequence of crystallisation, and on relative rates of crystallisation. But it is unjustifiable to infer, from the textural evidence presented, either a rapid cooling rate or the relative depth at which any mineral phase was precipitated. The supersaturation principally responsible for rapid crystallisation does not necessarily imply rapid cooling. Supersaturation, undercooling and 'quenching' are not synonymous, and a small degree of undercooling can yield a high degree of supersaturation (Wyllie 1963). The remarkable range in the immature forms of olivine in 12009 is probably a response to micro-scale environmental differences in the degree of supersaturation corresponding to different chemical diffusion gradients around each growing crystal (Drever and Johnston 1957; Chernov 1963). Associated with a rapid vectorial growth at one point there may be some cessation of growth, or even dissolution, at another. In many lunar basalts it is evident that the rate of crystallisation exceeded the rate of diffusion.

Interpretative igneous petrology requires a more rigorous approach to textural analysis, a standardisation of textural types and a rejection of antiquated terminology applied too loosely. In addition, comparatively new terms such as cumulate, which has a genetic connotation, should not be used in descriptive petrography. The occurrence of phenocrysts, even with preferred orientation, does not necessarily imply accumulation.

In applying the results of experimental phase petrology either to magmatic differentiation on a major lunar scale, or to lunar fractionation in a micro-environment, petrologists have recognised the need for reliable textural observations. But these observations are rarely rigorous enough, and attempts to correlate them with the sequence and composition of the liquidus phases may have limited validity.

References


Breccia sample 14321,184 contains many clasts which exhibit a variety of lithologies, held together by a fragmental matrix which is itself quite variable in composition. Basaltic clasts, some as large as 5 cm in greatest dimension, are present but are not common. Elemental abundances in the basaltic clasts we analysed are all similar to one another, but they do not match closely any of the previously-known compositional types. The low Ti contents (about 1.3% Ti) and the high Al (6.3 - 6.8% Al), Na (3900 - 4100 ppm Na), and K (about 1300 ppm K) contents of these basaltic clasts are noteworthy, as are their relative enrichments in the light rare earths. Numerous microbreccia clasts are present in 14321,184 and exhibit a considerable range in size, morphology, lithology, and composition. We distinguish three classes of microbreccia clasts. The oldest identified, microbreccia-1, consists of crystals of plagioclase and orthopyroxene incorporated in a matrix which appears to consist of fragments of similar minerals. Microbreccias of class 2 contain fragments of microbreccia-1 as well as clasts of norite and microgranite, and large xenocrysts of olivine, plagioclase, and ilmenite from rocks which apparently were more coarse grained than are mare basalts previously studied. Microbreccias of class 3 contain fragments of microbreccia-2 and norite and are characterized by abundant crystal fragments. They constitute the large microbreccia clasts readily discernible in 14321, which itself is a fourth-stage macrobreccia assembled from the clasts of microbreccias and igneous rocks previously mentioned, incorporated in a pale-colored matrix. The pale-colored matrix material appears on textural, mineralogical, and compositional grounds to be a mixture of fragments derived from basaltic and microbreccia clasts, generally with a higher proportion of basaltic material than of microbreccia material. Small proportions of anorthositic material, clasts of which have been observed in 14321, are also present in many samples of the matrix.

The three classes of microbreccias do not necessarily represent three different impact events because it is possible, for example, that microbreccias of class 2, which are frequently observed to have cores of large crystals, may have formed as accretionary lapilli within an impact ejecta cloud or base surge which settled as a hot ejecta blanket in which
were formed the materials that later gave rise to microbreccia-3 clasts.

We analysed 15 samples of basalt fragments, 5 of microbreccia-2, 16 of microbreccia-3, and 10 of pale-colored matrix, all extracted from sample 14321,184. Data were obtained by instrumental neutron activation analysis for 19 elements: Ti, Al, Fe, Na, K, Ba, La, Ce, Sm, Eu, Yb, Lu, Hf, Zr, Ta, Mn, Co, Sc, and Cr. Data for a few samples are given in Table 1, and copies of our complete data set are available on request. Compositional characteristics of the basaltic clasts have been mentioned above. Both microbreccias of class 2 and those of class 3 appear from their high K and rare earth contents to be compositionally dominated by KREEP-like components, although the microbreccia-2 fragments have a wider range of compositions than do the microbreccia-3 clasts in which they are incorporated. Two microbreccia-2 clasts in particular, 14321,184-15 and 14321,184-42, are very rich in KREEP components (184-15 contains almost 8000 ppm K and 260 ppm Ce; 184-42 contains about 6000 ppm K and 330 ppm Ce), and were used to represent KREEP end members in the mixing models discussed below. The composition of another microbreccia-2 fragment, 14321,184-49, seems to be dominated by plagioclase for it has high Na (1.16%) and Al (14.20%), relatively low K (3800 ppm) and low Fe (4.36%).

According to our calculations of multi-component mixing models, most of the microbreccia-3 samples can be approximated by mixtures of a high-KREEP microbreccia-2, anorthosite, and chondrite (type CC-1). The microbreccia-2 or KREEP component is usually about 80% by weight in these models. The residuals observed for these mixing models are somewhat higher than those obtained for previous mixing models for lunar clastic materials, indicating that at least one additional component is probably present in significant amounts in microbreccia-3 clasts. Potash microgranite and dunitic microbreccia clasts have been observed in 14321 and should be considered as additional components; however we have inadequate compositional data to introduce these components into our mixing models.

As mentioned above, the pale-colored matrix can be closely approximated by a mixture of basaltic clasts, microbreccia-3, and in many cases anorthosite. Probably the matrix was formed by physical abrasion of these types of clasts during formation of the microbreccia. Chondritic material beyond that already present as a component of microbreccia-3 is not required to approximate the composition of the matrix, indicating that the matrix is not related to "soils".

We should emphasize that the existence of several generations of breccias within breccias, as in 14321,184, implies that more than one major impact was involved in generating the rocks of the Fra Mauro formation. The alternative view, that only one impact was involved but that it had sufficiently complex internal structure to give rise to an object as complex as breccia 14321, is implausible.
### Table 1. Abundance Data for Samples of 14321,184

<table>
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<tr>
<th></th>
<th>1</th>
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<td>Fe%</td>
<td>12.95</td>
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<td>8.2</td>
<td>8.4</td>
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<tr>
<td>Sc</td>
<td>55.9</td>
<td>20.9</td>
<td>21.4</td>
<td>21.7</td>
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</tr>
<tr>
<td>Ti%</td>
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<td>1.08</td>
<td>1.04</td>
<td>1.20</td>
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<td>31.4</td>
<td>19.8</td>
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<tr>
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<td>33.4</td>
<td>32.1</td>
<td>38.7</td>
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</tr>
<tr>
<td>Ta</td>
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<tr>
<td>Sm</td>
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<td>41.8</td>
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<td>27.2</td>
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<td>1.55</td>
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<tr>
<td>Yb</td>
<td>6.9</td>
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<td>18.8</td>
<td>9.3</td>
</tr>
<tr>
<td>Lu</td>
<td>1.28</td>
<td>4.35</td>
<td>4.35</td>
<td>4.60</td>
<td>3.00</td>
<td>1.60</td>
</tr>
</tbody>
</table>

Abundances are in ppm unless otherwise indicated.

1: Average of 14 basaltic clasts from 14321,184
2: Average of 13 microbreccia-3 clasts from 14321,184
3: Microbreccia-2 sample 14321,184-15
4: Microbreccia-2 sample 14321,184-42
5: Microbreccia-2 sample 14321,184-49
6: Light matrix material from 14321,184; sample 14321,184-9A

Errors are as follows:

- $\pm$ 2%: Fe, Sc, Co, Na
- $\pm$ 5%: Ti, Cr, Mn, Hf, Al, La, Ce, Sm, Eu
- $\pm$ 10%: Ta, Yb, Lu
- $\pm$ 20%: K, Ba
The investigation and interpretation of the natural remanent magnetization (NRM) of lunar samples is made difficult because of the predominance of fine particles of metallic iron as a carrier of remanence and because for the most part the NRM was acquired in very weak fields. Thus, little of the basic data required for interpretation is available from paleomagnetic studies of the NRM of terrestrial rocks. Although the history of lunar samples is evidently complicated, it is likely that TRM has played some role in their magnetization. We have therefore investigated the TRM of lunar samples in very weak fields. Most of our work has been carried out with the breccia 10048-55(1), and the igneous rock 14053-48(2). We have also studied the TRM and NRM of a variety of possible analogues of lunar material, such as impactites, glasses and tektites. The magnetization of glass on 14047-47(2) is being compared with that of the rock in an attempt to distinguish between magnetization acquired at the time of formation of the rock and magnetization acquired subsequently.

Field Dependence of TRM: In order to interpret the NRM of lunar samples, it is important to know whether TRM is linear over the range of relevant fields. We have investigated the linearity of TRM of a variety of terrestrial samples and of the lunar breccia between $20 \gamma$ and $10^5 \gamma$ (1 oe). The samples were first heated to 800°C and then allowed to cool in the desired field. The experiment was carried out in a three-stage $\mu$-metal shield with Helmholtz coils inside the shield giving fine field control. To minimize sample degradation, heating was carried out in a hard vacuum. To monitor degradation, repeat determinations of TRM and control observations of saturation isothermal remanent magnetization (IRM) were made. The results showed that the TRM of all but one of the samples was essentially linear. They also revealed the progressive destruction of the carriers of remanence in the lunar breccia. Repeated heating reduced the magnitude of the TRM by approximately one-third and that of the saturation TRM by nearly two-thirds. The carriers which were destroyed were distributed evenly across the microscopic coercivity spectrum except for a disproportionately large amount in the very low coercivity range. Sample 14053-48 departs strongly from linearity in the range of fields investigated and departs from a simple power law relation in fields of a few tenths of an oersted. No destruction of magnetic carriers was observed during heating this rock.

Stability of TRM Against AF Demagnetization: AF demagnetization is a common method of analysis of NRM which serves as a means of eliminating undesirable soft contributions to remanence and as an indicator of the nature of the remanence. It is therefore of interest to obtain the AF demagnetization curves of TRM in a variety of fields and to compare them with NRM demagnetization curves. Preliminary results suggested that the TRM of the lunar samples in very weak fields was distinctively softer than TRM in higher fields.
However, further work has negated this as a useful diagnostic characteristic of very weak field TRM in 10048-55. By multiple demagnetization with up to ten measurements for each AF field value, we have been able to show that the relative stability of TRM in a wide range of fields is similar. However, the saturation TRM is less stable than TRM which suggests that remanence is carried predominantly by the fine grain iron. The stability of 14053-48 is so low that it seems probable that remanence is carried by multidomain iron, a suggestion which is consistent with the observation that in this sample TRM is less stable than saturation TRM.

Stability of TRM Against Thermal Demagnetization: Thermal demagnetization of NRM, like AF demagnetization, serves to eliminate noise and to give information about the nature of the remanence. It discriminates between components of remanence on the basis of their blocking temperatures, that is the temperature at which the relaxation time of remanence of a single particle becomes long compared with the experiment time. The sample is demagnetized by thermal cycling in field free space (+2\gamma). Components of remanence due to particles whose blocking temperatures are exceeded relax to give zero net moment. Remanence carried by particles with higher blocking temperatures is unaffected. Thermal demagnetization of TRM acquired in a variety of fields has been observed using the equipment described in the discussion of the field dependence of TRM. Demagnetization of 0.2 oe TRM in 10048-55 revealed that the TRM was fairly evenly distributed across the range of blocking temperatures with a slight enhancement at high temperatures near the Curie point. The same behavior was found in a 1000 \gamma TRM in this sample. In contrast, the blocking temperatures of TRM in 14053-48 are strongly field dependent. In a one oersted TRM, there is substantial blocking below 300° C, but one third of the magnetization is blocked between 700 and 800° C. In a 5000 \gamma field TRM, about one half of the remanence is blocked evenly across the temperature range up to 700° C and the other half is blocked between 700 and 800° C.

Comparison of TRM With NRM of Lunar Samples: We are able to compare the AF demagnetization of NRM and TRM of 10048-55 and the AF and thermal demagnetization of the NRM and TRM of 14053-48. The two AF demagnetization curves for 10048-55 are markedly dissimilar. Thus, the AF demagnetization of NRM decreases substantially in fields of tens of oersted. In contrast, the TRM decreases more slowly with increasing AF field and exhibits very little change in fields of less than 100 oe. It therefore seems unlikely that the NRM of this rock is entirely due to TRM. The AF demagnetization of the NRM of 14053-48 reveals that it is softer than TRM. The thermal demagnetization of NRM and TRM are somewhat similar, but a TRM of comparable magnitude to the NRM has more remanence blocked at very high temperature than does NRM. The field required to generate such a TRM is several tenths of an oersted, but it is unlikely that NRM is a simple thermoremanence. Thus, in the samples which we have studied the bulk of NRM is softer than TRM and in the one sample for which we have thermal demagnetization of NRM subtle differences between NRM and TRM are found. We have not seen any TRM which is as stable in demagnetization fields of a few hundred oersted as is the NRM of such specimens as 10047, 12002, 12017, 12021, 12038, 12051 and 12063.

Field Test of the NRM of Lunar Samples: In the face of the difficulty encountered in interpreting the NRM of the lunar samples, it is desirable to
design a test which places constraints on the time of acquisition of NRM. In principle, the occurrence of glass on certain samples should provide the possibility of one such test because of the difference of age between the rock and glass. Glass on 14047-47 is capable of carrying detectable remanence: the saturation IRM and the TRM acquired by the glass in a 0.1 oe field have been determined. Unfortunately, we have not yet been able to obtain a sufficiently large sample of glass to measure NRM. However, the NRM of the rock immediately below the glass is approximately one hundred times greater than that of the bulk NRM of the rock although its saturation IRM is similar to that of the rock(2). Moreover like the NRM of many lunar samples, it is more stable than IRM but less stable than TRM. Since the total mass of the sample is 15 mg of which between one fifth and one tenth is glass, the NRM is most probably a pTRM acquired at the time the glass cooled on the surface of the rock. The age of the glass is not yet known. However, it does appear that at the time the glass formed a substantial magnetic field was present at least in the immediate vicinity of the rock. This result is somewhat confused by the fact that the rock 14047 exhibits particularly strong viscosity. Nevertheless, by examining paleomagnetically suitable rocks with dated spatter important constraints may be placed on interpretations of the NRM of lunar samples.

Interpretation of the NRM of Lunar Samples: The lack of critical data precludes a satisfying interpretation of the NRM of lunar samples at this point. Yet, it does seem clear that it is not a simple TRM in the samples which we have studied. Moreover, without invoking implausibly large fields for IRM, it is in many instances difficult to explain the NRM in terms of a soft IRM superimposed upon a small stable TRM. We have therefore started to examine the characteristics of pTRM, i.e., the magnetization acquired by cooling through a limited temperature range below the Curie point in the presence of a magnetic field. In this way we have been able to obtain AF demagnetization curves similar to some observed in the lunar sample collection. Although we find difficulty in envisioning a process which could generate the necessary temperatures and fields it does appear that pTRM shows promise for duplicating the AF characteristics of NRM. It remains to be seen how well it duplicates thermal demagnetization curves. Another important possibility which should be investigated is the acquisition of pressure or shock remanent magnetization at elevated temperature. The ability to duplicate the basic characteristics of the NRM of the lunar samples in the laboratory could be an important aspect of our understanding of the phenomenon and until it can be done the implications of the NRM will inevitably remain somewhat equivocal.

(2) Nagata, T. et al., Abs. 3rd Lunar Science Conf.
(5) Strangway, D. W. et al., unpublished manuscript.
HIGH TEMPERATURE THERMOLUMINESCENCE OF APOLLO 12 LUNAR SAMPLES

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Thermoluminescence (TL), both natural and γ-ray induced, has been studied in unsorted fines 12070, 112 and in powdered samples from the interior rock chip 12051, 15, in the temperature range 20-600°C, using linear heating rates \( \beta = 5^\circ C/sec \) and \( 20^\circ C/sec \).

Figure 1 shows typical TL glow curves from the fines, both in their natural state and when subjected to artificial γ-ray doses (from a \( ^{60}\text{Co} \) source) ranging from 330 to 2700 krad. Artificial radiation causes a prominent new peak (I) near 170°C to be introduced, and the TL in the region 300°-400°C (II) to be appreciably enhanced, while having no significant effect above 400°C (III). The rock sample shows a broadly similar glow structure (Figure 2), but peak I grows at a much lower rate. A sharp hump (at \( \sim 225^\circ C \)) is observed from rock samples pre-heated to at least 500°C, which grows roughly proportionately with dose together with the main body of the TL curve. The spectral emission at temperatures below 350°C is in the green-yellow region (\( \sim 510-590 \) nm) as well, while that above 350°C is mainly in the violet to blue region (\( \sim 400-500 \) nm) only.

The initial rise method was used to determine the trap parameters. These are listed in the Table. The TL integrated over 300°-500°C from both fines and rock samples saturates at natural + \( \sim 3 \) Mrad artificial γ-ray dose. The natural dose is estimated to be \( \sim 2.5 \) Mrad in the rock chip and \( \sim 600 \) krad in the fines. The dose sensitivity (TL output per unit radiation dose) of a sample pre-heated to 500°C is enhanced by \( \sim 30-40\% \) for fines and by \( \sim 20\% \) for rock samples at a test dose of 920 krad.

Properties of peak III in lunar fines have been used in a preliminary
calculation (ignoring any non-thermal leakage) to show that if a dose rate of \( \sim 10 \) rad/yr is assumed for the lunar surface, the maximum temperature reached on the surface could not have exceeded \( \sim 105^\circ C \pm 20^\circ C \) during the \( 6 \times 10^4 \) yr needed to reach the equilibrium value of 600 krad.

Conversely, if the maximum surface temperature is taken to be \( 120^\circ C \), the \( \gamma \)-equivalent dose rate must have been \( \sim 10^2 \) rad/yr to within a factor of 10.

Table 1. Trap parameters obtained by the "initial rise method". (Dose given, 3.9 Mrad of \( \gamma \)-rays; rate of heating \( \beta = 5^\circ C/sec^{-1} \))

<table>
<thead>
<tr>
<th>Sample</th>
<th>Peak No</th>
<th>Temp. (°C)</th>
<th>Trap Depth, ( E ) (eV)</th>
<th>Frequency Factor, ( s ) (sec(^{-1}))</th>
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<tr>
<td>12070, 112</td>
<td>I</td>
<td>160</td>
<td>1.15 ± 0.05</td>
<td>( \sim 7 \times 10^{12} )</td>
</tr>
<tr>
<td></td>
<td>II</td>
<td>375</td>
<td>1.30 ± 0.05</td>
<td>( \sim 2 \times 10^{10} )</td>
</tr>
<tr>
<td></td>
<td>III</td>
<td>450</td>
<td>1.63 ± 0.10</td>
<td></td>
</tr>
<tr>
<td>12051, 15</td>
<td>I</td>
<td>175</td>
<td>1.05 ± 0.05</td>
<td>( \sim 2 \times 10^{11} )</td>
</tr>
<tr>
<td></td>
<td>II</td>
<td>400</td>
<td>1.72 ± 0.05</td>
<td>( \sim 2 \times 10^{12} )</td>
</tr>
</tbody>
</table>

UV light is found to fill empty traps below \( \sim 250^\circ C \), and drain the filled traps above this temperature, in natural samples. It progressively drains TL in \( \gamma \)-irradiated samples over the whole temperature range.

* On attachment from the Department of Physics and Astronomy Tel-Aviv University, Tel-Aviv, Israel.
Figure 1 Typical TL glow curves from the unsorted surface fines 12070, 112. The rate of heating was $\beta = 5^\circ C/sec$. A new sample of 2 mg was used for each irradiation. Grain size $d < 106 \mu m$.

Figure 2 Typical glow curves from (powdered) interior rock chip 12051, 15. Experimental conditions, and the scale of the ordinate, are as in Figure 1 (but grain size $d < 125 \mu m$). The rock sample had in all cases, an initial dose of $\sim 200$ krad of 160 MeV protons.
The Apollo 12, 14 and 15 magnetometer network has yielded unique information about the history and present physical state of the Moon. The measured remanent magnetic fields vary considerably from site to site: 38 ± 3 gammas at Apollo 12, 103 ± 5 and 43 ± 6 gammas at two Apollo 14 sites separated by 1.1 km, and 6 ± 4 gammas at Apollo 15. The strengths and variety of these field magnitudes imply that the field sources are local rather than global in extent. Sample analyses by other investigators indicate that a magnetizing field > 10^3 gammas existed when the source material at the sites cooled below the Curie temperature. The large field and sample remanence measurements indicate that material located below the randomly oriented regolith was uniformly magnetized over large areas. The Apollo 15 site lies near the edge of the mare Imbrium mascon basin; the fact that little or no remanent field exists at that site leads us to draw the preliminary conclusion that mascons are not highly magnetic.

The global magnetic response of the moon to solar and terrestrial fields varies considerably with the lunar orbital position. During times when the Moon is immersed in the steady geomagnetic tail field, the bulk relative lunar magnetic permeability is calculated to be \( \mu/\mu_0 = 1.03 \pm 0.13 \).

When the Moon is located in the free-streaming solar wind, measurements from a magnetometer on the nighttime side of the Moon can be analyzed to determine the lunar electrical conductivity profile. A three-layered model has been used to analyze Apollo 12 magnetometer data for solar wind magnetic field step-transient events. The results yield a thin non-conducting outer layer; an intermediate layer of conductivity \( \sigma_1 \sim 10^{-4} \) mhos/meter for a shell of radial thickness \( \Delta R = R_1 - R_2 \), where 0.95 \( R_{moon} \leq R_1 < R_{moon} \) and \( R_2 \sim 0.6R_{moon} \); and conductivity \( \sigma_2 \sim 10^{-2} \) mhos/meter for the core bounded by \( R_2 \). The temperature of the lunar interior can be calculated for assumed material compositions; for an olivine Moon, temperatures are calculated to be < 440°K for the crust, ~ 810°K for the intermediate shell, and ~ 1240°K for the core. Subsequently a continuous model of the lunar interior has been developed which determines more detailed radial profiles of internal conductivity and temperature.

Comparison of preliminary Apollo 15 data with the Apollo 12 data indicates that the lunar response to solar wind transients is similar at
the two sites and that Apollo 15 data, when fully processed and analyzed, will allow calculation of both horizontal and radial conductivity profiles.

Correlation with solar plasma measurements yields information concerning the plasma interaction with lunar remanent and induced fields. The 38-gamma remanent field at Apollo 12 is found to be compressed by the solar wind during times of high solar plasma density. Since the Apollo 15 remanent field is much lower than 38 gammas, detailed analysis of Apollo 15 data should allow investigation of the plasma interaction with induced lunar fields alone.

The noble gases He, Ne, Ar, Kr and Xe were measured in:
- Four ilmenite grain size fractions separated from the Apollo 11 breccia 10046.
- Seven bulk grain size fractions obtained from Apollo 12 fines 12001.
- Five ilmenite grain size fractions separated from Apollo 12 fines 12001.
- Two olivine fractions separated from Apollo 12 fines 12001.

In addition Sr, Zr, Ba, were measured in five and La, Ce, Sm, and Eu in three bulk grain size fractions.

The salient results are:

1) The \((\text{He}^4/\text{Ne}^{20})_{tr}\), \((\text{Ar}^{36}/\text{Kr}^{86})_{tr}\) and \((\text{Kr}^{86}/\text{Xe}^{132})_{tr}\) ratios in the four ilmenite fractions of 10046 are grain size independent and similar to the ratios found in the 10084 ilmenite (Eberhardt et al. 1970). The \((\text{Ne}^{20}/\text{Ar}^{36})_{tr}\) ratio, however, is highly grain size dependent, increasing from 10 for the coarsest fraction to 29 for the finest one. This may be due to an impurity (glass?) containing large amounts of Ar, Kr and Xe, but no He and Ne. This impurity may have been degassed from He and Ne during the breccia formation.

2) The average \((\text{He}^4/\text{He}^3)_{tr}\) ratio in the 10046 ilmenite is \(3060 \pm 160\). This value is distinctly higher than the corresponding ratio of \(2720 \pm 90\) found in the 10084 ilmenite (Eberhardt et al. 1970). The similarity of the \((\text{He}^4/\text{Ne}^{20})_{tr}\) ratios in the 10046 and 10084 ilmenites suggests that long time variations in the solar wind \(\text{He}^4/\text{He}^3\) ratio exist.

3) The \((\text{Ne}^{20}/\text{Ne}^{22})_{tr}\), \((\text{Ne}^{22}/\text{Ne}^{21})_{tr}\), and \((\text{Ar}^{36}/\text{Ar}^{38})_{tr}\) ratio in the 10046 ilmenite are very similar to the 10084 ilmenite. The \((\text{Ar}^{40}/\text{Ar}^{36})_{tr}\) in the 10046 ilmenite is a factor of two higher than in the 10084 ilmenite, probably reflecting time variations in the outgassing behaviour of the moon.

4) The isotopic composition of trapped Kr and Xe was evaluated from the five finest 12001 bulk grain size fractions using the \((\text{Kr}^M/\text{Kr}^N)_{m}\) versus \([\text{(Sr + 0.8 Zr)/Kr}^N]_m\) and \((\text{Xe}^M/\text{Xe}^N)_{m}\) versus \((\text{Ba}/\text{Xe}^N)_{m}\) correlation.
method (cf. Eberhardt et al. 1970). The inclusion of a correction for the variable target element chemistry in the different grain size fractions was important for the 12001 fines and led to improved values for the isotopic composition of trapped Kr and Xe (cf. Figure 1).

5) The isotopic composition of the trapped Kr in the 12001 fines (BEOC 12) is slightly different from the composition derived from the Apollo 11 fines 10084 (Eberhardt et al. 1970, Pepin et al. 1970). The BEOC 12 krypton agrees with slightly fractionated atmospheric Kr (−4.6‰ per mass unit fractionation), except for a possible excess of Kr⁸⁶ in atmospheric Kr. Such an excess of Kr⁸⁶ in atmospheric Kr could be due to fission Kr.

6) The isotopic composition of the trapped Xe in the 12001 fines (BEOC 12) agrees, within the error limits, with the composition derived from the Apollo 11 fines (BEOC 11) (Eberhardt et al. 1970). The precision of the BEOC 12 determination is superior to the BEOC 11 values. BEOC 12 xenon agrees with the SUCOR Xe isotopic composition obtained by Podosek et al. (1971) from a limited set of measurements. The abundance of the light isotopes agrees with average carbonaceous chondrite Xe (AVCC-Xe) (Eugster et al. 1967) and trapped chondritic Xe (Marti 1967) (cf. Figure 2). The heavy isotopes are less abundant and the difference can be explained as a fission component in AVCC-Xe with Xe¹³¹ : Xe¹³₂ : Xe¹³₄ : Xe¹³₆ = (52 ± 25) : (88 ± 33) : (77 ± 19) : 100.

This research was supported by the Swiss National Science Foundation.

References


Figure 1. \((\text{Xe}_{126}/\text{Xe}_{132})_m\) versus \((1/\text{Xe}_{132})_m\) and \((\text{Xe}_{126}/\text{Xe}_{132})_m\) versus \((\text{Ba}/\text{Xe}_{132})_m\) correlations. Plotting against \((\text{Ba}/\text{Xe}_{132})_m\) corrects for the variable chemical composition of the grain size fractions.

Figure 2: \(\delta\)-values, relative to average carbonaceous chondrite Xe, of the trapped Xe in Apollo 11 and Apollo 12 fine material.

The concentrations and isotopic composition of He, Ne, and Ar have been measured in a number of individual lithic fragments and glassy objects of the Apollo 11 fine 10084, and in 17 KREEP (1) fragments separated from the Apollo 12 fine samples 12033 and 12001. The size of the investigated single fragments was between 0.5 and 1 millimeter. In some cases also the isotopic abundances of Kr and Xe were determined. From these data, concentrations of trapped solar wind particles and their variability are derived, and cosmic ray exposure ages are calculated for the individual grains.

The exposure ages of 14 lithic fragments from sample 10084 are between 16 m.y. and 400 m.y. Their average exposure age of 120 m.y. is much smaller than the average exposure age of 500 m.y. of the bulk material of this lunar fine sample (2). Actually a breccia fragment and some glassy objects give higher exposure ages, and thus contribute to the high average.

The KREEP fragments were obtained by heavy liquid separation and subsequent hand-picking. Two groups were distinguished according to density (group fa 36: 2.88 < ρ < 2.96; and group fa 33: 2.96 < ρ < 3.11). Exposure ages were calculated from the Ne spall concentrations using a production rate of 0.11 x 10^-14 cm^3 Ne/gm yr. This production rate is based on a comparison of Kr spall ages and Ne spall concentrations in Apollo 12 rocks, and on the chemical composition of KREEP (3). A grouping of exposure ages near 200 m.y. is found for the 12033 fragments; the 12001 fragments appear to have a wider distribution (Figure 1). Funkhouser (4) has recently obtained similar Ne spall concentrations, although his 12033 fragments do not show as strong a grouping as the data presented here.

The average cosmic-ray produced Xe^{131}/Xe^{126} ratio in the 12033 KREEP fragments was 6 ± 1.

The trapped noble gas component represented by the Ne_{tr}^{20} concentration was systematically higher in the 12001 fragments than in the 12033 fragments.

Ar^{40}/Ar^{39} -ages were measured in two small samples of KREEP (fa 38: >150μ, 2.88 < ρ < 2.96; and fa 35: >150μ, 2.96 < ρ < 3.11) separated from 12033.
WHEN WAS THE APOLLO 12 KREEP EJECTED?

P. Eberhardt

The results are given in Figure 2. The samples show low temperature plateaus at 820 and 870 m.y., the plateau of sample fa 38 at 870 m.y. being quite well defined.

$^{38}\text{Ar}$ exposure ages of 200 and 210 m.y. were obtained for fa 35 and fa 38 respectively using a method similar to the one described by Turner et al. (5).

From the pile-neutron produced $^{37}\text{Ar}$ and $^{39}\text{Ar}$ we derive the following concentrations: fa 35: K = 0.6%, Ca = 9%; fa 38: K = 0.5%, Ca = 7%. The K concentrations give an indication of the degree of KREEP enrichment in the samples.

From our results we derive the following tentative life-history of the glassy KREEP material found at the Apollo 12 landing site:

1. A large impact (Copernicus) melts and ejects the KREEP material and transports it into the Apollo 12 area some 850 m.y. ago (from $^{40}\text{Ar}/^{39}\text{Ar}$ age).
2. Irradiation takes place at a depth of more than 50 g/cm$^2$ for several hundred million years (from spallation isotopes and the high $^{131}\text{Xe}/^{126}\text{Xe}$ ratio (6)).
3. Secondary processes (Head Crater?), some 50 to 100 million years ago, bring the material close to the surface.

This research was supported by the Swiss National Science Foundation.

References

WHEN WAS THE APOLLO 12 KREEP EJECTED?

P. Eberhardt

Figure 1. $^{21}\text{Ne}^{21}$ exposure ages of individual KREEP fragments separated from the 12033 and 12001 lunar fines.

Figure 2. $^{40}\text{Ar}^{40}/^{39}\text{Ar}^{39}$ release patterns of two KREEP samples separated from 12033. At left sample fa 35 (14.5 mg), at right sample fa 38 (10.8 mg).
OXYGEN AND OTHER MAJOR ELEMENTS IN APOLLO 14 ROCKS
AND SOME LUNAR SOILS, William D. Ehmann and David E. Gillum,
Department of Chemistry, University of Kentucky, Lexington,
Kentucky 40506.

Abundances of C, Si, Al, Mg and Fe have been determined
by 14 MeV neutron activation analysis in 3 soils and 13 rock
samples from the Apollo 14 mission. The analyses are essentially
non-destructive and the data are based on many replicate deter­
minations. In addition, 3 different sieve-size samples of
Apollo 11 soil 10084,50 and 2 samples of Luna-16 soil have been
analyzed for a number of major and minor elements by 14 MeV and
thermal neutron activation.

The majority of the Apollo 14 rocks have Si and O abundan­
ces which fall closely along a silicon-rich extension of the Si–
O regression line for the Apollo 11 crystalline rocks.

\[ \text{O}_\% = 0.98 \text{Si}_\% + 20.8 \] (Ehmann and Morgan, 1970).
The slope of this regression line is significantly different
than that for terrestrial igneous rocks. As has been found
previously for rocks from earlier missions, the Apollo 14 rocks
are depleted by approximately 1-2\% O with respect to terrestrial
igneous rocks of comparable Si contents. The Luna-16 soils
fall on a Si-O regression line similar to that for the Apollo
samples. Two Apollo 14 soils (14163,87 and 14259,65) have
higher O abundances than any other lunar material we have ana­
yzed with the exception of several portions of the unusual rock,
12013. These two soils have Si and O abundances very similar to
USGS standard rock, BCR-1. However, the high O abundances in
these soils may be due to terrestrial atmospheric contamination,
since the Office of the Lunar Sample Curator has noted that the
vacuum box in which these samples were returned did not seal
properly. Variations of less than 4 relative \% for C and Si and
10-20 relative \% for Al, Mg and Fe were obtained for data de­
ferred from analyses of 5 separate samples of the large rock,
14321. Hence, although this rock contains many clasts, the
specimens we analyzed did not exhibit gross compositional
differences for the elements we have determined to date. Based
on the data we have obtained to date, there appears to be a
direct Al-Si correlation in the Apollo 14 materials, but this
is not as clear as that we found for the Apollo 12 materials.
No distinct chemical groupings of the Apollo 14 rocks based
on Al abundances are obvious based on the data obtained to date.
Three different sieve fractions (>100 mesh, 100-200 mesh, 200-325 mesh) of Apollo 11 soil 10084,50 yielded no significant differences in their abundances of Si, Mg, Mn, Fe, or Ca. However, Al and O appear to be significantly enriched in the finest particle fraction. This is consistent with suggestions in the literature that the finest fines may be enriched in the anorthositic component. Additional analyses of the Apollo 14 materials are currently in progress and a more comprehensive interpretation of the data will be possible on completion of the studies. A summary of the data completed to date is given in Table 1.

Acknowledgement: This work has been supported by NASA Grant NGR 18-001-058.

Table 1. Major Elements in Lunar Rocks and Soils.

<table>
<thead>
<tr>
<th>Sample</th>
<th>O%</th>
<th>Si%</th>
<th>Al%</th>
<th>Mg%</th>
<th>Fe%</th>
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<tr>
<td>LS14003,13</td>
<td>43.2±0.2</td>
<td>23.2±0.1</td>
<td>9.2±0.1</td>
<td>4.6±1.0</td>
<td>8.4±0.2</td>
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<td>LS14027,29</td>
<td>41.3±0.4</td>
<td>23.0±0.2</td>
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<td>7.9±0.3</td>
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<tr>
<td>LS14053,42</td>
<td>42.1±0.2</td>
<td>22.2±0.1</td>
<td>6.8±0.1</td>
<td>5.0±0.4</td>
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<tr>
<td>LS14163,87</td>
<td>44.5±0.3</td>
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<td>9.1±0.1</td>
<td>4.9±0.2</td>
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<tr>
<td>LS14259,65</td>
<td>44.8±0.2</td>
<td>23.1±0.1</td>
<td>9.3±0.1</td>
<td>8.4±0.5</td>
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</tr>
<tr>
<td>LS14303,15(1)</td>
<td>43.5±0.4</td>
<td>23.1±0.4</td>
<td>8.6±0.1</td>
<td>7.2±0.3</td>
<td>8.6±0.1</td>
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<tr>
<td>LS14303,15(2)</td>
<td>43.6±0.4</td>
<td>23.1±0.3</td>
<td>8.8±0.2</td>
<td>6.1±0.1</td>
<td>7.9±0.2</td>
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<tr>
<td>LS14303,15-6(3)</td>
<td>42.9±0.4</td>
<td>23.0±0.2</td>
<td>8.8±0.1</td>
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<td>7.7±0.6</td>
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<tr>
<td>LS14305,76</td>
<td>41.1±0.4</td>
<td>20.4±0.2</td>
<td>7.5±0.1</td>
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<td>LS14310,113</td>
<td>43.1±0.3</td>
<td>21.9±0.1</td>
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<td>LS14311,66</td>
<td>42.9±0.3</td>
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<td>LS14321,64</td>
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<td>9.7±0.4</td>
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<td>21.5±0.2</td>
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<td>LS14321,171B</td>
<td>42.1±0.6</td>
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<td>8.9±0.1</td>
<td>5.0±0.4</td>
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<td>6.1±0.4</td>
<td>9.5±0.4</td>
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<td>LS14321,225B</td>
<td>42.8±0.6</td>
<td>21.5±0.1</td>
<td>6.7±0.1</td>
<td>4.9±0.4</td>
<td>12.0±0.2</td>
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<tr>
<td>Luna-16-A-5</td>
<td>40.2±0.2</td>
<td>20.5±0.2</td>
<td>9.2±0.6</td>
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<tr>
<td>Luna-16-G-5</td>
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<td>22.2±0.2</td>
<td>10.3±0.4</td>
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<td></td>
</tr>
<tr>
<td>LS10084,50-</td>
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<td></td>
<td></td>
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<td></td>
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<tr>
<td>(&gt;100 mesh)</td>
<td>41.0±0.1</td>
<td>20.4±0.1</td>
<td>7.0±0.1</td>
<td>4.6±0.3</td>
<td>12.3±0.4</td>
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<tr>
<td>(100-200 mesh)</td>
<td>40.3±0.3</td>
<td>19.9±0.1</td>
<td>7.2±0.1</td>
<td>5.0±0.3</td>
<td>11.8±0.2</td>
</tr>
<tr>
<td>(200-300 mesh)</td>
<td>43.4±0.4</td>
<td>20.3±0.3</td>
<td>8.3±0.2</td>
<td>4.9±0.4</td>
<td>11.1±0.4</td>
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<tr>
<td>Rock BCR-1</td>
<td>44.8±0.2</td>
<td>25.6±0.2</td>
<td>7.2±0.2</td>
<td>2.1±0.3</td>
<td>9.5±0.3</td>
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</table>

* These analyses are still in progress. Some of the data for Mg and Fe are subject to revision, since analyses are being repeated where poor precision was obtained.
OXYGEN AND OTHER MAJOR ELEMENTS IN APOLLO 14 SAMPLES

William D. Ehmann

Table 1. Notes: Data for O and Si represent 5 to 10 replicate determinations and data for Al, Mg and Fe represent 3 to 4 replicate determinations. Error limits are standard deviations of the means based on the replicate analyses. Soils 003, 163 and 259 are <1 mm fractions. Luna-16 soils are <0.125 mm fines. Samples 305, 310 and 311 were provided in the form of sawdust and these data may reflect contamination and/or dilution as compared to chunk analyses reported elsewhere. A dilution decrement of approximately 0.27 has been estimated for most elements in sawdust from rock 12013 (Showalter et al., 1971).

References:


Lunar rocks and soils are texturally complex. This complexity represents in many instances, the accumulated effects of a series of events. An effect of any of these events (such as fracture, whole or partial melting, crystallization, or metamorphism) if strong enough, may generate a characteristic texture and concomitantly completely erase any previous texture. Alternatively, subsequent events may only partially modify pre-existing texture, resulting in a sort of geological palimpsest—a composit texture composed of the superimposed records of a sequence of events. Therefore, textural analysis of lunar samples has strong potential as a major method for resolution of alternative process-models and process-sequences in lunar history. Analysis of texture in this light requires a higher degree of development and application of textural measurements than is usual in sedimentology or petrology.

Closed-form Fourier series approximation of grain shape is a suitable variable for this purpose. The harmonic amplitude spectrum represents partition of shape into a set of orthogonal components. Each harmonic amplitude represents the degree of contribution to the total shape of a known shape factor.

For an initial trial, the shapes of plagioclase grains in thin section were evaluated from Apollo 14 samples 14301, 14321, 14331, and 14310.

The Fourier amplitude spectra for harmonics two through eight was analyzed using a standard pattern recognition routine (ISODATA). At least six shape families are present. Variation in the relative proportion of grains in each family (Table 1) were examined.
SHAPE ANALYSIS PLAGIOCLASE APOLLO 14 ROCKS AND SOIL

Robert Ehrlich

Table 1

Proportional Membership in Shape Families of Plagioclase Grains

<table>
<thead>
<tr>
<th>Thin Section Number</th>
<th>14321</th>
<th>14301</th>
<th>14331</th>
<th>14310</th>
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<tr>
<td>Shape Family</td>
<td>Clast Number</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>X₁</td>
<td>C₁</td>
<td>E₂</td>
<td>B₂</td>
</tr>
<tr>
<td>1</td>
<td>35%</td>
<td>31</td>
<td>34</td>
<td>30</td>
</tr>
<tr>
<td>2</td>
<td>31</td>
<td>10</td>
<td>20</td>
<td>24</td>
</tr>
<tr>
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<td>5</td>
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<tr>
<td>6</td>
<td>14</td>
<td>18</td>
<td>11</td>
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<tr>
<td>Sample Size</td>
<td>80</td>
<td>39</td>
<td>35</td>
<td>80</td>
</tr>
</tbody>
</table>

Thin section 14321, 20 is a microbreccia - itself only slightly metamorphosed - containing a variety of clasts displaying a wide spectrum of groundmass alteration. Feldspar shapes were taken from seven such rock fragments. Regardless of matrix condition, however, proportionate shape family membership appears to be the same from clast to clast. Thus the parent material for these clasts was probably a single small volume - homogeneous in its heterogeneity - and the spectrum of clast types observed represents differential metamorphism of that parent.

Thin sections 14301, 16 and 14311, 90 are microbreccias with fused matrix. They resemble some of the clasts discussed above from 14321, 20. Their proportionate shape family memberships are almost identical. However, their proportionate shape family membership differs distinctly from any clast in 14321. These results raise the possibility that 14301 and 14311 may have a common origin distinct from 14321.

Thin section 14310, 26 is a basalt containing primarily needle-like plagioclase crystals oriented like jack straws enclosed in pyroxene. In addition, a very small proportion of plagioclase occurs as relatively large equant grains were examined. The proportionate shape family membership is similar to that of microbreccia 14321, 20. This result raises the possibility that 14310, 26 may either be a melted soil or at least contain refractory clastic "contaminants."

Analysis of Apollo 14 soil samples is now under way and will be discussed in a following abstract.
For the first time Apollo 15 carried, in the SIM bay, the sophisticated 3-inch Metric and 24-inch Panoramic Camera systems. Photographs by both cameras, when combined with those obtained from the Command Module by the Hasselblad Camera, constitute the most thorough photographic coverage of any Apollo mission to date. Due to the high inclination of Apollo 15 orbits, these photographs contain an unmatched wealth of data pertaining to numerous lunar surface features and processes.

Solutions to lunar geologic problems, both regional and local are best sought by examining returns of all three cameras. Metric Camera photographs at 25-30 m resolution provide the regional setting. Panoramic Camera photographs depict detail down to 1-2 m resolution, and Hasselblad Camera photographs (20-60 m resolution range of the 500 mm, 250 mm and 80 mm lenses) provide special views due to the obliquity of the photographs or the use of color film. Examples are given as they may apply to specific lunar geologic studies.

1. Geologic Mapping and Characterization of Surface Units

The regional structural and stratigraphic setting of the hilly and furrowed terrain southeast of Mare Smythii in the lunar farside is best depicted in Mapping Camera terrain photography. This unit resembles the hilly and furrowed terra west of both Mare Nectaris and Mare Humorum. It appears to be controlled by fault systems which trend nearly north-south. Oblique photographs taken with the Hasselblad Camera show distinct elongate vents, which appear to be volcanic in origin, throughout the unit. Panoramic Camera photographs depict details of these vents and the apparently volcanic ejecta around them. This occurrence indicates the large extent of terra volcanism on the moon, it also suggests the presence of a major tectonic trend, with which volcanism is associated, near the eastern limb of the moon.

Characterization of the Aristarchus Plateau is another example where the vertical as well as oblique Mapping Camera photographs display the regional geologic setting of the area.
Overall relationships between the larger features can be established by studying these photographs. Panoramic Camera photographs display the surface characteristics of the various units in great detail, which makes photogeologic interpretations of the origin and stratigraphic relationships possible. Color photographs taken with the Hasselblad Camera and the 80, 250, and 500 mm lenses provide additional data for delineating subtle textural differences between the mostly volcanic units. These photographs also provide useful panoramas of the numerous rilles associated with the Aristarchus Plateau, particularly of Vallis Schroeteri.

2. Study and Classification of Lunar Surface Features

Apollo 15 ground coverage included typical examples of nearly all classes of lunar rilles. Comparative studies of the straight, accurate, sinuous, and complex rilles are possible. Detailed studies of one class, e.g., sinuous rilles, will be enhanced by the capabilities inherent in the Metric Camera System. For the first time, it will be possible to generate accurate profiles along and across meandering channels such as Rima Hadley. The profiles, when combined with the high resolution panoramic photography will aid in resolving the controversy of sinuous rille origins.

Layering and banding on fault scarps (e.g., the Apennine Mountains), crater walls (e.g., the crater Picard), and other slopes (e.g., Tsiolkovsky central peaks) may also be studied in greater detail from the Apollo 15 photography. Relationships between the layers or bands and the structural setting can be established by tectonic maps based on the Metric Camera products. Panoramic Camera photographs show the local setting and extent of these structures.

3. Investigation of the Processes that Modify the Lunar Surface

The lunar surface appears to have been modified largely by comet and meteorite impacts of varying sizes. Apollo 15 photography provides coverage of numerous impact basins on both the far side and the near side of the moon. This photography is especially suited for comparative studies of these basins, their relative ages and stratigraphy. Detailed studies of the ejecta of smaller impact craters, especially those with ray-excluded zones (e.g., Proclus), are made feasible by the extensive coverage of the Metric Camera. Panoramic Camera photographs provide new insight into the process of regolith formation by meteorite impacts. This is particularly true because of the fact that high resolution stereo strip photography allows direct comparison
of crater density and textural characteristics of large segments of the lunar surface.

Volcanism appears to have played a significant role in the modification of the lunar surface features, both in the maria and in the highlands. The multitude of oblique photographs obtained by the Mapping and Hasselblad Cameras provide excellent material for studying surface flows of volcanic origin, e.g., in Mare Imbrium. Photographs taken with the Hasselblad Camera and the special black-and-white film at low-sun illumination angles (i.e., near the terminator) also display flow scarps and other volcanic features which are not decipherable in any other photography.

4. Unusual and Newly Observed Features

The Apollo 15 photographs portray a number of features which were not known from studying previous photography of the moon. Among these features are: 1) a fissure on the lunar far side which bounds what appears to have been a lava lake. The latter appears to have experienced a collapse of the central portion, leaving a frozen lava terrace and the surface markings of lava channels; 2) a cluster of large (over 20 km) domes in the vicinity of Rima Schroedinger in the southern highlands of the lunar far side; the domes appear to be associated with isolated mare-like units and a wrinkle ridge which cross-cuts old crater rims and surrounding terrain; and 3) a cluster of unique surface markings on a dark plateau-like area between the Haemus and Apennine Mountains. These irregular markings are unlike any familiar lunar surface features and are presently not understood.

Apollo 15 returned over 6,000 photographs taken of the lunar surface from orbit. The photographs constitute a major scientific return of the mission; they are essential to our understanding of the moon and its history by extrapolation of knowledge gained by surface exploration to larger areas.
Orbital photography and visual observations on Apollo 14 provided significant new results pertaining to numerous lunar surface features and processes. The data provided new clues to deciphering the role of volcanism in the formation and modification of the lunar highlands. Among these are the delineation of the following:

1. The smooth appearance of v-shaped fractures in the floors of two (unnamed) conjugate craters on the lunar farside. The larger of the two craters is 40 km across and portrays a concentric pattern of cracks; the smaller crater is 25 km in diameter and displays an alligator-hide pattern of fractures. The latter appear to have been cooling cracks in the lava fill in caldera-type floors.

2. The numerous elongate hills with summit vents in the area between Mare Smythii, on the eastern limb of the moon, and the crater Pasteur. The hills occur in clusters which are reminiscent of furrowed terra units west of both Mare Nectaris and Mare Humorum. Individual occurrences may portray a crater chain, an elongate depression or irregular vents surrounded by smooth but distinct rim deposits. They are interpreted as being volcanic in origin.

3. The detailed morphology of furrowed terra units in the vicinity of the crater Descartes and implications relative to the probable stratigraphic sequence of the geologic units of the Apollo 16 landing site.

The Hycon camera provided us with the first high resolution stereo strip photography of the lunar surface. The photographs covered a strip about 350 km long and 4 km wide which included the crater Theophilus and the Kant Plateau. Study of the photographs reveals that craters on the floor of Theophilus, on its ejecta blanket and rim deposits, and on the Kant Plateau are distinctly different in shape, depth and appearance. This suggests, in addition to differences in origin of some of the craters, a difference in thickness and other characteristics of the regolith in three units.
Photogrammetric reduction of the photographic data obtained on Apollo 14 revealed that the Kant Plateau is elevated by about 6 km relative to the mare surface to the east. This topographic rise confirms earlier earth-based measurements of the relative height of this portion of the central highlands. The upward slope averages about 6° and leads to an elevation that is higher than the top of the Apennine Mountains relative to the center of mass of the moon.

Other new results of the photography include: 1) Depiction of what is probably the youngest crater in the 20-40 km size range on the moon. The farside crater is 35 km in diameter and displays a bright halo about 100-150 km in diameter. Flow units in its floor display what appear to be drag fractures and are interpreted as somewhat viscous lava flows; 2) Providing the first photographs of a flow channel with multiple levels in the mare materials southeast of the crater Lansberg; 3) Establishing the utility of near-terminator photography in photogeologic interpretations of small scale topographic variations on the lunar surface.

Visual observations were made from the Command Module to complement the photography: Color characteristics of lunar surface units were compared; Tracking surface features at zero-phase illumination conditions were attempted and variations were noted between targets in the highlands and similar objects in the maria; Other features such as ejecta blankets of large impact basins and bright-haloed craters and their characteristics were described to aid in photogeologic interpretations of the processes which were responsible for the modification of the lunar surface.
Visual observations from lunar orbit constitute a significant complement to photography and other remote-sensed data. Utilization of the capabilities of a well-trained observer in lunar orbit is an illustration of the role of man in space flight. This is particularly significant when one considers the special characteristics of the human eye and the interpretive powers of the brain. The unaided eye resolves 20-30 m objects from 110 km orbit. It is also especially equipped, through its wide dynamic range, to distinguish subtle differences in brightness levels, color tones, topographic expressions and textural variations.

Fifteen lunar surface areas were studied in detail from lunar orbit on Apollo 15. The wide field of view allowed investigation of both the general and local settings at varying sun elevation angles and viewing directions. "On the scene" interpretations were made and later checked and confirmed on successive orbits. In certain cases, features and phenomena were compared to similar occurrences in other parts of the groundtracks. Results of the observation of eight targets are given and an example is provided of the geologic significance of observing cinder cones on the southeastern rim of Mare Serenitatis.

The more significant observations were made of the following features starting with the most easterly target: 1) Swirls of light colored markings on the floor of Mare Inginii on the lunar farside. No topographic expressions are associated with these sinuous markings which appear to have been the result of a "bleaching" mechanism. These markings are identical to those in northern Mare Marginis on the eastern limb of the moon; 2) An 80 x 70 km flow on the northwestern rim of the crater Tsiolkovsky was interpreted as a landslide bounded at the rim crest by two faults with about 10 km horizontal displacement. This lineated flow unit displays a larger population of small sharp craters than the floor material of the crater Fermi on which the landslide is superposed. Small flow units on the northeastern and southern parts of the crater rim display characteristics suggestive of somewhat viscous lava flows; 3) A brownish color tint
was detected in association with the crater Picard in western Mare Crisium. About 6 distinct layers were observed on the wall of the crater. Similar layering was also observed in the walls of the crater Peirce in the same region; 4) A fault zone appears to be associated with the ray-excluded zone of the crater Proclus on the western rim of Mare Crisium. The western part of the crater wall appears to be part of the fault plane and a displaced segment of the crater rim may have been responsible for ray-shadowing; 5) Cone-shaped hills with summit craters on the southeastern rim of Mare Serenitatis were interpreted as cinder cones, and as explained below they seem to have been the source of the dark smooth material which mantles both the highland and mare materials; 6) A subtle color difference appears to be associated with the flow scarp east of the Apollo 15 landing site in Palus Putredinis. A subsidence of a segment of the mare in this region may have been responsible for the "lava-mark" on Mount Hadley; 7) Numerous lava flows delineate a major flow front in western Mare Imbrium. Generations of flow were distinguished by their texture and subtle color tones. Lava flows in that area appear to have originated at a northwesterly wrinkle ridge in the mare; 8) Terminal portions of sinuous rilles in the Harbinger Mountains/Aristarchus Plateau regions appear to be filled by mare material. The younger (Eratosthenian) mare displays a brownish tint when compared with the gray tones of the older (Imbrian) mare materials.

Detection of what appears to be cinder cones on the southeastern rim of Mare Serenitatis has significant implications on lunar internal processes and the thermal history of the moon. The dark deposit is peppered with probable volcanic cones which surround the high hills of the southwestern Taurus Mountains. The unit mantles both plain-forming materials on the Serenitatis rim as well as the younger mare material and a wrinkle ridge in the southeastern corner of the mascon basin. This dark unit is interpreted as an ash deposit which came to the surface via volcanic conduits which formed the cinder cones. The deposit appears relatively smooth, which is confirmed by the lack of enhancements in Radar and IR maps of the area. These characteristics suggest a pyroclastic type material which originated at great depth and was extruded by explosive volcanic activity. The relative young age of the deposit may give additional information on lunar internal processes.
ABUNDANCES OF PRIMORDIAL AND COSMOGENIC RADIONUCLIDES IN APOLLO 14 ROCKS AND FINES, James S. Eldridge, G. Davis O'Kelley, and K. J. Northcutt, Oak Ridge National Laboratory, P. O. Box X, Oak Ridge, Tennessee 37830

Nondestructive gamma-ray spectrometry methods developed for Apollo 11 and 12 samples (1-3) were used for the determination of K, Th, U, $^{26}$Al, and $^{22}$Na in eight rock samples and three soils returned by the Apollo 14 mission. All the rock samples were "fragmental," and the soils were from the "top," "middle," and "bottom" of the Soil Mechanics trench. Results of these measurements are presented in Table I.

Our suite of samples is distinguished by its uniformity in primordial radielement content. Potassium concentrations ranged from 4000 to 5800 ppm, Th ranged from 10.9 to 15.6 ppm and U ranged from 3.1 to 4.5 ppm. The only samples from previous Apollo missions yielding this high level of primordial radielements (average rock values) were 12013 and 12034 from the Ocean of Storms. Two soils measured from the Apollo 15 collection show K, Th, and U contents 3 to 8 times lower than the Apollo 14 soils and breccias (4).

From our simple two-component mixing predictions in Fig. 2 of reference 2, we deduce that our Apollo 14 samples contain a range of 60 - 80% foreign component (KREEP), with the average value closer to 80%. This lends support to the predictions of many investigators that the Fra Mauro formation would be rich in KREEP.

All our Apollo 14 samples fit in a tight grouping in the K/U systematics we presented in our Apollo 11 and 12 studies (Fig. 1 of reference 2). This grouping shows that the Fra Mauro samples are very similar to the dark portion of sample 12013.

Breccia 14321, the largest rock returned, originally weighed 9 kg. Our sample 14321,38 is an 1100-g piece cut from one end of the rock. Sample 14321,256 is sawdust from cutting of 14321. The good agreement for the two samples of 14321 shows that the distribution of primordial nuclides within the whole rock is uniform despite its small-scale inhomogeneity.

Cosmogenic radionuclide determinations reported in Table I show little differences from those found in previous missions with the exception of the three soil samples, which deserve special discussion. The Soil Mechanics Experiment trench was planned to be a 60-cm-deep trench at station G with one vertical sidewall to provide a means for sampling at depth. The trenching did not yield a vertical side wall; sloping occurred with walls of 60°-80° and a maximum depth of 36 cm was achieved. Photograph AS14-64-9161 shows the degree of crumbling in the trench walls (5). Samples 14148, 14149 and 14156 shown in Table I were taken from the top, bottom, and middle, respectively of the trench and are all <1 mm sieved fractions. From the concentrations of K, Th, and U, it would appear that the soil at the trench site is uniform throughout its sampled depth of 0-36 cm. We expected to find pronounced decreases in the concentrations of the cosmogenic species $^{26}$Al and $^{22}$Na with depth.
Abundances of Primordial and Cosmogenic Radionuclides

James S. Eldridge

Instead, all three samples show a surprising uniformity in concentrations of these nuclides. We would predict values of ~40 and ~35 dpm/kg for $^{22}\text{Al}$ and $^{22}\text{Na}$ at the depth of 36 cm for sample 14l49 (6). Due to the uniform distribution of $^{22}\text{Al}$ and $^{22}\text{Na}$ and their high concentrations at depth, we must conclude that extensive mixing occurred and sample 14l49,62 is not representative of the soil at a 36-cm sampling depth. This also gives reason to question the uniformity of K, Th, and U concentrations in the different soil layers. In addition, the separation of the <1 mm fraction from the trench bottom samples has further emphasized the sampling defect since the bottom sample has a median grain size of 0.41 mm compared to 0.09 and 0.007 mm for the surface and middle trench samples (7).

Our studies with similar trench samples from Hadley Base (4) yielded the expected decrease in $^{28}\text{Al}$ and $^{22}\text{Na}$ content with increasing depth.

REFERENCES


Research sponsored by the U. S. Atomic Energy Commission under contract with the Union Carbide Corporation, through interagency agreements with the National Aeronautics and Space Administration.
Abundances of Primordial and Cosmogenic Radionuclides
James S. Eldridge

Table I. Primordial and Cosmogenic Nuclides in Apollo 14 Samples

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Weight, g</th>
<th>K, ppm</th>
<th>Th, ppm</th>
<th>U, ppm</th>
<th>26Al, dpm/kg</th>
<th>22Na, dpm/kg</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Clastic Rocks</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>169,0</td>
<td>78.66</td>
<td>5500 ± 300</td>
<td>14.2 ± 0.2</td>
<td>3.9 ± 0.1</td>
<td>82 ± 6</td>
<td>54 ± 7</td>
</tr>
<tr>
<td>170,0</td>
<td>26.34</td>
<td>5850 ± 300</td>
<td>14.9 ± 0.5</td>
<td>4.1 ± 0.1</td>
<td>88 ± 6</td>
<td>39 ± 9</td>
</tr>
<tr>
<td>265,0</td>
<td>65.79</td>
<td>4100 ± 200</td>
<td>10.9 ± 0.6</td>
<td>3.3 ± 0.2</td>
<td>102 ± 8</td>
<td>70 ± 7</td>
</tr>
<tr>
<td>271,0</td>
<td>96.58</td>
<td>5250 ± 250</td>
<td>15.6 ± 0.2</td>
<td>4.5 ± 0.3</td>
<td>118 ± 6</td>
<td>61 ± 5</td>
</tr>
<tr>
<td>272,0</td>
<td>46.20</td>
<td>4500 ± 200</td>
<td>11.3 ± 0.5</td>
<td>3.3 ± 0.2</td>
<td>94 ± 6</td>
<td>78 ± 9</td>
</tr>
<tr>
<td>273,0</td>
<td>22.40</td>
<td>4560 ± 200</td>
<td>11.7 ± 0.5</td>
<td>3.1 ± 0.2</td>
<td>73 ± 7</td>
<td>66 ± 8</td>
</tr>
<tr>
<td>321,38</td>
<td>1100.0</td>
<td>4050 ± 220</td>
<td>12.7 ± 0.5</td>
<td>3.9 ± 0.4</td>
<td>50 ± 20</td>
<td>35 ± 20</td>
</tr>
<tr>
<td>321,256</td>
<td>200.2</td>
<td>3900 ± 200</td>
<td>11.2 ± 0.5</td>
<td>3.2 ± 0.4</td>
<td>70 ± 7</td>
<td>42 ± 5</td>
</tr>
<tr>
<td><strong>Fines less than 1 mm</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>148,0</td>
<td>45.3</td>
<td>4150 ± 200</td>
<td>11.4 ± 0.5</td>
<td>3.3 ± 0.2</td>
<td>130 ± 10</td>
<td>74 ± 7</td>
</tr>
<tr>
<td>149,62</td>
<td>50.0</td>
<td>4650 ± 200</td>
<td>11.4 ± 0.5</td>
<td>3.2 ± 0.2</td>
<td>105 ± 10</td>
<td>66 ± 6</td>
</tr>
<tr>
<td>156,46</td>
<td>100.0</td>
<td>4410 ± 200</td>
<td>11.9 ± 0.5</td>
<td>3.3 ± 0.2</td>
<td>148 ± 12</td>
<td>68 ± 7</td>
</tr>
</tbody>
</table>
The opaque minerals observed in the Apollo 14 crystalline rocks 14053, 14072, 14073, and 14310 are ilmenite, chromium ulvöspinel, titanium-aluminum chromite, picotite, rutile, baddeleyite, schreibersites, native FeNi metal, troilite, mackinawite, chalcopyrrhotite, and tranquillityites. El Goresy et al. (1, 2, 3) presented preliminary reports on the opaque minerals from some of these rocks.

Of these four rocks, strong textural and mineralogical similarities were found between rocks 14310 and 14073 on the one hand and 14053 and 14072 on the other. The Ni content of the metals in samples 14310, 14072, and 14073 ranges between 1.5 and 37 wt% and 5.5 and 24 wt% Ni, respectively. The Ni content of the metal phase is dependent upon the mineral assemblage -- i.e., there is a preference of Ni for the metal phase associated with troilite. This relation is quite distinct in rock 14073 (Fig. 1). In samples 14053, 2 and 14072, 12, the Ni content of the metal phase ranges from <0.01 to 4.8 wt% and between 0.2 and 6.1 wt% Ni, respectively, and there is no apparent correlation between the Ni content of the metal phase and the mineral assemblage.

The presence of schreibersite in the Apollo 11 and 12 breccias and fines was used as criterion for meteoritic contamination. Rock 14310, however, contains schreibersite as small rounded inclusions within some of the FeNi metal grains. Thus, the criterion used previously is suspect, because in this rock the schreibersite is indigenous and not a result of contamination. In fact, two different compositions of this mineral coexist here. One is almost pure iron phosphide, whereas the other contains 28 wt% Ni, in addition to the Fe & P. The FeNi metal in which these schreibersites occur contains 16 wt% Ni and less than 0.1 wt% P.

Rock 14310, in addition to having a somewhat KREEPY composition, contains one of the highest amounts of ZrO$_2$ (0.13%) of the Apollo 14 samples. This ZrO$_2$ content is contributed mainly by the mineral baddeleyite which occurs with ilmenite, troilite, and ulvöspinel, and the amount of this mineral is far greater than in any sample from Apollo 11 or 12. In addition, ilmenites and ulvöspinels in this rock were found to contain detectable Zr. As is evident in Figure 2, there exists a strong geochemical preference of Zr for the ilmenite structure versus that of ulvöspinel. Rock 14310 also contains 2 optically distinguishable tranquillityites; one has a normal composition.

Fig. 1 Ni Content of FeNi Metal.
whereas the other contains 4 wt% less ZrO and 4 wt% higher combined TiO\(_2\) + SiO\(_2\).

Based on spinel analyses of more than 160 grains, the distinctions and similarities between the four crystalline rocks is again emphasized. In rocks 14310 and 14073, the only spinel phase observed is an optically anisotropic chromium ulvöspinel whose chemistry is similar to the Apollo 12 ulvöspinel occurring in the late-stage groundmass (i.e., rich in the Fe\(_2\)TiO\(_4\) mole). In rocks 14053 and 14072, both chromium ulvöspinel and titanium-aluminum chromite are common. Compared to Apollo 12 titanium chromites, these chromites are distinctly higher in Al\(_2\)O\(_3\) contents (13 to 22 wt% versus 9 to 12 wt% in Apollo 12 chromites). El Goresy et al. (3) reported compositional data for the spinels in some of these rocks, and Figure 3 shows spinel data for sample 14072,12. An unusual assemblage of spinels was observed in sample 14053,2. It consists of titanium-aluminum chromite, chromium ulvöspinel, and picotite (4.28% TiO\(_2\), 24.9% FeO, 11.3% MgO, 18.8% Cr\(_2\)O\(_3\), 38.2% Al\(_2\)O\(_3\), 0.22% V\(_2\)O\(_5\)).

SUBSOLIDUS REDUCTION - The Apollo 14 crystalline rocks, particularly 14053 and 14072, contain more abundant evidence for subsolidus reduction reactions than was present in any of the rocks returned by the earlier Apollo missions and emphasizes the extreme reducing environment present during the formation of these Fra Mauro rocks. Rock 14053 contains a unique texture associated with the breakdown of late-stage fayalite (Fa 86-96) to pure Fe metal + tridymite + SiO\(_2\) rich glass (El Goresy et al., 3). This texture and assemblage has never been reported as occurring in any other rock, either terrestrial or extra-terrestrial.

The subsolidus reduction of chromium ulvöspinel to ilmenite + native Fe was observed in many Apollo 12 rocks; however, the breakdown of ulvöspinel is developed to a much greater extent in the Apollo 14 crystalline rocks with all stages of reduction from incipient "exsolution" to complete breakdown. Study of the involved phases has resulted in the determination of compositional trends which have revealed
certain systematic changes in the spinel chemistry during the subsolidus reduction processes.

The compositional variations of spinels during crystallization of lunar magmas is shown in Figure 4b. In general, the spinels, both chromites and ulvöspinel, increase in the percentage of Fe₂TiO₄ mole as crystallization progresses. However, during subsolidus reduction, a reverse trend was noted for the composition of ulvöspinel (Fig. 4a).

Rocks 14053 and 14072 contain ulvöspinel which have undergone much subsolidus reduction. When the initial ulvöspinel composition contained a large amount of Fe₂TiO₄ mole, the reduction resulted in ilmenite + native Fe metal. In these cases, the ulvöspinel "exsolves" minerals containing FeO, TiO₂, and MgO and thereby becomes enriched in the FeCr₂O₄ mole and changes composition as shown diagrammatically in Figure 4a. (As a side point, the "exsolved" ilmenite always has a higher MgO content than the coexisting ulvöspinel and probably represents a closer approach to an equilibrium partitioning of MgO, based on the experimental studies of Johnson et al. (4)).

A second ulvöspinel breakdown assemblage was observed and includes titanium chromite, in addition to ilmenite + native Fe. When the initial composition of the ulvöspinel contains sufficient FeCr₂O₄ mole or when the reduction process proceeds to an advanced stage, a point is reached (approximately Usp32-Chr68 for 14072,12 on Fig. 4a) when the ulvöspinel becomes so supersaturated with chromite molecule that titanium chromite nucleates and is precipitated as distinct rims around the ilmenite and native Fe metal grains. Thus, the compositional change of ulvöspinel during subsolidus reduction is in an opposite direction to that occurring during the crystallization from the melt. They also indicate the presence of a solvus between chromium ulvöspinel and titanium chromite, as suggested by the primary spinel assemblages from both Apollo 12 and 14 rocks.

Thus, these factors, namely 1) the occurrence of lunar schreibersite, 2) the existence of the unique fayalite breakdown texture, and 3) the extreme reduction of ulvöspinel to chromite + native Fe + ilmenite, lead us to conclude that these Apollo 14 crystalline rocks formed under even more reducing conditions than those of Apollo 11 or 12.

References Cited
GEOLOGY OF THE REGION OF THE DESCARTES (APOLLO 16) SITE
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Arizona 86001; and C. A. Hodges and D. J. Milton, U. S.
Geological Survey, 345 Middlefield Road, Menlo Park, California
94025

The Apollo 16 landing site, known as the Descartes site (9°00'01"S;
15°30'59"E), was selected to sample and study the geology of probable volcanic
constructional units of the lunar highlands, units which have yet to be di­
rectly studied.

The landing site area is at the west edge of the Kant Plateau, which is
topographically the highest region on the near side of the moon, and which is
believed to consist mainly of volcanics. Rocks of the Cayley Formation and
of the Descartes highlands (a distinctive unit of the Kant Plateau) can be
sampled within short traverse distances of the landing site.

The Cayley Formation is the plains-forming unit of the highlands. It
is characterized by mostly smooth to undulating terrain and probably is com­
posed of rocks derived from fluid lavas and/or pyroclastic debris. The
Cayley fills old crater floors and topographic lows in the highlands. At
places well away from the Descartes region, the Cayley appears to be over­
lapped by the latest mare volcanic materials.

Materials of the Descartes highlands form hilly and mountainous regions
which stand topographically above the Cayley. However, Descartes materials
at places appear to be intergradational and thus contemporaneous with the
Cayley, and are interpreted to represent a more viscous phase of volcanism.
Samples of the plains- and mountain-forming units will contribute essential
information on the difference between mare and highland volcanism, and on the
chemical and thermal evolution of the moon.

Detailed geologic mapping was carried out using second-generation film
positives of Apollo 14 convergent 80 mm and 500 mm photographs analyzed in an
analytic stereoplotter; the photogeologic observations were recorded on a
planimetric base by means of a coordinograph. Fine details not recorded on
conventional photographic prints became available for structural and strati­
graphic analysis.

Geology of the landing site region.--The main mountain and hill areas are
divided into two units. Areas characterized by irregular hills and plateaus
and linear troughs are mapped as hilly, furrowed Descartes materials. Areas
characterized by smooth convex domes and linear troughs are mapped as domical
Geology of the region of the Descartes site

D. P. Elston

The plains of Cayley materials are mostly gently undulating, except for craters. One particularly smooth area has been mapped separately as smooth Cayley. Also mapped separately is a large area of irregular Cayley that includes several diverse types of topography not easily separable, and at places includes low, irregularly lobate hills. Apollo 16 is to land on irregular Cayley.

A transitional unit of undulating to low-hilly topography separates plains from hills and mountains in many places. This unit has been previously included in Cayley, but we prefer to call it a facies of Descartes materials, because its geologic features are commonly continuous into adjacent hill and mountain areas. Even so, the boundary between Cayley and Descartes is crossed by numerous structural lineaments and locally by apparent outcrop traces of stratiform rock bodies.

Outcrop traces of rock strata are recognized in a few places on the moderate-resolution 80 mm photographs on which the regional map is based.

Lineaments believed to be the traces of fractures and faults in the bedrock occur abundantly in all map units except the smoothest Cayley material, and locally constitute as many as four intersecting sets. The mapped density and orientations of the lineaments indicate multiple deformations. Some lineaments follow the bottoms or sides of the conspicuous linear troughs of the hill and mountain areas, supporting an interpretation that the troughs are of fault origin. Some such lineaments extend out of the troughs, which indicates that the lineaments are bedrock and not slump features. Zones of fine-parallel lineation occur at a few places. They locally transect apparent lines of stratification, and may be zones of foliation or shearing.

Numerous subcircular to elongate depressions, rimless or with rims notably lower than those of impact craters of comparable form, occur throughout the region, except in pre-Imbrian materials. These depressions are gradational in size and form into the linear troughs of the uplands, and are believed to be mostly of endogenetic (volcano-tectonic) origin.

Impact craters of Imbrian (?) to late Copernican age are scattered throughout the region. On both Cayley and Descartes materials, craters of sizes above the limits of resolution (about 50 m on the 80 mm model and 10 m on the 500 mm model), are of frequencies notably below saturation or steady-state levels. This indicates either a relatively recent age of formation of the rocks compared to those of previous Apollo landing areas, or an anomalously low impact frequency. However, two "island" areas of older landscape, bearing severely degraded craters (or ring structures) of inferred pre-Imbrian age, project through the cover of younger materials about 20-25 km southwest and northwest of the landing site. One of these areas appears to be saturated with craters of sizes far above the limit of resolution. The apparent overlap
relations suggest that pre-Imbrian materials may occur at a moderate depth beneath the landing site.

Geology of the traverse area.--The landing site and most of the traverse length lie within the unit of irregular Cayley materials. Traverse plans presently call for the examination and sampling of domical Descartes materials on Stone Mountain to the south during EVA 2, and on Smoky Mountains to the north during EVA 3.

A layered series of rocks that may include both Cayley and Descartes materials is exposed in the walls of North Ray crater. Strata in the crater locally appear highly deformed, as well as cut by a prominent N-S fault. Structural and stratigraphic relations between Cayley and Descartes materials may be exposed in the crater. A possible overturned flap of material derived from within the crater occurs on the western rim of the crater.

Sets of fine irregular lineaments are identifiable at many places in both the Cayley and Descartes materials. At some places, some of the mapped lines may have been produced by downslope movement of regolith. However, in most places the relationship of the lines to topography indicates that they represent the outcrop traces of sequences of finely stratified rocks that lie very close to, if not at, the surface. Actual bedrock outcrops perhaps occur only on cliffs such as those within North Ray crater. However, regolith in the other areas of inferred stratification cannot be more than a few meters (perhaps locally less than one meter) thick, so that the stratification pattern has not been destroyed by movement of regolith.

Several craters other than North Ray in and near the traverse area reveal that a layered series of rocks makes up the unit mapped as irregular Cayley. The layers visible on the photographs appear to range in thickness from about 10 to 40 m. At least four distinct layers are exposed in North Ray crater, two layers in Baby Ray crater, and three layers in South Ray crater. Stratification also has been recognized in parts of the walls of several older craters of comparable size. Only a most tentative correlation of strata between craters can be made at this time. Planned surface geologic and geophysical investigations will provide data that could lead to a reasonably accurate understanding of the stratigraphic and structural relations in the landing site area.
Lunar rocks vary in composition from basaltic to ultramafic and anorthositic, with traces of granophyre and "granite" (12013). These same rocks are common to the earth. The striking differences in composition between the known lunar and terrestrial rock analogues, as well as their similarities, are well documented. This is especially true for the basalts widespread on the moon and earth. Some interrelations between several lunar and terrestrial basalts of widely disparate ages, but which have many features in common, are shown in abbreviated form in Table 1. All of the rocks have been chosen from those analyzed in our laboratories to eliminate interlaboratory bias. These and all other petrochemical studies confirm the much lower ratio in the lunar rocks of many volatile to refractory elements, and the absence of water and ferric iron.

The significant differences in composition between lunar and terrestrial rocks indicate divergent environments of origin and emplacement. The major, early thermal and volcanic episodes on the moon between 3 and 4.6 aeons seem to require gravitational heat generated by accretion and impaction, probably augmented by radiogenic and perhaps solar heat. Enough heat was required at least in the lunar lithosphere to generate the floods of maria lavas and to further degas the lunar lithosphere and surface. The complex series of original and succeeding volcanic events have left much, if not all of the moon essentially dry and dead for the last 2.5 to 3 aeons.

In contrast, the "excess" volatiles characteristic of the continuously defluidizing earth are cited as proof of relatively cold terrestrial accretion. The complementary terrestrial thermal pulses, and differentiation into core, mantle, and crust, commonly have been ascribed largely to radioactivity with additional uncertain contributions from gravitational energy. But a binary origin of the moon and earth implies initial terrestrial impact and gravitationally generated (?) thermal episodes from 4.6 to at least 3.9 AE equal to, or probably in excess of, those inferred on the moon. The terrestrial events could have begun and culminated in a divergent geochemical fractionation in which the large amounts of accreted volatiles

This is especially true for the basalts widespread on the moon and earth. Some interrelations between several lunar and terrestrial basalts of widely disparate ages, but which have many features in common, are shown in abbreviated form in Table 1. All of the rocks have been chosen from those analyzed in our laboratories to eliminate interlaboratory bias. These and all other petrochemical studies confirm the much lower ratio in the lunar rocks of many volatile to refractory elements, and the absence of water and ferric iron.

The significant differences in composition between lunar and terrestrial rocks indicate divergent environments of origin and emplacement. The major, early thermal and volcanic episodes on the moon between 3 and 4.6 aeons seem to require gravitational heat generated by accretion and impaction, probably augmented by radiogenic and perhaps solar heat. Enough heat was required at least in the lunar lithosphere to generate the floods of maria lavas and to further degas the lunar lithosphere and surface. The complex series of original and succeeding volcanic events have left much, if not all of the moon essentially dry and dead for the last 2.5 to 3 aeons.

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EARTH-MOON PETROGENESIS

A. E. J. Engel

Table 1. Average compositions of basaltic rocks from Apollo 11, 12, and 14 compared with basaltic rocks from the earth.

<table>
<thead>
<tr>
<th></th>
<th>Apollo 11</th>
<th>Apollo 12</th>
<th>Apollo 14</th>
<th>Oceanic Tholeiitic Basalt</th>
<th>Archean Basalt</th>
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<tr>
<td>SiO₂</td>
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<td>45.10</td>
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<td>9.64</td>
<td>18.79</td>
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<tr>
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<td>nil</td>
<td>nil</td>
<td>nil</td>
<td>2.63</td>
<td>2.95</td>
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<tr>
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<td>20.23</td>
<td>9.97</td>
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</tr>
<tr>
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<td>.25</td>
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<tr>
<td>MgO</td>
<td>6.95</td>
<td>10.28</td>
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<td>7.42</td>
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<tr>
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<td>.13</td>
<td>.13</td>
<td>.13</td>
<td>.09</td>
<td></td>
</tr>
</tbody>
</table>

(in parts per million)

Ba  240  74  780  14  12
Co  16  35  25  32  36
Cr  2000 3600 1350 300 325
Cu  10  8  30  77  42
Li  17  7  30  9  14
Ni  6  36  230 97  86
Sc  90  52  28  61  49
Sr  130 120 140 130 140
Y   65  180 33  290 200
Yb  18  10  25  5  6
Zr  160 164 900 95  81

Apollo 11, average of 1004 and 10057.
Apollo 12, average of 12021, 12022, and 12075.
Apollo 14, average of 14301, 14310, and 14318.
Average of 14 oceanic tholeiites from the Atlantic, Pacific and Indian Oceans.
Average of 12 Archean basalts from South Africa, 3 to 3.5 aeons in age.
Major oxides by C. G. Engel. Spectrographic analyses by A. L. Sutton.
were retained.

Studies of the oldest igneous rocks of the earth (3-3.5 AE) indicate they are essentially identical in composition with their contemporary counterparts. They include appreciable combined water, sodium, and ferric iron, and far less Ba and Zr than lunar basalts, as indicated by the partial compositional data for recent and Archean basalts in Table 1, Columns 4 and 5. These Archean basalts formed at about the same time as the Porcellarium basalt. This overlap in time in the formation of terrestrial and lunar basalts with important discrepancies in compositions clearly indicates the major differences in their coeval rock-forming environments in the lunar and terrestrial lithosphere and surfaces.

The petrochemical differences indicated in coeval moon-earth systems and the physical constraints first detailed by Jeffreys now are incompatible with an earth-fission hypothesis, and seem to further complicate and obscure events in any binary planet hypothesis. Hence lunar studies to date resolve a few, but perpetuate many of the enigmas of the first aeon of earth history, its origin, and initial composition.

The similarities in composition of widespread Archean and recently erupted oceanic basalts, as well as other common species of terrestrial rocks suggest the earth was largely differentiated into proto-crust, mantle, and core by between 4.6 and 3.5 AE. Any prolongation of first-order terrestrial differentiative events from 3.5 AE in the earth's mantle -- from more to less "primordial" -- would be reflected in changes with time in the compositions of the more voluminous igneous rocks, especially the floods of oceanic basalts ubiquitous in time and space. In this context, the early, major differentiative events observed on the moon, presumably involving large sources of gravitational heat, appear to have at least crude analogues on earth.
Main constituents of fines 14003, 14230, 14259 and 14259.64 (in brackets: grain % of 14003 fractions 1000-500, 500-250, 250-125, 125-63, 20-63μm): crystalline rocks (44, 26, 21, 15, 8), agglomerates and regolith breccias (47, 53, 49, 46, 23), glasses (7, 19, 21, 24, 35), plagioclase (0, 2, 5, 7, 13), pyroxene (clino and ortho) and subordinate olivine (1, 2, 4, 7, 19). Fines below 4μm consist of predominant pyroxene and plagioclase with subordinate olivine, ilmenite, cristobalite, troilite and iron. Modal compositions of 14003 and 14259 are similar, 14230 contains more glasses and less crystalline rock fragments. Mare basalts and anorthosites are rare. Predominant are various rocks of the noritic family (plagioclase, clinopyroxene including pigeonite, orthopyroxene, ilmenite; minor constituents: olivine, Ti-poor Cr-spinel), exhibiting two main textural types: (a) medium-grained cumulates of euhedral or fragmental and often brecciated crystals; (b) large fragments of plagioclase, less pyroxene and recrystallized plagioclase in a fine-grained matrix of interlocked pyroxene and plagioclase, interspersed with randomly distributed ilmenite. By their textural resemblance with rocks from large terrestrial impact craters type (b) rocks are interpreted as recrystallized base surges or impact melt rocks. Type (a) rocks may be either of impact melt origin or represent magmatic cumulates. Transitional textures between (a) and (b) make distinction of both types sometimes difficult. Apollo 14 soils differ from 11 and 12 soils by their higher content of crystalline rocks, mostly of noritic character, and by the lower pyroxene : plagioclase ratio (about 1 : 1).

Glasses which we analyzed from soils and breccias belong to six types, see Table 1: Type (1) glasses represent shock produced diaplectic plagioclase glasses. Types (2) to (6) are interpreted as shock produced quenched melts, originating from alkali-poor mare basalts (2), alkali-rich basaltic rocks and/or local soil (3), anorthositic rocks (4), mafic (5) and granitic (6) rocks. 17 analyzed glasses from soil 14259 belong to types (1), (3) and (4).

nD of glasses increases with decreasing (SiO2+Al2O3)- and increasing (FeO+TiO2)-contents. Annealing of two colored melt glasses at 630°C resulted in increase of nD (1.60 to 1.63) within 1.5 hours and unmixing after 3.5 hours, indicating a very rapid cooling of the melts on the lunar surface. Annealing of a diaplectic plagioclase glass at 800°C resulted in decrease of nD (1.574 to 1.569) within two hours and no further changes for longer times, in accordance with the behavior of maskelynite from Shergotty meteorite and diaplectic glass from Ries crater, but contrary to the increase of nD observed with Apollo 11 and Manicouagan diaplectic glass.

Shock effects found in soils are fragmentation, undulatory extinction, deformation lamellae in pyroxene and ilmenite, plagioclase with isotropic lamellae, diaplectic plagioclase glass in single grains and shocked rock fragments. Recrystallized plagioclase grains are interpreted as devitrified diaplectic glasses.

Breccias belong to three types (I, II, III):
Eye glasses: plagioclase ratios are equal to or larger than those found in the soils. 17 analyzed glasses from 14307 belong to types (3), (4) and (5), 24 analyzed glasses from 14049 to types (1), (2), (3) and (4). 14307 is surrounded by vesicular type (3)-glass, interpreted as melted soil from a younger impact.

Glass-poor breccia produced by an impact which penetrated solid rocks below the regolith (14082,11 and 13). Rock fragments: noritic rocks of types (a) and (b) as described above, plagioclase breccias, often with shock features, mare basalts absent. Mineral fragments: plagioclase (large fragments), clinopyroxene, orthopyroxene, ilmenite. Pyroxene : plagioclase = 0.5. Subordinate fluidal, slightly devitrified glass fragments.

Glass-poor breccias with crystalline matrices, interpreted either as recrystallized base surges or as impact melt rocks, identical to type (b) rock fragments in soils and breccias (14312,12; 14314,12; 14319,16; 14320,5; 14006,8; 14066,50). Rocks 14312, 14314, 14319, 14320 and 14066 are composed of a light and a dark component. The dark material contains less and smaller inclusions and forms irregular blebs and lapilli within the light mass. These rocks may be interpreted as mixtures of two hot masses or melts produced by the same impact from two different source rock units. Predominant rock inclusions exhibit impact melt or cumulate textures and are of noritic or even anorthositic composition. Mare basalts are absent. Mineral fragments: plagioclase, recrystallized plagioclase, clinopyroxene (including pigeonite), orthopyroxene and ilmenite. Minor constituents (in 14312, 14314 and 14319 preferably within the dark matrix): olivine and Ti-poor Cr-spinel (see Table 1). Only 14066 contains more pyroxene than plagioclase, the others more plagioclase than pyroxene. Glasses are rare and more or less devitrified. Slightly devitrified, brownish glass fragments in 14312, 14314 and 14319 are of rhyolitic composition (see Table 1). Fissures in 14312, 14314, 14319 and 14320 are filled with light brown type (3) glass, apparently produced by the impact which excavated the rock from a deeper level.

Shock effects as described from soil samples have been observed in breccia types (I) and (II) and are lacking or very rare in type (III) breccias, due to recrystallization.

Interpretation: Fra Mauro formation as represented by breccia types (II) and (III) is composed of breccias originating from single or multiple pre-mare impacts which incorporated plagioclase-rich rocks, characterized by Ca-poor pyroxene and high alkali contents, exhibiting indications of magmatic differentiation, ranging from spinel-bearing SiO$_2$-poor rocks to rhyolitic residuals. Soils of Apollo 14 were mainly produced by local impacts and consist of debris from Fra Mauro rocks with a small admixture of mare material.
Table 1

<table>
<thead>
<tr>
<th></th>
<th>(1)</th>
<th>(2)</th>
<th>(3)</th>
<th>(4)</th>
<th>(5)</th>
<th>(6)</th>
<th>Sp.</th>
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<tr>
<td>SiO₂</td>
<td>45-46</td>
<td>39-45</td>
<td>45-53</td>
<td>37-45</td>
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<td>75</td>
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<td>2-11</td>
<td>1-5</td>
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<td>23-31</td>
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<td>4-12</td>
<td>3-9</td>
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<td>0.08</td>
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</tr>
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<td>0</td>
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<td>0.04-0.1</td>
<td>0-0.1</td>
<td>0.1</td>
<td>0.03</td>
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<tr>
<td>P₂O₅</td>
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<td>n.d.</td>
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<td>0-0.6</td>
<td>0.06</td>
<td>0.06</td>
<td>0</td>
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</tbody>
</table>

Number of analyses: (3) (6) (35) (13) (1) (1) (1)

Glass types:
1. diaplectic plagioclase glass (colorless)
2. basaltic, alkali-poor (brownish)
3. basaltic, alkali-rich (brownish)
4. anorthositic (colorless - green)
5. mafic (opaque)
6. rhyolitic (light brown)

Sp.: Spinel \((\text{Al}_{0.79}\text{Cr}_{0.21}\text{Ti}_{0.005})_2(\text{Mg}_{0.46}\text{Fe}_{0.54}\text{Mn}_{0.003})_4\)
The Apollo 14 and 15 samples analyzed to date exhibit the same $^{18}O/^{16}O$ and $Si^{30}/Si^{28}$ relationships previously found on Apollo 11 and 12 samples. The most striking feature of all these data is that the whole-rock oxygen and silicon isotopic composition of the lunar crust is remarkably uniform, and is identical to that of terrestrial basalts. No significant distinctions can be made between the mare basalts and the non-mare basalts with respect to either whole-rock or mineral isotopic compositions, and no isotopic evidence has been found for metamorphism of the Apollo 14 breccias. This suggests either (1) these portions of the Imbrium ejecta blanket did not undergo a slow, prolonged period of cooling, or (2) $H_2O$ was absent, thus inhibiting isotopic exchange. The Apollo 14 and 15 breccias and fines are enriched in $^{18}O$ and $Si^{30}$ with respect to the lunar basalts, similar to the Apollo 11 and 12 results. Partial fluorination experiments similar to those previously reported by us have been undertaken on some Apollo 14 and 15 soils; some of the data are presented in Fig. 1. Plotted in Fig. 1 are all samples for which we have both partial fluorination data and hydrogen concentrations (shown in parentheses in terms of μmoles of $H_2$/g of sample). Note the excellent correlation between the surface $^{18}O$ enrichments and the amounts of solar wind $H_2$. A completely analogous diagram holds for the silicon isotope effects. The almost total lack of solar wind $H_2$ and of any isotope effects in rock breccia 14321 is in striking contrast to the situation in soil breccias such as 10061. This is compatible with the fact that our sample is an interior fragment from 14321 (about 3 cm inward from the surface); sample 14321 has been shown by other workers to possess a short exposure age (~25 m.y.) and to have probably been derived from about 80 meters depth in Cone Crater. The 0.01 μmoles/g of $H_2$ present in 14321 can be attributed to ~10 to 20 MEV protons from solar flares, because ordinary solar wind hydrogen would not penetrate so deeply into the rock. The $H_2$ contents of interior portions of certain lunar samples thus may serve to monitor past solar flare activity. Partial fluorination experiments on several other Apollo 14 and 15 surface soil samples typically give results similar to those shown in Fig. 1 for 14422. Note that samples 14149 and 12033 are both bottom-trench samples, so it is reasonable that they exhibit smaller effects. The first 8% of material removed during the fluorine stripping
experiment on 14422 has an average $\delta^{18}O$ about 5 per mil higher than the lunar basalts. This material is much more depleted in oxygen than in silicon, with O/Si ratios as low as 0.8 to 1.3; the whole-rock samples have O/Si ratios of 1.8. The above data indicate that the enrichment of $^{18}O$ and Si$^{30}$ in the grain-surface coatings of the lunar fines is a widespread phenomenon on the moon. There must be a ubiquitous fractional vaporization/condensation process that has removed the lighter isotopes $^{18}O$ and Si$^{28}$ preferentially from the moon. An appreciable fraction of material in the lunar fines may have been added to the soils as a fractional condensate from a transient lunar atmosphere that in part contained SiO, O, and/or O$_2$. Such a transient atmosphere could form during vaporization of silicate material during meteorite, micrometeorite, or particle bombardment. The same processes may have produced the enrichment of "old, parentless" lead in acid-leach fractions of the lunar fines observed by L. T. Silver, as well as the higher contents of other volatile "condensates" such as Hg and Bi in the soil. They may also have produced the enrichment of C$^{13}$ and S$^{34}$ in the lunar fines as compared with the lunar basalts.

We found that by cycling O$_2$ gas over the samples we can markedly decrease the $\delta^{13}C$ and increase the amounts of carbon produced during pyrolysis of the lunar samples. These effects are due to oxidation of stopcock grease present in the vacuum extraction system. In the light of these experiments, we decided to continue to use the pyrolysis method on Apollo 14 and 15 soil samples, and have obtained similar variations in carbon concentration and $\delta^{13}C$ as were obtained on Apollo 11 and 12 soil samples (see Fig. 2). These data reemphasize the qualitative correlation previously pointed out, that $\delta^{13}C$ tends to increase with increasing carbon concentration. The sample with the lowest $\delta^{13}C$ value on Fig. 2 is a rock breccia, 14321, which also has the lowest solar wind H$_2$ content of any breccia or soil sample we have analyzed. This low $\delta^{13}C$ value is similar to values obtained by Kaplan et al. on lunar basalts. The relationship shown on Fig. 2 indicates mixing between a low-C$^{13}$ lunar carbon and a high C$^{13}$ extra-lunar carbon; the latter is more abundant in the fines and soil breccias. The shape of the curve suggests that it does not represent a simple two-component mixture and that some or all of the C$^{13}$ enrichment could be due to the mechanisms described above for Si$^{30}$ and $^{18}O$ enrichment of the lunar fines. Plausible sources of the extra-lunar carbon are the solar wind and/or infall of carbonaceous meteorites.

Hydrogen isotope data on the Apollo 14 and 15 samples are similar to results obtained from Apollo 11 and 12. The soils contain about 20 µmoles/g of extractable hydrogen gas, having $\delta^D$ values of -750 to -850 per mil. The samples containing smaller amounts of hydrogen gas tend to be more enriched in deuterium, showing the importance of contamination of the solar wind hydrogen gas by the small amounts of H$_2$O (<10 µmoles/g) present in the samples. A plot of the $\delta^D$ of lunar hydrogen gas against the amount of water in the samples is shown in, 

\[ \text{Fig. 2} \] 

Lunar Carbon

\[ \text{Carboxy (ppm)} \]

\[ 8 C^{13} (\text{%o}) \]
Fig. 3. The extrapolation of a least-squares line indicates that in the absence of water, the extracted hydrogen would have a $\delta D$ value of -995. Considering the errors, this result is compatible with our previous estimate of -980 per mil for "lunar" hydrogen gas. The latter value was obtained by another technique in which we eliminated the deuterium from the contaminating water by isotopic exchange with deuterium-free water. The "lunar" hydrogen probably represents deuterium-free solar wind $H_2$ containing about 3 ppm of spallation deuterium. A further suggestion of the presence of spallation deuterium is indicated by the unusually high $\delta D$ values of -250 to -255 obtained for the hydrogen gas extracted above 1200°C (i.e., during melting) of soil samples 14240 and 14420. The deuterium contents of the two gas fractions can be wholly accounted for from spallation processes. An effort was made to extract hydrogen gas from the lunar soil sample 14422 utilizing the method used for helium extraction by Heyman and Yaniv; namely, by pounding the soil with a metal anvil in a vacuum. It was hoped that this method, by possibly releasing hydrogen gas at low temperatures, would avoid the cross-contamination of lunar hydrogen gas by hydrogen from the "lunar" water, thus permitting a direct measurement of $D$ in uncontaminated lunar hydrogen gas. Unfortunately, although we confirmed that rare gases were in fact released by this method, no measurable hydrogen gas was released. Thus, at low temperatures He is preferentially released relative to $H_2$ from the lunar fines, whereas the opposite is true at higher temperatures (>500°C). The $O^{18}/O^{16}$ ratio of "lunar" $H_2O$ has been measured in two samples, 14321 ($O^{18}/O^{16} = 18.2$) and 15301 ($O^{18}/O^{16} = 5.9$). The $\delta D$ values of $H_2O$ in these samples are similar to those obtained previously from Apollo 11 and 12 samples (-200 to -250). The "lunar" $H_2O$ is thus similar in $\delta O^{18}$ and somewhat lower in $\delta D$ than terrestrial atmospheric $H_2O$ vapor in the southern United States. If one makes allowance for the cross-contamination with solar-wind $H_2$ that we know occurs during extraction of the "lunar" $H_2O$, the corrected isotopic analyses are close to those expected for meteoric waters. The so-called "lunar water" has a relatively uniform concentration of about 8 to 17 μmoles/g and is more readily extracted during heating than either the rare gases or the $H_2$. It is therefore likely that most of this "lunar water" resides on or within the surface layers of the grains. If this water is of true lunar origin, it is surprising that its $\delta D$ and $\delta O^{18}$ values are so similar to that of terrestrial waters. This is particularly so in the light of the enormous enrichments and fluctuations in the isotopic composition of other constituents (such as $O^{18}$, $Si^{30}$, and $C^{13}$) in the surface coatings of lunar soil grains. Therefore, although we cannot as yet be certain that some truly lunar $H_2O$ is not present in the Apollo samples, it must be present in lesser amounts than the absorbed terrestrial $H_2O$. 

![Fig. 3](image-url)
INTER-ELEMENT RELATIONSHIPS BETWEEN THE MOON AND STONY METEORITES WITH PARTICULAR REFERENCE TO SOME REFRACTORY ELEMENTS.
A.J. Erlank, J.P. Willis, L.H. Ahrens, J.J. Gurney and T.S. McCarthy, Department of Geochemistry, University of Cape Town, Private Bag, Rondebosch, Cape, South Africa.

We have analysed one Apollo 15 and five Apollo 14 specimens by XRF analysis. We use these data, together with other data obtained in this laboratory on lunar samples and stony meteorites, to discuss some selected interelement relationships in these materials. Other interelement relationships will be discussed elsewhere.

Zr and Nb: Abundances of Zr and Nb in lunar materials examined by us vary by a factor of 10, and are higher than those measured in howardites and eucrites, as indicated in the following table.

<table>
<thead>
<tr>
<th>Sample description</th>
<th>Zr, ppm</th>
<th>Nb, ppm</th>
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</thead>
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<td>16.4</td>
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<td>10084,173 Fines</td>
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<td>21.3 ± 1.0</td>
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<td>7.9 ± 1.3</td>
<td>12.9</td>
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<td>12038,77 &quot;</td>
<td>182 ± 2</td>
<td>10.4 ± 1.0</td>
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<td>12053,24 &quot;</td>
<td>133 ± 2</td>
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</tr>
<tr>
<td>12063,52 &quot;</td>
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<td>7.1 ± 1.3</td>
<td>18.0</td>
</tr>
<tr>
<td>12032,38 Fines</td>
<td>705 ± 2</td>
<td>47.8 ± 1.3</td>
<td>14.8</td>
</tr>
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<td>12070,88 &quot;</td>
<td>523 ± 2</td>
<td>36.9 ± 1.3</td>
<td>14.2</td>
</tr>
<tr>
<td>14053,43 Rock</td>
<td>215 ± 2</td>
<td>17.4 ± 1.2</td>
<td>12.4</td>
</tr>
<tr>
<td>14310,117 &quot;</td>
<td>852 ± 2</td>
<td>60.6 ± 1.0</td>
<td>14.1</td>
</tr>
<tr>
<td>14305,121 Breccia</td>
<td>1158 ± 2</td>
<td>78.6 ± 1.1</td>
<td>14.7</td>
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<tr>
<td>14163,56 Fines</td>
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<td>68.8 ± 1.0</td>
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<td>15101,67 &quot;</td>
<td>313 ± 2</td>
<td>19.8 ± 1.0</td>
<td>15.8</td>
</tr>
<tr>
<td>Chaves Howardite</td>
<td>27.5 ± 0.9</td>
<td>1.7 ± 0.5</td>
<td>16.3</td>
</tr>
<tr>
<td>Malvern &quot;</td>
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<td>2.5 ± 0.5</td>
<td>14.5</td>
</tr>
<tr>
<td>Haraiya Eucrite</td>
<td>36.2 ± 1.0</td>
<td>2.1 ± 0.5</td>
<td>17.1</td>
</tr>
<tr>
<td>Sioux County &quot;</td>
<td>42.3 ± 1.0</td>
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<td>15.3</td>
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<tr>
<td>Cachari &quot;</td>
<td>45.1 ± 1.0</td>
<td>2.8 ± 0.5</td>
<td>16.3</td>
</tr>
<tr>
<td>Juvinas &quot;</td>
<td>45.6 ± 1.0</td>
<td>2.7 ± 0.5</td>
<td>16.8</td>
</tr>
<tr>
<td>Bereba &quot;</td>
<td>52.0 ± 1.0</td>
<td>4.0 ± 0.5</td>
<td>13.1</td>
</tr>
<tr>
<td>Pasamonte &quot;</td>
<td>52.5 ± 1.0</td>
<td>3.5 ± 0.5</td>
<td>15.2</td>
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<tr>
<td>Macibini &quot;</td>
<td>54.3 ± 1.0</td>
<td>3.4 ± 0.5</td>
<td>15.8</td>
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<tr>
<td>Stannern &quot;</td>
<td>87.1 ± 1.0</td>
<td>6.3 ± 0.5</td>
<td>13.8</td>
</tr>
</tbody>
</table>

The Zr-Nb relationship in lunar materials is exceptionally well developed and the variation in Zr/Nb is smaller than that shown for well-known geochemically coherent pairs such as K/Rb,
K/Ba and La/Yb; the Zr/Nb coherence approaches that shown by Th and U. The lack of Zr/Nb fractionation in all types of lunar materials, and specifically in lunar soils where Zr/Nb varies from 14-16, attests to similar behaviour of Zr and Nb during partial melting/crystal fractionation processes. Other workers report the presence of K, REE, Zr and P rich phases in late stage residua in Apollo 11, 12 and 14 rocks and high concentrations of Nb are present in some late Zr minerals (1). Published data (2) show that lunar ilmenites contain up to 3000 ppm Zr, and either they are not important as liquidus phases (Apollo 11) or Zr/Nb is similar to that of the co-existing liquids (Nb data are not available).

KREZP (KREEP) and other similar materials must have higher Zr and Nb contents, but similar Zr/Nb ratios. We draw attention also to the relative constancy of the K/Zr ratio (a volatile/refractory pair) in lunar materials. This observation indicates similar behaviour of K and Zr (and Nb) in lunar magmatic processes, it argues against large-scale volatilization of the heavier alkalies from lunar lavas, and it allows calculation of Zr (and Nb) in KREZP materials. Average KREZP glass (Apollo 12) contains ~ 9000 ppm K (3). Using K/Zr = 4.5, Zr/Nb = 15, KREZP glass should contain about 2000 ppm Zr and 130 ppm Nb. These abundances are far in excess of those measured by us for terrestrial basaltic rocks, and indicate that very small degrees of partial melting and/or large amounts of crystal fractionation must have been operative during the production of KREZP basaltic materials.

The similarity of Zr/Nb in all types of surface lunar materials implies a similar ratio for the moon as a whole and the ratio is, within experimental error, apparently the same as in the eucritic and howarditic achondrites, for which we report the first Nb measurements. The only available chondrite data are for Allende where Zr/Nb = 10 (4). Terrestrial oceanic and continental basaltic rocks have Zr/Nb varying from 4 - 40; in kimberlitic peridotites the ratio ranges from 2 - 7.

In terms of fundamental properties of atoms (ionic radius, nature of bond and crystal field effects), it is not clear why Zr and Nb should preserve such a close coherence in a wide variety of extra-terrestrial materials. In terrestrial rocks these elements are exclusively present as Zr$^{4+}$ and Nb$^{5+}$. The radius of Zr$^{4+}$ is 0.80 Å and that of Nb$^{5+}$ 0.69 Å, but if Nb is in a reduced state, as is possible in the eucrites and the moon, we should consider the existence of Nb$^{4+}$ (0.75 Å) and of Nb$^{3+}$ (~ 0.8 Å). The latter value is a new estimate based on ionization potential considerations (5) which also support the existence of Nb$^{3+}$ as a stable oxidation state (6). It is noteworthy that Nb in tranquillityite has been reported as Nb$_2$O$_3$ (7).

Ca and Al. The importance of the Ca - Al relationship in stony
meteorites and the lunar surface has been stressed by us (8). We have determined these two elements in 23 additional carbonaceous chondrites, howardites and eucrites. With the exception of the CIII chondrites, Coolidge and Leoville and the eucrites Serra de Mage and Moore County, this information confirms that a Ca/Al ratio of about 1.10 is typical of these stony meteorites. We have stated that this ratio may be typical of most extra-terrestrial material, including the sun and the earth. However, the mesosideritic value of 0.87 should be borne in mind because such a ratio appears to be typical of Apollo 14 fines. Our interpretation of the Ca/Al data in the fines from all areas sampled is that it varies from \( \sim 0.86 \) to a little more than 1.2 with an average close to that of the stony meteorite value. In contrast specific lunar rocks and impact glasses show a wider variation in Ca/Al; nevertheless, the range observed varies almost equally on either side of the stony meteorite and lunar fines average, with mare materials tending to have higher and non-mare materials tending to have lower ratios.

Inter-element slopes and condensation processes. The possible importance of inter-element slope relationships (rates of increase of concentration) of the refractory elements between various meteorite types and the lunar surface has been discussed before (8,9). Slope might have a bearing on the nature of condensation processes including the possibility of placing an element in a condensation sequence such as that proposed by Lord, Larimer, Anders and others. Examples will be given.

References:
Several investigators have previously shown that composite samples of lunar dust contain rare gases, N\textsubscript{2}, H\textsubscript{2}, and CH\textsubscript{4}. The purpose of the present work is to provide a technique whereby measurements can be made on single particles. Particular attention is given to glassy particles, selected from Apollo XI and XIV fines.

The particles in question are typically 50 \mu m in diameter and weigh about 0.3 \mu g. The absolute amount of dissolved or occluded gas per particle is of the order of 1 nl (STP). This is beyond the capabilities of the gas chromatograph or even of mass spectrometers operated in the ordinary dynamic mode. However ample sensitivity is provided by the omegatron mass spectrometer operated in a static mode. A baked all glass ultra high vacuum system provides the low background necessary for static operation.

Gases are extracted from the particles by disintegrating them one at a time with pulses from a neodymium glass laser. The particles are not vaporized by a 5 joule pulse, but form a fine spray of molten glass which provides a very satisfactory release of dissolved gases.

The gas produced in largest amount is O\textsubscript{2}. This gas is of no interest, since it arises from thermal decomposition of the glass. To provide for its removal, it is sufficient to expose about 0.1 g of Pt black in the vacuum system.

Oxides of carbon are also produced, probably by oxidation of methane and other occluded carbon compounds in the particle. No attempt has been made to measure carbon in this form. CO is effectively removed by adsorption on the Pt black, which simplifies the quantitative determination of nitrogen.

A series of exploratory runs on both glassy and crystalline particles has been completed. Gases positively identified are He, Ar, N\textsubscript{2}, CH\textsubscript{4}, and CO\textsubscript{2} in addition to O\textsubscript{2} and CO already
DISSOLVED GAS IN LUNAR PARTICLES

F. M. Ernsberger

mentioned. The amounts of the gases are large enough for quantitative determination, though quite variable from particle to particle.

The presence of He complicates the procedure, because large losses of this gas would be expected during the preliminary bake out of the system. A method for maintaining the sample temperature lower than that of the rest of the vacuum system has been worked out.

It is intended to use this technique to determine whether or not the dissolved gases are localized near the surface of the particles, by comparing the gases recovered from etched and unetched glassy microspheres. This part of the work should be complete by the date of the conference.
INFRARED AND RAMAN SPECTROSCOPIC STUDIES OF STRUCTURAL
VARIATIONS IN MINERALS FROM APOLLO 11, 12 and 14 SAMPLES

Morgantown Energy Research Center, U. S. Department of the Interior,
Bureau of Mines, Morgantown, West Virginia, 26505, U.S.A.

Infrared and Raman vibrational spectroscopic data, yielding direct
information on molecular structure, have been obtained for single grains
(>150 μm) of minerals, basalts and glasses isolated from Apollo 11, 12 and 14
rock and dust samples, and on grains in polished butts from Apollo 14 samples.
The spectral data show both similarities and differences between the lunar
maria and highland samples that we studied: 4 crystalline rocks (12018-26,
12020-26, 12021-24, 14310-93); 6 dusts (10085-46, 12001-60, 12057-57,12070-24,
14163-80, 14259-15); 1 breccia (14321-108) and 2 potted butts (14310-2, 14321-
97). The application of microsampling techniques has furnished new data on
grain-to-grain structure variations and has made possible the identification
of some trace accessory minerals. No hydrated minerals were found in any of
the samples.

STRUCTURE DETERMINATION OF MINERAL GRAINS

Single grains of the following minerals were isolated and characterized
from spectral-structure correlations derived from synthetic and terrestrial
standards: K-feldspar, α-quartz, plagioclase, pyroxene, a pyroxenoid,
olivine, ilmenite and chromite. Raman and infrared reflection techniques
were employed to obtain in-situ structure determination of plagioclase,
pyroxene and olivine grains in the polished butt samples.

Feldspars. The possibility that primary granitic rocks may occur on the moon
is further enhanced by the identification of 7 granitic grains from the Apollo
14 breccia. The biminarlic composition of these white granular grains varied
from predominantly α-quartz to predominantly K-feldspar. The same K-feldspar
was further isolated in pure single grains from the breccia and the infrared
data indicated a random distribution of Si/Al over the lattice positions,
similar to terrestrial sanidine. The occurrence of this disordered high-
temperature polymorph commonly found in terrestrial volcanic rock is consist-
tent with the belief that impact processes formed this rock type. We further
verified the presence of α-quartz in Apollo 14 samples by isolation of a pure
grain from dust 14259-15. Single grains of the abundant plagioclase feldspars
from the Apollo 14 samples exhibited a wide range of morphologies and spectra
of these were similar to those from Apollo 11 and 12 samples for equivalent
anorthite contents. The Apollo 14 plagioclase grains ranged An_{81} - An_{90} and
showed grain-to-grain compositional variation within single samples.
Pyroxenes. Lunar pyroxenes showed more variations in both infrared and Raman spectra than those of any of the other lunar mineral classes. These included both orthopyroxenes (Fs3 – Fs4) and clinopyroxenes, with highland sample clinopyroxenes ranging from pigeonite to subcalcic augite (Fs18 – Fs40) and maria sample clinopyroxenes ranging from pigeonite to augite (Fs20 – Fs85). We isolated orthopyroxenes from the Apollo 14 samples only (2 dusts, breccia, crystalline rock) and determined specific ferrosilite contents for these, utilizing a determinative curve derived from an infrared absorption band which shifts linearly from 396 to 350 cm\(^{-1}\) for Fs11 to Fs86 in a series of 22 analyzed terrestrial orthopyroxenes. The intensity of an infrared absorption band in orthopyroxene spectra at 450 cm\(^{-1}\) appears to be directly correlatable with Fe\(^{2+}\) distribution over the nonequivalent octahedrally coordinated sites, M1 and M2, as detected by Mössbauer spectroscopy. In studies of heated, shocked and meteoritic orthopyroxene samples for which Mössbauer data show substantial cation disorder, we observed the intensity of the 450 cm\(^{-1}\) band to be substantially reduced, relative to that in terrestrial metamorphic orthopyroxenes with equivalent ferrosilite contents. The weak intensity of the 450 cm\(^{-1}\) band in lunar orthopyroxene spectra was thus interpreted as an indication of substantial cation disordering from either high temperatures and rapid cooling or shock events. Spectra of pigeonite and augite pyroxenes were generally very similar for both our maria and highland samples, and these matched better with a series of seven synthetic standards than with terrestrial compositional equivalents. Specific ferrosilite contents for lunar clinopyroxenes were therefore obtained from determinative curves derived from the synthetic samples, utilizing for infrared the same analytical absorption band: near 396 cm\(^{-1}\) (Fs18) as used for orthopyroxenes, and for Raman the systematic frequency shift of the strong line near 1013 cm\(^{-1}\) (Fs18).

Olivine. The Apollo 14 olivine grains that we studied occurred in a wider range of morphologies and compositions (Fa17 – Fa45) than those of Apollo 11 and 12 (Fa28 – Fa44) and exhibited considerable within-sample grain-to-grain compositional variations. These were all in the forsterite-fayalite olivine series and for a given composition, spectra were very similar for grains isolated from maria and highland samples. These all compared well with spectra of terrestrial and synthetic compositional equivalents. Determinative curves for fayalite contents were used for an infrared absorption band which shifts linearly from 418 to 356 cm\(^{-1}\) in a series of 9 synthetics from Fa8 to Fa100, and for a Raman doublet which shifts from 863 to 847 cm\(^{-1}\) and from 830 to 820 cm\(^{-1}\) for the same fayalite contents.
Opaque Oxides. Ilmenite grains were isolated from both maria samples (dust 10085-46, rock 12018-26, rock 12021-24) and from highland samples (dust 14259-15, breccia 14321-108). In spectra of terrestrial ilmenites from a variety of sources and with different formation histories, we have observed gradational frequency shifts for the predominant absorption bands, indicating subtle and continuously variable structural differences. To describe these variations, we have distinguished two end-members, Type I and Type II, according to the position of the lowest frequency absorption band. This diagnostic band varies from 305 cm$^{-1}$ in Type I to 275 cm$^{-1}$ in Type II. Spectra of lunar ilmenites similarly exhibited these same frequency variations and could be classified according to the terrestrial scheme. There were both sample-to-sample spectral variations within the maria and highland samples and grain-to-grain spectral variations within single samples. However, we observed a distinct trend that absorption bands in the spectra of ilmenite from the highland samples generally occurred at higher frequencies (closer to Type I) than those from the maria samples (closer to Type II). Two unique ilmenite grains from the Apollo 14 samples showed frequencies higher than those observed in any terrestrial sample. These frequency shifts in ilmenite spectra might be related to stoichiometry deviations, disorder in the lattice, or possibly to variations in some element, such as Mg or Zr, which have been reported to show major compositional variations. Studies are in progress to determine if any of these structural parameters can be correlated with the observed frequency shifts.

From the groundmass of rock 12021-24 we identified a single grain of a spinel-group mineral. Infrared spectral correlations, relating frequency shifts to various cation substitutions into the spinel lattice, indicate that this grain is a chromite with minor Ti and Al substitution, as previously reported for other spinel grains from this rock.

DETERMINATION OF BULK COMPOSITIONS OF COMPOSITE SAMPLES AND GLASSES

Spectra of basaltic grains and screened fractions isolated from the two Apollo 14 dusts were very similar. Most showed a strong predominance of plagioclase over pyroxene, in a fairly constant ratio. These spectra were distinctly different from maria composite sample spectra, previously shown to be predominant in pyroxenes. For composite samples and glasses from the highland dusts, the frequency of the strong infrared band in the Si-O stretching region was generally slightly higher than in those from the maria dusts, indicating slightly less basic compositions. However, these were still within a basaltic composition range (38-54 wt% SiO$_2$). Infrared and Raman spectra of glasses from Apollo 14 samples were similar to those from Apollo 11 and 12 samples, exhibiting 3 principal absorption bands assignable to Si-O stretching and Si-O bending vibrations. As previously determined for the Apollo 11 and 12 glasses, none of the spectra of Apollo 14 glasses matched those of tektite spectra in frequencies or band shapes. The detection of plagioclase absorption bands in spectra of various colors of irregularly shaped glass fragments from the Apollo 14 samples suggests the presence of crystalline phases. However, we detected little or no fine structure in either the infrared or Raman spectra of glass beads to indicate recrystallization. This research was sponsored by NASA under contract T-1760A.
Li, Be AND B ABUNDANCES IN FINES FROM THE APOLLO 11, APOLLO 12, APOLLO 14 AND LUNA 16 MISSIONS

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91-ORSAY, France

This paper reports determinations of Li, Be and B abundances in the Apollo fine samples 10084, 12070, 14141 and 14259 and in sample L 16-19 No. 118 belonging to the fraction "Regolith C" of the core tube returned by the automatic probe Luna 16. The results for the concentrations of Li, Be and B in lunar fines are listed in Table 1.

Table 1. Concentrations of Li, Be and B (ppm) in lunar fines.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Li</th>
<th>Be</th>
<th>B</th>
</tr>
</thead>
<tbody>
<tr>
<td>Apollo 11</td>
<td>10084</td>
<td>12.8</td>
<td>1.6</td>
</tr>
<tr>
<td>Apollo 12</td>
<td>12070</td>
<td>19.1</td>
<td>4.6</td>
</tr>
<tr>
<td>Apollo 14</td>
<td>14259</td>
<td>35.6</td>
<td>4.4</td>
</tr>
<tr>
<td>Apollo 14</td>
<td>14141</td>
<td>40.7</td>
<td>3.1</td>
</tr>
<tr>
<td>Luna 16-19</td>
<td>No. 118</td>
<td>9.8</td>
<td>1.2</td>
</tr>
</tbody>
</table>

For the fines of the four different lunar sites the Li and B concentrations are correlated (see Fig. 1).

Fig. 1:
Correlation of B vs. Li for lunar fines.
Relative to chondritic abundances of Li, Be and B the lunar ones are strongly enriched. The lunar abundance patterns for Li, Be and B are remarkably similar. In particular, the abundances of these elements in the fines from Luna 16 and Apollo 11 are almost the same, differing by factors of 1.2 to 1.3.

The Li, Be and B abundances of the terrestrial standard basalt BCR-1 resemble those of average lunar material, whereas the Be and B abundances in basaltic achondrites are about 10 times lower than the lunar ones. Although the Apollo 12 fines come from a typical mare region as is the case for the Luna 16 and Apollo 11 fines, the Li, Be and B abundances in the Apollo 12 fines are about a factor of 2 higher than in the fines from the other mare sites. A comparison of the concentrations of these elements in the soil sample 12070 with those calculated for an average basaltic rock of the Apollo 12 site shows that the soil is enriched by a factor of 3 to 5 relative to the rocks. An enrichment of the same magnitude of K, rare earth elements and P (KREEP component) has been discovered and investigated by Hubbard et al. (2) and others. Possible origins of the KREEP component suggested by Hubbard et al. are the admixture of highland or pre-mare material to the Apollo 12 soil.

In fact the fines from Apollo 14, which were collected at a pre-mare site, are enriched in the KREEP elements as well as in Li, Be and B. In the following discussion an attempt is made to relate the soils from different lunar sites with each other by means of a two-component mixing model with one end member being the average Apollo 12 basaltic rocks and the other one the fines 14141. Similar two-component models have been proposed by Hubbard et al. (2) and Wänke et al. (3) for the Apollo 12 fines. Fig. 2

![Fig. 2: Li and B mixing lines for a two-component model.](image-url)
Li, Be, B ABUNDANCES
O. Eugster

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shows the mixing lines for Li and B. Applying this model to the Li and B concentrations causes most of the data points to lie close to the mixing line. The mixing ratios of the two end members in the samples 12070 and 14259 can be obtained from Fig. 2 where the q-values indicate the percentage of 14141 type material in the fines. We derive 36\% and 83\% for the fines 12070 and 14259, respectively. The percentage of the foreign end member obtained here for the fines 12070 is almost identical to that calculated by Wänke et al. (3) which means that the mixing model proposed by those authors and the one in this paper are virtually the same and that the chemical composition of the fines 14141 probably is similar to that of the KREEP component in the Apollo 12 soil. This suggests that large quantities of Fra Mauro material have been transferred to the Apollo 12 mare site.

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STRUCTURE OF LUNAR GLASSES BY RAMAN AND SOFT X-RAY SPECTROSCOPY.

The surface fines from all Apollo landing sites, particularly those of Apollo 11 and 15, have been found to contain small spheres and shards of glass. Some are iron and titanium-rich and are nearly opaque. Others are brown, green, and, rarely, nearly clear. At least two categories, the KREEP glasses and the normal basaltic glasses, have been recognized on the basis of their chemical composition. The objective of the present study was to determine as much as possible about the structural state of the glasses using Raman spectroscopy and soft X-ray spectroscopy. Both measurements were made of individual particles of glass extracted from bulk lunar fines. Thus many measurements were made on each sample and indeed Raman measurements (using a 100 μm spot from an ionized argon laser as a source) were obtained from different areas of the same particle. Raman spectra were also measured on a suite of natural glasses (tektites, fulgarites, and obsidians) and some synthetic glasses to provide a reference for comparison.

Raman spectra exhibit two principal regions: broad, rather weak bands occur at high frequencies and are characteristic of short-range order structures. Sharp, often intense, fine structure occurs mainly at low frequencies (less than 400 cm\(^{-1}\)). The glasses, both lunar and terrestrial, are very heterogeneous. Different details appear in spectra measured at different spots on the same particle. Similar heterogeneity is found in the soft X-ray spectra.

Glasses containing alkaline earths, transition elements, and alumina exhibit fairly sharp and well defined vibrational modes in the high-frequency region. High-silica glasses such as the fulgarites give only weak ill-defined spectra. The lunar glasses show surprising bands in the region of 1200 to 1250 cm\(^{-1}\) and 900 to 950 cm\(^{-1}\) where stretching motions of the SiO\(_4\) tetrahedra are expected. Minerals such as olivine with no bridging oxygen have Si-O stretching frequencies in the range of 950 cm\(^{-1}\), but such un-bridged linkages are usually only associated with high-alkali glasses. Feldspars containing only bridging oxygens have Raman frequencies near 1100 cm\(^{-1}\). The unusually high vibrational frequencies, with related high force constants, also occur in a few of the tektites.

### Selected Raman Frequencies from Cuts of 14162

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Major &quot;glass&quot; bands (cm(^{-1}))</th>
<th>1152</th>
<th>596</th>
<th>492</th>
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</thead>
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<tr>
<td>14162, 47, 1</td>
<td>1250 980 745 630</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14162, 47, 2B</td>
<td>933 610 592</td>
<td>596</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14162, 47, 2C</td>
<td>1250 750 638</td>
<td>492</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14162, 61, 1</td>
<td>1250 980 745 630</td>
<td>596</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
The low-frequency regions of many lunar glasses and tektites exhibit a complicated fine structure. Dozens of closely spaced bands are observed, some with half-widths normal to crystal spectra (10 to 20 cm$^{-1}$) and some much narrower (2 to 6 cm$^{-1}$). The origin of much of this fine structure is in question at this writing. Some appears to arise from incipient crystallization (below the detection limit for X-rays) of feldspar-like material. The sharp intense high-frequency bands expected from incipient crystallization of ferromagnesian minerals has not been observed.

Similar measurements to be made on Apollo 15 glassy material will also be reported.
CORRELATED ELECTRON MICROSCOPY AND DIFFRACTION STUDIES
OF CLINOPYROXENES FROM APOLLO 14 ROCKS
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Department of Biophysics & Pritzker School of Medicine
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Correlated electron microscope and diffraction studies have been carried out on yellow pigeonite separates from Apollo 14 rock 14053,47-P1 separated by K. Schürmann. Spot analyses of this pigeonite separate (ca. 50 grains) were made with a computer automated electron microprobe by D. Virgo. The average molecular percent end-member composition is Wo 11.6(+2.9), En 58.1(+6.1), Fs 30.3(+3.9). The pigeonite analyses show the presence of octahedral Al, indicating a Tschermak-type coupled substitution within the specimens. This is distinctly different from the coupled substitutions involving Al, Ti and Cr in the groundmass clinopyroxene from 14053 also being studied.

Specimens were prepared by controlled micro-manipulation procedures and cleavage, and were examined at 200 kV with a modified Hitachi HU-200E high voltage microscope under conditions of reduced radiation damage using a liquid nitrogen anti-contamination device. The results are based on a preliminary evaluation of several thousand plates.

Most of the electron micrographs show submicroscopic exsolution lamellae ca. 200A to 600A wide, oriented along (001) or (100) and other planes, with interband spacings of ca. 400A to 1500A. However, exsolution lamellae are absent in some of the crystals. In addition, certain specimens were examined by electron microscopy and selected-area diffraction using a new form of high vacuum evaporation decoration technique in an attempt to visualize directly the magnetic domains possibly related to the band structures within the specimens. The preliminary findings indicate an enhanced concentration of iron within these domains.

Several experimental modifications are being carried out in collaboration with S.S. Hafner and his associates, including heating in a vacuum chamber in the range of 1000° to 1100° C. The results are now being evaluated and will be reported in a separate publication.

The exsolution phenomena in these pyroxenes are discussed in terms of the intracrystalline distributions as suggested by L.W. Finger, S.S. Hafner, D. Virgo, et al.

The possibility of the existence of an appreciable concentration of low energy electrons above the lunar surface due to photoemission and secondary electron emission processes has previously been considered theoretically by a number of authors\textsuperscript{1-5}. Direct evidence for the photoelectron component has been provided by the charged particle lunar environment experiment (CPL EE) set up during the Apollo 14 mission\textsuperscript{6}. The low energy electron spectrometer showed quite clearly an electron flux due to solar irradiation in the energy range from 200 down to 35 eV. We have undertaken measurements of the yield and the energy distribution of photoelectrons due to solar irradiation, together with the diffuse reflectance, for sample 14259.116. Data were taken in the range from the photoelectric threshold near 2500 Å to 500 Å wavelength, corresponding to a photon energy range of 5 eV to 20 eV. Future measurements using a synchrotron light source will extend this data up to about 100 eV and allow direct comparison with the CPL EE data taken in the region of overlap. Calculations have been performed on the basis of those data, which give the density and energy distribution of the electrons in the lunar photoelectron layer. Together with data on the secondary electron emission\textsuperscript{7} due to solar wind particles a complete description may be given of the low energy electron layer and the electrostatic potential of the sunlit lunar surface.

The samples were handled in a dry nitrogen atmosphere of less than 10 ppm impurities during sealing into the ultrahigh vacuum chamber. Both photoelectric yield and the electron energy distributions were measured using a hemispherical collector. The resolution of the electron energy analyzer was 300 mV.

Figure 1 shows preliminary results on the photoelectric yield of the sample in the wavelength range 500 to 2500 Å. The yield is found to be about 7\% in its maximum at 900 Å wavelength. This yield is much lower than expected for insulating materials, which reach yields of 30\% and higher\textsuperscript{6}. The lower yield is probably due to the fact that the material is in the form of a fine powder, thus increasing the probability of reabsorption of a photoemitted electron. For an insulator one would also expect the yield to increase further for higher photon energies between 12 and 20 eV. Instead a decrease is observed, dropping to 1.2\% at 584 Å wavelength (21.2 eV photon energy). The measured yield is even lower than that of common metals\textsuperscript{8}, which typically reaches peak values of about 10\%.

A set of photoemission energy distribution curves is shown in Fig. 2 as measured with different photon energies. Those curves show a high contribution of low energy electrons, while the number of electrons emitted with the highest possible energies, namely photon energy minus work function, is small. The average electron energy therefore will be small compared to the photon energy.
PHOTOEMISSION FROM LUNAR SURFACE FINES.
B. Feuerbacher

Fig. 1 Photoelectric yield per incoming photon of lunar sample 14259.116.

Fig. 2 Energy distribution of photoelectrons emitted from lunar sample 14259.116 for various photon energies.
PHOTOEMISSION FROM LUNAR SURFACE FINES.

B. Feuerbacher

Computer calculations have been performed to study the properties of the lunar photoelectron sheath on the basis of the data given in Fig. 1 and 2. The results give the saturation photoelectron flux as \( 2.8 \times 10^{13} \text{ electrons/m}^2 \times \text{sec} \) or \( 4.5 \mu\text{A/m}^2 \). This is an order of magnitude less than for aluminium, due partly to the low yield and partly to the high work function of 5 eV. The density of electrons in the sheath is calculated to be 130 per cm\(^3\) at the surface, and the shielding length 78 cm.

A direct comparison with the data measured by the CPLEE\(^6\) on the lunar surface is not possible with the present measurements due to the limitation in light energy used. However, one can say that if these results are typical, the lunar photoemission is small compared to that of common materials\(^6\), and the average energy is very low. Therefore care has to be taken in the interpretation of low energy electron measurements taken on the lunar surface, since a considerable amount of the measured electrons might be due to contamination of the local photoelectron sheath by the instrument itself.

References:

Measurements of $^{236}\text{U}$ ($2.4 \times 10^7$ y) in seven lunar and four terrestrial samples are shown in Table 1. There is considerable variation of the $^{236}\text{U}$ content in Apollo samples. Soil sample 12070 as a whole exhibits a 7.5 times greater $^{236}\text{U}:^{238}\text{U}$ ratio than does its coarsest fraction (obtained by settling faster than 0.17 cm/sec in water). There appears to be some correspondence between the $^{236}\text{U}$ and the $^{26}\text{Al}$ ($7.4 \times 10^5$ y) content in a given sample, suggesting that $^{236}\text{U}$, like $^{26}\text{Al}$, is produced in nuclear reactions by energetic solar flare protons. The total neutron exposure, measured from gadolinium isotope ratios [1] in 12070 precludes there having been enough thermal neutrons to make 233 ppb $^{236}\text{U}$ from the $^{235}\text{U}$ in the sample. Crude estimates indicate that the $^{236}\text{U}$ produced from $^{238}\text{U}$ by solar flare protons, probably was not formed at a uniform rate, but was generated in larger quantities more than 10 million years ago than at the present rate. All other radioactive products of solar proton irradiation are shorter lived than $^{236}\text{U}$ by an order of magnitude or more.

Table 1. $^{236}\text{U}$ in Lunar and Terrestrial Samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>$^{236}\text{U}/^{238}\text{U}$ ($\times 10^{-9}$)</th>
<th>$^{26}\text{Al}$ (Literature) d/m/Kg</th>
</tr>
</thead>
<tbody>
<tr>
<td>10084-75</td>
<td>&lt; 3</td>
<td>107 - 147</td>
</tr>
<tr>
<td>12070-91</td>
<td>233 ± 15</td>
<td>146 - 171</td>
</tr>
<tr>
<td>12070-91</td>
<td>31 ± 6</td>
<td></td>
</tr>
<tr>
<td>Coarsest fraction of &lt; 1 mm soil</td>
<td></td>
<td></td>
</tr>
<tr>
<td>12013-10,42</td>
<td>9.4 ± 2</td>
<td>115</td>
</tr>
<tr>
<td>Rosholt's Uranium fraction</td>
<td></td>
<td></td>
</tr>
<tr>
<td>12073,34 - Breccia</td>
<td>47 ± 10</td>
<td>110 ± 10</td>
</tr>
<tr>
<td>14163,135 - Soil</td>
<td>&lt; 3</td>
<td>78</td>
</tr>
<tr>
<td>14305,80 - Breccia</td>
<td>4 ± 2</td>
<td>85 ± 17</td>
</tr>
<tr>
<td>Uraninite, Shinkolobwe, Katanga, Congo [11]</td>
<td>0.23 ± 0.08</td>
<td></td>
</tr>
<tr>
<td>Shroeckingerite, Lost Creek, Wyo. [11]</td>
<td>&lt;0.2</td>
<td></td>
</tr>
<tr>
<td>Thucolite, Parry Sound, Ont. [11]</td>
<td>0.08 ± 0.15</td>
<td></td>
</tr>
<tr>
<td>Pitchblend, Great Bear Lake, Port Hope Refinery, Ont. [11]</td>
<td>0.62 ± 0.22</td>
<td></td>
</tr>
</tbody>
</table>
The protons reacting with $^{238}U$ can be expected to produce $2.2 \times 10^6$ $^{237}Np$. The neptunium fraction in the chemical separations of the various Apollo samples had previously been isolated, and tracer experiments revealed the chemical yield of neptunium in these fractions. In such a fraction, pure $^{239}Np$ tracer was added, and after further purification, the $^{237}Np/^{239}Np$ ratio was measured with our 100" mass spectrometer. There is $1.1 \pm 0.3 \times 10^{-13} \text{g}^{237}Np$ per g of 12070, '91. This is 0.28 as much as the $^{236}U$ content; if one assumes a steady equilibrium bombardment for the last 108 years the cross section for formation of $^{237}Np$ would be 3.2 times the cross section for production of $^{236}U$.

The isotopic ratios of $^{235}U/^{238}U$ in five samples are shown in Table 2. Although deviation from terrestrial standards as much as 2.6 σ are reported the lunar and terrestrial uranians appear to have the same ratio.

### Table 2. ($^{235}U/^{236}U \times 10^6$)

<table>
<thead>
<tr>
<th>Sample</th>
<th>$^{235}U/^{236}U$ x $10^6$</th>
</tr>
</thead>
<tbody>
<tr>
<td>10084,75</td>
<td>7233 ± 15</td>
</tr>
<tr>
<td>12070,91</td>
<td>7153 ± 20</td>
</tr>
<tr>
<td>12073,34</td>
<td>7390 ± 200</td>
</tr>
<tr>
<td>14163,135</td>
<td>7230 ± 40</td>
</tr>
<tr>
<td>14305,80,41</td>
<td>7294 ± 15</td>
</tr>
</tbody>
</table>

Terrestrial Uranium (Assumed) 7257

$^{239}Pu$, $^{244}Pu$, Uranium and Thorium The results of isotopic dilution analysis $[2,3,4]$ for two Apollo 14 samples are shown in Table 3. Two adjacent 1/3 g fragments of 14305,80,41 (taken from an outer portion of the breccia) exhibited varying amounts of thorium: the 17.5 ppm value leads to the unusual thorium/uranium ratio of 4.16, while the 15.6 ppm value leads to the more usual ratio, 3.78. The thorium/uranium ratio in 14163 is 3.88.

### Table 3. Apollo XIV Samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>14163,135</th>
<th>14305,80,41</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thorium</td>
<td>13.2 ± .4 ppm</td>
<td>15.6 ± .5 ppm</td>
</tr>
<tr>
<td>Uranium</td>
<td>3.4 ± .1 ppm</td>
<td>4.13 ± .12 ppm</td>
</tr>
<tr>
<td>$^{239}Pu$</td>
<td>&lt;4.5 x $10^{-9}$ ppm</td>
<td>&lt;4 x $10^{-10}$ ppm</td>
</tr>
<tr>
<td>$^{244}Pu$</td>
<td>4 x $10^{-9}$ ppm</td>
<td>&lt;2 x $10^{-10}$ ppm</td>
</tr>
<tr>
<td>$^{238}U/^{234}U$ activity ratio</td>
<td>1.01 ± .01</td>
<td>1.01 ± .02</td>
</tr>
</tbody>
</table>
The report by Hoffman et al. of $8.2 \times 10^7$ year $^{244}$Pu in terrestrial bastnèsite [5] is supported by some unpublished evidence at Argonne National Laboratory for $^{244}$Pu in terrestrial gadolinite [6]. In each case the observed $^{244}$Pu appears to be too abundant to have survived from the amount estimated to have been present during the condensation of the solar system. Despite the absence of $^{244}$Pu in five lunar samples, the lunar surface remains a promising collector of any $^{244}$Pu that might come from accretion of extra-solar sources.

The limits to $^{239}$Pu ($244 \times 10^3$ years) in Table 3 imply that recent thermal neutron fluxes were less than 33 n/(cm$^2 \cdot$sec) and 370 n/(cm$^2 \cdot$sec) for samples 14305, 80 and 14163.

Anomaly We have reported earlier [2] a $^{238}$U/$^{234}$U ratio of 1.06 ± 0.02 in 12070, and Rosholt and Tatsumoto [7] have reported many cases where the $^{230}$Th content of a sample was also less than its equilibrium value. This suggests that some of the lunar material has existed in space for thousands of years, within the last million years, in the form of particles so small that the 4 $\mu$g/cm$^2$ alpha decay recoil range would be a significant fraction of the particle's diameter. Ideally the diameter of such a particle would be $2.7 \times 10^{-6}$ cm, its weight would be $3 \times 10^{-17}$ g (20 million atomic mass units). Such a particle might be generated by the interaction of energetic heavy elements in solar flares with the lunar surface [8,9,10].

References:
4. A new value of $(1.585 \pm 0.018) \times 10^5$ years for the half-life of $^{233}$U has been used. Y. A. Ellis, Nuclear Data Sheets 6, 257 (1971).
ORTHOPYROXENE, CLINOPYROXENE, AND PLAGIOCLASSE SEPARATES FROM THE TWO BASALTIC ROCKS 14053 AND 14310 (1) WERE STUDIED USING ELECTRON MICROPROBE, MÖSSBAUER SPECTROSCOPY, AND X-RAY DIFFRACTION TECHNIQUES.

**Pyroxenes**

The composition of our pyroxene separates are shown in Tab. 1. The pigeonite and the Ca-rich clinopyroxene from 14053 are distinctly different in coupled Al-Ti-Cr substitutions. Whereas the Ca-rich clinopyroxene has a systematic deficiency in the Al-Ti-Cr balance, the pigeonite shows Tschermak type substitution. Orthopyroxene crystals from 14310 indicate a somewhat higher content of Tschermak's molecule. This does not necessarily result from a high pressure crystallization (Kushiro, pers. commun.). X-ray diffraction studies of single crystals are being carried out. Some orthopyroxene crystals (a=18.326 Å, b=8.863 Å, c=5.235 Å) from 14310 have intergrown pigeonite (a=9.89 Å, b=8.90 Å, c=5.23 Å, β=108.2°) on (100). Using the b-Å nomogram (2) this corresponds to a composition WO(10)En(60)Fs(30).

The Fe$^{2+}$ Mössbauer doublets at M1 and M2 were well resolved in each spectrum. The deviations of the absorption peaks from Lorentzian shape were small. The data can therefore be interpreted as precise average values of our separates. The results are shown in Tab. 2.

The standard Gibbs free energy difference for the Mg$^{2+}$,Fe$^{2+}$ exchange between M1 and M2 in the 14053 pigeonite at 1000°C is $\Delta G^0=4.7$ kcal per mole $\text{M}_2\text{Si}_2\text{O}_6$. This energy has been previously found to be nearly invariant between 1000 and 500°C(3). Thus, the natural cation distribution corresponds to an equilibrium temperature of $T_n=840°C$. Heating of a natural sample at 550°C for a few hours reveals ordering, and the steady state distribution attained in this ordering experiment corresponds to an equilibrium temperature $T_c=650°C$.

$\Delta G^0$ of the orthopyroxene separated from 14310 is 3.8 kcal per mole. This value is somewhat higher than that known for homogeneous, terrestrial orthopyroxenes (3.6 kcal per mole, cf. ref.3). This may be due to the presence of a small amount of clinopyroxene in the separate. The natural sample corresponds to an equilibrium temperature $T_n=550°C$, and heating at 550°C yields neither ordering nor disordering.

Our data on rock 14053 suggest that the first-formed Ca-poor pigeonite crystallized from an undercooled near-surface magma (at approximately 1200°C) in which crystallization of plagioclase was
Distinct cooling histories
L.W. Finger

The natural cation distribution of pigeonite shows that rock 14053 was cooled extremely rapidly. $T_n=840^\circ\text{C}$ is the highest temperature observed in terrestrial and extraterrestrial pyroxenes until this time. The cooling rate was so rapid that it cannot be attributed to the initial subsolidus cooling at the surface of a lava flow as simple heat flow considerations show (Gray, pers. commun.). The presence of extensive exsolution lamellae in the Ca-poor pigeonite cores (Fernandez-Moran, pers. commun.) indicates that the original cooling rate was in fact quite slow. After its solidification and cooling to low temperatures this rock was reheated (possibly for a very short time) and then was cooled extremely rapidly. The temperature peak in the reheating process was in excess of $840^\circ\text{C}$. It was probably associated with a relatively recent impact event, possibly in connection with the creation of Cone crater (4).

The cooling rate of rock 14310 was fairly slow compared with rock 14053. It was similar to the rate of basalt 12021 from Oceanus Procellarum, but slower than the rate at the surface of some Hawaiian lava flows. There is no indication of reheating at a later event. Temperatures $T_n$ of the same magnitude were previously found in orthopyroxenes from Japanese basalts. However, $T_n$ is significantly higher than the values of orthopyroxenes from terrestrial metamorphic rocks.

<table>
<thead>
<tr>
<th>Table 1. Chemical composition of pyroxene separates</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average Standard Number of spot composition deviations analyses</td>
</tr>
<tr>
<td>Wo En Fs Wo En Fs</td>
</tr>
<tr>
<td>14053,47 pigeonite 66.0 58.1 30.3 2.6 6.1 3.9 50</td>
</tr>
<tr>
<td>14053,17 augite 23.0 36.0 42.0 5.6 13.8 13.0 22</td>
</tr>
<tr>
<td>14310 orthopyroxene 6.3 69.7 23.7 3.6 7.2 54.3 25</td>
</tr>
</tbody>
</table>

$^a$ Na, Mg, Al, Si, Ca, Ti, Cr, Mn, Fe determined.

<table>
<thead>
<tr>
<th>Table 2: $\text{Mg}^{2+},\text{Fe}^{2+}$ distributions over $M_1$ and $M_2$ in pyroxenes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pyroxene Composition Heat Area ratio Distribution</td>
</tr>
<tr>
<td>Wo En Fs treatment $\text{Mg}/(\text{Mg}+\text{Fe})$ constant</td>
</tr>
<tr>
<td>14053,47 pigeonite 12 58 30 natural 0.236 0.247 0.120</td>
</tr>
<tr>
<td>550,12h 0.176 0.194 0.0755</td>
</tr>
<tr>
<td>550,3days 0.156 0.199 0.0893</td>
</tr>
<tr>
<td>1000,1day 0.273 0.283 0.158</td>
</tr>
<tr>
<td>14310,116 orthopyroxene 6 70 24 natural 0.186 0.181 0.0971</td>
</tr>
<tr>
<td>550,3days 0.141 0.166 0.102</td>
</tr>
</tbody>
</table>

$^a$ absorbers held at $77^\circ\text{K}$.
$^b$ from products peak height times width.
$^c$ from peak heights; $k$ was calculated using $c$. 
Distinct cooling histories
L.W. Finger

Plagioclase

The $^{57}$Fe hyperfine spectrum of Ca-rich plagioclase depends on the thermal history of the crystals. Some data of natural and heated lunar and terrestrial plagioclases are shown in Tb. 3. Whereas the plagioclases of the lunar basalts exhibit complex spectra which consist of at least two Fe$^{2+}$ doublets, the plagioclases of anorthosites from Stillwater and Lake Superior show only doublet C (assignment cf. Fig. 1, ref. 5).

The nuclear quadrupole splittings and peak widths of plagioclase from 14053 correspond to the values of the Stillwater plagioclase heated at 1000°C. The widths of the 14310 plagioclase are somewhat smaller and probably reflect a somewhat lower temperature than 1000°C.

The center of gravity of the total resonance absorption area of the $^{57}$Fe spectrum is tentatively interpreted in terms of ratios Fe$^{3+}$/Fe$_{tot}$ (cf. Tb. 3). Unfortunately, the precision of the data is limited because of poor signal-to-background ratios. The experimental errors of the data in col. 4 of Tb. 3 are large. Fe$^{3+}$ peaks were not resolved in the spectra.

We thank Barbara Janik and NASA grant NGR-14-001-173.

Table 3: $^{57}$Fe-data of plagioclases

<table>
<thead>
<tr>
<th>Rock</th>
<th>Anorthite mol.%</th>
<th>Heating treatment</th>
<th>Quadrupole splittings</th>
<th>Peak widths (FWHM)</th>
<th>Center of gravity</th>
<th>(Fe$^{3+}$/Fe$_{tot}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>C mm/sec</td>
<td>A mm/sec</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>B mm/sec</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>C mm/sec</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1004, 25</td>
<td>88-92</td>
<td>natural</td>
<td>2.01</td>
<td>1.94</td>
<td>0.52</td>
<td>0.35</td>
</tr>
<tr>
<td>1202, 150</td>
<td>88-90</td>
<td>natural</td>
<td>2.09</td>
<td>1.99</td>
<td>0.51</td>
<td>0.36</td>
</tr>
<tr>
<td>1405, 47</td>
<td>77-90</td>
<td>natural</td>
<td>2.06</td>
<td>1.55</td>
<td>0.59</td>
<td>0.36</td>
</tr>
<tr>
<td>1405, 47</td>
<td>1123</td>
<td>1000°C</td>
<td>2.04</td>
<td>1.58</td>
<td>0.65</td>
<td>0.38</td>
</tr>
<tr>
<td>14310, 116</td>
<td>92-93</td>
<td>natural</td>
<td>2.04</td>
<td>1.48</td>
<td>0.46</td>
<td>0.45</td>
</tr>
<tr>
<td>Lake Superior anorthosite</td>
<td>74</td>
<td>natural</td>
<td>2.63</td>
<td>-</td>
<td>0.44</td>
<td>-</td>
</tr>
<tr>
<td>Hawaii Kilauea lava flow</td>
<td>56</td>
<td>natural</td>
<td>2.63</td>
<td>-</td>
<td>0.71</td>
<td>-</td>
</tr>
</tbody>
</table>

a. absorbers held at 205°C.
b. assignment and symbols cf. Ref. 5.
c. of total resonant absorption area (referred to metallic iron).
d. tentative determination neglecting nonequivalent recoilless fractions and assuming 0.14 mm/sec for Fe$^{2+}$ and 0.22 mm/sec for Fe$^{3+}$ as isomer shifts.
e. Ref. 6.
f. determined from two-line fits (peak widths very broad).

References:
RADIOACTIVITIES IN APOLLO 14 AND 15 MATERIALS; E. L. Fireman, J. D'Amico, J. DeFelice, G. Spannagel, Smithsonian Astrophysical Observatory, Cambridge, Massachusetts 02138.

Study of the depth variation of Ar$^{37}$, Ar$^{39}$, and H$^3$ in rock 12002, a crystalline rock exposed to a significant solar flare 3 weeks before recovery, showed that the Ar$^{37}$ and Ar$^{39}$ were approximately constant and tritium decreased over 6.4-cm depth. The radioactivities were interpreted in terms of solar and galactic cosmic-ray interactions. Occasionally, a lunar sample was observed to release tritium at low temperature, the source of the low-temperature tritium is not known.

These studies were extended with further measurements, mainly in rocks 14321 and 15555. Rock 14321, a breccia approximately 13 cm thick, was documented on the lunar surface and subjected to a significant solar flare 2 weeks before recovery. Rock 15555, a larger documented crystalline rock, was not subjected to significant flares for several months before recovery.

Because of its half-life, Ar$^{37}$ activity is only influenced by solar flares that occur within 3 months before recovery. Table 1 gives the Ar$^{37}$ activities at various depths in the 3 rocks, together with the calcium contents. The Ar$^{37}$ activity in the top sample of 15555, $13.4 \pm 3.1$ dpm/kg, is lower than in the top samples 14321 and 12002, $38.5 \pm 2.5$ and $30 \pm 3$ dpm/kg, respectively. The 1/24/71 flare produced $400 \pm 60$ dpm of Ar$^{37}$/kg of calcium near the top of 14321; the 11/2/69 flare produced $355 \pm 70$ dpm of Ar$^{37}$/kg of calcium near the top of 12002. On the basis of Ar$^{37}$ cross-section measurements with protons between 50-160 Mev-energy (D'Amico et al., 1971), we estimate that the 1/24/71 flare had $(1.07 \pm 0.17) \times 10^7$ protons ($> 50$ Mev cm$^{-2}$ sr$^{-1}$) and the 11/2/69 flare had $(0.95 \pm 0.25) \times 10^7$ protons ($> 50$ Mev cm$^{-2}$ sr$^{-1}$), if the excess Ar$^{37}$ was produced by protons of greater than 50-Mev energy with a $1/E^3$ differential energy spectrum. Van Allen (letter of 12/14/71) obtained $0.49 \times 10^7$ protons ($> 55$ Mev cm$^{-2}$ sr$^{-1}$) for the 1/24/71 event and $0.32 \times 10^7$ protons ($> 55$ Mev cm$^{-2}$ sr$^{-1}$) for the 11/2/69 event with 20% uncertainty from Explorer 35. Part of the discrepancy is caused by the fact that some Ar$^{37}$ is produced by protons below 50-Mev energy.

Reedy and Arnold (private communication, 1971), by use of an exponential rigidity spectrum ($R_0 = 50$ MV), obtain that $12 \times 10^7$ protons ($> 10$ Mev cm$^{-2}$ sr$^{-1}$) are required to produce the excess Ar$^{37}$ in 14321; Van Allen obtained $6.3 \times 10^7$ ($\pm 50\%$) protons ($> 10$ Mev cm$^{-2}$ sr$^{-1}$) for the 11/2/69 flare from Explorer 35.

Table 2 gives the Ar$^{39}$ activities at various depths in the three rocks. The Ar$^{39}$ increases with increasing depth in 14321, a high potassium rock, decreases slightly with increasing depth in 15555, a low potassium rock, and is approximately constant in 12002, another low potassium rock. The depth dependence of Ar$^{39}$ appears to be related to the action of fast neutrons on potassium. The comparison between the measured Ar$^{39}$ activities and the Ar$^{39}$ activities expected from Reedy and Arnold's (private communication, 1971) calculation for galactic cosmic rays is given in Table 2. The comparison leads to the conclusion that solar flares during the past 1000 yr have increased the Ar$^{39}$ activities at shallow depths by a factor of 2.
Radioactivities in Apollo 14 and 15 Materials
E. L. Fireman et al.

Table 3 shows a decrease of tritium as a function of depth in 14321, evidence for tritium produced by solar-flare interactions. In three of the four samples, no tritium was released at 275°C; a small amount, approximately 10%, was released at 275°C from the fourth. To examine the room-temperature weepages of tritium and radon, samples of Apollo 15 soil, crystalline rock, and a breccia were sealed in closed containers with air and hydrogen carriers on 8/9/71 shortly after the sample box was opened. The tritium, argon, and radon radioactivities in the gas were measured about 3 months later. No tritium weeped from the crystalline rock. Tritium weepage from the breccia was small, between 5 and 10% of the total, and that from the soil sample, approximately 20%. Tritium weeped from material on the moon is probably reimplanted on the lunar surface.

Reference:

Table 1. $\text{Ar}^{37}$ activities and recent solar flares.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Depth (cm)</th>
<th>$\text{Ar}^{37}$ (dpm/kg)</th>
<th>Ca (%)</th>
<th>$\Delta \text{Ar}^{37}$ (dpm/kg Ca)</th>
<th>Solar protons ($&gt;$50 Mev/cm² sr)</th>
<th>$\text{Ar}^{37}/\text{Ar}^{39}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crystalline 15555.98</td>
<td>0-1.0</td>
<td>13.4 ± 3.1</td>
<td>6.7</td>
<td>0 ± 45</td>
<td>-</td>
<td>1.40 ± 0.35</td>
</tr>
<tr>
<td>Crystalline 15555.80</td>
<td>~ 6</td>
<td>31.2 ± 3.3</td>
<td>6.7</td>
<td>-</td>
<td>-</td>
<td>4.2 ± 0.6</td>
</tr>
<tr>
<td>Breccia 14321.81</td>
<td>0-0.5</td>
<td>38.5 ± 2.5</td>
<td>6.4</td>
<td>400 ± 60</td>
<td>(1.07 ± 0.17) × 10^7</td>
<td>4.35 ± 0.37</td>
</tr>
<tr>
<td>Breccia 14321.81</td>
<td>0.5-1.0</td>
<td>37.6 ± 2.5</td>
<td>6.4</td>
<td>387 ± 60</td>
<td>(1.03 ± 0.17) × 10^7</td>
<td>4.43 ± 0.47</td>
</tr>
<tr>
<td>Breccia 14321.81</td>
<td>1.0-1.5</td>
<td>35.0 ± 4.0</td>
<td>6.4</td>
<td>-</td>
<td>-</td>
<td>3.69 ± 0.49</td>
</tr>
<tr>
<td>Breccia 14321.95</td>
<td>~ 12</td>
<td>30.0 ± 2.7</td>
<td>6.4</td>
<td>-</td>
<td>-</td>
<td>2.30 ± 0.34</td>
</tr>
<tr>
<td>Crystalline 12002.57</td>
<td>0-0.8</td>
<td>30 ± 3</td>
<td>5.4</td>
<td>355 ± 70</td>
<td>(0.95 ± 0.25) × 10^7</td>
<td>3.7 ± 0.7</td>
</tr>
<tr>
<td>Crystalline 12002.57</td>
<td>0.8-3.1</td>
<td>25.0 ± 1.5</td>
<td>5.4</td>
<td>-</td>
<td>-</td>
<td>3.05 ± 0.26</td>
</tr>
<tr>
<td>Crystalline 12002.59</td>
<td>4.9-6.4</td>
<td>27.5 ± 2.5</td>
<td>5.4</td>
<td>-</td>
<td>-</td>
<td>3.45 ± 0.41</td>
</tr>
<tr>
<td>Soil 15271.17</td>
<td>0-5</td>
<td>21.4 ± 2.0</td>
<td>8.1</td>
<td>0 ± 25</td>
<td>-</td>
<td>2.14 ± 0.34</td>
</tr>
<tr>
<td>Soil 15281.14</td>
<td>~ 30</td>
<td>36.4 ± 3.3</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>5.15 ± 1.13</td>
</tr>
<tr>
<td>Soil 10017.14</td>
<td>0-5</td>
<td>21.0 ± 2.0</td>
<td>7.4</td>
<td>24 ± 30</td>
<td>-</td>
<td>1.28 ± 0.14</td>
</tr>
<tr>
<td>Soil 10004.24</td>
<td>0-5</td>
<td>27.2 ± 2.2</td>
<td>8.8</td>
<td>48 ± 35</td>
<td>-</td>
<td>2.25 ± 0.22</td>
</tr>
</tbody>
</table>

* Activity at time of recovery.
† Solar-flare-produced $\text{Ar}^{37}$ from calcium in top sample =

\[
\Delta \text{Ar}^{37} (\text{rock}) = \frac{\text{Ar}^{37} (\text{rock})}{\% \text{Ca}} - \frac{\text{Ar}^{37} (15555)}{\% \text{Ca}}
\]

\[
\Delta \text{Ar}^{37} (\text{soil}) = \frac{\text{Ar}^{37} (\text{soil})}{\% \text{Ca}} - \frac{\text{Ar}^{37} (15271)}{\% \text{Ca}}
\]

** $\frac{\% \text{Ca}}{\text{Ar}^{37}} = 37 ± 3 \text{ mb for } 85 \text{ Mev} \leq E_p \leq 158 \text{ Mev}$ Based on simulated target with 6.7% Ca bombarded by 50 - 158 Mev protons (D'Amico et al., 1971) and a 1/E² differential energy spectrum.
Table 2. Ar$^{39}$ activities and estimated excess Ar$^{39}$ from solar flares during the past 1000 yr

<table>
<thead>
<tr>
<th>Sample</th>
<th>Depth (cm)</th>
<th>K (%)</th>
<th>Ar$^{39}$ (dpm/kg)</th>
<th>(Ar$^{39}$/Ar$^{39*}$)</th>
<th>Ar$^{39*}$ (dpm/kg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crystalline 15555.98</td>
<td>0-1.0</td>
<td>0.025</td>
<td>9.5 ± 0.5</td>
<td>(2.1 ± 0.1)</td>
<td>4.6</td>
</tr>
<tr>
<td>Crystalline 15555.98</td>
<td>~ 0</td>
<td>0.025</td>
<td>7.5 ± 0.5</td>
<td>(1.0 ± 0.1)</td>
<td>7.2</td>
</tr>
<tr>
<td>Breccia 14321.81</td>
<td>0-0.5</td>
<td>0.47</td>
<td>8.3 ± 0.5</td>
<td>(1.8 ± 0.1)</td>
<td>4.8</td>
</tr>
<tr>
<td>Breccia 14321.81</td>
<td>0-1.0</td>
<td>0.47</td>
<td>8.3 ± 0.7</td>
<td>(1.6 ± 0.1)</td>
<td>5.2</td>
</tr>
<tr>
<td>Breccia 14321.81</td>
<td>1.0-1.5</td>
<td>0.47</td>
<td>9.5 ± 0.8</td>
<td>(1.7 ± 0.2)</td>
<td>6.6</td>
</tr>
<tr>
<td>Breccia 14321.267</td>
<td>~ 5</td>
<td>0.47</td>
<td>12.1 ± 1.0</td>
<td>(1.3 ± 0.1)</td>
<td>9.1</td>
</tr>
<tr>
<td>Breccia 14321.85</td>
<td>~ 12</td>
<td>0.47</td>
<td>14.8 ± 1.0</td>
<td>(1.3 ± 0.1)</td>
<td>11.6</td>
</tr>
<tr>
<td>Crystalline 12002.57</td>
<td>0-0.8</td>
<td>0.045</td>
<td>8.0 ± 0.7</td>
<td>(2.0 ± 0.2)</td>
<td>4.0</td>
</tr>
<tr>
<td>Crystalline 12002.57</td>
<td>0-3.1</td>
<td>0.045</td>
<td>8.2 ± 0.5</td>
<td>(1.7 ± 0.1)</td>
<td>4.8</td>
</tr>
<tr>
<td>Breccia 12002.59</td>
<td>4.9-6.4</td>
<td>0.045</td>
<td>8.0 ± 0.6</td>
<td>(1.3 ± 0.1)</td>
<td>6.2</td>
</tr>
</tbody>
</table>

* Based on Reedy and Arnold's (1971, private communication) calculation for galactic cosmic-ray production of Ar$^{39}$ in 14321.

Table 3. Tritium activities and weepage.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Wgt. (g)</th>
<th>Depth (cm)</th>
<th>Temperature (°C)</th>
<th>H$^3$ (dpm/kg)</th>
<th>Total H$^3$ (dpm/kg)</th>
<th>Extraction date</th>
</tr>
</thead>
<tbody>
<tr>
<td>Breccia 14321.81 (10.08)</td>
<td>surface (0-0.5)</td>
<td>275</td>
<td>275-Melt</td>
<td>&lt;3</td>
<td>270 ± 12</td>
<td>280 ± 14</td>
</tr>
<tr>
<td></td>
<td>0.5-1.0</td>
<td>275</td>
<td>275-Melt</td>
<td>5 ± 3</td>
<td>197 ± 6</td>
<td>208 ± 8</td>
</tr>
<tr>
<td></td>
<td>0.5-1.0</td>
<td>275</td>
<td>Remelt</td>
<td></td>
<td></td>
<td>6 ± 3</td>
</tr>
<tr>
<td></td>
<td>1.0-1.5</td>
<td>275</td>
<td>275-Melt</td>
<td>6 ± 3</td>
<td>104 ± 6</td>
<td>198 ± 11</td>
</tr>
<tr>
<td></td>
<td>1.0-1.5</td>
<td>275</td>
<td>Remelt</td>
<td></td>
<td></td>
<td>28 ± 5</td>
</tr>
<tr>
<td>Breccia 14321.257 (4.7)</td>
<td>~ 5</td>
<td>275</td>
<td>275-Melt</td>
<td>18 ± 5</td>
<td>161 ± 7</td>
<td>176 ± 11</td>
</tr>
<tr>
<td></td>
<td>~ 5</td>
<td>275</td>
<td>Remelt</td>
<td>&lt;7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Breccia 14321.95 (10.08)</td>
<td>~ 12</td>
<td>275</td>
<td>275-Melt</td>
<td>3 ± 3</td>
<td>128 ± 5</td>
<td>145 ± 8</td>
</tr>
<tr>
<td></td>
<td>~ 12</td>
<td>275</td>
<td>Remelt</td>
<td></td>
<td></td>
<td>9 ± 4</td>
</tr>
<tr>
<td>Soil 15271.17 (1.97)</td>
<td>0-5</td>
<td>275</td>
<td>Melt</td>
<td>45 ± 7</td>
<td>235 ± 10</td>
<td>300 ± 15</td>
</tr>
<tr>
<td></td>
<td>0-5</td>
<td>Melt</td>
<td>Remelt</td>
<td>&lt;7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Soil 15261.14 (2.03)</td>
<td>~ 30</td>
<td>275</td>
<td>275-Melt</td>
<td>20 ± 5</td>
<td>220 ± 15</td>
<td>240 ± 15</td>
</tr>
<tr>
<td></td>
<td>~ 30</td>
<td>275</td>
<td>Remelt</td>
<td>&lt;4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Breccia 15555 (weepage) (9.7)</td>
<td>surface</td>
<td>25</td>
<td>3 months with air</td>
<td>12 ± 1.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Crystalline 15555 (weepage) (18.7)</td>
<td>surface</td>
<td>25</td>
<td>4 months with air</td>
<td>&lt;3</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Soil 15021.2 (weepage) (10.0)</td>
<td>-</td>
<td>25</td>
<td>3 months with air</td>
<td>51 ± 4</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
"Dating" of mechanical events by deformation-induced erasure of particle tracks is made possible by virtue of the fact that natural, fine scale plastic deformation can fragment pre-existing charged particle tracks in lunar crystals, as sketched in Figure 1. Many examples of such effects have been identified in electron micrographs of etched lunar soil grains. In one case, sample 12028, 111, 30.3, 2, slip lines cover the field of view and are correlated with the track lengths: Where the slip spacing is relatively wide (≈0.7 microns) the tracks are typically ≈0.7 microns long; where the slip spacing is reduced to ≈0.07 microns, the track length is correspondingly less. In the light microscope (Leitz Ortholux at 1350X magnification) no tracks are visible in the region of fine slip, and recognizably short tracks are just discernible in the region of coarser slip. Superimposed on the approximately 3 x 10^8/cm^2 short tracks are 3 x 10^5 longer tracks that are uniformly distributed with no differences between the regions of fine and coarse slip. These longer tracks presumably correspond to irradiation of the sample by cosmic rays subsequent to the event which fragmented the
original tracks and hence should allow the deformation history to be inferred.

Since deformation markings are abundant features in lunar soil, their effects provide a powerful tool for constructing deposition histories for stratified soil samples. When many crystals are examined in a given soil layer, it is likely that several of these have been "reset" by the impact that laid down that layer. If we assume that the lowest track density observed in a layer came from near the bottom of that layer, we can compute a surface exposure time. For the long Apollo 12 core (12025 + 12028), we compute intervals ranging from 2 to 60 million years for various sublayers, with an accumulated time of 220 million years for samples for which Arrhenius et al. inferred a 310 million year surface exposure. The difference may be attributed to Arrhenius et al.'s using the lower quartile of their track density distribution instead of the lowest track density observed. Their assumption of no predepositional surface residence for these samples yields a higher age. A similar analysis, assuming that Apollo 11 core 10005 is composed of three layers, gives an integrated surface age of 40 to 75 million years for its 10.5 cm depth.

CRYSTAL WITH TRACKS

CRYSTAL WITH TRACKS AFTER DEFORMATION
Apollo sample 14313 was collected at STA G1 approximately 150 meters east of the landing site in the Fra Mauro region. It is a coherent polymict microbreccia with a complex history of comminution and reagglomeration. At least four generations of breccia (i.e., breccias within breccias) are recognizable. The most "recent" brecciation is represented by large clasts (5-10 mm in diameter) set in a matrix of finer lithic clasts, monomineralic fragments, and fragments of brown glass. These large clasts are rounded and sharply defined megascopically. They are slightly darker brown in color than the matrix, apparently due to higher glass content. However, they appear very similar in composition to the matrix and contain the same types of clasts (including "older" microbreccias). They are heterogeneously distributed and are not a significant constituent in all of the thin sections we have examined. They are most conspicuous in sample 14313,1 where they comprise about 40% by area of the thin section. For the purpose of this paper, clasts are defined as those fragments > 0.25 mm in diameter and, so defined, make up 30% of the rock. The following types of clasts are recognized: (1) igneous clasts, (2) monomineralic fragments, (3) breccia clasts and (4) glassy fragments including glass spherules. There is a complete size gradation (up to 10 mm) in the clasts.

Petrographic examination reveals varying degrees of shock damage ranging from unshocked through shock-melted fragments. Most of the shock-induced features described by Scar (1971) for the Apollo 12 microbreccias have been observed in 14313. Subparallel microfractures which break across grain boundaries are present but overall porosity is low.

The percentages of the types of clasts present are shown in Table 1. Monomineralic, glassy, and lithic clasts occur in the numeric proportions of 3:1:1. Thus, numerically individual crystal fragments (primarily plagioclase, An80-95) are the most abundant clasts present. These fragments range up to 0.5 mm. Intense deformation is illustrated by composite crystalline plagioclase glass fragments in which schlieren are present. The only other major monomineralic clasts are clin- and orthopyroxenes. They have a variable grain size and many exceed 0.3 mm.

Volumetrically the lithic component is the most important clast type. Igneous clasts range up to 4 mm with smaller particles (< 0.25 mm) representing an important component of the matrix. Many of these igneous clasts are reocrystallized to varying degrees. Others retain original subophitic to intergranular textures. They are highly variable in the proportions of the minerals present and exhibit a gradation in composition from pyroxenite to norite to anorthosite. Of these "micronorites" predominate. They appear to be similar to the norites described by Wood et al. (1971) and the orthopyroxene plagioclase fragments of Fuchs (1970). They are composed of subequal amounts of
orthopyroxene and plagioclase with minor amounts of olivine, clinopyroxene and ilmenite. The feldspar laths in the unrecrystallized clasts average 0.15 mm long. Rare phenocrysts of plagioclase exceed ½ mm. Pyroxene and olivine form equant grains 0.05-0.10 mm in size. Preliminary microprobe analyses indicate the following compositions: plagioclase, An75-An95; orthopyroxene En65Wo4Fs31 - En85WoFsl; clinopyroxene En45Wo35Fs20; olivine Fa23-Fa35. Other rock clasts which appear different from the micronorites include mare basalt, and rare clasts of peridotite and dunite composition. Clasts of pyroxenite mineralogy are not infrequent and are texturally similar to the recrystallized micronorites. The small size of the igneous clasts makes any classification of rock types extremely tenuous. Because of the large variation in composition of the noritic clasts, it is conceivable that the pyroxenite clasts may have come from an inhomogeneous norite source. Accordingly, most of them have been included with the micronorites in Table 1. In most of these igneous rock clasts large pyroxene crystals are present, and occasionally contain oriented plates of ilmenite.

Both homogeneous glass (fractured spherules) and heterogeneous glass fragments are present. Colorless glass is most abundant although fragments of yellow and brown glass occur within the matrix.

Based upon our observations, we conclude: (1) the prime igneous rock has noritic affinities and mare-type basalts are extremely rare; (2) texturally there are four generations of breccia and one interpretation is that a maximum of four cratering events is recorded in our samples; (3) the occurrence of abundant glass in the matrices of the two youngest breccias indicates that the last two events were of low intensity; (4) if the assumptions of the origin of the Fra Mauro formation are correct, namely, that it is Imbrian ejecta sampling to depths of 5 km, then we can conclude that noritic rocks were a major near-surface pre-Imbrian rock type.
Table 1. Numeric and Percent Abundance of Clast Types in Apollo Rock 14313.

<table>
<thead>
<tr>
<th>Thin Section</th>
<th>Generationa of Breccia</th>
<th>Monomineralic</th>
<th>Glass</th>
<th>Breccia</th>
<th>Igneous</th>
<th>[B]</th>
<th>N</th>
<th>A</th>
<th>U</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>14313,41</td>
<td>1</td>
<td>144 (46)c</td>
<td>68</td>
<td>36</td>
<td>63</td>
<td>4</td>
<td>58</td>
<td>1</td>
<td>1</td>
<td>311</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>11 (41)</td>
<td>8</td>
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</tr>
<tr>
<td></td>
<td>3</td>
<td>2 (100)</td>
<td>--</td>
<td>--</td>
<td>--</td>
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<td></td>
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<td>2</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>--</td>
<td>--</td>
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<td>--</td>
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<td></td>
<td></td>
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</tr>
<tr>
<td>14313,42</td>
<td>1</td>
<td>21 (38)</td>
<td>14</td>
<td>10</td>
<td>10</td>
<td>10</td>
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<td></td>
<td></td>
<td>55</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>112 (39)</td>
<td>114</td>
<td>26</td>
<td>40</td>
<td>2</td>
<td>37</td>
<td>1</td>
<td></td>
<td>292</td>
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<td>55 (70)</td>
<td>13</td>
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<td>9</td>
<td>8</td>
<td></td>
<td></td>
<td></td>
<td>78</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>11 (53)</td>
<td>7</td>
<td>0</td>
<td>3</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td>21</td>
</tr>
<tr>
<td>14313,43</td>
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<td>87 (38)</td>
<td>68</td>
<td>16</td>
<td>58</td>
<td>4</td>
<td>43</td>
<td>10</td>
<td>1</td>
<td>229</td>
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<tr>
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<td>2</td>
<td>18 (41)</td>
<td>7</td>
<td>7</td>
<td>12</td>
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<tr>
<td></td>
<td>3</td>
<td>7 (64)</td>
<td>4</td>
<td>--</td>
<td>--</td>
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<tr>
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<td>4</td>
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<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td></td>
<td></td>
<td></td>
<td>--</td>
</tr>
</tbody>
</table>

aArranged according to textural age (1 → 4, "youngest" → "oldest").
bB = mare basalt, N = norite, A = anorthosite, U = ultramafic.
cNumber of fragments with percentage of that generation in parentheses.
ANALYSIS OF ORGANOGENIC COMPOUNDS IN APOLLO 11, 12, AND 14 LUNAR SAMPLES.
Flory, D.A., Wikstrom, S., Gupta, S., Gibert, J.M., and Oro, J., Department of Biophysical Sciences, University of Houston, Houston, Texas 77004.

In our work we have placed particular emphasis on obtaining information to allow us to identify which source(s) is the principal contributor for the compounds evolved by thermal and acid treatment. Quadrupole mass spectrometry (QMS) and gas chromatography-mass spectrometry (GC-MS) were used to analyze gases released by thermal treatment. Gases released by acid treatment were analyzed by GC-MS. Sample 14003 was also extracted with benzene-methanol and the extract analyzed by GC-MS with no extractables found at the 10^-9 g/g level.

ACIDOLYSIS. The gases generated upon treatment of the samples with acid were swept into and separated in a Porapak Q (2.5 mm I.D. x 2 m) packed column following our previously described method. DCl was used in order to distinguish hydrocarbons generated by reaction between the acid and the sample from hydrocarbons present as such in the sample. At the time of injection the column temperature was maintained at -50°C for 6 min. and then programmed at a rate of 8°C/min. to 150°C. These conditions proved to give a better separation of the different components evolved from the samples than did our previous work, although separation of all components was not achieved. Samples examined by this technique were solid and crushed aliquots of Apollo 11 breccia 10059, solid and crushed aliquots of Apollo 14 clastic rock 14311, Apollo 12 fines samples 12023 and Apollo 14 fines 14422. In order to distinguish between deuterated and undeuterated hydrocarbons, to resolve C2H2 from C2H6, and to resolve C3H6 from C3H8 semicontinuous mass scans were recorded to provide mass spectrometric evidence for the presence of the different species. Multiple ion plots were then constructed to improve the gas chromatographic separation as reported earlier.

The DCl treatment released H2, N2, CO, CO2, CH4, C3H6, C3H8, C2H4, C2H6 and H2S from all samples. Incomplete GC separation precluded any quantitation of the H, N2 and CO. CO2 ranged from about 2 ppm in sample 14311 to 50 ppm in 10059. CH4 ranged from about 0.1 ppm in 14311 to 40 ppm in 10059. Total hydrocarbons released ranged from about 1 ppm in 14311 to 55 ppm in 10059. CO2 generally correlated with total (deuterated included) hydrocarbons. The ratio of deuterated to nondeuterated hydrocarbons ranged from 2.1 in 10059 to 5.4 in 14311 and generally increased with decreasing hydrocarbon content. Evidence was obtained by multiple ion plotting for CH4, CH3D, CH2D2, CHD3 and CD4 in the methane GC peak of samples 14003 and 10059. Results of a DCl-CH4 methane exchange experiment ruled out any significant D-H exchange during the sample runs.

The deuterated hydrocarbons produced by acid treatment can be considered reaction products with active or "carbide like" carbon atoms in the lunar material while the nondeuterated species evolved should be present as such in
the sample if there is no significant D-H exchange during the times involved in the procedure. The nondeuterated species can then be truly indigenous to the sample or contamination from terrestrial exposure. The deuterated: non-deuterated ratio does not allow one to distinguish between these two sources, but the higher amounts of nondeuterated hydrocarbons found in the older, more exposed samples suggest contamination may be a factor. This ratio, along with the amounts detected, does indicate that there are substantially less hydrocarbons present as such in Apollo 14 samples than in earlier samples. The presence of the several deuterated species of methane may be interpreted as evidence for the presence of some partially hydrogenated carbon atoms which react with the DCl to produce partially deuterated methane. Alternately the hydrogen evolving from the sample could be competing with the deuterium from the DCl for the active sites.

**VOLATILIZATION-GC-MS.** The experimental procedure for GC-MS analysis of volatiles released by stepwise heating of the sample to 980°C was similar to that described in our previous analyses of Apollo 11 and 12 samples. The procedure involves the use of a quartz pyrolysis tube (4 mm I.D. x 25 cm) that can be heated to 980°C which is connected in line to a modified LKB-9000. The physical construction of the system and the gas chromatographic conditions have been improved to give a better separation of the different components evolved. Several samples (14003, 14240, 14311 crushed, and 14422) of both fines and rocks from Apollo 14 have been analyzed by this volatilization GC-MS system. Sample 14240 (SESC, fines) which should be the cleanest and driest sample provided to us was given special handling. This sample was loaded into the volatilization apparatus in a dry nitrogen cabinet at the NASA Manned Spacecraft Center, exposed to D_2O vapors for 72 hours and then volatilized into the GC-MS in a way which excluded any exposure to the terrestrial atmosphere. The special handling of this sample was intended to aid in determining amounts of light hydrocarbons which may be produced by hydrolysis of indigenous material due to terrestrial atmospheric exposure. A duplicate run on 14240 was made according to this same procedure. This sample was also rerun after several exposure to atmosphere.

The volatilization-GC-MS technique provided evidence for the presence of N_2, CO, CO_2, CH_4, C_2H_4, C_2H_6, C_3H_6, C_3H_8 and H_2O (or D_2O in 14240) in all samples. A duplicate run of sample 14422 under evacuated (He carrier gas pumped away between heating steps) conditions produced only about 10 percent of the amounts of all volatiles, with the exception of H_2O, compared to the pressurized (He carrier gas remaining at 40 psi between heating steps) conditions used for the other runs. Methane evolved ranged from a trace in 14311 crushed to 10 ppm in 14422 and generally correlated with CO_2 content. The maximum for hydrocarbon release occurred in the 500-700°C range for all samples. Sample 14422 produced nine times more total hydrocarbons compared to acidolysis while 14311 crushed produced approximately the same total hydrocarbons as did acidolysis. The total water evolved by the samples not intentionally exposed to D_2O ranged from 50-150 ppm with 10-20% being released at temperatures above 700°C. CO was evolved in the 200-600°C range in several samples but the greatest portion came off in the 800-1000°C range. Sample 14240 gave unique results compared to other fines samples in that 10-100 times less vola-
tiles were released even though qualitatively the release pattern was very similar. No CH₄ was detected up to 200°C and only a trace at higher temperatures. The second volatilization after exposure released several times larger amounts of CO₂ and H₂O at 200°C. Trace amounts of methane, ethylene, propane and propylene were released at higher temperatures but the mass spectra intensities indicated the hydrocarbons released at all temperatures were significantly lower than for the unexposed sample.

The results obtained for sample 14240 indicate that hydrolysis of active "carbide-like" substances by terrestrial water does not contribute significantly to the amounts of methane found as such in the DCl treatment. The very low amounts of volatiles, especially methane and the other hydrocarbons, can be attributed to either true differences in the sample or the effects of exposure to the terrestrial environment. The data from the second run of the already pyrolyzed sample 14240, the general observation of increased hydrocarbons at higher temperatures, and the increased yields under pressurized conditions indicate that synthesis (such as Fischer-Tropsch type) could be an important factor in pyrolysis. The presence of CO at the lower temperatures is probably not due to reaction of indigenous minerals (e.g. Fe₂O₃ + 3C → 2Fe + 3CO) as occurs at higher temperatures and may represent a product of reaction between "active carbon" and terrestrial atmospheric components.

VOLATILIZATION-QMS. Direct quadrupole mass spectrometry of gases released by heating to 950°C has been carried out in a manner similar to that described in our previous work with the addition of controlled heating to slowly increase the temperature of the sample. Sensitivity was also increased by approximately 50 as a result of ion source modification. Samples 10086, 12023, 14003, 14156, 14311, and 14422 were analyzed by this technique.

Overall the results were similar to those obtained in our previous work. The presence of water, carbon monoxide, nitrogen, carbon dioxide and argon was confirmed in all of the samples. The temperatures of maximum evolution of the various gases was found to vary considerably from sample to sample. In some cases the release patterns for Apollo 14 samples bear a closer resemblance to samples from earlier missions than to other Apollo 14 samples. The outstanding feature of these results is the absence of hydrocarbons with the one exception of methane in sample 14003. Two samples evolved compounds which we feel represent lunar module exhaust contamination. Sample 14003, a sample of fines collected very near to the LM gave mass spectra for CH₂CO, C₃H₃, HCN, and HNCO which have been reported as lunar exhaust products. A crushed sample of rock 14311 also gave mass spectral evidence for CH₂CO and C₃H₃. Two other samples of fines collected at 50 and 1200 meters from the lunar module did not contain any detectable amounts of these compounds. Evidence was obtained which indicates most of the H₂O and CO₂ evolved is terrestrial contamination. Nitrogen ions found to be evolved in conjunction with CO at 800–1000°C perhaps indicate the rupturing of closed vesicles or traps of some type. Terrestrial N₂ is not significant at such high temperatures in exposed prepyrolyzed samples.
At atmospheric pressure anorthite is liquidus silicate phase, joined by spinel, then by olivine when c. 30% crystalline (the chance of this happening in a random high pressure liquid is $> 0.67$) and subsequently by pyroxene (fig. 1). The composition of 14310 is far removed from atmospheric pressure cotectic liquids and develops iron as a separate phase near or above the silicate liquidus; had this happened and the iron been separated, the relatively high Ni content of 14310 would not have been maintained. 14310 cannot have suffered appreciable fractional crystallization as a system with the present water and alkali contents within, say, 1 km of the lunar surface, nor as a system with its present oxygen content under any conditions. This, and its high $\frac{Mg}{Mg + Fe}$ relative to Apollo 11 and 12 basalts, show 14310 is a more primitive material from the lunar interior than any previously investigated by us. As pressure is increased, however, anorthite, spinel or garnet are the liquidus oxide or silicate phases; none of these phases is suitable as the crystalline residuum from partial melting of the lunar mantle, hence 14310 is not a primary magma. In the presence of water vapour the plagioclase liquidus is rapidly suppressed and replaced by spinel. Olivine succeeds plagioclase as the first silicate at c. 3 kb, equivalent to c. 55 kms depth. The 8% plagioclase phenocrysts in 14310, therefore, indicate derivation from strictly limited pre-eruption temperature and pressure conditions (fig. 2). Restoration of volatilized alkalis (Brown and Peckett 1971) and probable losses of H$_2$O, O$_2$, to the present composition of 14310 leads to compositions similar to terrestrial high alumina basalts. Restoration of alkalis lowers the plagioclase liquidus and in the presence of water vapour at c. 0.25-0.5 kb (5-10 kms) the resultant composition is nearly cotectic with respect to olivine, spinel and plagioclase (spinel-troctolite), and could have been derived from a fractionating magma chamber (Brown and Peckett 1971) precipitating those crystalline phases, whose liquid would contain c. 0.25-0.5% H$_2$O. Derivation of such a calc-alkaline magma from a terrestrial-type lunar upper mantle, with $\frac{Mg}{Mg + Fe} > 0.87$, cannot be excluded.

The compositions B, C, D and E, identified by Reid (1971) as constituents of lunar soil, resemble 14310 and also have plagioclase as liquidus phase at atmospheric pressure, most displaying extensive feldspar crystallization before the appearance of olivine. Chemical differences between the compositions are not, however, explicable solely in terms of plagioclase fractionation. Any effective fractionation scheme to link these compositions presupposes circumstances under which the plagioclase liquidus field is much
All dry runs above 1 bar are self-buffered and most contain small opaque droplets, identified as metallic iron, even above the liquidus.

14310 + 1.5% Na2O + 0.5% K2O
Note slipped P&T scales

Interpretations of run data
14310
(plagioclase-phyric basalt)

The field of chrome-spinel and olivine are restricted to low pressures.

All runs fit phase field shown
- dry experiments
- wet experiments
solid lines-dry equilibria, solvity
certainty
Broken lines-water saturated equilibria

All runs at low oxygen fugacity
(see text)
1 bar runs in Mg
Wet runs in Ag90Pd10 (1150°C) or Pt
Dry runs in Fe; a few in Pt
reduced - i.e. much higher pressure (>300 kms!); or presence of water combined with a moderate increase of pressure (>10 kms); or some combination of these factors. Restoration of alkalis to compositions B, C, D and E does not greatly change the situation, plagioclase being retained as a near liquidus phase in C, D and E and spinel crystallization enhanced (cf. 14310) with additions of 0.5% K₂O + 1% or 2% Na₂O. In B the near simultaneous appearance of plagioclase and olivine is replaced by an olivine liquidus.

Using glass (1400°C) as starting material, near liquidus plagioclase in 14310 nucleated in sprays and could be misinterpreted as quench growths; olivine and pyroxene were encountered at slightly higher temperatures than when rock powder was used as starting material, but essentially similar atmospheric pressure quenching data were obtained in each case. The higher temperature entry of olivine in the glass starting material was eliminated by increasing run lengths from 5 hours to 3 days.

The possibility of elutriation of observed plagioclase phenocrysts from a plag+sp+ol+t situation has NOT been taken into account here.
Lunar fines from Apollo 11, Apollo 12, and Apollo 14 have been analyzed for amino acid precursors. When the aqueous extracts are hydrolyzed, the various samples provide close to a common pattern, consisting of dominant glycine (all amino acids by RT), usually alanine or glutamic acid as the second amino acid, and smaller amounts of two or three others such as aspartic acid and serine. The patterns observed are, in both the unhydrolyzed and hydrolyzed forms, markedly different from those due to typical terrestrial contamination. When the samples and controls were prepared with precautions and by a method comparable to those employed in this study, two other groups of analysts (one by ion exchange chromatography and one by gas liquid chromatography of derivatives of amino acids) obtained results concordant with those found by this group. This concordance was in part observed within the same laboratory at the Ames Research Center, on a specially collected sample from Apollo 14. The concordant results from Apollo 14 (Table I) agree with those obtained by this group from Apollo 11 and Apollo 12 fines (Table II; 1). The finding of the same amino acids by two methods, one using free amino acids and the other using derivatives, virtually assures the identity of the compounds.

The essential procedure for the analysis included extraction of the lunar fines with hot water, hydrolysis of the aqueous extract instead of direct hydrolysis of the lunar fines, and full cleanroom precautions in a laminar flow hood in the subsequent analyses. Glassware had to be specially cleaned (more recently without dichromate) and rinsed, and then baked at 550°C (2). The proportions of amino acids found after hydrolysis have been in the range of 5-70 mg/g. The method developed is believed to yield a minimal recovery and is probably subject to much improvement. All samples are corrected for contamination in reagents, which is in most cases barely measurable. Any contribution from oxidized rocket fuel should be ascertained with samples from Apollo 15.

The significance of these findings may be related to the origin of life and the history of the Moon, as follows. First, even when the content of carbon-nitrogen organic matter is very low, compounds removed from evolutionarily significant amino acids by a single (geologically relevant) step of hydrolysis are found to have been formed extraterrestrially. Secondly, the results suggest that the pathway from cosmic organic matter to life on the Earth could have been pursued on the Moon to the stage of amino acid precursors, and then have been terminated for lack of sufficient water (3).
the third context, the results make use of amino acids as markers. One model of the Moon's history that is consistent with the results is that in which the Moon was baked free of organic compounds, after which organic compounds accumulated in small amounts due to onfall, which may be continuous. The question of how small amounts of amino acid precursors are stabilized requires identifying those compounds and the material from which they in turn came. Various candidate sources of intermediates are solar wind (4), organic interstellar clouds (5), meteorites (6), or comets' tails (7). Further research may permit distinction between these possibilities or others, and may also provide useful data for comparison with eventual analyses of the Martian soil.

Aided by NASA Grant no. 10-007-088. Contribution no. 210 of the Institute of Molecular Evolution.

Table 1. Amino acid contents of SESC samples from Apollo 14 mission by two assay procedures (from precursors)

<table>
<thead>
<tr>
<th>Amino acid</th>
<th>Ion exchange chromatography</th>
<th>Gas-liquid chromatography</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aspartic acid</td>
<td>0.03</td>
<td>0.01</td>
</tr>
<tr>
<td>Serine</td>
<td>0.01</td>
<td>0.03</td>
</tr>
<tr>
<td>Glutamic acid</td>
<td>0.02</td>
<td>0.01</td>
</tr>
<tr>
<td>Glycine</td>
<td>0.17</td>
<td>0.12-0.15</td>
</tr>
<tr>
<td>Alanine</td>
<td>0.04</td>
<td>0.01-0.02</td>
</tr>
</tbody>
</table>

Miami Sample (14240,2)

<table>
<thead>
<tr>
<th>Amino acid</th>
<th>Ion exchange chromatography</th>
<th>Gas-liquid chromatography</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aspartic acid</td>
<td>0.02</td>
<td>0.01</td>
</tr>
<tr>
<td>Serine</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>Glutamic acid</td>
<td>0.04</td>
<td>0.01</td>
</tr>
<tr>
<td>Glycine</td>
<td>0.12</td>
<td>0.08-0.12</td>
</tr>
<tr>
<td>Alanine</td>
<td>0.03</td>
<td>0.004</td>
</tr>
</tbody>
</table>

\(a/\) Corrected for contents in the reagents.
Amino Acid Precursors
S. W. Fox

Table 2. Apparent amino acid contents of lunar samples from Apollo 11 and Apollo 12 (in molar percent)*

<table>
<thead>
<tr>
<th>Amino acids in hydrolyzate</th>
<th>Apollo 11</th>
<th>Apollo 12</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Sample 1</td>
<td>Sample 2</td>
</tr>
<tr>
<td>---------------------------</td>
<td>-----------</td>
<td>-----------</td>
</tr>
<tr>
<td>Aspartic acid</td>
<td>5</td>
<td>5</td>
</tr>
<tr>
<td>Threonine</td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td>Serine</td>
<td>9</td>
<td>10</td>
</tr>
<tr>
<td>Glutamic acid</td>
<td>9</td>
<td>11</td>
</tr>
<tr>
<td>Glycine</td>
<td>50</td>
<td>52</td>
</tr>
<tr>
<td>Alanine</td>
<td>25</td>
<td>19</td>
</tr>
<tr>
<td>Valine</td>
<td>&lt;1</td>
<td>&lt;1</td>
</tr>
<tr>
<td>Isoleucine</td>
<td>&lt;1</td>
<td>&lt;1</td>
</tr>
<tr>
<td>Leucine</td>
<td>&lt;1</td>
<td>&lt;1</td>
</tr>
<tr>
<td>Tyrosine</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Phenylalanine</td>
<td></td>
<td></td>
</tr>
<tr>
<td>BAA**</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Total content of amino acids in hydrolyzate (parts per billion)
53 37 19 69

* Calculated without ammonia. Values for the water blank have been subtracted in each analysis.

** A dominant peak in the basic amino acid (BAA) region, which does not correspond to a proteinous amino acid, with an RT between those of phenylalanine and histidine.

REFERENCES

In Table 1 the average normative composition of 127 glasses from soil samples 14148, 14149, 14156, 14163, 14259 and 14260 and 36 glasses from breccia 14318 are compared with our data for Apollo 11 and 12. The Apollo 14 glasses are enriched in salic components and depleted in olivine and ilmenite. Glasses with more than 5% normative olivine (~50%) fall into three very distinct groups: 1) high FeO (~18%) with low K and Al, and wide variation in Mg and Ti, 2) high Al₂O₃ (~26%) with low Fe and K and, 3) intermediate FeO (~11%) and Al₂O₃ (~17%) with rather wide variations in K. Silica-saturated (<5% olivine and <2% quartz in the norm) or quartz normative (>2%Q) glasses mostly show intermediate Al and Fe content but are more variable in other elements, especially K (0.1 to 1.5% K₂O) than the olivine normative glasses. The Apollo 14 soil glasses show neither the simple mixing of basaltic minerals as in the Apollo 11 samples nor the rather distinct grouping of the Apollo 12 glasses. Instead, the Apollo 14 glasses show relatively smooth, broad variations, but with some preferred compositions (1). The Apollo 14 glasses also seem to be enriched in KREEP, although K and REE are not always positively correlated, which indicates that only some of the K-rich glasses are related to the "norite" (or "magic" or "cryptic") component whereas others may originate from preferential melting of the K-Si mesostasis of some local basalt. The average compositions of these various groups are given in Table 3.

In contrast to the soil glasses all but two of those in breccia 14318 are silica-saturated or quartz normative. Furthermore, they show a negative Al-Ti correlation similar to the Apollo 11 glasses although with a lower absolute TiO₂ content. Apparently these glasses could have formed from local "basalt" (3) by impact melting of various proportions of the constituent minerals. A few glasses with high silica (~51%), high FeO (~18%), high and variable (0.7-1.4%) K₂O and relatively high TiO₂ are likely to be melt products of interstitial materials in these basalts.

The relatively restricted variations in pyroxene composition in breccias 14318 and 14321, Fig. 1, as compared to pyroxenes from the soil samples, indicate that they are not lithified local soil, and support the hypothesis of relatively restricted source materials for these breccias. Also, the 14321 olivines have a fayalite content of 15 to 20% while olivines from the soil center around 40 mol% Fa. The petrography of rock 14321, particularly the lack of glass, indicates some recrystallization. It seems possible that it formed a part of an earlier regolith perhaps buried and heated by Imbrium ejecta and then ejected in the Cone Crater event. This is consistent with
the short exposure age and the probability that it was excavated from a depth of more than 10 m (4).

The breccias 14315 and 14318 contain an abundance of chondrule-like bodies. We have separated similar, single particles from the soil samples and attempted to find structural analogs in chondrites. The "barred olivine chondrule" from 14259,33-id shown in Fig. 2 is (so far) a rarity in the lunar soils; similar chondrules are abundant in most chondrites. On the other hand, glass or plagioclase chondrules abundant in the lunar breccias and soils are rare in meteorites, but have been found in Vigaran (5) and Hedjaz (6) respectively. The differences in the abundance of chondrules and chondrule types in the breccias and in chondrites can easily be explained as due to gross mechanical and chemical differences in the respective parent materials. The manifest structural resemblance between impact-generated lunar spherules and breccias and chondrules and chondrites is demonstrated and provides strong independent evidence for the possibility that the latter were also impact-generated, i.e., that they are impact-ignimbrites.


Table 1. Average normative composition of lunar glasses

<table>
<thead>
<tr>
<th>No. of Grains</th>
<th>Q</th>
<th>Or</th>
<th>Ab</th>
<th>An</th>
<th>Di</th>
<th>Hy</th>
<th>Ol</th>
<th>I1</th>
</tr>
</thead>
</table>
| Soil
| Apollo 11 69 | 0.8| 0.6 | 2.2 | 40.9| 17.3| 17.8| 9.2 | 9.2 |
| Apollo 12 78 | 1.4| 1.5 | 3.5 | 37.6| 17.6| 24.3| 9.5 | 4.8 |
| Apollo 14 127| 2.0| 3.1 | 4.5 | 41.9| 10.5| 24.9| 7.2 | 3.5 |
| Breccia1)    | Apollo 14 36 | 2.6| 3.0 | 4.7 | 45.9| 11.0| 28.0| 0.5 | 2.1 |

1) Section 14318,6; "chondrule" glasses not included.

Table 2. Phase composition (wt.%) in lunar chondrule, Fig. 2.

<table>
<thead>
<tr>
<th>SiO2</th>
<th>Al2O3</th>
<th>TiO2</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na2O</th>
<th>K2O</th>
</tr>
</thead>
<tbody>
<tr>
<td>Olivine</td>
<td>43.3</td>
<td>0.7</td>
<td>0.2</td>
<td>3.2</td>
<td>53.6</td>
<td>0.6</td>
<td>-</td>
</tr>
<tr>
<td>Glass</td>
<td>55.9</td>
<td>22.6</td>
<td>1.5</td>
<td>0.6</td>
<td>6.8</td>
<td>9.6</td>
<td>2.7</td>
</tr>
</tbody>
</table>
Table 3. Chemical compositions of various groups of Apollo 14 soil glasses

<table>
<thead>
<tr>
<th>Abundance</th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
</tr>
</thead>
<tbody>
<tr>
<td>High FeO</td>
<td>16%</td>
<td>44.3</td>
<td>2.2</td>
<td>8.0</td>
<td>23.1</td>
<td>11.5</td>
<td>9.1</td>
<td>0.2</td>
</tr>
<tr>
<td>Intermediate FeO</td>
<td>22%</td>
<td>44.9</td>
<td>1.8</td>
<td>16.5</td>
<td>11.3</td>
<td>10.5</td>
<td>10.7</td>
<td>0.5</td>
</tr>
<tr>
<td>Low FeO</td>
<td>15%</td>
<td>43.8</td>
<td>0.3</td>
<td>25.3</td>
<td>6.3</td>
<td>8.4</td>
<td>14.8</td>
<td>0.1</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>FeO</th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
</tr>
</thead>
<tbody>
<tr>
<td>Intermediate FeO</td>
<td>33%</td>
<td>47.1</td>
<td>2.4</td>
<td>16.2</td>
<td>10.5</td>
<td>8.8</td>
<td>10.3</td>
<td>0.8</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>FeO</th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
</tr>
</thead>
<tbody>
<tr>
<td>Intermediate FeO</td>
<td>7%</td>
<td>49.9</td>
<td>2.5</td>
<td>15.2</td>
<td>10.8</td>
<td>6.2</td>
<td>9.8</td>
<td>0.8</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>FeO</th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
</tr>
</thead>
<tbody>
<tr>
<td>High Al₂O₃</td>
<td>3%</td>
<td>45.9</td>
<td>0.2</td>
<td>32.4</td>
<td>0.9</td>
<td>2.5</td>
<td>17.0</td>
<td>0.8</td>
</tr>
<tr>
<td>High Na₂O</td>
<td>2%</td>
<td>51.0</td>
<td>0.1</td>
<td>30.0</td>
<td>0.2</td>
<td>0.2</td>
<td>13.2</td>
<td>3.6</td>
</tr>
<tr>
<td>High K₂O</td>
<td>2%</td>
<td>76.8</td>
<td>0.7</td>
<td>11.6</td>
<td>2.0</td>
<td>0.1</td>
<td>1.1</td>
<td>0.5</td>
</tr>
</tbody>
</table>

Fig. 1. Pyroxene composition in 14318 (+) and 14321 (*). Zoning indicated by tieline. Range of soil pyroxenes.

Fig. 2. Lunar chondrule. Skeletal olivine with constant composition, Fa₃, in glass (see Table 2). This is an analog to a common type of meteoritic chondrule. Length of section, 0.2 mm.
On March 7th, 1971, the Apollo 14 ALSEP Suprathermal Ion Detector Experiment (SIDE 14) reported an intense, prolonged series of ion bursts. The SIDE mass analyzer (MA) showed the mass per unit charge of these ions to be characteristic of the water vapor group if singly ionized.

The conceptual design of the SIDE MA is as follows: The instrument is basically a neutral gas mass spectrometer for the lunar exosphere where photoionization by solar U.V. and charge-exchange processes with the solar wind are allowed to play the role of the ionizer and the interplanetary electric field associated with the motion of the solar wind plays the role of the ion accelerating field (analyzed by Manka and Michel, 1970) in a conventional mass spectrometer. The SIDE then contains simply an ion velocity filter (Wien filter) and energy per unit charge filter in tandem followed by an ion detection system consisting of a post-analysis acceleration field and a channel-electron-multiplier operated as an ion counter. The velocity filter and energy/q filter are sequenced through a series of steps to insure coverage of the appropriate mass/q and energy/q ranges. The efficacy of this system has been demonstrated by numerous observations by the SIDE of mass spectra associated with the LM rocket exhaust gases, LM cabin venting, and gas emission from the S-IVB and LM lunar impact events. For coarse (eyeball as opposed to computer) examination of the data the sensitivity of the instrument at 18 AMU/q and the 48.6 ev/q energy level is approximately 0.05 ions/cm$^3$ and the resolution ($\Delta M/M$) is approximately 0.1. The SIDE also includes a total ion energy-per-unit-charge spectrometer (TID) that can be used as a check on the MA.

The event on the 7th of March, 1971, first attracted our attention for several reasons. First, its peak intensity (approximately $3 \times 10^6$ ions/cm$^2$-sec-ster-ev) exceeded any other observed with the MA, including a year of data from the Apollo 12 SIDE, by over one order of magnitude. Moreover, this peak intensity was
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well over two orders of magnitude above the background level. Second, the mass/q spectrum showed a distinct absence of the masses above 18 AMU/q that are characteristic of LM exhaust spectra we had seen previously. Until the March 7th event, all mass spectra examined in detail that were clearly not LM exhaust spectra could be easily traced to a man-made event by their close time proximity to that event. Further, such events were always of short duration, lasting only a few minutes.

A more detailed examination of the March 7th data revealed the following:
1. The event had a total time duration of at least 14 hours. It could be roughly divided into three parts separated by quiet intervals. The first part provided a rather prolonged but unobtrusive beginning. The second part consisted of a very intense burst near the middle of the total event. The event remained strong during the third part. (It should be noted that the significance of time variations should be treated with caution because of the narrow field of view of the detector and possible changes in the interplanetary field direction.)

2. The observed mass/q spectrum showed a good fit to that obtained in laboratory calibrations when the water vapor group (H2O+, OH+ and O+), predominantly H2O+, were the main constituents of the beam. Mass/q 20 and heavier ions were not present in significant abundances during this event.

3. The ions were found only in the highest energy/q channel of the MA, 48.6 ev/q, and were exceedingly monochromatic. These ions were also found in the 50 ev/q channel of the TID. Occasionally their energy would increase so that they could then be seen in the 70 ev/q energy channel of the TID. Rarely, if at all, did they go above this energy. No ions were observed in the next lower energy channel in the MA, 16.2 ev/q. The ions were occasionally seen as low as 30 ev/q in the TID, however. These energies suggest a source for the ions some 50 km distant, assuming acceleration by a 1v/km solar wind-induced electric field.

4. The same event was apparently detected at nearly the same time by the Apollo 12 SIDE some 180 km to the west. This statement is based on data from the SIDE 12 TID which resembles that of the SIDE 14 TID. Mass identification at the 12 site must remain uncertain because of the higher background counting-rate exhibited by the SIDE 12 MA during lunar day.
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Items 1, 3 and 4 above are suggestive of a distant rather than local (i.e. at the 14 site) source for the parent gas. Item 2 together with the unprecedented intensity of the event suggests that the water vapor ions are not LM exhaust products. Further, mass/q spectra characteristic of LM exhaust gases have been seen by the Apollo 12 SIDE instrument at time intervals from a lunar landing greater than the time from the Apollo 14 landing to the March 7th event. Therefore, this rules out the possibility of preferential evaporation from the lunar surface of a single component of the exhaust gases as the explanation for this event.

Water dumped from the CSM in lunar orbit has been mentioned as a possible explanation for this event. However, calculations indicate that the expected counting rates due to this source would be about three orders of magnitude less than those actually observed during Part II of the event.

Neither the seismic signals (G. Latham, private communication) nor the SIDE energy spectra seen at the time of the event are characteristic of meteoroid impact events (believed to have been seen on other occasions). Therefore, we arrive at the conclusion that the most likely explanation for the March 7th event is the natural emission of water vapor from beneath the lunar surface. Such a conclusion could explain numerous smaller scale events seen by the SIDE 12 as late as a year after its deployment. This conclusion also appears to be supported by data from the passive seismic experiment (G. Latham, private communication). Furthermore, there appears to be no reason to suspect that water may not be trapped beneath an ice blanket below the lunar surface (as suggested by Gold, 1960 and considered more recently by Schubert et al., 1970) and liberated occasionally by the flexing or fracture of the lunar crust.

References:
AN EXPLANATION OF TRANSIENT LUNAR PHENOMENA FROM STUDIES
OF STATIC AND FLUIDISED LUNAR DUST LAYERS

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For two centuries observers have claimed to have seen transient brightenings and colour changes on the lunar surface. Over 700 observations of various degrees of authenticity have been catalogued, those of more recent events being listed by Moore (1971). Various workers (Blair and Edgington, 1970; Geake et al., 1970; and notably Nash and Greer, 1970) have shown that TLP's cannot be due to luminescence emission from the lunar surface because of its very low efficiency. We have thought for some time that TLP's might be associated with flows or disturbances of lunar dust which raise the albedo. TLP's seem to be most frequently observed in the vicinity of certain craters and around the perimeters of the maria. They appear to be most prevalent when the moon is at perigee and are possibly associated with moon- quakes. To test our hypothesis we constructed a vibrator unit which could agitate a layer of lunar dust with vertical oscillations of about 0.07 mm amplitude at a frequency of 200 Hz. A parallel beam of light from a tungsten lamp illuminated the sample at a selected angle of incidence, and the reflected light was measured by a photomultiplier viewing the surface through a system of apertures to select a particular angle of reflection.

Under static conditions a rough dust surface showed the expected low reflectivity with an albedo of around 7%. However, when fluidised by the vibrator the reflectance rose within a second or so to as much as 50 to 100% above the reflectance produced by a static rough surface. The effect was markedly dependent on the viewing angle but relatively independent of the angle of incidence, as is shown in the polar plots of fig. 1. The brightening effect is a maximum for reflection angles of about 20° to the plane of the dust surface on the opposite side of the normal from the incident light. Tests on terrestrial basalt and sand layers showed smaller changes in albedo under flow conditions. In addition to these measurements we also studied the effect of fluidising a dust layer on the diffuse reflection spectrum. No significant spectral dependence of the incremental brightness was found, but the sieved nature of lunar dust samples precludes any detailed correlation with the particle size distributions existing on the lunar surface and thus with effects of size.
on spectral scattering. We also need to look at spectra during the initial rise of reflection since lunar dust flows may not be as rapid as those in our experiments. As an important addendum to this work we are making a system to measure changes in the polarization of the reflected light when the sample is fluidised. We expect significant changes to occur in the form of the polarization curves previously reported (Geake et al, 1970).

Our conclusion is that the 'fairy castle' structure of dust on the lunar surface (and the simulated version in our experiments under static conditions) which gives rise to the low albedo is destroyed temporarily by dust flow. This makes the optical scattering properties similar to those of terrestrial basalts and sands of similar composition (but having higher albedos). Such changes will be a maximum at an angle which maximises the shadowing effects of the static dust structure. TLP's arising from this mechanism would therefore be more readily observable in regions where the lunar surface is inclined at a certain angle to the line of sight of the observer. We would therefore expect a correlation between TLP's and those locations on the lunar surface (crater walls, perimeters of maria etc.) for which the angular criterion is satisfied. It is in these locations that dust movements are most likely to occur during lunar seismic activity.

References


Fig. 1. Variation of lunar dust sample reflectance (sample no. 14259.56) with viewing angle for two different angles of incidence and under static and 'fluidised' conditions.

A. Incidence angle 0° to normal: (i) Static dust layer, (ii) 'fluidised' dust layer.

B. Incidence angle 60° to normal: (i) Static dust layer, (ii) 'fluidised' dust layer.

Broken line shows viewing angle for maximum brightening effect.

Samples returned from the lunar surface provide clear evidence that the moon underwent chemical differentiation by internal melting and upward movement of silicate liquids during the first billion years of lunar history. Whole rock Rb/Sr ages, particularly on KREEP basalts, clearly indicate that this differentiation started with the formation of the planet or very shortly (100-200 million years) thereafter (1).

The chemical composition of silicate liquids brought to the surface have two common characteristics; i.e. all are depleted in volatile elements relative to chondrites, as is evidenced by their low K/U ratios; and all are depleted in siderophile elements. In addition, the dark-colored basalts of the mare region appear to be readily distinguished from other basaltic rocks or their derivatives by a number of chemical characteristics. Mare basalts consistently have: 1) high FeO contents and high FeO/MgO ratios, 2) low, but variable, Al₂O₃ contents, 3) high and variable TiO₂ contents, and 4) low, but variable, large ion lithophile element contents.

To a first approximation, these basalts appear to be formed by moderate (3-10%) to extensive (20-30%) partial melting of a pyroxene-rich source. In addition, there is strong trace element evidence for the occurrence of plagioclase in this source.

The number of samples from highland regions is limited, and the chemical and textural character of these samples is inferred rather than directly observed. Nevertheless, it is clear that the highlands probably contain a wide variety of basaltic rocks. Inferred primary basaltic liquids differ widely in the range of trace element characteristics. All, however, appear to be rich in plagioclase. In addition, plagioclase-enriched rocks (anorthosites) may be common in highland areas. It may be inferred that the rocks from the highlands are, in general, higher in Al₂O₃ and lower in FeO than the mare basalts. The difference in Al₂O₃ content is dramatically demonstrated by the x-ray fluorescence experiment (2).

The variation in LIL element abundances in very old (approximately 4.5 billion years) lunar basaltic liquids poses a major problem for the origin of these liquids. The KREEP basalts appear to require small degrees of partial melting of a plagioclase-rich source. The LIL element abundance patterns in these basalts imply that they are derived from a source with U, Th, Ba, REE, and Sr concentrations at least 5 times those of average chondrites. The origin of LIL element-poor liquids is less well-defined. If they are formed by very extensive, perhaps complete, melting of the outer shell of the moon, they require that this shell be extensively enriched in Ca, Al, and refractory LIL elements. These inferred concentrations and the inferred abundance of plagioclase suggests that the moon had an outer shell rich in refractory
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It is suggested that this shell was produced by the accretion of materials that were enriched in refractory elements during condensation from a hot solar nebula. The depletion in volatile elements obviously is concordant with this suggestion.

It has been shown (3,4) that the whole moon cannot be as rich in aluminum as the outer shell here inferred. The inversion of Ca/Al-rich assemblages to eclogite at modest pressures thus requires that the primitive moon had a more ferro-magnesium interior composition. It is suggested that this interior accreted from material that separated from the solar nebula at lower temperatures. It may, in fact, approach chondrites in many of its chemical characteristics. The high Fe/MgO ratio of the mare basalts, as well as the LIL element abundance patterns, provide further evidence for the chemical heterogeneities already inferred.

The chemical structure for the primitive moon inferred from the surface chemistry can also explain the relatively high heat flow (5) and thermal history inferred from the ages of igneous rocks (1). Extensions of earlier thermal model calculations (6) using uranium concentrations that decrease from 0.1 ppm in the outer 200 kilometers to 0.02 ppm in the inner 1,000 kilometers and initial temperatures that decrease from 1200°K at the surface to 273°K at the center of the primitive moon predict that melting of the interior of the moon begins 100-200 million years after formation of the moon beginning at a depth of ~100 kilometers, proceeding to depths of 200-400 kilometers 1-1.5 billion years after formation of the moon. However, the present predicted heat flow for these models is still less than 3/4 of that reported for the Apollo 15 site. The present temperature predicted by these models does not exceed the basalt solidus but is substantially higher than that inferred by Sonett, et al. (7).

In summary, we suggest that many of the chemical and thermal characteristics of the moon can be consistently explained if the moon accreted as a heterogeneous body in which refractory elements are enriched in the outer 30-20% of the mass of the moon.

REFERENCES:

TOPOTACTIC REACTIONS IN SOME LUNAR PYROXENES AND OLIVINES

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Oriented products developed within a host single crystal provide evidence of post-crystallisation reactions within a mineral. These products can sometimes be seen optically, and single crystal diffraction methods are particularly powerful in studying such topotactic reactions, for they can detect very small amounts of product phases which are often sub-microscopic in the early stages of a reaction. For pyroxenes and olivines diffraction techniques have been successfully applied both to laboratory-induced reactions and to those found in natural terrestrial processes (1,2,3). Results of the examination of pyroxenes and olivines from Apollo 11 and 12 rocks have already been reported (e.g. 4,5); the former show only pyroxene-pyroxene exsolution relationships (sometimes quite complex), and there were no detectable topotactic reactions in the olivines. The identification of additional oriented products in the same minerals from a fragmental Apollo 14 rock suggests that reactions hitherto unsuspected in lunar rocks may have occurred.

This rock (14321) shows a wide variety of rock and mineral clasts set in a partially recrystallised fine-grained matrix. Two thin sections (19 and 240) and a rock chip (106) were available for this study.

The rock is of a markedly heterogeneous texture, and it is almost impossible to correlate directly crystals removed from the chip with a particular occurrence of the same mineral as seen in thin section.

Both clinopyroxene and orthopyroxene crystals have been removed for X-ray examination, and all show pyroxene-pyroxene exsolution effects in their diffraction patterns. For most clinopyroxenes the host has either the C2/c or P$n_\text{c}$ type structure with exsolution of the other phase on (001); the proportions of the two phases are variable from crystal to crystal, and nearly all exhibit the peculiar complexities already reported for lunar pyroxenes. A few crystals show no pigeonite but only a C2/c host carrying exsolved orthopyroxene crystals show exsolution of twinned augite-type structures on (100), often in such small amounts that they can only be detected on heavily exposed photographs. On such strong patterns additional reflections corresponding to other oriented phases can be found; these correspond to ilmenite and spinel structures with the orientations described for terrestrial pyroxenes (1). In some crystals both phases are present, but in others only one of them can be detected. Re-examination of clinopyroxene crystals shows that in some of the crystals in which orthopyroxene is exsolved faint traces of similar oxide
phases can be found. The spinel phase is exsolved only in the m,l orientation as in natural terrestrial hypersthenes and in heated iron-bearing orthopyroxenes (1). This suggests that the exsolution of the spinel phase (and possibly the ilmenite phase) is associated with the presence of orthopyroxene as host or exsolution product. Oriented ilmenite lamellae in orthopyroxene crystals in Apollo 14 soils have already been reported (6), and both coarse and fine lamellae of ilmenite can be seen optically in some orthopyroxene crystals in the thin sections of 14321. In some of these crystals there are fine droplets of a green spinel (probably incorporated during growth), and occasionally, in addition to ilmenite, there are discontinuous lamellae of an opaque phase, possible spinel, in two orientations.

Many of the olivine crystals removed from the rock chip show diffraction patterns typical of the composition range 20-40%Py with occasional traces of superposed powder arcs due to misoriented inclusions very similar to those described for other lunar olivines. Careful examination of heavily exposed photographs for some crystals, however, show weak symmetry related maxima additional to those of the main olivine pattern. These are not always sharp and are sometimes extended around constant 0 loci, as has been reported for exsolved phases in terrestrial olivines. They are not always the same in different crystals which, coupled with their generally weak intensities, makes their interpretation as particular reaction products very difficult; indeed it seems probable that different products may develop in different crystals. One set of extra spots occurs more consistently than any other, and it accords with the development of an oriented spinel-type structure (2). So far it has proved impossible to identify with certainty any other products; in one crystal there may be a trace of a wüstite-like phase whilst in another crystal the spinel phase is in the alternative orientation found in some heated terrestrial olivines (3). One or two crystals also show weak diffuse maxima contiguous to the sharp olivine maxima, very similar to those reported in terrestrial olivines at the earliest stages of the "iddingsitization" process (2). In some olivine crystals which have been X-rayed there are optical inhomogeneities in the form of erratic thread-like trails. In thin section, many of the clasts of olivine have micron-wide thread-like loops of a pale material, somewhat darker in colour than the olivine and of a higher refractive index, occasionally with a beaded appearance. Such inhomogeneities are common in section 19 but only rarely present in 240; they bear some resemblance to the curving rod-like inclusions seen in some olivines from 12018 and tentatively identified as a chrome-spinel (7).

It is clear that these unexpected reactions in lunar olivines and pyroxenes are only weakly developed and likely to be very localised. Their significance in a lunar environment is uncertain; in terrestrial rocks their presence would suggest oxidation and possibly hydration reactions. There is considerable evidence that such conditions are unlikely generally on the moon, but it is conceivable that comparable conditions might exist locally. Alteration has already been drawn to the development of goethite rims around metallic iron
fragments in a comparable Apollo 14 fragmental rock, and to the parallels between lunar impactites and their terrestrial analogues in which oxidised and hydrated minerals are formed when glasses containing "frozen-in" water are devitrified (8). On the other hand it may be argued that the slight but significant compositional variations of the lunar minerals compared to similar terrestrial pyroxenes and olivines are responsible for the particular exsolution products, especially since the lunar rocks in which they occur certainly show evidence of re-heating. The precise constitution of the exsolved phases could be of significance in resolving the nature of the reaction processes, but these are difficult to obtain. For the present this problem is unresolved.

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References


LUMINESCENCE OF LUNAR MATERIAL EXCITED BY PROTONS OR ELECTRONS

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Two samples of Apollo 14 fines (14163.51 and 14259.56) show luminescence emission spectra similar to those for Apollo 11 and 12 fines previously investigated (Geake et al., 1971); their efficiencies are between those for 'dark' and 'light' (plagioclase-rich) Apollo 12 fines. Excitation by 60KeV protons or 10KeV electrons has been used, with similar results. Decay-times are now being measured, using a pulse-modulated electron beam, and a phase-sensitive detector with adjustable phase-shift.

Luminescence emission photographs have been taken, in colour, of two Apollo 14 rock samples (14310.206 and 14312.3); these both show only the characteristic plagioclase luminescence observed for Apollo 11 and 12 rocks. Other luminescent components were observed in earlier samples, and work is in progress on the identification of these.

As with Apollo 11 and 12 material, the main luminescent component of the Apollo 14 samples appears to be plagioclase. This shows three main emission peaks at about 4500, 5600 and 7800A respectively; our previous investigations (Geake et al., 1971) indicated that Mn$^{2+}$ in Ca$^{2+}$ sites is the cause of the brightest emission peak, at 5600A. The 4500A peak is generally ascribed to the strain defects common in silicates. Interest now centers on the IR peak at 7800A, which is much stronger for terrestrial than for lunar plagioclase, a difference which probably contains information about their different conditions of formation. We are now investigating the cause of this IR emission, using lunar samples and doped samples of terrestrial and synthesised plagioclase, and there is now some evidence that the cause may be ferro iron, Fe$^{3+}$, in Ca$^{2+}$ sites.

Van Doorn and Schipper (1971) have recently found that sodalite (Na$_4$Al$_3$Si$_3$O$_{10}$Cl) doped with Mn$^{2+}$ gave a green emission band at about 5400A; this is similar to the one we found for plagioclase doped with Mn$^{2+}$. However, they also found that sodalite doped with Fe$^{3+}$ gave a broader emission band with its peak at about 7000A; this peak resembles, in both position and width, the IR peak which we find to be strong in terrestrial plagioclase, but weak in lunar plagioclase, and may provide a clue as to its cause. We have therefore doped some terrestrial plagioclase (labradorite) with iron by heating it to 1200°C in air with ferrous sulphate, in the expectation that at this temperature, in air, most of the iron would go in as Fe$^{3+}$. Fig.1 shows un reduced spectral scans of the emission under electron excitation for doped and undoped samples: the IR
peak has evidently been enhanced in relation to the green and blue peaks by adding iron. However, the 'doped' scan is shown at 5 times the gain of the undoped one, so the actual effect of doping has been a reduction everywhere, but less of a reduction for the IR peak than for the others. There are at least two possible explanations for this, but both are consistent with Fe$^{3+}$ as the cause of the IR peak. One is that the iron actually went in partly as Fe$^{2+}$ and partly as Fe$^{3+}$; Fe$^{2+}$ is a powerful quencher of luminescence, so there might be an overall reduction in intensity, superimposed on some enhancement of the IR peak alone by the Fe$^{3+}$ present. However, attempts to produce IR emission by iron doping of synthesised plagioclase, and also of CaO and wollastonite, have not so far been successful.

Another possible explanation is that the spectral changes shown in Fig.1 are simply due to differential quenching by iron. We have now made some decay-time measurements which tend to support this explanation, but even these results, together with all the other features of the IR peak, seem consistent with theoretical predictions as to the properties of Fe$^{3+}$ as a luminescence centre. If we compare the probable effects of Mn$^{2+}$ and Fe$^{3+}$ as impurity ions, the expected luminescence transitions are very similar, in that they are both spin-forbidden and both have the same d$^5$ electron configuration. They therefore have the same energy level diagram, as shown in Fig.2, and differ only in the value of the crystal field splitting parameter $\Delta$; $\Delta$ is greater for the trivalent ion than for the divalent one, so a peak due to Fe$^{3+}$ would be expected to occur at a longer wavelength than for the green Mn$^{2+}$ peak, because the energy gap decreases as $\Delta$ increases. Also, the slope of the upper level is greater at the Fe$^{3+}$ position, so the energy spread is greater for a given $\Delta$ spread due to lattice vibrations, and the Fe$^{3+}$ peak should be broader than the Mn$^{2+}$ peak, as the IR peak is indeed observed to be. The luminescence decay-times for the two peaks should be similar, and fairly long (of the order of 1ms), as the same spin-forbidden transition is involved – but nevertheless slightly shorter for Fe$^{3+}$ as the energy jump is less, giving a greater chance of non-radiative de-activation. Our measurements show that the green peak for the iron-doped plagioclase (attributed to the Mn$^{2+}$ already there) has a decay-time of about 5ms, whereas for the IR peak it is 1 - 2ms, which is consistent with it being due to Fe$^{3+}$. The blue peak has a decay-time at least an order of magnitude shorter, which is to be expected if it is caused by strain defects. As the IR peak has a slightly shorter decay-time than the green one, it should be somewhat less affected by a quenching agent such as Fe$^{2+}$, and this alone could qualitatively explain the spectral change produced by iron doping.

There is thus a good deal of circumstantial evidence that the IR peak for plagioclase is caused by Fe$^{3+}$, and this is consistent both with theoretical predictions and with our experimental observations. Other transition metal ions could have rather similar properties, except that the most likely ones would probably produce emission much further into the IR than that observed.

Our tentative explanation of the much weaker IR emission from lunar
plagioclase, as compared with terrestrial plagioclase, is therefore that the lunar material probably has most of its iron in the form of Fe$^{2+}$, owing to a shortage of oxygen during its formation, whereas abundant oxygen would produce dominant Fe$^{3+}$ in the terrestrial material. This further indication of oxygen shortage when the lunar surface material was formed is consistent with geological evidence, and also with our earlier work on the decomposition of heated lunar ilmenite.

Fig. 1 Labradorite

Recorder deflection

Fe-doped

no Fe added

4000 5000 6000 7000 8000
Wavelength Å

Fig. 2 Schematic energy level diagram for transition metals of d$^5$ electron configuration in a cubic crystal field. The 4$^T_1$ level is triple; only the emitting state is shown.

References


The rationale for the lunar halo search is that halos in terrestrial minerals serve as pointers to localized radioactivity and thereby make possible analytical studies that bear importantly on the general problems of isotopic dating and mode of crystallization of the host mineral. For example, several minerals such as mica, fluorite and cordierite often contain unusual halos originating with fairly short half-life alpha radioactivity, e.g., $^{218}\text{Po}$ ($t_1/2 = 3\text{ min}$), and it is not clear how the presence of these halos can be reconciled with a hydrothermal, magmatic, or metamorphic origin of the host minerals (1). Therefore, a further aim was to pursue such ancillary studies on terrestrial halos and on certain samples of special origin (e.g., tektites, meteorites).

With respect to the first aim, the MSC thin section collection from the Apollo 11-15 suite of rocks has been personally scanned for halos on several occasions with completely negative results. It is difficult to estimate the probability that this means a real absence of halos because the only basis for such an estimate, their terrestrial occurrence, is exceedingly spotty. Further, terrestrial halos are most numerous in rock types that are generally conspicuously absent from the lunar samples thus far available for study. The absence of halos is not due to a lack of radioactive inclusions in lunar samples, since such have been well identified in rocks such as 12013. Perhaps the high radiation environment indigenous to the lunar surface effectively anneals the halos as they are being formed.

In parallel with the search for halos in lunar material, certain types of terrestrial halos have been examined for information on the general field of radioactive chronology. In particular, we have used the ion-microprobe mass spectrometer to determine lead isotope ratios in the microscopic inclusions within certain terrestrial radiohalos that were expected to possess unusual Pb isotope abundances (1). Specifically, one inclusion showed $\text{Pb}/\text{U}$ and $\text{Pb}/\text{Th}$ ratios $> 5000$ with $^{206}\text{Pb}/^{204}\text{Pb} > 1000$, i.e., no $\text{U}$, $\text{Th}$ or $^{204}\text{Pb}$ detected. In another case, the $^{206}\text{Pb}/^{207}\text{Pb}$ ratio was $\approx 60$, an impossibly high value if the Pb isotopes had been derived from uranium decay. Clearly, any attempt to date this sample in the usual way would be impossible, and it appears desirable to establish the extent of such anomalies in lunar and other terrestrial samples.

While their exact origin and history is uncertain, iron meteorites are important in radioactive chronology as reference material for defining primordial Pb isotope ratios. Over the last decade, several analyses have shown modern radiogenic lead in addition to the primordial mixture in the troilite phase of a number of iron meteorites (2). These anomalous results have usually been attributed to terrestrial contamination (3), but in order to settle the question unambiguously, we have initiated ion-probe Pb isotope studies of this class of samples. To evaluate the potential of the ion-probe to perform accurate Pb isotope ratios on the meteorites, NBS common lead was utilized as a
standard in test runs. A comparison of ion-probe (IP) runs at Oak Ridge with
the NBS reference data for their sample yielded the following results (4):
\[ \text{Pb}_{204}/\text{Pb}_{206} = 0.0591 \pm 0.0074 \text{ (IP), 0.059042} \pm 0.000037 \text{ (NBS)}; \]
\[ \text{Pb}_{207}/\text{Pb}_{206} = 0.909 \pm 0.025 \text{ (IP), 0.91464} \pm 0.00033 \text{ (NBS)}; \]
\[ \text{Pb}_{208}/\text{Pb}_{206} = 2.174 \pm 0.035 \text{ (IP), 2.1681} \pm 0.0008 \text{ (NBS)}. \]

Another part of the study was to have been a comparison of lunar and ter­
restrial U and Th halo sizes in relation to the fundamental question treated
more than 30 years ago by Dirac (5) and more recently by Hoyle and Narlikar (6),
i.e., that of the possible time variation of the fundamental physical constants
and in particular of the radioactive transformation rate over geological time.
Due to the absence of lunar specimens comparative studies were not possible,
yet terrestrial halo investigations have yielded some data that may bear con­
siderably on lunar radioactive chronology. As background information I note
that in the context of the Dirac hypothesis, Gerling and Ovchinikova (7) have
recently reported differences in rock ages as measured by various age-dating
techniques, a situation which they attribute to a variation in the alpha-beta
branching ratio in the Precambrian era. A few years ago Gamow (8) proposed
using the ring structure of uranium halos to check branching ratios in the
Precambrian, not realizing that Wilkinson (9) had earlier interpreted the same
data (somewhat erroneously) in favor of confirming presently accepted values.
Interestingly, the same data which for decades have been utilized to establish
decay rate invariance and branching ratio constancy are now being interpreted
to imply a variation in the radioactive decay rate (10). Clearly the same data
cannot support both conclusions.

From \( \alpha \) -decay theory it has been argued previously that the near agree­
ment between uranium and thorium halo ring radii and computed alpha ranges in
the same mineral was proof of an invariant decay rate (11). To see this is not
exactly true we use the \( \alpha \) -decay theory notation given in (12) to compute the
fractional change in the decay constant (\( \lambda \)), arising from the fractional change
in ring radius (\( r \)). In this case \( \ln T \approx 3Z^2E^2 - 4ZE - \frac{1}{T} \lambda = \nu T \) where \( T \)
is the transmission probability, \( E \) the energy of the emitted \( \alpha \) -particle, \( Z \) the
atomic number of the daughter nucleus, \( \nu \) is the frequency with which the \( \alpha \)
particle strikes the barrier, and \( R \) is the nuclear radius. It then follows
that in the appropriate units \( d\lambda /\lambda \approx \left[ \frac{3}{2} Z^2 E^2 - 1 \right] (dR/R) + \left[ \frac{1}{R} + \frac{2Z^2}{E^2} \right] (dE/E) \).
A particle of mass \( m_1 \) and charge \( Z \) has a range \( r \) (halo radius) given
approximately by the expression \( r = \text{const.} \times Z^2 E^2 m_1 \). Subsequent calculations
lead to the expression \( d\lambda /\lambda \approx 43(dR/R) + 46(dE/E) \). Since the minimum
uncertainty in measuring halo radii is about \( \Delta r = 0.1 \) \( \mu \)m, then the resulting frac­tional uncertainty in \( \lambda \) for \( ^{238}\text{U} \) (halo radius 12.5 \( \mu \)m) is
\( d\lambda /\lambda \approx 46(0.2/12.5) \approx 1/2 \) assuming \( dR = 0 \). Since some uncertainties are involved in the conver­sion of air to mineral ranges, Van de Graaff \( \text{He}^{++} \) ion beams of varying energy
were utilized to induce coloration bands in the mineral containing the halos.
In examining scores of U and Th halos and in attempting to match their ring
radii with the corresponding ranges obtained from the experimentally determined
range-energy curve, I have found occasional rather than systematic differences
in certain halo ring radii. Since the Van de Graaff induced ranges correlated
closely with halo radii at 4.2 Mev (13 ± 0.5 \( \mu \)m V.d.G. vs 12.7 ± 0.3 \( \mu \)m for
halo radii) and 5.3 Mev (19 ± 0.5 \( \mu \)m V.d.G. vs 18.9 ± 0.5 \( \mu \)m for halo radii),
at present I do not attribute such differences to an actual change in \( E_\alpha \).
LUNAR RADIOACTIVE CHRONOLOGY

R. V. Gentry

An unanswered question is whether \( \lambda \) variations might occur without changes in halo radius (\( \Delta R = 0 \)). Wilkinson (9) has suggested that the physical constants might vary in some unknown fashion without affecting \( \lambda \). The contrary of this suggestion would imply the correlation of measured \( \alpha \)-ranges and halo radii is not a test of \( \lambda \) invariance. Variations in \( \lambda \) might then be detected by noting age differences in samples determined by radiometric, geologic, or archeologic methods. As a case in point, there is a wide discrepancy between the radiometric (13,14) ages (0.7 my and 4 my) and the geologic and stratigraphic (15,16) age (several thousand years) of the australites. That two cosmic events separated by several million years would result in similar strewn-field tektite patterns is in itself most unusual irrespective of whether the tektites are of lunar or terrestrial origin (17). Evidence suggesting a common link between the H/Na (4 my) australites and the general australite population comes from recent Th/U determinations at ORNL (18), i.e., Th/U = 5.9 (AN-87) and 8.3 (P-192) in the H/Na group compared to Th/U values of 4.8 - 8.5 previously reported for the general group (19).

It has been remarked that the fission track age of the new H/Na group of australites may be high for some reason (17). This is quite significant in that it is universally agreed that fission tracks can form only after the tektites have cooled. Unless the australite fossil fission tracks originated with a spontaneous fissioning nuclide besides \(^{238}\text{U}\), the only remaining possibility for a "high" fission track age is a hiatus in the decay rate due to causes presently unknown. While this explanation is unusual, I do not think it can be definitely ruled out by simply appealing to the regularity of radiohalo data.

Acceptance of this hypothesis would not necessarily have implied significant changes in the other fundamental constants but would affect age dating.

*Research sponsored by the U. S. Atomic Energy Commission under contract with Union Carbide Corporation, and by NASA (NASA/T-90277) and by Columbia Union College through NSF grant GP-29510.

Clinopyroxenes from Apollo 12 and 14: Exsolution, Cation Order and Domain Structure; Subrata Ghose, George Ng and L. S. Walter, Goddard Space Flight Center; Code: 644, Greenbelt, Maryland 20771.

Clinopyroxenes from 14310, 155 are mostly fine grained pigeonite, showing two sets of augite exsolution lamellae parallel to (001) and (100). Δβ is equal to 2°30'. The corresponding pigeonite and augite X-ray diffraction spots are connected by faint diffuse streaks and the b-type pigeonite reflections are sharp. 14053, 44 clinopyroxenes are pigeonite-cores with augite rims. Δβ ranges from 2°25' to 2°50' and the b-type reflections range from diffuse to sharp. Degree of cation disorder is fairly high in pigeonite from 14053, indicating that this rock apparently has been quenched quickly above 1000°C. The degree of cation order in the mixed pigeonite-orthopyroxene phase from 14310 is higher than that in pigeonite from 14053. This evidence along with the large domain size (ΔX ≈ 1500 Å) in pigeonite from 14310 indicate a slow cooling rate. A pink subcalcic augite from the soil sample 14162 does not show any exsolution lamellae.

Core pigeonite from 12053, 72 shows no exsolution, but very diffuse b-type reflections stretched parallel to a*, indicating the presence of very fine-scale domains. Rim augite from 12053, 72 shows the smallest Δβ (1°35') and very diffuse exsolved pigeonite spots, connected by diffuse streaks with corresponding augite spots. Subcalcic augite from 12038, 72 shows similar behavior, but has larger Δβ (2°17') and sharper exsolved pigeonite spots. Both 12053 and 12038 subcalcic augites show a second generation of very fine exsolution lamellae, rotated about 1° (with b as the rotation axis) with reference to the first generation exsolution lamellae. These X-ray diffraction observations are in complete agreement with high-voltage electron microscopic observations on exsolution and domain structure in these pyroxenes (Christie et al., 1971, Proceedings, 2nd Lunar Sc. Conf., p. 69-89). Pigeonites from both 12053 and 12038 show high degree of cation order. These rocks following crystallization from the melt, must have been held at ~1200°C for some time, when most of the exsolution took place; this stage was followed by rapid cooling to ~700°C, where the rock was briefly annealed.

In contrast, 12021 pigeonites show two sets of augite lamellae parallel to (001) and (100) with sharp augite spots (Δβ = 2°55'), sharp b-type reflections and a fairly high degree of cation order, indicating continuous cooling to ~700°C or lower at a slower rate. Pigeonites from 12040 show diffuse exsolved augite spots only (a_A ≠ a_P), indicating that the rock after crystallization was cooled very quickly to ~700°C, where it was briefly annealed. This observation supports Green et al.'s (1971, Proceedings, 2nd Lunar Sc. Conf., p. 601-615) interpretation of the history of the rock 12040. The very strong diffuse streaks connecting augite and pigeonite reflections in 12053 and 12038 pyroxenes indicate the presence of continuous structural and chemical variation from one lamellar region to the next. This evidence suggests the spinodal decomposition mechanism for the very early stages of exsolution.
in some clinopyroxenes. In the early stages of $c_A^* \neq c_B^*$ ($a_A^* = a_B^*$) and compositional fluctuations develop parallel to (001), followed by fluctuations parallel to (100), when $a_A^* \neq a_B^*$. However, the available X-ray diffraction and electron microscopic evidence does not prove the spinodal mechanism conclusively. The size of the domains in pigeonite (as observed through the electron microscope by imaging through the $b$-type reflections) increases progressively in the order: 12053 (50-100Å), 12038 (~500Å), 12021 (~1000Å), 14310 (~1500 Å). The shape of the domains change from rounded irregular in 12053 to blocky columnar (parallel to the $c$-axis) in 12021.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Pyroxenes</th>
<th>$a$ (Å)</th>
<th>$b$ (Å)</th>
<th>$c$ (Å)</th>
<th>$\beta$ (°)</th>
<th>Wo-En-Fs</th>
</tr>
</thead>
<tbody>
<tr>
<td>14310, 115</td>
<td>Pigeonite</td>
<td>9.67</td>
<td>8.90</td>
<td>5.23</td>
<td>108°40'</td>
<td>6, 34, 31</td>
</tr>
<tr>
<td></td>
<td>(001) Augite</td>
<td>9.69</td>
<td>8.96</td>
<td>5.26</td>
<td>106°10'</td>
<td>42, 42, 16</td>
</tr>
<tr>
<td>14053, 44</td>
<td>Pigeonite</td>
<td>9.74</td>
<td>8.96</td>
<td>5.27</td>
<td>105°05'</td>
<td>40, 30, 30</td>
</tr>
<tr>
<td></td>
<td>(001) Augite</td>
<td>9.76</td>
<td>8.96</td>
<td>5.25</td>
<td>108°55'</td>
<td>2, 45, 53</td>
</tr>
<tr>
<td></td>
<td>Augite</td>
<td>9.78</td>
<td>8.99</td>
<td>5.27</td>
<td>106°00'</td>
<td>40, 20, 40</td>
</tr>
<tr>
<td></td>
<td>Pigeonite</td>
<td>9.71</td>
<td>8.94</td>
<td>5.22</td>
<td>108°50'</td>
<td>4, 52, 45</td>
</tr>
<tr>
<td></td>
<td>(001) Augite</td>
<td>9.71</td>
<td>8.94</td>
<td>5.26</td>
<td>106°00'</td>
<td>45, 34, 21</td>
</tr>
<tr>
<td>14162, 34</td>
<td>Augite</td>
<td>9.76</td>
<td>8.88</td>
<td>5.30</td>
<td>106°45'</td>
<td>35, 52, 13</td>
</tr>
<tr>
<td>12053, 72</td>
<td>Pigeonite</td>
<td>9.67</td>
<td>8.90</td>
<td>5.23</td>
<td>108°25'</td>
<td>10, 63, 27</td>
</tr>
<tr>
<td></td>
<td>Augite</td>
<td>9.74</td>
<td>8.88</td>
<td>5.27</td>
<td>106°40'</td>
<td>38, 51, 11</td>
</tr>
<tr>
<td></td>
<td>(001) Pigeonite</td>
<td>9.74</td>
<td>8.88</td>
<td>5.25</td>
<td>108°15'</td>
<td>10, 70, 20</td>
</tr>
<tr>
<td>12038, 72</td>
<td>Augite</td>
<td>9.71</td>
<td>8.90</td>
<td>5.31</td>
<td>106°35'</td>
<td>38, 45, 18</td>
</tr>
<tr>
<td></td>
<td>(001) Pigeonite</td>
<td>9.71</td>
<td>8.90</td>
<td>5.23</td>
<td>108°50'</td>
<td>2, 66, 32</td>
</tr>
<tr>
<td>12040, 24</td>
<td>Pigeonite</td>
<td>9.67</td>
<td>8.88</td>
<td>5.21</td>
<td>106°30'</td>
<td>7, 71, 22</td>
</tr>
<tr>
<td></td>
<td>(001) Augite</td>
<td>9.74</td>
<td>8.88</td>
<td>5.26</td>
<td>106°15'</td>
<td>34, 53, 13</td>
</tr>
<tr>
<td></td>
<td>Augite</td>
<td>9.68</td>
<td>8.89</td>
<td>5.25</td>
<td>105°25'</td>
<td>40, 46, 14</td>
</tr>
<tr>
<td></td>
<td>(001) Pigeonite</td>
<td>9.68</td>
<td>8.89</td>
<td>5.20</td>
<td>108°50'</td>
<td>4, 68, 28</td>
</tr>
</tbody>
</table>
Table 2. Cation distribution in lunar pigeonites and augites determined by $^{57}$Fe Mössbauer resonance spectroscopy

**a. Pigeonites**

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Composition</th>
<th>Site Occupancy</th>
<th>Fe</th>
<th>Mg</th>
<th>Ca</th>
<th>Fe</th>
<th>Mg</th>
<th>K_D</th>
</tr>
</thead>
<tbody>
<tr>
<td>14310, 115†</td>
<td>(Ca$<em>{18}$Fe$</em>{58}$Mg$_{12}$)$Si_2$_O$_6$</td>
<td></td>
<td>.096</td>
<td>.904</td>
<td>.181</td>
<td>.435</td>
<td>.384</td>
<td>.094</td>
</tr>
<tr>
<td>14053, 44</td>
<td>(Ca$<em>{22}$Fe$</em>{58}$Mg$_{12}$)$Si_2$_O$_6$</td>
<td></td>
<td>.142</td>
<td>.858</td>
<td>.224</td>
<td>.439</td>
<td>.337</td>
<td>.127</td>
</tr>
<tr>
<td>12053, 72</td>
<td>(Ca$<em>{21}$Fe$</em>{58}$Mg$_{12}$)$Si_2$_O$_6$</td>
<td></td>
<td>.112</td>
<td>.888</td>
<td>.208</td>
<td>.471</td>
<td>.321</td>
<td>.086</td>
</tr>
<tr>
<td>12038, 72</td>
<td>(Ca$<em>{22}$Fe$</em>{58}$Mg$_{12}$)$Si_2$_O$_6$</td>
<td></td>
<td>.159</td>
<td>.841</td>
<td>.263</td>
<td>.486</td>
<td>.251</td>
<td>.030</td>
</tr>
<tr>
<td>12021, 21</td>
<td>(Ca$<em>{22}$Fe$</em>{58}$Mg$_{12}$)$Si_2$_O$_6$</td>
<td></td>
<td>.112</td>
<td>.882</td>
<td>.216</td>
<td>.456</td>
<td>.328</td>
<td>.097</td>
</tr>
</tbody>
</table>

**b. Augites**

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Composition</th>
<th>Site Occupancy</th>
<th>Fe</th>
<th>Mg</th>
<th>Ca</th>
<th>Fe</th>
<th>Mg</th>
<th>K_D</th>
</tr>
</thead>
<tbody>
<tr>
<td>14053, 44</td>
<td>(Ca$<em>{64}$Fe$</em>{52}$Mg$_{32}$)$Si_2$_O$_6$</td>
<td></td>
<td>.253</td>
<td>.747</td>
<td>.638</td>
<td>.314</td>
<td>.048</td>
<td>.052</td>
</tr>
<tr>
<td>12053, 72</td>
<td>(Ca$<em>{59}$Fe$</em>{52}$Mg$_{32}$)$Si_2$_O$_6$</td>
<td></td>
<td>.140</td>
<td>.860</td>
<td>.577</td>
<td>.366</td>
<td>.057</td>
<td>.014</td>
</tr>
<tr>
<td>12021, 21</td>
<td>(Ca$<em>{45}$Fe$</em>{55}$Mg$_{32}$)$Si_2$_O$_6$</td>
<td></td>
<td>.342</td>
<td>.658</td>
<td>.450</td>
<td>.519</td>
<td>.031</td>
<td>.031</td>
</tr>
</tbody>
</table>

*+ determined by microprobe analysis on more than 20 grains and averaged.*

* K_D = X_M1^Fe\cdot (1 - X_M2^Fe)/X_M2^Fe\cdot (1 - X_M1^Fe), where X_M = Fe/(Fe + Mg).*

† Mixed pigeonite and orthopyroxene phases.
VOLATILE ELEMENT DEPLETION INVESTIGATIONS ON APOLLO 11 AND 12 LUNAR BASALTS VIA THERMAL VOLATILIZATION. Everett K. Gibson, Jr. and Norman J. Hubbard, NASA Manned Spacecraft Center, Houston, TX 77058

Thermal volatilization studies on two lunar basalts (10017 and 12022) have been carried out in order to measure the depletion of selected elemental abundances during heating under vacuum. Samples of lunar and terrestrial basalts were heated in a step-wise fashion to temperatures of 1400°C and the sample residues analyzed via stable isotope dilution analysis and atomic absorption spectroscopy methods for changes in the elemental abundances of Na, K, Rb, Li, Ba and selected rare earth elements. The data were used to evaluate previous hypotheses about volatile element depletion in lunar samples and the approximate boundary conditions within which selected volatilization may have produced observed elemental abundance patterns in the lunar samples.

The measured depletions for the elemental abundances resulting from thermal volatilization experiments on lunar basalts are listed in Table 1. Lunar basalts and terrestrial basalts were heated for two hours under a vacuum of 10^-6 torr during the volatilization experiments. The measured elemental depletions are consistent with Brewer's (1953) and DeMaria et al. (1971) data for the volatilization of elemental oxides and for lunar basalt 12022. The greater depletion in K and Rb concentration relative to Na concentration found by us for 12022 basalt reflects the higher vaporization temperature of Na2O (~1050°C) as compared to K2O (~877°C) and Rb2O (~827°C). The initial K

<table>
<thead>
<tr>
<th></th>
<th>10017</th>
<th>12022</th>
</tr>
</thead>
<tbody>
<tr>
<td>ppm</td>
<td>Initial</td>
<td>After</td>
</tr>
<tr>
<td>K</td>
<td>2610</td>
<td>641</td>
</tr>
<tr>
<td>Rb</td>
<td>5.63</td>
<td>1.05</td>
</tr>
<tr>
<td>Na</td>
<td>3800</td>
<td>500</td>
</tr>
<tr>
<td>Li</td>
<td>18.1</td>
<td>19.3</td>
</tr>
<tr>
<td>Ba</td>
<td>----</td>
<td>59.5</td>
</tr>
<tr>
<td>Ce</td>
<td>17.4</td>
<td>----</td>
</tr>
<tr>
<td>Nd</td>
<td>14.8</td>
<td>----</td>
</tr>
<tr>
<td>Sm</td>
<td>5.59</td>
<td>----</td>
</tr>
<tr>
<td>Eu</td>
<td>1.28</td>
<td>----</td>
</tr>
<tr>
<td>Gd</td>
<td>7.90</td>
<td>----</td>
</tr>
<tr>
<td>K/Rb</td>
<td>464</td>
<td>610</td>
</tr>
<tr>
<td>Na/K</td>
<td>1.46</td>
<td>0.78</td>
</tr>
</tbody>
</table>
abundance for 12022 was reduced from 536 ppm to 449 ppm at 950°C and to 360 ppm at 1050°C. The depletion of Na did not begin to occur until above 1050°C, in agreement with the data of Brewer (1953). The large depletion of Rb in 12022 during the heating at 950°C and 1050°C also reflects the lower vaporiza-
tion temperature of rubidium oxide.

Potassium and Rb do not behave identically during volatilization. Rubidium was depleted to 30% of the initial concentration after thermal volatilization at 1050°C, whereas K was depleted to 67% of the initial amount. The K/Rb ratio of 12022 changed from 723 in the unheated sample to 932, 1259, 1041 and 1035 at 950°C, 1050°C, 1200°C and 1400°C respectively, further depicting differences between the volatilization behavior of K and Rb.

Refractory elements such as Li, Ba and the rare earth elements were not depleted during the heating, reflecting their high vaporization temperatures. Significantly, the size of the europium anomaly was unchanged under the conditions of these experiments. The data reported here and in DeMaria et al. and Naughton et al. (1971) restrict depletion by thermal volatilization from lunar lava flows to the alkali elements and Pb (Silver, 1971) and other elements with low vaporization temperatures. These data also demonstrate that alkalis can be lost from regolith material that attains a temperature as low as 550°C and ambient vapor pressures of 10⁻¹ torr.

Thermal volatilization may be important in the generation of some lunar glasses, although appreciable loss is probably limited to those glasses heated well above liquidus temperatures. Loss of Pb by thermal volatilization from lunar samples has been demonstrated by Silver (1971) and for some Apollo 12 soils rich in KREEP glass (Tatsumoto, 1971). By extension other equally volatile elements (C, S, Bi etc.) have been lost during formation of lunar glasses.

The effects of lunar gravity, lithostatic pressure and the vapor pressures of the species of interest must all be taken into consideration when considering volatilization processes within a lunar basalt flow. We find that any volatilization process which could occur would be only a "skin-effect" depletion of the volatile element unless stirring occurred. The lithostatic pressure below 0.1 cm in the lunar basalt exceeds the vapor pressure of all volatile species found in the lunar basalts. For any large scale volatile loss to occur some means of mixing or exposing new material for vaporization must be found. Such a mechanism could involve gas bubbling within the lava, convection, splattering or fountaining. With the release of the volatile phase from the lava the pressure above the lava flow is increased, partially offsetting any gain in the overall vaporization rate.

Alkalis and other volatile elements such as lead can be lost from a hot, but unmelted and even unsintered regolith if the ambient vapor pressure is lower than the partial vapor pressure of any volatile element or compound and there is mass transfer of volatilized material to regions of lower ambient vapor pressure. That is, volatilization of material from a uniformly hot regolith requires that the regolith have extensive interconnecting pore spaces, i.e. not be well compacted. Volatilization from a nonuniformly hot
Volatile Element Depletion Investigations...

Everett K. Gibson, Jr.

Regolith still requires extensive interconnecting pore spaces. This restricts volatilization losses from regoliths to relatively shallow depths where sufficient temperatures are less likely.

Basalts from Apollo 11 and 12 show a range in alkali contents; similar alkali abundances were produced in our experiments. However, Apollo 11 and 12 basalts also show a wide range in refractory elements such as rare earth elements, strontium, and barium. Variations in refractory and alkali elements are consistent with separate sources and differing degrees of partial melting rather than with alteration by selective volatilization of a once homogeneous liquid, although some variation in alkali concentrations may be due to volatilization loss. The low Na content of lunar regoliths and lunar basalts and variations in K and Rb concentrations are not post-eruption volatilization features in the majority, if not all, cases.

It is currently considered by some authors (Hubbard and Gast, 1971; Urey, 1971) that the outer portion of the moon was partially or completely molten very early in its history (before 4.0 b.y.). This may be the event that generated the low volatile element concentrations characteristic of lunar basalts, and thus their source regions. Volatile element loss at this period in lunar evolution may have been very efficient due to stirring by intense bombardment. If this mechanism was operative and resulted in alkali loss from the source areas for both mare and non-mare basalts (Hubbard and Gast, 1971), this event must be followed by a large scale differentiation of the upper portions of the moon into Al rich and Fe poor and Al poor and Fe rich zones.

References

Trace element concentrations in lunar samples have for the most part hitherto been determined for whole rock or average soil samples. Our present technique allows us to determine with moderate accuracy a very wide range of trace element contents in an individual mineral grain or glassy fragment obtained from lunar rock, breccia or soil.

Some glasses may show trace element contents related closely to those of mineral grains from which they were derived by direct melting, some may correspond primary glasses, and some may be of hybrid character through mixing while in the molten state. Effects of selective volatilisation and its variation with degree of impact melting can be looked for, and contrasts might be expected between fused glasses, condensation glasses and diaplectic glasses.

A study of the above possibilities, many of which interact, is being undertaken with the Apollo 12 and 14 samples made available to us, but only very few specimens have so far been examined.

**Method of analysis**

Mineral grains (< 0.01 mg) from rocks and glass particles hand-picked from soils have been analysed using spark source mass spectrography (Nicholls et al., 1967) with the modification that tipped electrodes were used instead of bar electrodes, and no internal standard was used. Individual grains were made the pointed tips of compressed silver rod electrodes and the spark was struck between a sample electrode and a counter-pointed silver rod electrode. To obtain maximum sensitivity exposures were continued until the sample had been completely consumed. Sparking parameters used previously on deep sea basalts were found to be suitable for lunar material.
Mass Spectrometric Analysis...  
H. N. Giles

Results

So far only two minerals, olivine and pyroxene, have been examined from Apollo 12 igneous rocks and five glass particles from Apollo 14 soils. In the five glasses the contents of B, F, Ni, Cu, Ga, Br, Sr, Zr and Ba generally exceed their contents in the two minerals. Ga, Sr and Ba may have higher concentrations in plagioclase, so far not examined, and Ni might derive from nickel-iron or ilmenite. Cu and Zr might derive from other minerals but B, F and Br require some other explanation.

Of the two fragments from soil sample 14257, 11, (both dark grey) one was highly vesicular while the other was virtually free from gas bubbles. The former is, as expected, depleted in B, F and Br relative to the latter, but it is also depleted in Sc, Ba, HF and Bf one of which would be expected to be lost readily on volatilization. Ga shows a particularly high depletion. Some elements are enriched by factors of two or more in the highly vesicular glass, (e.g. Co, Sr and Y). The relation between these two fragments is thus not a simple one of differing degree of melting, though this process may well have contributed.

Similar difficulties are encountered in comparing honey colored vesicular and non-vesicular particles from sample 14258, 37. Elements depleted in the vesicular particle are not only B, F and Br, but Zn, Ga, Ge, Se, Zr, Te, HF, Th and U. Of these neither Zr nor HF would be expected to be lost in a volatile phase. Elements enriched in the vesicular particle include Sc, Co, Ni, Cu, Sr and Ba. The relative enrichment (or depletions) of Rb and Cs, and of Sr and Ba, are contrary to those expected for a simple hypothesis of differential loss of volatiles.

A yellow-green non-vesicular glass from the same soil sample 14, 258, 37 shows more affinities with the non-vesicular honey colored glass, but also significant differences (e.g., Ni, Cu, Ga, Sr and Ba). It is not immediately apparent how such differences could be related to mineralogical variation in the parent material modified by subsequent differential loss of volatiles.
It is apparent that the results so far raise more problems than they solve. Clarification may emerge when analyses are acquired from a greater variety of mineral grains and perhaps from some assemblages of glasses and minerals which are physically more closely related. It is felt that the technique is one which can be usefully applied to study various problems of lunar particle origin and lunar surface processes.

BULK AND RARE EARTH ABUNDANCES IN THE LUNA 16 SOIL LEVELS A AND D

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Abundances of the major, minor and trace elements (SiO₂, O, TiO₂, Al₂O₃, FeO, CaO, Na₂O, K₂O, Cr₂O₃, MnO, Sc, V, Co, La, Sm, Eu, Yb and Lu) have been determined via sequential INAA (instrumental neutron activation analysis) in two Luna 16 soils (< 0.125 mm fines and 17 mg each) from the A level (~ 7 cm) and D level (~ 30 cm depth). Aliquants for these soils were also analyzed by Laul et al., the University of Chicago group, for volatile and other trace elements. Abundances of the bulk elements in Luna 16 soils generally agree with the values reported by Vinogradov. Elemental abundances of both bulk and trace elements are nearly the same for the two A and D soil levels. The overall chemical composition of Luna 16 soil (Sea of Fertility) is more similar to Apollo 11 soil (Sea of Tranquillity) than to soils from Apollo 12 (Oceanus Procellarum) and Apollo 14 (Fra Mauro). However, differences are observed between Luna 16 and Apollo 11 soil, e.g., TiO₂ in Luna 16 is ~ 50% of the TiO₂ in Apollo 11, and ~ 30% more alkali and REE (rare earth elements) abundances and ~ 20% lower Cr₂O₃ are found in Apollo 11 soil. Although the TiO₂ abundances are essentially equally abundant in Luna 16 soil and the average Ap. 12 soil within ± 15%, lower SiO₂ and Sc and higher MgO, Na₂O, K₂O, V and REE have been observed in the average Ap. 12 soil, relative to Luna 16 soil. Of all the K₂O values reported in various Ap. 12 soils, only the K₂O in 12032-40 (see Table 1) most closely approximates the K₂O in Luna 16 soil. Most other K₂O abundances in Ap. 12 soils are 2-5 times higher.

Consistent with findings for Apollo 11 and 12 rocks and soil, the Luna 16 fines are depleted in oxygen with respect to terrestrial igneous rocks of comparable Si content. The low O/Si ratio is certainly in part due to the presence of ilmenite (31.6% O) and reduced species, such as Ti (III).

The chondritic normalized REE distribution patterns of Luna 16 soils are similar to that of Apollo 11 soil and these in turn are quite different from the light REE enriched patterns of Apollo 12 and 14 soils. The REE abundances and the Sm/Eu ratio in Luna 16 soils are significantly different from the higher values of Apollo 11, 12 and 14 soils. Abundances of Sm, which are approximately proportional to the total REE abundances, at 9.0 ppm in Luna 16 soils are compared to 12.6, 16-31 and 31 ppm in Apollo 11, 12 and 14 soils, respectively. A Sm/Eu ratio of 3.9 observed in Luna 16 soils may be compared...
BULK AND RARE EARTH
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to 7.4 ± 0.4, 9.7 ± 0.3 and 11.6 ± 0.2 in Apollo 11, 12 and 14 soils, respectively. Relative to the Sm/Eu ratio of 2.7 in chondritic meteorites, Luna 16 soil has the lowest Eu depletion of all lunar soils measured to date.

Abundances of the siderophilic element Co are rather uniform in all lunar soils and vary from 31-40 ppm. Scandium abundances range from 22 to 59 ppm in lunar soils and Sc appears to correlate directly with FeO and inversely with the REE abundances. MnO and FeO correlate directly in lunar soils.

Acknowledgements: This study was supported by NASA grants NGL 18-001-058, 38-002-039 and 38-002-020.
Table 1  Elemental abundances in Luna 16 soil samples

<table>
<thead>
<tr>
<th>Element</th>
<th>A-level (17.1 mg)</th>
<th>A-level Vinogradov</th>
<th>D1 level (17.1 mg)</th>
<th>D1 level Vinogradov</th>
<th>Apollo 11 10084 soil</th>
<th>Apollo 12 12037-40 soil</th>
<th>Apollo 14 Average of 3 soils</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO2(%)</td>
<td>43.9</td>
<td>41.7</td>
<td>47.5</td>
<td>41.3</td>
<td>43.2</td>
<td>41.8</td>
<td>---</td>
</tr>
<tr>
<td>O(%)</td>
<td>40.2</td>
<td>---</td>
<td>43.0</td>
<td>---</td>
<td>40.8</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td>TiO2(%)</td>
<td>3.3</td>
<td>3.39</td>
<td>3.2</td>
<td>3.42</td>
<td>7.5</td>
<td>3.3</td>
<td>1.9</td>
</tr>
<tr>
<td>Al2O3(%)</td>
<td>15.32</td>
<td>15.32</td>
<td>16.32</td>
<td>16.15</td>
<td>13.8</td>
<td>13.1</td>
<td>18.8</td>
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<tr>
<td>FeO(%)</td>
<td>16.8</td>
<td>16.80</td>
<td>16.90</td>
<td>16.90</td>
<td>16.3</td>
<td>16.3</td>
<td>10.8</td>
</tr>
<tr>
<td>MgO(%)</td>
<td>11</td>
<td>12.20</td>
<td>11.55</td>
<td>11.9</td>
<td>11.1</td>
<td>11</td>
<td></td>
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<tr>
<td>CaO(%)</td>
<td>0.343</td>
<td>0.37</td>
<td>0.370</td>
<td>0.28</td>
<td>0.445</td>
<td>0.458</td>
<td>0.71</td>
</tr>
<tr>
<td>Na2O(%)</td>
<td>0.12</td>
<td>0.10</td>
<td>0.11</td>
<td>0.11</td>
<td>0.134</td>
<td>0.102</td>
<td>0.57</td>
</tr>
<tr>
<td>Cr2O3(%)</td>
<td>0.234</td>
<td>0.31</td>
<td>0.330</td>
<td>0.26</td>
<td>0.260</td>
<td>0.352</td>
<td>0.207</td>
</tr>
<tr>
<td>MnO(%)</td>
<td>0.212</td>
<td>0.21</td>
<td>0.203</td>
<td>0.22</td>
<td>0.207</td>
<td>0.217</td>
<td>0.127</td>
</tr>
<tr>
<td>Sc(ppm)</td>
<td>54</td>
<td>27</td>
<td>55</td>
<td>25</td>
<td>59</td>
<td>40</td>
<td>22</td>
</tr>
<tr>
<td>V(ppm)</td>
<td>76</td>
<td>64</td>
<td>83</td>
<td>55</td>
<td>64</td>
<td>130</td>
<td>53</td>
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<tr>
<td>Co(ppm)</td>
<td>37</td>
<td>66</td>
<td>34</td>
<td>61</td>
<td>31</td>
<td>40</td>
<td>38</td>
</tr>
<tr>
<td>La(ppm)</td>
<td>12.0</td>
<td>7.3</td>
<td>12.4</td>
<td>7.2</td>
<td>15</td>
<td>38</td>
<td>68</td>
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<tr>
<td>Sm(ppm)</td>
<td>8.8</td>
<td>5.6</td>
<td>9.2</td>
<td>6.8</td>
<td>12.6</td>
<td>18.3</td>
<td>30.7</td>
</tr>
<tr>
<td>Eu(ppm)</td>
<td>2.3</td>
<td>1.6</td>
<td>2.3</td>
<td>1.4</td>
<td>1.8</td>
<td>1.7</td>
<td>2.66</td>
</tr>
<tr>
<td>Yb(ppm)</td>
<td>6.3</td>
<td>3.5</td>
<td>6.7</td>
<td>3.5</td>
<td>11.0</td>
<td>14.5</td>
<td>23</td>
</tr>
<tr>
<td>Lu(ppm)</td>
<td>0.96</td>
<td>0.28</td>
<td>0.98</td>
<td>0.3</td>
<td>1.6</td>
<td>1.96</td>
<td>3.3</td>
</tr>
</tbody>
</table>

(a) Si, O and Al were determined by 14 MeV neutron activation at the University of Kentucky; all other elements, and also Al, were determined by thermal neutron activation at Oregon State University. For convenience, all common major and minor elements, except O, are expressed as oxides. Estimated errors are: Si and O ±1%; Na, Cr and Mn ±3%; Ti, Fe, Sc, Co, La, Sm, Yb and Lu, ±5%; Ca, K and Eu ±8%; V, ±12%.

(b) Thermal neutron activation result with an error of about ±2%.

(c) 14 MeV neutron activation result with an error of about ±12%.

(d) Vinogradov [1].

(e) For Si and O, values were taken from Ehmann and Morgan [7]; abundances for Ti, Al, Fe, Mg, Ca and Na were the average values obtained by seven different groups cited in their Table 4 by Wakita et al. [15]; abundances of Cr and Mn were obtained by Wakita et al. [15]; K and Ba by Tera et al. [22] and other trace elements by Wakita et al. [15].

(f) Wakita et al. [10, 16]. Abundances of Al2O3, CaO, Na2O, MnO, V and the REE are the averages for the two 0.22 g samples that were analyzed.

(g) Wakita et al. [19].
APOLLO 14 GLASSES; B.P. Glass, Dept. of Geology, University of Del., Newark, Del. 19711.

The 0.149 to 1.000 mm size fraction of samples 14148; 14149; 14156; 14163; 14230.75 and 14230.82 contain from 13 to 27% glass fragments and droplets. In general, the finer the soil the higher the percentage of glass that it contains. Glass spherules (spheres, dumbbells, etc.) make up approximately one percent of the samples. Scanning electron microscope studies show that the glass particles have surface features similar to those observed on Apollo 11 and 12 glasses (i.e. splash silicate glass, metallic beads and impact pits).

The chemical compositions of 214 glass particles (149 fragments and 65 spherules) from samples 14163, 14148, 14156 and 14230,82 have been determined by electron microprobe analysis. The glasses exhibit a wide range in compositions; however, histograms of oxide abundances and various oxide plots show that the analyses tend to cluster into several groups. One hundred sixty-three of the glass particles appear to be homogeneous. Analyses of these particles (103 fragments and 55 spherules) were used to divide the glasses into six groups (Table 1).

GROUP 1. The four glass particles (two fragments and two spherules) in this group are monomineralic with anorthite compositions. The particles are transparent colorless to pale yellow-green. Similar glasses have been reported in Apollo 11 and 12 samples.

GROUP 2. The glasses in this group are characterized by their high Al₂O₃ (>22%) and CaO (>14%) contents. This group has been divided into two subgroups based on TiO₂ content. The glasses in group 2A have TiO₂ contents less than 0.5% and the glasses in group 2B have TiO₂ contents greater than 0.9%. The glasses in group 2A are mostly transparent pale green fragments or translucent ropy fragments covered with rock flour. Group 2B glasses are colorless to yellow-green fragments. Similar glasses are a minor but conspicuous component of Apollo 11 and 12 fines (1, 2). These glasses have compositions equivalent to feldspathic basalt or anorthositic gabbro and may be derived from highland areas (3, 4).

GROUP 3. The glasses in this group are distinguished by their high FeO contents (>15%). They have been divided into three subgroups based on their Ti and Mg contents. The glasses in group 3A have high TiO₂ (>3.5%) contents. The glasses in this group are generally red spherules. The glasses in group 3B have TiO₂ contents between 2.7 and 3.5% and low MgO contents
Apollo 14 Glasses

B.P. Glass

**GROUP 3.** These glasses are generally yellow-brown to red-brown to red fragments. The **group 3** glasses have low TiO$_2$ (<1.5%) and high MgO (>14%) contents. These glasses are generally transparent yellow-green fragments without crystalline inclusions or bubble cavities. The glasses in group 3 are similar in composition to the Apollo 12 and Luna 16 fines and glasses. Thus these glasses are probably derived from a mare area. It has previously been estimated that mare-derived glasses make up 11.1% of sample 14259 (5).

**GROUP 4.** Nearly half of the analyzed glasses fall in this group. These glasses are characterized by their low FeO (<14%) and high K$_2$O (>0.4%) contents. The glasses in this group have been divided into three subgroups based on their MgO contents (Table 1). These three subgroups are similar to types B, C and D as defined by the Apollo Soil Survey (5). According to the Apollo Soil Survey glasses of this composition make up about 58% of the glass particles analyzed from sample 14259,26. The group 4 glasses (especially group 4B) are similar in composition to the soil from the Apollo 14 site (Table 1). The major difference is the higher alkali content of the glasses. Glasses with similar compositions are common in Apollo 12 fines (2, 3). Material of this composition in Apollo 12 and 14 samples has been variously called KREEP, norite, "gray mottled" fragments and Fra Mauro basalt (5).

**GROUP 5.** These glasses are distinguished from the group 1 and 2 glasses by their low Al$_2$O$_3$ contents (<21%), from group 3 glasses by their low FeO contents (<11%) and from group 4 glasses by their low K$_2$O contents (<0.4%). The glasses in this group are generally yellow, yellow-brown or brown spherules. These glasses are similar in composition to soil samples taken at the base of Hadley Delta during the Apollo 15 mission.

**GROUP 6.** Only one glass particle is placed in this group. It is distinguished by its high SiO$_2$ (73.9%) and K$_2$O (4.9%) contents. According to the Apollo Soil Survey (5) glasses of this composition make up about 1.6% of the glasses in sample 14259,26. They point out that the composition is similar to potash granite.

All but about 2% of the homogeneous glasses analyzed, and nearly all of the heterogeneous glasses can be assigned to one of the above groups.

Almost half of the analyzed glasses have compositions similar to the rocks and soil from the Apollo 14 site. Most of the rest are highly feldspathic or have compositions similar to mare-derived material.

**References.**


### Apollo 14 Glasses

**B.P. Glass**

### Table 1. Modal compositions of glass types in Apollo 14 soils.

<table>
<thead>
<tr>
<th></th>
<th>Group 1 (Anorthite)</th>
<th>Group 2 (Feldspathic)</th>
<th>Group 3 (Mare-Derived)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(4)</td>
<td>A(36)</td>
<td>B(3)</td>
</tr>
<tr>
<td>SiO$_2$</td>
<td>46.</td>
<td>45.</td>
<td>46.</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.1</td>
<td>0.4</td>
<td>1.1</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>35.5</td>
<td>25.</td>
<td>24.</td>
</tr>
<tr>
<td>FeO</td>
<td>0.3</td>
<td>6.0</td>
<td>5.</td>
</tr>
<tr>
<td>MnO</td>
<td>0.01</td>
<td>0.07</td>
<td>0.07</td>
</tr>
<tr>
<td>MgO</td>
<td>0.09</td>
<td>8.6</td>
<td>6.5</td>
</tr>
<tr>
<td>CaO</td>
<td>18.5</td>
<td>15.</td>
<td>14.8</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>1.3</td>
<td>0.27</td>
<td>0.3</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>0.15</td>
<td>0.11</td>
<td>0.15</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>Group 4 (Fra Mauro Basalts)</th>
<th>Group 5</th>
<th>Group 6 Fines</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO$_2$</td>
<td>48.</td>
<td>49.</td>
<td>52.</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>2.0</td>
<td>2.2</td>
<td>1.8</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>16.6</td>
<td>17.</td>
<td>17.</td>
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<tr>
<td>FeO</td>
<td>10.9</td>
<td>10.2</td>
<td>9.8</td>
</tr>
<tr>
<td>MnO</td>
<td>0.12</td>
<td>0.11</td>
<td>0.12</td>
</tr>
<tr>
<td>MgO</td>
<td>9.9</td>
<td>7.6</td>
<td>5.0</td>
</tr>
<tr>
<td>CaO</td>
<td>10.4</td>
<td>10.6</td>
<td>10.7</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.95</td>
<td>1.3</td>
<td>1.4</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>0.7</td>
<td>0.9</td>
<td>1.0</td>
</tr>
</tbody>
</table>

(\()\)Refers to number of analyses

* LSPET (1971)
NITROGEN ABUNDANCES IN LUNAR SAMPLES BY NEUTRON ACTIVATION ANALYSIS, P.S. Goel and B.K. Kothari, Indian Institute of Technology, Kanpur, India.

Total nitrogen concentration in several rocks and fines has been measured by neutron activation analysis using the reaction $^{14}\text{N}(n,p)^{14}\text{C}$. The method, originally developed for studying iron meteorites(1) was improved and modified for silicate materials.

Lunar samples (0.2 to 2 g) along with NBS standard steel (33d and 12h) as monitors, were sealed under vacuum and irradiated in a reactor. Fragments of broken irradiated rocks were picked up for a run. The powdered sample, mixed with a carrier, was fused with oxidizing flux in a stream of purified oxygen at about 1100°C. The outgoing gases were bubbled through a set of purifying traps and the radiocarbon dioxide was absorbed in NaOH. It was precipitated as BaCO$_3$ which was purified and counted in solid form against a thin wall low background Geiger counter. The NBS monitors were also treated in a similar way. Appropriate blanks and control experiments were done. The data, given in Table 1, are assumed to have a conservative 10% error.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Description</th>
<th>Mass, mg</th>
<th>Nitrogen, ppm</th>
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<tbody>
<tr>
<td>14163,117</td>
<td>Fines</td>
<td>74</td>
<td>78</td>
</tr>
<tr>
<td></td>
<td></td>
<td>61</td>
<td>80</td>
</tr>
<tr>
<td></td>
<td></td>
<td>19.2</td>
<td>82</td>
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<td></td>
<td></td>
<td>16.3</td>
<td>90</td>
</tr>
<tr>
<td></td>
<td></td>
<td>15.5</td>
<td>72</td>
</tr>
<tr>
<td>14049,34</td>
<td>Fragmental</td>
<td>109</td>
<td>71</td>
</tr>
<tr>
<td>(Soil Clod)</td>
<td></td>
<td>106</td>
<td>70</td>
</tr>
<tr>
<td>14305,78</td>
<td>Fragmental</td>
<td>103.5 (one piece)</td>
<td>20</td>
</tr>
<tr>
<td></td>
<td></td>
<td>73 (white clast)</td>
<td>26</td>
</tr>
<tr>
<td></td>
<td></td>
<td>41 (five pieces)</td>
<td>20</td>
</tr>
<tr>
<td>14321,224</td>
<td>Fragmental</td>
<td>110 (dark clasts)</td>
<td>23</td>
</tr>
<tr>
<td></td>
<td></td>
<td>70</td>
<td>25</td>
</tr>
<tr>
<td>14310,121</td>
<td>Basaltic</td>
<td>105</td>
<td>22</td>
</tr>
<tr>
<td></td>
<td></td>
<td>126</td>
<td>21</td>
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</table>

Uncertainty, ±10%
There exist several measurements of nitrogen in lunar materials (2,3), some undoubtedly suffering from atmospheric contamination. Samples of powdered meteorites irradiated and analysed by our technique did not show any significant nitrogen contamination. Excess nitrogen (60 ppm) in lunar fines obtained by us is therefore not due to laboratory contamination or atmospheric adsorption. Moore et al. (2) have suggested, from data on previous missions, that lunar fines contain nitrogen due to solar wind implantation. Present data support this. This conclusion is further strengthened by the parallel behaviour of carbon (4) and nitrogen results in the analysed samples and the closeness of the excess C/N ratio to the solar ratio (5).

Measurements on irradiated and sieved fines are reported in Table 2. The excess nitrogen shows a d^{-1} dependence on mean grain diameter 'd' except for the smallest sieve fraction (<38 μm) which is significantly lower as compared to the expected d^{-1} value of 180 ppm. This could happen if particles of the smaller size were exposed to solar wind for a comparatively shorter period. Perhaps during thousands of years of direct exposure of fines on the lunar surface some sorting takes place burying the smaller grains below the larger ones.

Table 2. Grain size dependence of nitrogen in fines, 14163

<table>
<thead>
<tr>
<th>Grain size (μm)</th>
<th>Mass recovered (mg)</th>
<th>Mass analysed (mg)</th>
<th>Nitrogen (ppm)</th>
<th>Mean (ppm)</th>
<th>Excess (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt;38</td>
<td>112.2</td>
<td>63.7</td>
<td>101</td>
<td>95</td>
<td>73±10</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>58.5</td>
<td>88</td>
<td></td>
</tr>
<tr>
<td>38 - 75</td>
<td>201.2</td>
<td>96.5</td>
<td>92</td>
<td>82</td>
<td>60±9</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>114.7</td>
<td>72</td>
<td></td>
</tr>
<tr>
<td>75 - 151</td>
<td>96.3</td>
<td>28.5</td>
<td>47</td>
<td>48</td>
<td>26±5</td>
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<td></td>
<td></td>
<td></td>
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<td>50*,47*</td>
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<tr>
<td>151 - 270</td>
<td>93.5</td>
<td>93.5</td>
<td>42</td>
<td>42</td>
<td>20±5</td>
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<td></td>
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<td></td>
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<td>44*,41*</td>
<td></td>
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<tr>
<td>&gt;270</td>
<td>62.1</td>
<td>62.1</td>
<td>20</td>
<td>25</td>
<td>5±3</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>27*,27*</td>
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</tr>
</tbody>
</table>

Sieved, 625.0 mg; Recovered 585.3 mg * Separate Irradiation
NITROGEN ABUNDANCES
P. S. Goel

Stepwise removal of nitrogen (converted to radiocarbon) from irradiated fines was attempted in a series of leaching experiments. The scheme followed for nitrogen removal and the results obtained at various steps are presented in Figure 1.

The results show that the major portion of nitrogen gets removed by acid attack. Some losses occur, perhaps due to transfer and handling. Similar data on meteoritic powders show that not much nitrogen is removed by acid dissolution.

References
GRAIN SIZE ANALYSIS, OPTICAL REFLECTIVITY MEASUREMENTS AND DETERMINATION OF HIGH FREQUENCY ELECTRICAL PROPERTIES FOR APOLLO 14 LUNAR SAMPLES. T. Gold, E. Bilson and M. Yerbury, Center for Radiophysics and Space Research, Space Sciences Bldg., Cornell Univ., Ithaca, N.Y. 14850

Grain Size Analysis. The particle size distribution of the Apollo 14 lunar fines has been determined by the method of measuring the sedimentation rate in a column of water. This method was described in the Apollo 12 report; it is very useful for the purpose of comparison.

Figure 1a compares the differential particle size distribution of the Apollo 14 bulk fines with that of two core samples taken from different depths. Among these samples there seems to be no significant variation in the grain size distribution. This result could either imply that the 14230 core sample is intrinsically homogeneous and that the core fines have the same origin (down to a depth of 20 cm) as the surface bulk fines collected nearby, or that this core tube sample was perhaps highly disturbed.

![Fig. 1a](image1.png)  

![Fig. 1b](image2.png)
Figure 1b compares the differential particle size distribution of 3 surface fine samples: one from each Apollo mission. From these curves it would appear that A14163 is finer grained than either the Apollo 11 or Apollo 12 fines.

Reflectivity Measurements, Surface Darkening by Simulated Solar Wind. It has been proposed (first by Gold in 1960) that the low albedo of the lunar soil is due to sputtering effect of the solar wind. In order to further investigate this possibility, lunar fines from different Apollo missions have been subjected to simulated solar wind in this laboratory. The albedo of these samples was determined before and after irradiation. As can be seen in Figure 2, the albedo of both the Apollo 11 bulk fines and the more reflectant Apollo 14 bulk fines decreased significantly after being subjected to approximately 1.5x10^5 years of solar wind. The albedo of a finely ground terrestrial olivine basalt sample decreased from 34.3% (at 10° phase angle and ε=0° viewing angle) to 22.8% after being irradiated with approximately 5 x 10^3 years of solar wind in the same experimental conditions.

Figure 2: Reflectivity of an Apollo 11 and an Apollo 14 powder sample before and after being subjected to simulated solar wind (2 keV, 0.16 mA/cm^2 proton beam, irradiation times: 13.5 hours and 14.75 hours respectively).

Electrical Properties. Two surface fines samples (A14163 and A14003) and one rock sample (14310.161) have been analyzed. The dielectric constant measurements are reported in Figure 3, the power absorption length in these samples is shown in Figure 4. (Some earlier Apollo 11 and 12 data are also shown on these figures.) The striking result is the very long absorption length, 28 wavelength, observed for the Apollo 14 rock sample, 14310.161 at 430 MHz compared to only 1.7 wavelength for the Apollo 12 rock sample, 12063.89. The density and dielectric constants of these two rock samples were both similar. The power absorption of rock 12063.89 is typical for other Apollo 11 and 12 rock samples as well.

The Apollo 14 dust samples are also less absorbent than the Apollo 11 and 12 samples. At a packing density of 1.6g/cm^3 the absorption length is approximately 34 wavelengths or 2.2 times greater than the absorption length in A10084 or 3.5 times
The low absorption observed adds new significance to lunar radar observations. Thus at a wavelength of 10 meters reflections from a depth of over 100 meters may well contribute very significantly to the radar echoes. If any soil exists that is the pulverized version of rock 14310 then a radar penetration of 70 wavelengths would be expected. Materials of such low absorption are not a common constituent of the crust of the earth.

Evidence from topography seen in Orbiter pictures, the Apollo seismic evidence, the long wave radar investigations, and the detailed survey of the mascons all give a clear indication that the moon has generally a deep deposit of fairly compact but finely divided material. While there is some layering apparent in some areas, there has been no indication of the widespread presence of subsurface rock, and such rock would indeed be incompatible with radar data at a depth of less than 50 meters, and with the seismic data at a depth less than several kilometers. Many steep sided craters or rills show very little general layering, and it therefore seems most likely that the dusty soil on the surface is similar to the medium that extends down to a depth of several kilometers.

Topographic evidence shows clearly that surface flows of powdered rock have taken place on a substantial scale and with rather unusual detailed properties. All steep slopes show characteristic patterns and erosion features associated with them. The evidence from Apollo core samples showing a delicate layered structure makes clear that surface flows rather than ballistic effects have dominated the deposition of the present mare surface.

The long wave radar evidence indicates that rocks, possibly similar to the ones distributed over the surface, exist also throughout the deep soil deposit. The density of such rocks is regionally very variable, and appears to be lowest in Mare Procellarum. This low density of internal rocks seems characteristic of mare surfaces that are not mascons.

The distribution of rocks in a deep layer of soil is also indicated by the fact that many old and eroded slopes show a higher concentration of rocks, as if the fines had been removed leaving the coarse material behind, and the same is true of ridges which sometimes show a concentration of rocks along their crest.

We have experimented with electrostatic processes that may be capable of transporting the lunar soil. Electron bom-
bordment with electrons in the range of 200 volts to 2 kilo-
vols has been found to be most effective through a variety of
interesting local electrostatic instabilities that are produced.
Very fast surface flows can be generated in circumstances of
electron bombardment that actually occur on the moon. The
strongest source of kilovolt electrons striking the moon is in
the magnetic tail of the earth and not in the free solar wind,
and this gives immediately a reason for the striking difference
between the near and the far side of the moon. The far side
evidently contains many basins that are unfilled, while on
the front all low-lying areas appear to be filled.
Metallic particles were separated magnetically and by hand from two 15 g samples (14003,18 and 14163,165) of the <1 mm Apollo 14 soil. Over 250 particles, containing more than 90% metal, were thus obtained, as well as a smaller number of particles that were predominantly silicate with minor metallic inclusions. The weight fraction of metal particles in the >354 µm fraction of 14163, for example, was 0.5 wt%. The yield of metal particles was over five times that obtained in similar studies from the Apollo 11 and 12 sites. This observation may be interpreted as indicating a larger amount of meteoritic impact-cratering activity around the Apollo 14 site and is consistent with the high percentage of breccia rocks found at the site.

All of the metallic particles were mounted in transparent epoxy and prepared for metallographic and microprobe examination. The vast majority of the metallic particles are associated with greater or less quantities of silicate phase and the external form of the composite particle is distinctly irregular. A small number of particles (13) were encountered with little or no silicate associations. In all these 13 particles the structure of the metal suggested a history of remelting. It is convenient to refer to these as remelted metal as distinct from ordinary metal-(silicate) particles.

The range of structures of ordinary metal-(silicate) particles, is very wide and extends from the occasional shocked silicate with minor metallic inclusions to the occasional free standing metal fragment entirely without silicate associations. However the majority of particles may be described either as II) silicate with included metal or as I) metal with associated peripheral silicates. In type II, the silicate assemblage may be crystalline rock or may be altered in structure. In the limit, the silicate may be glassy in appearance. In the highly altered silicate the structure and distribution of metal is quite similar to the vein material of shocked chondrites. However the sulfide phase which is so abundant in the chondrite veins is usually not found in the analogous lunar soil samples. In type I, the silicates may again take all morphological forms from undamaged crystalline to highly metamorphosed. In most of these cases the metal proved essentially chemically and structurally heterogeneous.
METALLIC PARTICLES IN THE APOLLO 14 LUNAR SOIL

J. I. Goldstein

Bulk chemical analyses for Ni, Co, P, S, and Fe for over 205 ordinary metal-(silicate) particles were measured. For type II, silicate with included metal, the chemical composition ranged from <0.02-11.2 wt% Ni, <0.020-0.75 wt% Co, ≤0.02 wt% S and <0.02-0.3 wt% P. For type I, metal with associated silicates, bulk chemical analyses for the chemically and structurally homogeneous samples ranged from <0.02-8.3 wt% Ni, ≤0.05-1.4 wt% Co, ≤0.02-0.05 wt% S, ≤0.02-1.5 wt% P. Bulk chemical analyses could not however be obtained for the small number of heterogeneous samples.

To determine whether the lunar metal is meteoritic, the Ni-Co criterion (Ni > 4.5 wt%, Co ~ 0.2-1.0 wt%) developed by Goldstein and Yakowitz was used. For type I, 65% of the included metal is of meteoritic compositions. All the remelted particles, and 70-75% of the type I metal-(silicate) particles are meteoritic in composition. These results are to be contrasted with the Apollo 12 results where the same relative amounts of meteoritic metal were found in the soil but where less than 20% of the type II metallic inclusions in the lunar soil particles were of meteoritic composition. The large percentage of meteoritic metal both in the soil and within the soil particles indicates a significant amount of shock impact at the site caused by chondritic and iron meteorites.

In the iron phases of chondritic meteorites the weight fraction of P is low, <0.01 wt%. However, reheated chondrites have been reported to contain greater P contents. This increased P content has been explained in terms of the decomposition of phosphates under reducing conditions and the penetration into the metal of the phosphorus so released. Our P measurements show that for metal of meteoritic composition, 65% contain more than 0.05 wt%. The relatively high P content of the metal particles in our samples of Apollo 14 soil may have arisen in part from projectiles that were composed of reheated chondrites. However the high P content more probably formed from a lunar reheating event that occurred either from hot ejecta from an impact on the lunar surface or by the proximity to a lava flow, a volcanic vent, or an ash flow.

Of the 13 remelted particles that were studied the 3 non-spherical ones have either fragmented from an original sphere or solidified in non-spherical form on the lunar surface. The bulk composition of the remelted particles varies from 6-16 wt% Ni, 0.3-0.75 wt% Co, 0.2-7.7 wt% P and 0.15-4.2 wt% S. Two of the spherules have a fanning-shaped internal structure of fine dendrites radiating from a nucleus on the surface of the sphere, similar to that encountered in some enstatite chondrules. The dendrites in these metallic "fibrous-enstatite" chondrules are 1-2 μm in size and appear to have formed at cooling rates of 10^5-10^6°C/sec. Two other particles have structures similar to those encountered in spheroids from the Canyon Diablo crater with cooling rates of about 2000°C/sec. Five particles are of similar composition (2.2-2.7 wt% P, 1.7-2.7 wt% S) and contain large individual dendrites (20-30 μm) surrounded by a phosphide eutectic and troilite. Phosphide (schreibersite) is observed as a solid state precipitate within the metal dendrites.
METALLIC PARTICLES IN THE APOLLO 14 LUNAR SOIL

J. I. Goldstein

The presence of these precipitates show that a period of either slow cooling or reheating occurred during the time that the spherules were on the lunar surface.

In discussing the origin of the remelted structures, it is noteworthy that a number of globules appeared to have fragments of remelted troilite adhering to their outer surfaces. A highly suggestive indication of the remelting mechanism may be contained in the observation of one metallic globule that was completely mantled by an outer husk of troilite-rich remelt. The metallic globule contains all of the phosphorus and small proportions of sulphur whereas the sulphide mantle is essentially free of massive metal. Studies of shocked iron meteorites and the vein material of chondrites suggest that in a shock event it is the troilite-sulphide assemblages that are most easily melted and once melted act as a solvent for phosphide, carbide and metal phases. Thus the origin of the sulphide-mantled globule may be ascribed to shock melting of sulphide by the impact of the projectile in the lunar environment, followed by solution of phosphorus and metal in the molten sulphide. On cooling the sulphide-metal-phosphide melt first separated into two immiscible liquids, typically with a metal-phosphide core and a sulphide husk. The metal-phosphide globule later solidified in a dendritic form.

One can assume that some of the metal globules in the present study were originally encased by a relatively brittle sulfide rim. If this is the case it will be extremely difficult to determine whether the meteorite projectiles which created the lunar craters were originally an iron meteorite or a chondrite.

It is possible to identify the following sources of metal in the Apollo 14 soil:

1) metal of meteoritic composition that fell onto the Mare Imbrium site and was incorporated into the lunar surface at the Imbrium site prior to the excavation of the Imbrium basin,

2) metal of meteoritic composition and structure that fell and produced local cratering activity on the Fra Mauro formation after the excavation of the Imbrium basin,

3) metal excavated from lunar Mare regions near the Fra Mauro site,

4) a small number of exotic particles. It is not easy to make an unambiguous identification of metal from the impacting planetesimal that excavated the Imbrium basin. This metal may be incorporated with that listed under 1) above.
Previous analysis of returned lunar samples has revealed significant concentration of uranium and thorium in lunar surface material. Both elements are unstable against radioactive decay and are the first members of two distinct highly complex decay series which terminate in stable isotopes of lead. Unstable isotopes of radon gas are produced as intermediate products of these series. Uranium produces $^{222}\text{Rn}$ and thorium produces $^{220}\text{Rn}$. Radon is a rather special component of the decay series because of its gaseous nature. There is a possibility that the radon will diffuse above the lunar surface where it remains trapped in an exceedingly rare atmosphere by the moon's gravity. As a result, the radioactive decay of the radon isotopes and their daughter products would have the effect of enhancing the radioactivity levels upon the surface of the moon in an important way. The alpha particle spectrometer was designed to measure the energy of alpha particles being emitted from the lunar surface and hence detect the presence of any alpha activity from radon or its daughter products.

An early estimate of the amount of radon diffusion by Kraner, Schroeder, Davidson and Carpenter, 1966\(^1\) assumed terrestrial conditions for the diffusion coefficient and concentrations. When the actual concentrations of uranium and thorium are used, their model predicts a rate of two disintegrations per sec cm\(^2\) for $^{222}\text{Rn}$ and about $10^{-2}$ disintegrations per sec per cm\(^2\) for $^{220}\text{Rn}$. Observations of alpha emission from the moon indicate that if the radon is present, the activity levels are considerably smaller than this. A measurement by Yeh and Van Allen\(^2\) from lunar orbiting Explorer 35 found no indication of alpha particle emission and set an upper limit that was about one-tenth of the value predicted by Kraner et al. Turkevich et al.\(^3\) reporting on background data obtained in the Surveyor 5, 6 and 7 alpha backscatter experiments cited evidence for a radioactive deposit at Mare Tranquillitatis (Surveyor 5) with an intensity of $0.09 \pm 0.03$ alpha disintegrations per sec per cm\(^2\). At the other two sites, Sinus Medii (Surveyor 6) and rim of Tycho (Surveyor 7) they report only upper limits to the alpha activity that are about a factor of two or three lower than Mare Tranquillitatis.
There are two other indirect measurements of alpha activity that look for radioactive deposits on returned samples that have been exposed to the lunar environment. Lindstrom, Evans, Finkel, and Arnold, 1971 (4) looked for an excess of the radon daughter $^{210}\text{Pb}$ in Apollo 11 samples. They fail to find an excess to within 3%, which implies that the effect of the radioactive deposit is less than $10^{-4}$ predicted by Kraner et al. However, there is a possibility that all or nearly all of the active deposit which resides entirely in the first micron of surface material could have been blown away by the action of the LM descent engine. A similar measurement was made by Economou and Turkevich upon the Surveyor 3 camera visor (5) which was returned to Earth from Oceanus Procellarum by the Apollo 12 astronauts. They found no evidence for the deposit and can set an upper limit that is about six times smaller than the value reported by Turkevich et al. for Mare Tranquillitatis.

From the analysis of 22 hours of Apollo 15 quick look data, we conclude that the alpha particle emission of the moon in the energy band 4.7-9.1 Mev. is at most equal to the observed total count rate, 0.004 counts/cm$^2$-sec-ster, to a statistical precision of one percent. No statistically significant difference between dark and light sides was observed. Assuming that the average lunar concentrations of uranium and thorium are correctly given by typical Apollo 11 and 12 fines the alpha particle activity of the moon due to radon emanation is less than about 0.016 the value predicted by the terrestrial model of radon diffusion of Kraner et al. (1). From the analysis of 15 minutes of final telemetry data corresponding to one-eighth of an orbit, the average count rate in restricted energy regions corresponding to alpha particles emitted by $^{222}\text{Rn}$ is less than $10^{-4}$ counts/cm$^2$-sec-ster. Assuming radioactive equilibrium between $^{210}\text{Pb}$ and $^{222}\text{Rn}$ the implied $^{210}\text{Po}$ disintegration rate is less than $6 \times 10^{-4}$ dis/cm$^2$-sec which is about a factor of fifty smaller than the value reported by Turkevich et al. (3) for the Surveyor 5 landing site.

A more extensive analysis of the orbital data will be presented.

References:
(2) Yeh, R. S.; and Van Allen, J. A., Science 166, 370 (1966).
Alpha Particle Spectrometer Experiment
Paul Gorenstein

P. 323

CESEMI STUDIES OF APOLLO 14 AND 15 FINES
Herta Görz, Eugene W. White, Gerald G. Johnson, Jr, and Mary W. Pearson
Materials Research Laboratory, The Pennsylvania State University, University Park, Pa. 16802

The CESEMI (Computer Evaluation of Scanning Electron Microscope Images) technique (Görz et al., 1971) permits the size, shape, and chemical analysis of very fine particles in the micron and submicron range. The medium for dispersing the grains on a slide (C-coated glass, LiF crystals and metallic Be) was a eutectic mixture of naphthalene and camphor (Thaulow and White, 1971). The samples under study are six lunar fines including 14163,158; 15031,44; 15041,50; 15231,49; 15501,25; 15531,40. For purposes of illustration, only the results of 15501,25 will be presented.

The optical microscopical studies of the six fines indicate an upper size limit of 600 μm, the shape of the particles ranging from spherical (seldom, brown glass) to angular with sharp edges.

Samples were processed on the SEM at magnification 1250 and 300X. For the particular sample 15501,25, equivalent circular diameters range from a low of .361 μm for a magnification of 1250X to a high of 32.126 μm for a 300X magnification. Merging the equivalent circular diameter data produces a graph that shows smaller particles in this Apollo 14 and 15 study than in the earlier Apollo 11 and 12. Particles ranging in size from .75 μm to 7.5 μm of the 1250X magnification are merged with particles of the 300X magnification greater in equivalent circular diameter than 7.5 μm and these particles produce the merged plot. Since a field magnified at 300X represents approximately 17 times in area a field of 1250X magnification, the particles at 1250X magnification are weighted accordingly.

The equivalent circular diameter versus probability plot provides two useful statistical results—median size (μm) and slope of a fitted linear relation of the size distribution—besides the visual aid of a graph. In order to see a more accurate fitting, the upper and lower 5% of the curve is discarded and fit by means of a least squares approximation of a straight line. The results of sample 1550,25 are shown in Figure 1 and given in numerical form in Table 1.

<table>
<thead>
<tr>
<th></th>
<th>300X</th>
<th>1250X</th>
<th>Merged</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fitted Straight Line</td>
<td>Y=1.0777X-1.96098</td>
<td>Y=1.0X-1.08962</td>
<td>Y=1.41436X-3.01448</td>
</tr>
<tr>
<td>Median</td>
<td>2.0530 μm</td>
<td>1.0076 μm</td>
<td>1.4599 μm</td>
</tr>
</tbody>
</table>
CESEMI STUDIES OF APOLLO 14 AND 15 FINES

Herta Görz

fines were found, but those angular particles with sharp, perhaps broken, edges were prevalent.

REFERENCES

   Particle size and shape distributions of lunar fines by CESEMI.

2. THAULOW N. and WHITE E.W. (1971)
   General method for dispersing and disaggregating particulate samples
   for quantitative SEM and optical microscope studies.
   (In press - Powder Technology).
The minerals identified are: pyroxene (greenish brown, sharp edges, sometimes isotropic), olivine (green to brown, rounded), opaque, plagioclase (optical character negative, sometimes twins), glasses with various colors, shapes and refractive indices. The microscopical study of the Apollo 14 and 15 fines shows a higher-amount of glass than found in the Apollo 11 and 12 fines studied. It is remarkable that the appearance of the glass fragments regarding color and shape resembles very often those of clinopyroxene. Only very few glass spheres have been found.

The median size for the elements studied (Fe, Ti, Si, and Ca) using the same technique is given in Table 2.

<table>
<thead>
<tr>
<th>Element</th>
<th>Median Size (μm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fe</td>
<td>1.65</td>
</tr>
<tr>
<td>Ti</td>
<td>1.53</td>
</tr>
<tr>
<td>Si</td>
<td>1.21</td>
</tr>
<tr>
<td>Ca</td>
<td>1.35</td>
</tr>
<tr>
<td>2nd electron</td>
<td>1.45 μm</td>
</tr>
</tbody>
</table>

Conclusions

The samples studied show the particles for Apollo 14 and 15 are smaller than those studied at this laboratory using the same techniques for Apollo 11 and 12. Computer analysis of individual mineralogical samples was made and corresponding size of chemical species was determined. Very few simple shaped...

With the return of the Apollo 14 rocks it has been necessary to closely examine the magnetization in breccias since most samples were of this type. In general, the breccias have the possibility of carrying a significant viscous remanent magnetization when exposed to magnetic fields. The two samples which we have studied in detail, rocks 14313 and 14321, appear to exemplify two limiting cases which can be clearly related to the iron distribution present. The VRM of sample 14313 (Fig. 1) has a definite spectrum of relaxation times associated with a sharp cutoff between 100 and 1000 minutes. The shape of the decay curve of the magnetization is that expected for a uniform distribution of grain sizes. The cutoff corresponds to a grain diameter of about 150Å and coincides with the transition from superparamagnetic to stable single domain size. These extremely fine-grained particles probably occur in the matrix of the breccia.

A different kind of VRM which follows the classical log t relationship is found in sample 14321 (Fig. 2). The time dependence of this VRM is typical of multidomain iron with a wide distribution of relaxation times and is probably related to the larger interstitial iron blebs greater than a few microns in size seen under the microscope. Igneous rocks, on the other hand, acquire little or no viscous magnetization (Fig. 1, sample 14310) suggesting that the spherical iron particles which are usually associated with troilite and are typically around 1 micron in size behave almost like single domain particles. This agrees with the very stable remanence found in these rocks.

The grain size distribution inferred from the VRM tests is well substantiated by hysteresis loop measurements. Typically, igneous rocks have a ramp-like hysteresis curve (Fig. 3) indicative of equant multidomain grains. The breccias, on the other hand, generally have rounded hysteresis curves reflecting superparamagnetic and equant single domain particles.

Furthermore, the analysis of hysteresis loops yields the concentration of metallic iron. Fig. 4 shows a histogram of the iron concentrations in igneous and fragmental rocks as well as in soils. The values represent our data as well as data from the literature obtained by magnetic and Mossbauer effect methods. The igneous rocks generally have a metallic iron concentration of about 0.1 wt% as compared to about 0.5 wt% for the soils and breccias. This five fold enrichment is more than can be accounted for by simple addition of meteoritic iron. Experiments with simulated lunar glass indicate that heating to about 900°C in a strongly reducing environment - conditions to be expected during impact - is an effective mechanism to produce native iron in quantities and grain sizes such as found in many breccias.
There is an indication that the time dependence of the magnetization of the breccias can be correlated with their metamorphic grade. In sample 14313, a low grade metamorphic sample, the VRM is dominated by superparamagnetic grains; the VRM of sample 14321, which is of medium metamorphic grade, is mainly carried by multidomain grains a few microns in size. Sample 14312 is of high metamorphic grade and magnetically totally unstable indicating a dominance of rather large multidomain grains.

In spite of the viscous effects there seems to be little question that some of the breccias carry a stable remanent magnetization and can therefore be used to reconstruct the history of the lunar magnetic field.
ON THE MAGNETIC PROPERTIES OF LUNAR BRECCIAS

W. A. Gose

Figure 3

Figure 4
LUMINESCENCE OF APOLLO 14 LUNAR SAMPLES. Norman N. Greenman and
H. Gerald Gross, McDonnell Douglas Astronautics Company, Huntington Beach,
California, 92647.

The objectives, experimental procedures, and instrumentation in our
current luminescence studies are given in our previous paper (1). We report
here the results of studies of the following Apollo 14 rock chip samples:
14301,50 (fragmental, bottom, exterior), 14310,155 (basaltic crystalline, top,
exterior), and 14321,264 (fragmental, interior).

In the irradiation with protons of 5 keV energy the samples displayed a
luminescence intensity decline over a time interval of the order of minutes.
In our earlier studies we found that, in the 1000-4000 Å luminescence band
with 100 keV proton irradiation, an Apollo 11 breccia (10048,36) showed a
lower initial intensity and a lower rate of intensity decline than an Apollo
14 (10044,53) and two Apollo 12 (12002,114; 12020,55) crystalline igneous
rocks. All three Apollo 14 rocks, however, had lower decline rates than the
Apollo 11 and 12 samples, and the crystalline rock rate was somewhat lower
than the rates for the fragmental rocks. In part, this may be due to the fact
that the two sets of samples are not directly comparable. The rates for the
Apollo 11 and Apollo 12 samples were calculated from 100 keV proton data on
peaks in the middle and near u.v. whereas those of the Apollo 14 samples were
calculated from 5 keV proton data on peaks and portions of the band in the
visible wavelengths. Also, because of the time durations of the respective
runs, the middle and near u.v. data points could be taken at about three
minute intervals, whereas the data points in the visible band had to be taken
at about seven minute intervals. It is also possible that the decline
characteristics of the u.v. luminescence may yield better diagnostic infor­
mation than those of the visible. Nash and Greer (2) have also reported
luminescence decline characteristics associated with lunar rock type and
exposure history, and Sippel and Spencer (3) have reported what are probably
related features, intensity and spectral differences in luminescence between
shocked and unshocked feldspars. These decline characteristics with proton
excitation, therefore, promise to yield important information on geologic
history of the lunar rocks. Data on the 5 keV proton irradiation with
luminescence recorded in the band 3500-7500+ Å, are as follows:

<table>
<thead>
<tr>
<th>Sample</th>
<th>Peak Wavelength (Å)</th>
<th>Bandwidth (FWHM) (Å)</th>
<th>Total Band Efficiency (ergs/erg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lunar:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14301,50</td>
<td>4730, 5510 declining to broad band</td>
<td>2250</td>
<td>$1 \times 10^{-6}$</td>
</tr>
<tr>
<td>14310,155</td>
<td>5230</td>
<td>1990</td>
<td>$1 \times 10^{-6}$</td>
</tr>
<tr>
<td>14321,264</td>
<td>5300 (broad)</td>
<td>2270</td>
<td>$1 \times 10^{-6}$</td>
</tr>
<tr>
<td>Terrestrial:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gabbro</td>
<td>4720</td>
<td>1620</td>
<td>$5 \times 10^{-6}$</td>
</tr>
</tbody>
</table>
LUMINESCENCE OF APOLLO 14 LUNAR SAMPLES

Norman N. Greenman

These values are for runs made after the luminescence intensity had levelled off, not for the first, higher intensity runs.

Irradiation with X-rays from a tungsten target at 70 kV, 45 ma, produced distinct to prominent near u.v., blue, and green luminescence peaks in all samples. A red peak was distinct in granite but barely discernible, though present, in gabbro and the lunar samples. Granite and gabbro also showed a small middle u.v. peak; this was absent from the lunar samples. Significant features of the spectra are: 1) the curves for the two fragmental rocks are similar, with the blue peak most intense and the green next, about half the intensity of the blue, 2) the curve for the basaltic crystalline rock differs from these, showing the blue and green peaks highest and of equal intensity, 3) the curve for terrestrial gabbro is identical to that for the lunar basaltic crystalline rock with respect to the relationship of the blue and green peaks, and 4) the granite curve differs from all the others in that the green peak is most intense and in that the red peak is at a longer wavelength. These features seem to indicate that both lithologic character (granitic vs. gabbroic) and geologic history and processes (unshocked lunar igneous vs. shocked lunar breccia) are reflected in the luminescence characteristics.

Data on the X-ray irradiation are as follows:

<table>
<thead>
<tr>
<th>Sample</th>
<th>Wavelength (A)</th>
<th>Bandwidth (A) FWHM</th>
<th>Efficiency (ergs/erg)</th>
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</thead>
<tbody>
<tr>
<td>Lunar</td>
<td>14301,50</td>
<td>3330, 4420, 5500, 6600</td>
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</tr>
<tr>
<td></td>
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<td>3330, 4350, 5530, 6680</td>
<td>400, 890, 780, 580</td>
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<td></td>
<td>14321,264</td>
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<td>470, 880, 810, 440</td>
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<td>Terrestrial</td>
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<td>3410, 4230, 5760, 7180</td>
<td>580, 810, 610, 1220</td>
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<tr>
<td></td>
<td>Gabbro</td>
<td>3370, 4260, 5690, 6470</td>
<td>480, 850, 820, 740</td>
</tr>
<tr>
<td></td>
<td>Willemite</td>
<td>3540, 4160, 5330</td>
<td>410, 460, 420</td>
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</table>

Far Ultraviolet (1100-2200Å) Irradiation

The terrestrial and lunar samples were first investigated with two vacuum monochromators but no measurable results were obtained because of the low intensity of the light source (~10^4 photons/sec). The procedure was changed to provide the excitation spectrum by scanning the light source spectrum through the sample luminescence was viewed directly by the Bendix BX754A channeltron photon counter (window cutoff: 3000Å) without a spectrometer. Hence only luminescence beyond 3000Å could be sensed. An appropriate series of Corning filters was used to study the luminescence spectrally. The most intense irradiation consisted of the group of lines in the band 1200-1400Å and a continuum from about 1800Å into the near u.v. The excitation spectrum showed (1) two major excitation spectrum features, with maximum excitation at about 1370Å and 1700Å in all samples; (2) approximately five smaller bands produced in the interval 1400-1800Å; and (3) the possibility of lines produced by excitation at around 1230, 1375, 1522, 1569, 1800, 1870, and 1879Å in lunar sample 14310,155, with similar sets for each of the earth samples (data on the other lunar samples are not yet available). Also luminescence was found in granite from around 3000Å to 8500Å. This luminescence may be similar in spectral character to that found with X-ray.
irradiation. Efficiencies determined on the basis of the first maximum at 1370Å, are as follows (given as ranges because of the uncertainty as to the spectral character of the luminescence):

<table>
<thead>
<tr>
<th>Sample</th>
<th>Total Band Efficiency Range(ergs/erg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lunar:</td>
<td></td>
</tr>
<tr>
<td>14310,155</td>
<td>3 x 10^{-2} to 5 x 10^{-3}</td>
</tr>
<tr>
<td>Terrestrial:</td>
<td></td>
</tr>
<tr>
<td>Granite</td>
<td>3 x 10^{-2} to 5 x 10^{-3}</td>
</tr>
<tr>
<td>Gabbro</td>
<td>5 x 10^{-4} to 8 x 10^{-4}</td>
</tr>
<tr>
<td>Willemite</td>
<td>4 x 10^{-1} to 6 x 10^{-2}</td>
</tr>
</tbody>
</table>

The major excitation peak at around 1700Å proved to be the most intense in all cases. The light source does not show any significant irradiation intensity in the band 1400-1800Å. Either this is produced by a lower intensity band around 1600Å, or by higher orders of the extreme u.v. In either case, higher efficiencies are implied than the values given above.

References and Notes
4. Supported by NASA contract NAS9-11679. We wish to thank W. M. Hansen, T. H. Mills, and R. R. Carlen for their assistance.

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Introduction: Polymict breccia 14321 contains 3 major lithologic components: basaltic fragments, rounded clasts of microbreccia, and a light colored friable matrix. A study of 7 thin sections further reveals the chronological order in which the lithic components of 14321 were assembled (summarized in fig. 1).

Microbreccia Clasts: Three generations of breccia are recognizable in 14321. Breccia 1 is present as strongly recrystallized and well rounded fragments included in breccia 2. The latter appear as distinct clasts, recognizable by virtue of a preponderance of very fine grained dark matrix. Breccia 3 incorporates 1 and 2 along with additional crystal fragments and matrix and is richer in microscopically resolvable crystal fragments than breccia 2. Breccias 2 and 3 are also recrystallized but not as strongly as breccia 1. This is supported by data for Mg-Fe partitioning between coexisting pyroxenes which indicate an average temperature of recrystallization of 1050°C for breccia 1 and only 700°C for breccias 2 and 3. Most of the breccia clasts have been well rounded by abrasion. An analysis of the average breccia matrix is given in table 1. The relatively high K and P' are due mainly to small grains of K-feldspar and phosphates. It is interesting to note that many of the individual crystal fragments of the breccias are large compared to the grain size of typical mare basalts or norites. Igneous rocks of coarser texture than those sampled to date are apparently an important component of breccias. The mineralogy of the crystal fragments includes: stoichiometric plagioclase, An58-99, with shock and thermal metamorphic features; low Ca pyroxene (orthopyroxene and pigeonite) and less abundant Ca-rich pyroxenes, many with exsolution lamellae and reaction rims; olivine, Fa8-64, with polysynthetic twinning (shock induced); ilmenite, 1-3 wt % Mg; Ti-chromite; Fe metal; troilite; zircon; fluorapatite, K-feldspar; silica. We have also analyzed several grains of pleonaste spinel, (Fe.25Mg.75)(Al1.80Cr.20)O4, pink-red in transmitted light and tentatively identified as garnet in the LSPET report. Numerous lithic fragments of norites are incorporated in the breccias. These contain low-Ca pyroxenes and plagioclase (An5-95) in varying proportions, are low in opaque phases and contain curious zircon-plagioclase and whitlockite-plagioclase intergrowths. A range of textures from igneous to recrystallized fragmental can be seen in these clasts. Much of the KREEP compositional character of the breccias is due to a large admixture of norite. "Granitic" fragments also occur in a variety of textures from partially devitrified glasses with vesicles, through complex intergrowths of plagioclase (An50-54), K-feldspar and silica, to micrographic and granoblastic (analyses 4 and 5, table 1).

Basalt Fragments: The majority of basalt clasts are sub-rounded with medium
grained sub-ophitic textures, but vitrophyric and variolitic fragments are also present. Shock features are restricted to fracturing and the development of twin lamellae in ilmenite. Olivine (Fa24-46) is normally zoned and shows evidence of resorption in some basalt fragments, but is completely absent in others. The typical mineralogy common to all crystalline basalts includes: clinopyroxene, commonly composed of pigeonite cores and augite rims with extensive iron enrichment; plagioclase, An80-95, with moderate departures from ideal stoichiometry; ilmenite (Mg from 0.1 to 2.4 wt% depending on the mineral and textural association of the individual grains); Ti-chromite; Fe metal; troilite; silica; cryptocrystalline mesostasis rich in K and Si and containing small grains of zircon, phosphates and sphene (this is the first report of sphene in lunar rocks; see table for analysis).

Light Colored Matrix: Seams of light colored material form a matrix between the breccia and basalt clasts. The composition of discrete crystal fragments in this material reveal a strong similarity to the basalt mineralogy. Only clinopyroxenes are present and the plagioclase crystals show a range of departure from ideal stoichiometry similar to that found in the basalt clasts. There is little, if any, textural evidence of recrystallization of the light matrix material. The larger fragments of 14321 that we have seen indicate that much of the light matrix material is friable and only loosely cements the aggregate.

Summary: Petrographic and bulk sample examination of 14321 indicate a complex and multistage formational history, but textural criteria and microprobe analyses allow a reconstruction of the sequence of inputs into this polymict breccia. This sequence has been summarized schematically in fig. 1. Space limitations preclude any detailed discussion, but the salient features may be summarized as follows. The earliest recognizable components of 14321 are breccia 1, "norite" and microgranite. All three lithic fragments were subjected to severe thermal metamorphism as indicated by intense recrystallization and high Mg-Fe partitioning temperatures for coexisting pyroxenes. Some of this material (e.g., norite and breccia 1) may have been excavated in the event that formed the Imbrium basin, and some (e.g., the breccias) may have been formed in a hot ejecta blanket deposited by the Imbrium event. Subsequently this material, along with crystal fragments, finer grained matrix, and some lithic fragments were incorporated first into microbreccia 2 and then 3. Both of these later stages were subjected to less severe thermal metamorphism. Finally, mare type basalt fragments and breccia 3 were relatively loosely bonded by the light matrix material into polymict breccia 14321. The light matrix material is predominantly composed of mare basalt type material. The time scale for the flow chart is necessarily relative but we consider it likely that at least three distinct stages are involved in the formation of 14321.
Table 1. Microprobe analyses of several components of 14321

<table>
<thead>
<tr>
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<th>1</th>
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<td>SiO₂</td>
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<td>46.01</td>
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<td>Al₂O₃</td>
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<td>TiO₂</td>
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<td>38.93</td>
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<tr>
<td>FeO</td>
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<tr>
<td>MgO</td>
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<tr>
<td>CaO</td>
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<tr>
<td>Cr₂O₃</td>
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<td>0.09</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>MnO</td>
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<td>0.10</td>
<td>0.11</td>
<td>0.04</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.25</td>
<td>0.97</td>
<td>*</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>BaO</td>
<td>0.15</td>
<td>0.24</td>
<td>--</td>
<td>0.58</td>
<td>0.54</td>
<td>--</td>
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<tr>
<td></td>
<td>98.78</td>
<td>97.97</td>
<td>98.78</td>
<td>99.51</td>
<td>100.49</td>
<td>100.04</td>
</tr>
</tbody>
</table>

Analyses 1-5 are defocused beam analyses. Total variation among 5 clasts is 0.0 to 0.93 depending on presence or absence of large phosphate grains.

Fig. 1
Electron spin resonance (ESR) absorptions have been observed in a suite of mm-sized particles, selected for their glassy appearance from the Apollo 11 bulk sample 10085, an Apollo 14 core 14230 and an Apollo 15 fines sample 15302. Individual particles ranging in color from clear to green to brown to opaque were examined both at Ka-band (35 GHz) and X-band (9.5 GHz) at temperatures between 77°K and 573°K. Two principal types of signals are discerned at X-band: type I, an asymmetric resonance centered near \( g \approx 2.1 \) with a linewidth and temperature dependence similar to those of the "characteristic" resonance (1-3) of lunar fines, and type II, a broad resonance with hysteresis effects having a first-derivative zero-crossing near \( g \approx 8 \) (observable only after cycling to \( \geq 5 \) K Gauss) and an associated shoulder near \( g \approx 1 \) (Fig. 1b). No statistically reliable correlation has been established between these ESR signals and the color of the samples, although the best examples of the latter type were found in dark brown glass particles.

The sources of these resonances are inferred from an investigation of a simulated lunar glass (4) of a composition close to the average for Apollo 12 red-brown-ropy and greenish-yellow-streaky glasses (5). The simulated glass as delivered (4) exhibited a weak resonance peak at \( g \approx 4.3 \), which is familiar (6) as arising from \( \text{Fe}^{3+} \), and a weaker, broad line at \( g \approx 2 \) (Fig. 1a). Remelting in an oxidizing hydrogen-oxygen flame followed by rapid quenching altered these resonances only slightly as did heating a dry-box-ground powder sample in vacuo at 650°C for 100+hours. However, heating the powdered sample in air at 650°C resulted in a 200-fold enhancement in the \( g \approx 2 \) resonance; after 200 hours this resonance strongly resembled the "characteristic" resonance of Apollo 14 fines (Fig. 1a)(7). Similar heat treatments performed on an otherwise identical control glass wherein the FeO was suppressed in favor of \( \text{CaO} + \text{MgO} \) resulted in no significant changes in the ESR spectrum, nor did \( \text{Co} \) irradiations up to \( 4 \times 10^7 \) R performed on the simulated glass. Since the only potentially paramagnetic elements in the simulated glass were iron (mainly as \( \text{Fe}^{2+} \)) and titanium (the latter of which was also present in the control sample) it is concluded that the \( g \approx 2 \) resonance results from the conversion of \( \text{Fe}^{2+} \) to \( \text{Fe}^{3+} \) in the solid state under the influence of an oxidizing atmosphere. This conclusion has been corroborated by Mössbauer spectroscopy (8). Studies of ESR intensity and linewidth as functions of temperature show that type-I resonances in both simulated and actual lunar materials are superparamagnetic, i.e., they must arise from magnetically ordered phases.
having an average grain size $\leq 150\mu\text{m}$. In view of (i) the high similarity of the two $g \approx 2$ resonances in Fig. 1a, (ii) the clear involvement of Fe$^{3+}$ in the case of the simulated lunar glass, (iii) the high fraction of glass in lunar soils, and (iv) the well-known tendency of high-iron silicate glasses to precipitate ferrite phases in the range 600-700°C (9), it is suggested that the "characteristic" resonance arises from ferrite phases in lunar glasses, and not from spherical metallic iron particles as suggested by others (2).

Samples of simulated lunar glass remelted in a reducing hydrogen-oxygen flame displayed ESR responses very similar to the second category of lunar glass signals. (Fig. 1b). On the basis of the observed hysteresis effects and the method of production, these signals are ascribed to the ferromagnetic resonance of iron precipitates within the simulated and actual lunar glasses. It would appear that many lunar glasses, like the rocks from which they were derived, must have been highly reduced initially, but that the finer fractions have subsequently undergone a degree of oxidation (Fe$^{3+}$=1-5% of total iron) followed or accompanied by crystallization. The possibility of ferrite formation in returned lunar materials as a result of terrestrial contamination is ruled out by the high temperatures required and by the findings of others (3); the sources of lunar-surface oxidants are currently unknown.

4. The simulated glass sample was prepared by C. Schott at Owens-Illinois. Iron was entered as Fe$^{2+}$ in synthetic fayalite and melting was carried out under nominally reducing conditions. All paramagnetic elements present in the average lunar composition (5) in amounts <1 wt. % were excluded for purposes of this experiment. The remaining transition metal oxides were 11.6% FeO and 2.4% TiO$_2$.
7. The linewidth of the "characteristic" resonance was found to be narrower for Fra Mauro fines than that reported (1-3) for fines from Tranquility Base (600 Gvs. 900 G, at room temperature); an intermediate width (720 G) was noted for Apollo 15 fines. Similar resonances having widths in the range 300-750 G were achieved in the simulated lunar glass by varying the atmosphere, temperature, or duration of the heat treatment.
ESR OF IRON IN LUNAR GLASSES

D. L. Griscom

Fig. 1. X-band ESR spectra of a simulated lunar glass are compared with (a) the "characteristic" resonance of Apollo 14 fines and (b) the ESR spectrum of a single, mm-sized brown-glass chip of homogeneous appearance selected from 10085. All spectra were obtained at 9.52 GHz and room temperature at a microwave power level of 10 mW. Signal intensities may be understood in terms of the product of spectrometer gain times sample mass. Normalizing to unity for the simulated glass traces of (a), this product for the simulated glass traces of (b) is 3; for the Apollo 14 fines 0.01; and for the brown glass chip 0.7.
The purpose of this study is to determine the manners in which various gases interact with lunar materials and the interaction differences, if any, between lunar and terrestrial materials. The importance of such studies lies in the fact that the lunar and terrestrial materials have had differing histories, the former being directly exposed to the various space radiations and the latter being exposed to the earth’s atmosphere, in both instances for a considerable period of time. Information provided by such studies could provide knowledge toward the understanding of the interaction of lunar surface material with gases of various origin. The work reported here is concerned with surface area measurements of Apollo 14 and 15 samples plus gas adsorption/desorption studies of high vacuum Apollo 12 and N\textsubscript{2} stored 14 samples.

The surface area measurements were made utilizing a mechanical pumped/oil diffusion system (volumetric) with research grade Krypton as the measuring gas. The samples for surface area were 14259,93 (sample weight 0.9640 gm.) and 15401,48 (sample weight 1.4551 gm.), both lunar fines. They were baked initially at 110°C for about two hours, then lowered to liquid nitrogen temperature. The recorded surface areas were 0.61 m\textsuperscript{2}gm\textsuperscript{-1} for the Apollo 14 sample and 0.43 m\textsuperscript{2}gm\textsuperscript{-1} for Apollo 15. The results are in the same magnitude range observed for samples returned by previous Apollo missions, plus terrestrial fines of about the same grain size distribution.

The Apollo 12 high vacuum sample (12001,118), which was maintained under an ultrahigh vacuum condition ($\leq$10$^{-8}$ torr) for about one year, was transferred under ultrahigh vacuum conditions, using special techniques to a high vacuum gravimetric electrobalance system (see Grossman et al., Proc. First and Second Lunar Sci. Conf. 1970 and 1971). This sample when exposed for the first time to dry N\textsubscript{2} at room temperature initially showed high adsorption. The entire weight gain occurred at pressures less than 0.5 torr. The weight then remained essentially constant up to 760 torr (1st N\textsubscript{2} experiment, Figure 1). On desorption there was no net N\textsubscript{2} retention within the experimental error. These results were reproducible in two subsequent runs. After short exposure ($\sim$5 torr seconds) to a pressure less than 2 torr, the N\textsubscript{2} adsorption isotherm changed and the knee moved to about 25 torr (2nd N\textsubscript{2} experiment, Figure 1) with a larger total weight gain. Argon also exhibited similar (Type I) adsorption. These results strongly suggest that the surface characteristics of ultrahigh vacuum lunar samples change even upon very brief exposure to air. Our previous studies of Apollo 11 and 12 samples exhibited high reactivity when exposed to oxygen, water vapor and their mixtures.
Adsorption isotherms were obtained for Apollo 14 samples 14259,93 and 14259,80, the former at liquid nitrogen temperature utilizing oxygen then nitrogen, and the latter at room temperature using nitrogen, oxygen, and water vapor. For the low temperature studies the same system was utilized as that for the surface area measurements. The room temperature (25±0.3°C) values were obtained with gravimetric electrobalances.

The low temperature oxygen adsorption/desorption showed a definite hysteresis which was not found for prior studies on Apollo 11 samples (see Fuller et al., Proc. Second Lunar Sci. Conf., 1971). At room temperature there seems to be a slight hysteresis, although the extent of hysteresis is close to the standard deviation of our measurements.

Since the sample (Apollo 14) used for low temperature studies was exposed to air for about twenty minutes (as compared to about one week exposure by Fuller et al.) the hysteresis and its shape for the oxygen adsorption isotherm were unexpected. Nitrogen adsorption isotherms, which were obtained using the same oxygen exposed sample at low temperature, showed dramatically the effect of oxygen (Figure 2). The first N₂ run showed gas release with the increase of relative pressure to about 0.15 and a significant gas retention on desorption. The second and third runs showed even larger effects at slightly higher relative pressures. We ascribe this behavior to be due to a group of processes which could include irreversible and reversible opening of micropores in lunar material grains. As a result, some of the gases, which could have been solar, lunar, terrestrial or otherwise trapped, might have been released. All low temperature runs were indicative of a non-terrestrial type behavior of the Apollo 14 sample, probably as a result of radiation damage.

In our room temperature runs of Apollo 14 samples, exposed to the gases N₂, O₂, H₂O vapor, we found that (a) N₂ showed no hysteresis, (b) O₂ showed a slight hysteresis (with reversibility) although the extent of hysteresis is close to the standard deviation and (c) H₂O vapor showed a high retention upon desorption (cf. Fuller et al.). The above findings indicate that the degree of adsorption is highly dependent upon surface nature, and that sequential use of an inert gas, reactive gas and followed by the inert gas may prove to be a valuable technique for probing the nature of adsorbent surfaces. In any case, it would appear that these differences can provide information relating to the history of the lunar surface materials and hopefully their evolution.
Figure 1 (left)  Nitrogen and Argon Adsorption on Apollo 12 High Vacuum Sample

Figure 2 (right) Nitrogen Adsorption on Apollo 14 Sample After Oxygen Exposure.
SUBSOLIDUS REDUCTION AND COMPOSITIONAL VARIATIONS OF LUNAR SPINELS.
S.E. HAGGERTY, GEOLOGY DEPT. UNIV. OF MASS. AMHERST, MASS. 01002.

(A) Reduction: Ulvospinel-ilmenite-iron intergrowths, reported in a restricted number of samples from Apollo XI and XII, are present in basalts from XIV, XV, and Luna 16. Phase intergrowths are of two types: Type 1: Ilmenite \((\text{Fe}_{97}\text{Mg}_{0.1}\text{Mn}_{0.1}\text{Cr}_{0.04}\text{Al}_{0.02}\text{Ti}_{0.9903})\) develops in well defined lamellae along \(\{111\}\) planes in Al-Ti-chromite \((\text{Fe}_{8.0}\text{Mg}_{2.0}\text{Mn}_{1.2}\text{Cr}_{7.85}\text{Al}_{4.12}\text{Ti}_{1.89}\text{O}_{32})\); coarse, particulate \(\text{Fe}\) is present and is intimately associated with the Widmanstatten texture. Type 1 intergrowths are present in basalt 14053 and in basaltic fragments in breccia 14321. Type 2: Ilmenite, compositionally similar to Type 1, is present along pseudo-octahedral planes, and in internal and external composite textures in Cr-Al-ulvospinel \((\text{Fe}_{15.63}\text{Mg}_{0.6}\text{Mn}_{0.07}\text{Cr}_{0.35}\text{Al}_{0.81}\text{Ti}_{7.20}\text{O}_{32})\); to \(\text{Fe}_{13.65}\text{Mg}_{3.6}\text{Mn}_{1.1}\text{Cr}_{2.37}\text{Al}_{1.03}\text{Ti}_{6.03}\text{O}_{32}\). These intergrowths contain lower concentrations of free metallic iron and ilmenite than those of Type 1, and troilite may be an additional minor phase. Type 2 intergrowths are present in samples 15058, 15076, 15495, 15535, 15555, and in Luna 16. Type 1 clearly results from subsolidus reduction of primary chromian-ulvospinel, and is accompanied by the decomposition of fayalite to \(\text{Fe}+\text{cristobalite}\) in sample 14053. The origin of Type 2 is problematical; such intergrowths may result from contemporaneous crystallization of ulvospinel and ilmenite, but slow cooling and less intense reduction are favored. More intense reduction (Type 1) results in a progressively enriched Cr-spinel host, and bulk compositions are maintained by the concomitant development of proportionately larger concentrations of \(\text{Fe}+\text{ilmenite}\). Ilmenite and primary Mg-Al-Ti-chromite remain unaffected by the reduction process. Thermodynamic calculations for the equilibrium dissociation of ulvospinel, ilmenite, hercynite and chromite, show that this is the order of increasing stability, as a function of \(T\) and \(f_0^{2}\). The reaction \(\text{Usp}=\text{TiO}_2+2\text{Fe}+\text{O}_2\) (a), for complete reduction of \(\text{Fe}\) from \(\text{Fe}_2\text{TiO}_4\), is close to that of \(\text{Ilm}=\text{TiO}_2+\text{Fe}+\frac{3}{2}\text{O}_2\) (b), and below 1000°K these curves are coincident (Fig 1). Thus in the presence of coexisting \(\text{Ilm}\), the partial reduction of \(\text{Fe}\) from Cr-bearing Usp in lunar basalts is above the univariant curves determined for reactions (a) & (b), but below the univariant curve for the observed reaction \(\text{Usp}=\text{Ilm}+\text{Fe}+\frac{3}{2}\text{O}_2\). The FQI buffer curve falls between these limits, provides confirmatory evidence for the observed reaction \(\text{Fe}=2\text{Fe}+\text{SiO}_2+\text{O}_2\) in sample 14053, in the coexisting region of \(\text{Ilm}\) stability and Usp instability, and therefore permits determination of prevailing \(f_0^{2}\) which at 1000°C=10^{-15} to 10^{-16} atmos. Intensity of reduction is related to
the modal ratio of chromite:ilm+Fe, and to the degree that secondary (reduced) chromites approach Fe/Ti and Fe/Cr ratios of primary chromites. These parameters suggest that the basaltic fragments in 14321 are slightly more reduced than 14053, and that 14053 is intensely reduced with respect to samples from XI, XII, XV and Luna 16. The intensely reduced samples contain veinlets of Fe (+FeS); reduction is probably post-deuteric and may have resulted from contact with volatilization of meteoritic iron.

(B) Compositional Variations: Total integration of spinel compositions from the 5 lunar landing sites (Fig 2) shows that a complete solid solution series is present between .25 FeAl2O4-.75FeCr2O4 and Fe2TiO4. Each site has a unique compositional characteristic: XI compositions (12 analyses) are restricted to intermediate members of the series; those from XII(106 anal.) show a discontinuity in the series and also show enrichment of MgCr2O4 in low Ti-chromites; XIV spinels are of three compositional groups (75 anal.):(a) Usp-rich phases in 14310(Fe14.84Mg.24Mn.1Cr.8...), (b) primary and secondary Al-Ti-chromites (14053, AII.ITi6.87032), (c) this enrichment is similar to the XII trend and is identical to the continuous trend shown by XV, (c) Mg-enrichment as Ti decreases; this enrichment is similar to the XII trend and is identical to the continuous trend shown by XV, (c) MgCr2O4, MgAl2O4, and FeAl2O4 are dominant components with the following typical compositions:Fe5.21Mg3.08Mn.11Cr9.81Al5.52Ti.19032;Fe5.63Mg3.54Mn.08Cr6.43Al7.2Ti1.1032;Fe9.06Mg2.1Mn.1Cr5.83Al3.48Ti3.36032; these spinels show maximum variation of Cr/Al as a function of Fe/Mg, and 4 distinct trends are present:(1) towards FeCr2O4,(2) towards the join FeCr2O4-.5MgCr2O4,(3) towards the join FeAl2O4-.5MgAl2O4 and (4) towards Fe2TiO4. The common overall trend, from FeAl2O4-FeCr2O4 towards Fe2TiO4, present at each of the sites is indeed surprising in view of the variety of rock types and the extreme differences in bulk chemistry. Other outstanding features are Al-enrichment in Luna 16 spinels, discontinuities in the hercynite-chromite-ulvospinel series and in the hercynite-spinel-chromite series, and the fact that Mg2TiO4 plays little or no part in spinels observed to date. Solid solution within the Her-Chr-Usp series is
accomplished by substitution of Fe+Ti for Cr+Al; extensive and highly variable ionic substitution, either independently or simultaneously, of Al for Cr and Mg for Fe is present as a function of Fe/Ti. Trends for Al/Cr are well defined (Figs 3,4) whereas Fe/Mg trends are generally incoherent (Fig 5) or show considerable overlap (Fig 6) as a function of Fe/Ti. Maximum variation of Cr/Al and Fe/Mg takes place in low Ti-chromites, limited substitution, except for Luna 16, is present for intermediate members, and these ratios are constant and high for Ti-rich chromian-ulvospinelis. Isobaric planes, extrapolated from Fig 1, and shown schematically in Fig 2 for the spinel compositional prism, indicate the minimum fO2 values required for stable crystallization, and suggest that the lunar spinel paragenetic sequence (Mg-Al-chromite, Al-Ti-chromite, Ti-Al-chromite, Cr-ulvospinel and ulvospinel), and the degree of ionic and coupled substitution are controlled not only by liquid composition but also by fO2. In conclusion, these data suggest that the substitutional behavior in spinel changes during crystallization but these changes cannot be fully evaluated in the absence of pertinent experimental data.
THE URANIUM DISTRIBUTION IN LUNAR SOILS AND ROCKS 12013 and 14310 (1)

The U distribution in lunar samples 12013, 14; 12013, 15 and 14310, 6 and in soils 10084, 8; 12070, 5; 14259, 97 and 15221, 46 was studied by fission track (2) and electron microprobe mapping techniques. The major U-bearing phases are whitlockite, apatite, zircon, baddeleyite, thorite and Zr-Ti rich phases. Zr-Ti rich phases have been reported by many authors (3,4) but the extremely high U, Th and Pb concentrations are rare. A Zr-Ti rich phase in rock 12013 contains major Ca, Fe, Nb, Y, REE and up to 4.2 wt% PbO, 4.7 wt% ThO₂ and 3.6 wt% UO₂ (3). Thorite was identified in soil sample 14259, 97.

Grains containing more than 10⁹ U atoms were recorded and identified with the electron microprobe. They contain from 1.8 to 29.9 atomic percent of the total sample U. Tables 1 and 2 list the minimum atomic percentages of the total sample U contained in the major U-bearing phases. The percentages ascribed to "unidentified hosts" collectively refers to major phases with very low U concentration and/or very small grains of the U-bearing phases containing less than 10⁹ U atoms and/or U adsorbed on the grain surfaces.

Lunar rock 12013 is distinguished among the rocks returned by the Apollo mission by its lithic heterogeneity and by its high abundance of Si, K, Rb, Y, Zr, Nb, Ba, REE, Th and U. It consists of a heterogeneous mixture of a "dark" clastic lithology and a "light" K-feldspar-rich lithology. In the dark lithology much of the U resides in phosphate and zircon and is relatively uniformly distributed (Table 1). The light lithology contains rare grains of a Zr-Ti rich phase. Although rich in U, Th, and Pb, these grains apparently do not dominate the U, Th, Pb systematics as was suggested earlier (3). Zircon and phosphates are also present in the light lithology, and while their U and Th concentrations are much lower than those of the Zr-Ti-rich phase, their larger modal abundances make their U contribution to the total rock U budget as significant as the Zr-Ti-rich phase (Table 1).

Concentrations of U, Th and Pb are sufficiently high in the Zr-Ti rich phase of rock 12013 to permit calculation of ages from electron microprobe analyses of these three elements. These ages average 4.0 ± 0.1 AE (3). Reasonable agreement with the Rb-Sr and Ar³⁹/Ar⁴⁰ ages suggests that these phases represent closed systems with respect to U, Th, and Pb.

Rock 14310 was investigated because it is the most U-rich crystalline rock known. The identified U-bearing phases include baddeleyite, phosphate, zircon and Zr-Ti rich phases (Table 1) but 95 atomic percent of the U resides in grains with less than 10⁹ atoms of U per grain. If this U is present in these U-rich phases, their grain size must be ≲ 1 μm.

Haines et al. (3) suggested that the Zr-Ti-rich phase may significantly affect the U, Th, and Pb systematics of the lunar soils. Table 2 gives U distribution data in soils from all the lunar missions. The data suggest that
a Zr-Ti rich phase is an important U-bearing phase, and contains more of the soils' U than either zircon or phosphate. However, in no case do these coarser grained (10^8 U atoms/grain) U-rock phases dominate the soils' U, Th, and Pb budget. However, in sample 14259,97 the one grain of thorite does dominate the U content of the sample, thus a few grains of thorite in a soil sample would seriously alter the total U content of the soil and the U, Th, Pb systematics.

References: (1) This research was supported by the Director's Discretionary Fund, Jet Propulsion Laboratory and by the National Aeronautics and Space Administration, contracts NAS 7-100 and NAS 9-8074.


### Table 1. Uranium Distribution in Rocks 12013 and 14310 (atomic %)

<table>
<thead>
<tr>
<th></th>
<th>14310,6</th>
<th>12013,14</th>
<th>12013,15</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Unidentified hosts</strong></td>
<td>95.6</td>
<td>89.0</td>
<td>95.1</td>
</tr>
<tr>
<td><strong>Phosphate minerals</strong></td>
<td>2.2(12)*</td>
<td>9.5(8)</td>
<td>1.9(10)</td>
</tr>
<tr>
<td><strong>Zr-Ti rich phases</strong></td>
<td>0.9(13)</td>
<td>0.9(7)</td>
<td>2.0(22)</td>
</tr>
<tr>
<td><strong>Zircon</strong></td>
<td>0.5(7)</td>
<td>0.6(7)</td>
<td>1.0(12)</td>
</tr>
<tr>
<td><strong>Baddeleyite</strong></td>
<td>0.7(6)</td>
<td>&lt;0.1(0)</td>
<td>&lt;0.1(0)</td>
</tr>
</tbody>
</table>

* (1) number of grains identified

### Table 2. Uranium Distribution in Lunar Soils (atomic %)

<table>
<thead>
<tr>
<th></th>
<th>10084,8</th>
<th>12070,5</th>
<th>14259,97</th>
<th>15221,46</th>
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<tbody>
<tr>
<td><strong>Unidentified hosts</strong></td>
<td>98.2</td>
<td>89.1</td>
<td>70.0</td>
<td>93.5</td>
</tr>
<tr>
<td><strong>Phosphate minerals</strong></td>
<td>0.2(1)*</td>
<td>2.4(5)</td>
<td>1.5(13)</td>
<td>2.2(6)</td>
</tr>
<tr>
<td><strong>Zr-Ti rich phases</strong></td>
<td>1.5(1)</td>
<td>7.3(7)</td>
<td>5.3(23)</td>
<td>2.5(5)</td>
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<tr>
<td><strong>Zircon</strong></td>
<td>0.1(1)</td>
<td>0.7(2)</td>
<td>1.2(7)</td>
<td>1.8(3)</td>
</tr>
<tr>
<td><strong>Thorite</strong></td>
<td>&lt;0.1(0)</td>
<td>&lt;0.1(0)</td>
<td>21.6(1)</td>
<td>&lt;0.1(0)</td>
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<tr>
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<td>&lt;0.1(0)</td>
<td>0.5(1)</td>
<td>0.4(3)</td>
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</tbody>
</table>
THE WHITE ROCK GROUP AND OTHER BOULDERS OF THE APOLLO 14 SITE: A PARTIAL RECORD OF FRA MAURO HISTORY
M. H. Halt, U. S. Geological Survey, 601 East Cedar Avenue, Flagstaff, Arizona 86001

The "white rocks" (Refs. 1 and 2) and other large boulders on the rim and ejecta blanket of Cone crater reveal the following sequential history of the Fra Mauro Formation:

1. The Pre-Imbrium terrane in the source area of the Fra Mauro Formation is represented in the large boulders by a heterogeneous assemblage of fragments of dark and light rocks. These fragments are mainly 10 cm or less in size with just a few compound clasts ranging up to 1 1/2 m. Wilshire and Jackson (Refs. 3 and 4) show that the compound clasts contain at least 4 generations and extend to as small as 0.1 mm.

2. The Imbrium event is represented in the boulders by the incorporation of this heterogeneous fragmental material into dark and light layers that were contorted during transportation, deposited in linear ridges and consolidated into rock types seen in the boulders and samples. These are the F3 and F4 fragmental rocks of Jackson and Wilshire (Ref. 4).

3. After the Imbrium event, the rocks underlying the linear ridge at the Apollo 14 site were fractured and jostled by cratering that preceded the Cone crater event.

4. The Cone crater event excavated the ridge generated in (2) and emplaced the "white rocks" and other large boulders in their present positions. The post-Cone crater history has included rounding and etching of the large boulders by micrometeorite bombardment.

References and Note
3. Wilshire, H.G., and Jackson, E.D., Petrology of the Fra Mauro Formation at the Apollo 14 Landing Site, this conference.
4. Jackson, E.D., and Wilshire, H.G., Classification of the Samples Returned From the Apollo 14 Landing Site, this conference.
5. Work done under NASA Contract Number T-65253-G
THE CARBON COMPOUNDS
IN APOLLO 14 LUNAR SAMPLES

by

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M. A. Jabbar Mohammed, Lois A. Nagy and Bartholomew Nagy
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The University of California at San Diego, La Jolla, Cal., 92037

Amino acids—a few other heteroatomic compounds and hydrocarbons were found in trace quantities. They were identified by ion exchange chromatography and vacuum pyrolysis-mass spectrometry, respectively. A far greater abundance of carbon was present in the carbonaceous gases CO₂, CO and CH₄, which were also liberated by vacuum pyrolysis.

The amino acids were determined in the SESC sample and in fines 14421, 9 by refluxing for a period of 24 hours, followed by hydrolysis with 6 N HCl and the hydrolysates were analyzed by ultra high resolution, high sensitivity ion exchange chromatography. Glycine > serine > aspartic acid > glutamic acid were the more prominent constituents of the SESC sample. In the other lunar sample glycine > aspartic acid > serine were the more important amino acid constituents. Methylamine was also present in both samples. It should be noted, however, that individual amino acids in the lunar samples were present in less than 1 nanomole per gram concentrations. Extreme care was taken to exclude terrestrial contaminations. In addition an astronaut's glove used for simulation and training was extracted for its amino acid constituents for the purpose of identifying sources of potential contaminations.
The Carbon Compounds in Apollo 14 Lunar Samples

Paul B. Hamilton

P. 355

The gas analyses have shown that the total carbon from CO, CO$_2$ and CH$_4$ varied between 76.0 and 161.3 ppm, as established by high resolution mass spectrometry. The carbon content of the SESC sample was similar to the contingency sample and it was almost identical to that of one of the other fines. Interestingly, the rock samples showed higher methane content among all samples that were analyzed, but the interior rock chip contained approximately as much carbon as the fines. The results of the gas analyses are shown in Tables I and II.

Vacuum pyrolysis-mass spectrometric analyses for hydrocarbons have shown the presence of ions of m/e = 78, 91, 104 as well as lower molecular weight aliphatic hydrocarbon fragments. SO$_2$, CS$_2$ and COS were prominent constituents of the fines and of the exterior rock. CN and HCN were also detected. The fines and the exterior rock chip showed approximately the same hydrocarbon and heteroatomic ion composition. The interior rock chip contained less carbon compounds but showed strong m/e = 78 and 104 ions. A control experiment involving pyrolysis of a small sample of the surface of the astronaut’s training glove yielded mass spectra which were entirely different from those of the lunar samples.

In order to attempt to find the locales where the carbonaceous substances are held in the lunar samples transmitted light and scanning electron microscopy were employed. The results showed that the Apollo 14 fines are morphologically similar to Apollo 11 and 12 fines. The concentration of the glass beads and dumbbells did not appear to be homogeneous among different samples and large inclusions were less common in the Apollo 14 fines than in those from the previous missions. Glass beads after vacuum pyrolysis at 1000$^\circ$C showed no inclusions or blow holes, although some lens-shaped openings were observed occasionally. Scanning electron microscopy revealed a number of wrinkled, flaky sheets in the SESC sample. It was established by control experiments that these sheets were not artifacts from the electron microscopic preparation. These particles were very unstable under the heat of the electron beam when they were not coated with the Au-Pd alloy. Scanning electron microscopy of the glove material showed morphologically different particles than the sheets found in the lunar samples.
The Carbon Compounds in Apollo 14 Lunar Samples
Paul B. Hamilton

Table I

Apollo 14 - Results of Gas Analyses

<table>
<thead>
<tr>
<th>Sample</th>
<th>ppm C from the Gases</th>
<th>Total ppm C</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CO</td>
<td>CO₂</td>
</tr>
<tr>
<td>14240 (SESC)</td>
<td>82.1</td>
<td>31.2</td>
</tr>
<tr>
<td>14003,55 (contingency)</td>
<td>72.1</td>
<td>28.3</td>
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<tr>
<td>14421,9</td>
<td>88.6</td>
<td>22.6</td>
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<tr>
<td>14163,101</td>
<td>57.7</td>
<td>17.1</td>
</tr>
<tr>
<td>14298,6</td>
<td>86.1</td>
<td>18.7</td>
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<tr>
<td>14047,18 (rock)</td>
<td>94.2</td>
<td>58.1</td>
</tr>
<tr>
<td>14311,58 (rock)</td>
<td>57.0</td>
<td>42.3</td>
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</table>

Table II

Apollo 14 - Results of Gas Analyses

<table>
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<th>Sample</th>
<th>% C from Gases</th>
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<th></th>
<th></th>
</tr>
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<td></td>
<td>CO</td>
<td>CO₂</td>
<td>CH₄</td>
<td></td>
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<tr>
<td>14240 (SESC)</td>
<td>71.8</td>
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<td>14003,55 (contingency)</td>
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<td>75.9</td>
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<td>81.1</td>
<td>17.6</td>
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<td>36.0</td>
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<td>14311,58 (rock)</td>
<td>54.8</td>
<td>40.6</td>
<td>4.6</td>
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</table>
MAGNETIC PROPERTY MEASUREMENTS ON SEVERAL APOLLO 14 ROCK SAMPLES, R.B. Hargraves and N. Dorety, Dept. of Geol. and Geophys. Sci., Princeton University, Princeton, N J. 08540

The change in orientation of NRM on A.F. demagnetization of "oriented" lunar samples 14310.57, (2.41 gm), 14303.28 (1.93 gm), 14321/194A-1 and 2 (contiguous pieces, 11.1 and 9.1 gm) are shown in Figure 1 lower hemisphere, equal area projection; the change in remanent intensity of these, plus unoriented sample 14063.44 are shown in Figure 2. Js/T curves on powder from 14321.194A, run at fields of 3800 and 8100 gauss, in vacuum of $3 \times 10^{-5}$ tor are also shown in Figure 2.

Technique involved repeated demagnetization at each A.C. field (vectors joined in Fig. 1) until the scatter became excessive. Oriented samples 14310 and 14303 contained remanent vectors stable up to >200 oe; measurements on 14321/194A are incomplete. Remanent vector orientations appear unrelated; however unequivocal orientation of the 14303 fragment provided was not possible - either orientation is as shown, or a 90° clockwise rotation is needed. If latter is correct, then all three samples would seem to possess a stable remanent vector north seeking pole down, and to the north east. Saturation (8000 gauss) remanence moments range from $2 \times 10^{-3}$ to $7 \times 10^{-4}$ emu/gm decreasing to 20% of initial value after 500 oe. A.F. demagnetization.

Major inflections in Js/T curves at $\sim 625^\circ$ on heating and $\sim 575^\circ$ on cooling is tentatively attributed to hysteresis in $\alpha$ to $\gamma$ phase transition in NiFe. Inflection at $\sim 275^\circ$ in cooling curve remains unexplained.
Magnetic Property Measurements...
Hargraves

Figure 1
Figure 2
Apollo 14 breccias, igneous rocks, and soils and two Apollo 15 soils have been analyzed by means of cosmic ray tracks. The most abundant Apollo 14 rocks, the breccias, have a mixture of high and low track densities at most positions in their interior. The observed track abundances make it clear that most of the tracks are inherited from the parent ingredients of the breccias. Measurement of the minimum track density at a known depth allows a maximum surface exposure time at that depth to be calculated. Since shock – the probable agent for producing these breccias – does erase tracks in some of the crystals, it is likely that in most cases the minimum densities are in fact true values for the number of tracks created since formation of the breccias. The observed maximum surface residence times (see Table I), 0.05 to 8.2 m.y. with a median of 1.35 m.y., are typically a factor of ten less than those observed for Apollo 11 and 12 igneous rocks. The low surface exposures appear to be the natural result of the friable nature of these rocks, which allows more rapid large scale erosion and more catastrophic break-up from impacts.
The only igneous Apollo 14 rock of interest is 14310. Our data on a section extending from the center of the rock to the bottom would be compatible with 1 m.y. surface exposure of the bottom followed by a 20 m.y. exposure in the upright position. Data from other members of the 14310 consortium, however, make it clear that a more complicated history must have obtained. One possibility is that the major surface exposure occurred over a longer time with the present rock 14310 as the interior of a considerably larger rock (at least 20 cm in radius, for at least 400 m.y.). Three igneous rocks in the size range 2-4 mm give surface ages of 3, 3, and 5 m.y.

Examination of gradients in a group of soil samples reveals variable slopes, most of which are artificially low because the samples were cut at random - i.e., without knowledge of the direction of the original nearest, space-exposed surface. The steepest slope is consistent with the spectrum inferred from the Surveyor III filter glass, and yields a surface resistance time of 4500 years (±35%).

Soils are extremely variable - median track densities ranging over at least a factor of 200. Soil from the bottom of the trench at site G (Apollo 14) has track densities typically a factor of 20 to 60 less that of nearby surface soils 14259 and 14163. The youngest soil we have examined is 15401, which is rich in clear, defect-free green glass spherules and ellipsoids.
Median track densities in the green glass and pyroxenes suggest that most of this soil is unusually young. Assuming it was scooped from depths ranging from 0 to 3 cm, we infer a deposition not more than $10^6$ years ago.

### SURFACE AGES OF BRECISAS

<table>
<thead>
<tr>
<th>ROCK NUMBER</th>
<th>MIN. TRACK DENSITY $[\times 10^6/\text{cm}^2]$</th>
<th>MAXIMUM DEPTH $[\text{cm}]$</th>
<th>MAXIMUM SURFACE AGE $[\times 10^6 \text{ yr}]$</th>
</tr>
</thead>
<tbody>
<tr>
<td>14047, 42</td>
<td>1.1</td>
<td>0.5</td>
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<td>14055, 1</td>
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<td>14066, 22</td>
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<td>0.7</td>
<td>0.49</td>
</tr>
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<td>14301, 33</td>
<td>0.27</td>
<td>0.5</td>
<td>0.34</td>
</tr>
<tr>
<td>14311, 36</td>
<td>0.73</td>
<td>${3.5$</td>
<td>3.1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.5</td>
<td>1.1</td>
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<td>14321, 270</td>
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<td>0.08</td>
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<td>15233, 5,16</td>
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<tr>
<td>15233, 5,17</td>
<td>3.5</td>
<td>0.13</td>
<td>1.3</td>
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</table>

**MEDIAN** 1.35

**MEDIAN FOR APOLLO 11 AND 12** 13

Investigators of sub-microscopic impact craters (1 to 200 microns in diameter) on lunar materials have suggested that the craters were formed by the impact of secondary lunar particles ejected during a larger impact event (1, 2, 3, 4). Other workers assume craters in this size range are due to the primary impact of extralunar material (5). Investigators of craters (100 microns to over 1 cm in diameter) on rock surfaces conclude that the dominant cratering process is the primary impact of extralunar particles (6). We offer the following evidence and arguments that all craters displaying glass-lined pits in the 1 micron to cm range are caused by primary impact:

1. In laboratory cratering experiments using aluminum projectiles greater than 0.4 mm in diameter and impact velocities of up to 7 km/sec, no glass-lined pits were produced in cold soda-lime glass targets (7). Cratering experiments using microparticle accelerators and micron-sized projectiles produced flow phenomena or melting at impacts of about 2 km/sec (8, 5). Accordingly, minimum impact velocities required to produce glass-lined pits of micron and mm size are about 2 km/sec and 7 km/sec respectively. In contrast to laboratory-produced pits, the geometry of crater lips surrounding micron-sized lunar impact pits indicate that a larger amount of melt was produced which in turn is evidence that even the micron-size lunar craters are generated by impact velocities significantly larger than 2 km/sec.

2. In general, rock surfaces exposed to space (based on lunar surface photography) always display a population of microcraters whereas buried rock surfaces often display no craters. Obviously, most cratering of rocks occur while they rest on the lunar regolith. Therefore the distribution of glass-lined pits on the entire exposed surface of a rock (9) is incompatible with a secondary origin, because secondary (lunar) particles generally cannot impact at velocities higher than 2.4 km/sec, the lunar escape velocity.

3. Rock 12054 is a prime example for accumulating such a primary crater population on its exposed parts while at rest on the lunar surface. Craters with diameters as low as 50 microns were observed optically on a glassy surface on this rock. The morphology of these craters is identical to the same size craters on glassy materials from the lunar fines permitting the conclusion that the larger impact craters on glass spherules from lunar fines are also primary. Smaller craters on glass spherules, those with pits generally less than 5 microns, differ in morphology in that no spall zones surround the pits. However, the larger pit-plus-spall craters and the smaller pit-only craters form a genetic continuum based on the observation of concentric fractures representing incipient spallation around pit-only craters (10). Thus the entire size range of craters showing evidence of shock produced melting is apparently the result of primary impacts.

4. Based on laboratory cratering experiments for a single event, one should expect a negative correlation between the amount of material ejected and the ejection velocity of that material. The bulk of crater ejecta will not be accelerated above 2 km/sec (11); indeed most ejecta should travel at velocities below 1 km/sec (12). Therefore, individual ejecta particles from one large impact traveling at relative velocities in excess of 2 km/sec have a low probability of occurrence. For each hypervelocity impact of genuine
The Origin & Significance of Lunar Microcraters

J. B. Hartung

On secondary nature, a large population of genuine low speed secondary craters (without melt phenomena) should be detected. Since this is not the case, it is taken as evidence that secondary impact is not the dominating cratering process.

5. Based on a summary of satellite-borne meteoroid detection experiments a differential mass spectrum of interplanetary dust has been derived (13). Cratering experiments using a microparticle accelerator provide a basis for calibrating observed microcraters in terms of impacting particle mass (8, 5). Based on observations of lunar microcraters, a differential mass spectrum of impacting particles has been obtained and agrees very well with that based on satellite-borne experiments. Thus, the observed size distributions of impact craters correspond to the previously known mass distribution of primary, interplanetary particles.

The origin of mm-sized impact pits due to primary micrometeoroids seems to be established according to the above arguments. However, problems for the origin of micron-sized craters by primary impact are as follows:

1. Material of definite lunar composition was observed lining impact pits on a metallic particle and interpreted as projectile remnants (2). It is not likely that primary impact alone would produce this result. however, the possibility exists that subsequent secondary deposits which are omnipresent in the lunar fines may have contaminated the interior of the impact crater.

2. Some investigators have shown that phenomena normally associated with high velocity impact may be produced at significantly lower velocities if the target material is heated short of melting (7). However, their data do not show conclusively that impact at moderate velocities (~2 km/sec) can produce the characteristic pit-plus-spall craters.

3. Statistically, relatively more glass spherules possess populations of impact craters than do crystalline fragments (3, 7). One might expect relatively as many craters on crystalline fragments as on glass spherules from a random, primary-impact process. For optical studies there is significantly greater difficulty in identifying impact craters on crystalline surfaces than on glass surfaces (6). This situation may also exist for scanning electron microscope studies.

Pertinent exposure age measurements of the surfaces investigated are still missing and therefore conclusions about the absolute flux of micrometeoroids based on lunar microcrater population statistics are not possible. However, these statistics may be used to derive independently the differential mass spectrum of interplanetary dust particles. The important characteristic of this spectrum is that more mass impinges on the lunar surface in the form of 10^-5 to 10^-7 g particles than in any other equivalent size class.

For example, in a given time, over a given area, for every impact of a single one-gram projectile, there may be expected 1 million impacts of 10-microgram particles. The mass and the kinetic energy involved in the 10-microgram impacts would be 10 times that for the one-gram event, assuming equal impact velocities.

It is well known that high-velocity impact causes ionization, vaporization, and melting of solid materials. If the mass of material affected by these processes during a single impact is linearly related to the impacting particle mass, then we may conclude that more mass is ionized, vaporized, and melted by mm-sized cratering events than by craters of any other equivalent size.

Small scale cratering events must also be considered as a possible dominant mechanism for the horizontal transport of lunar surface materials. Similarly, the erosion and destructive fracturing of rocks is the result of the microcratering process. Finally, relatively greater amounts of the extralunar component in the lunar regolith presently arrive as exceedingly small particles than as large meteorites.
The Origin & Significance of Lunar Microcraters

J. B. Hartung

Fig. 1. Cumulative densities of microcraters obtained by optical studies on glass-coating 12054. Total crater counts: 967. Open and closed symbols represent independent counts of two observers. The difficulty to record quantitatively the very small craters is illustrated by the 3 branches at 20x, 40x and 100x magnification. Scanning Electron Microscope investigations on a glass spherule (10) indicate significantly greater crater densities for .01-.05 mm craters than indicated here.

Fig. 2. Differential mass spectrum showing the relative amounts of mass impacting the lunar surface in the form of different sized particles. The curve is based on a summary of data obtained by microcrater statistics (10) and satellite-borne particle detection experiments (13) with smoothing to eliminate an artificial discontinuity introduced by the mathematical representation of the data.

REFERENCES:
10) Hartung, J. B. et al., to be published.
RARE EARTHS AND OTHER TRACE ELEMENTS IN APOLLO 14 LUNAR SAMPLES
by Philip A. Helmke and Larry A. Haskin, Chemistry Dept., University of Wisconsin, Madison 53706

Concentrations for trace elements in samples from the Apollo 14 mission, determined by neutron activation analysis, are given in Table 1. Other values referred to in the discussion below are preliminary.

Three igneous rocks were analyzed. Rock 14072 has a smaller Eu depletion than 14053 and only 0.5 to 0.8 times the REE concentrations of 14053. On a graph of Sm/Eu versus Sm, points for 14072 and 14053 fall among those for Apollo 12 basalts. Rock 14072 contains more Co, Mn, and Ni, but less Hf, Ga, and Cs than 14053, the same trends as observed among Apollo 12 basalts. From the trace element data and our experience with Apollo 12 rocks (1), we suggest that 14072 and 14053 could have come from the same parent, but 14072 would contain more olivine, probably cumulus olivine. The preliminary examination (2) revealed olivine phenocrysts in 14072 but not in 14053.

The REE concentrations of 14072 and 14053 resemble those of Apollo 12 basalts, e.g., 12020 (2), but the Apollo 14 rocks are not as depleted (compared with chondrites) in the lighter REE. Concentrations of Ba, Ga, Sc, and Hf in 14053 are near the top of the range found for Apollo 12 basalts; concentrations for Mn, Co, and Ni are near the bottom. Concentrations of those elements in 14072 fall nearer the middle of the ranges for Apollo 12 rocks.

REE concentrations in 14310 are the highest yet observed for a large fragment of basaltic rock. The relative abundances of the REE in 14310 resemble those for KREEP (3), but the REE concentrations are lower by factors of 0.5 to 0.2. The relative REE abundances in 14310 resemble those in fines sample 14163, but the concentrations in 14310 are lower, ranging from 0.8 for La to 0.95 for Yb. Concentrations for most of the other trace elements in 14310 more closely resemble those of 14163 than those of the other igneous rocks. The major element concentrations in 14310 and 14163 are quite similar also. The concentration ratios for several elements between 14310 and 14163 are as follows: Cr, Cs, Hf, Rb, Sc, 0.8 to 0.9; Co, Ga, Ni, 0.46-0.60; Zn, < 0.2. Does 14310 represent fused soil similar to 14163, or does 14163 represent the product of crushing of rocks similar to 14163? The similarity of Ga abundance to those of Co and Ni instead of those of Cs-Sc is hard to rationalize in terms of an igneous process. Co, Ni, and Ga can be added to lunar soils by infalling meteorites (4). The relative deficiency of Co, Ga, Ni, and especially Zn in 14310 argue against an origin for that rock by fusion of material similar to 14163.

The relationship of 14310 (if any) to 14053 and 14072 is unclear. The point for 14310 on the Sm/Eu versus Sm graph extends the line drawn through the points for Apollo 11 and 12 basalts to higher values. The concentrations of Hf, K, and Ni are greater, and those of Co, Cr, Mn, and Sc are less in
14310 than in 14053 and 14072. The trace element concentrations, in light of the mineralogy (2) do not suggest any simple genetic relationship between these rocks.

Table 1. Trace element concentrations in ppm.

<table>
<thead>
<tr>
<th></th>
<th>14053*</th>
<th>14072</th>
<th>14310</th>
<th>14163</th>
<th>breccia, fines</th>
<th>14063*</th>
<th>14063</th>
<th>14063</th>
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<tbody>
<tr>
<td>La</td>
<td>12.8±.2</td>
<td>6.76</td>
<td>57</td>
<td>70</td>
<td>19.4±.3</td>
<td>15.9</td>
<td>2.56</td>
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</tr>
<tr>
<td>Ce</td>
<td>36.4±.8</td>
<td>17.9</td>
<td>135</td>
<td>157</td>
<td>47±1</td>
<td>49</td>
<td>5.0</td>
<td></td>
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<tr>
<td>Nd</td>
<td>22±1</td>
<td>12.7</td>
<td>93</td>
<td>101</td>
<td>36±8</td>
<td>----</td>
<td>----</td>
<td></td>
</tr>
<tr>
<td>Sm</td>
<td>6.50±.06</td>
<td>3.93</td>
<td>25.6</td>
<td>30.8</td>
<td>9.2±.1</td>
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<td>1.06</td>
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<tr>
<td>Eu</td>
<td>1.23±.03</td>
<td>0.88</td>
<td>2.08</td>
<td>2.57</td>
<td>2.56±.02</td>
<td>2.89</td>
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<tr>
<td>Gd</td>
<td>8.5±.2</td>
<td>4.2</td>
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<td>36</td>
<td>12±2</td>
<td>8.3</td>
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<tr>
<td>Tb</td>
<td>1.62±.08</td>
<td>0.98</td>
<td>5.3</td>
<td>6.4</td>
<td>2.0±.2</td>
<td>1.49</td>
<td>0.22</td>
<td></td>
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<tr>
<td>Dy</td>
<td>1.11±.4</td>
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<td>56</td>
<td>45</td>
<td>12.0±.3</td>
<td>8.3</td>
<td>1.52</td>
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<tr>
<td>Ho</td>
<td>2.1±.2</td>
<td>1.5</td>
<td>6.7</td>
<td>8.6</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td></td>
</tr>
<tr>
<td>Er</td>
<td>----</td>
<td>3.5</td>
<td>20</td>
<td>25</td>
<td>7±3</td>
<td>----</td>
<td>----</td>
<td></td>
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<tr>
<td>Yb</td>
<td>6.1±.3</td>
<td>4.05</td>
<td>18.6</td>
<td>21.6</td>
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<td>4.7</td>
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<td>Lu</td>
<td>0.89±.01</td>
<td>0.61</td>
<td>12.4</td>
<td>3.16</td>
<td>0.199±.02</td>
<td>0.71</td>
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<tr>
<td>Co</td>
<td>25±1</td>
<td>32</td>
<td>16.0</td>
<td>27</td>
<td>21±3</td>
<td>15</td>
<td>----</td>
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<tr>
<td>Cr</td>
<td>2860±80</td>
<td>3880</td>
<td>1440</td>
<td>1570</td>
<td>----</td>
<td>----</td>
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<tr>
<td>Cs</td>
<td>0.218±.008</td>
<td>0.140</td>
<td>1.71</td>
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<tr>
<td>Ga</td>
<td>4.8±.2</td>
<td>3.8</td>
<td>4.3</td>
<td>7.5</td>
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<td>----</td>
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<tr>
<td>Hf</td>
<td>9.8±.4</td>
<td>6.6</td>
<td>32</td>
<td>39</td>
<td>17±2</td>
<td>20</td>
<td>0.6</td>
<td></td>
</tr>
<tr>
<td>Mn</td>
<td>1360±20</td>
<td>1840</td>
<td>705</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td></td>
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<tr>
<td>Ni</td>
<td>14±3</td>
<td>31</td>
<td>150</td>
<td>330</td>
<td>----</td>
<td>----</td>
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<td></td>
</tr>
<tr>
<td>Rb</td>
<td>3.2±.3</td>
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<td>23</td>
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<td>----</td>
<td></td>
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<tr>
<td>Sc</td>
<td>55±1.5</td>
<td>47</td>
<td>19</td>
<td>21</td>
<td>14.2±.5</td>
<td>12.6</td>
<td>4.0</td>
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<tr>
<td>Zn</td>
<td>3.4±.3</td>
<td>8</td>
<td>&lt;7</td>
<td>34</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td></td>
</tr>
</tbody>
</table>

*Reported uncertainties are based mainly on counting statistics. When converted to percents, the uncertainties given for 14053 correspond approximately to those for 14072, 14310, and 14163. Similarly, the uncertainties listed for 14063, <200 mesh may be used for the other materials from 14063.

+ Relative values only.

Six small fragments of igneous rock (total weight 3.55 mg) were separated from breccia 14063. The relative REE abundances in these fragments are enriched in the light REE over the heavy REE (relative to chondrites) by a factor of 2, similar to 14310. Their REE concentrations are at about the level of those in 14053. Unlike these other rocks, the igneous fragments from 14063 have a slight positive Eu anomaly. Concentrations of Co and Sc in these fragments are closer to those of 14310 than to those of 14053 or 14072; the concentration of Hf lies below the range for the other rocks. The point for these fragments on the Sm/Eu versus Sm diagram falls well below the line for the points for Apollo 11 and 12 rocks. It is risky to interpret data based on such small samples. We suggest that the fragments might represent feldspar-rich cumulates, but are more likely representatives of a new "strain" of lunar igneous rocks.
Similar small samples of igneous rocks and fines from the Luna-16 mission were analyzed (5). The relative REE abundances of the REE in the rock fragments and the fines were identical, and the REE concentrations in the fragments from the A level of the L-16 core were identical with those in rock 12038. The Eu depletion is small and the points for these fragments lie well below the line for Apollo 11 and 12 rocks on the Sm/Eu versus Sm diagram.

Perhaps the high concentrations of REE and the severe Eu depletions to which the rocks and fines from the Apollo missions have accustomed us are characteristics of rocks derived from sources in the Imbrium basin, rather than general characteristics of the lunar surface or even just of lunar mare. REE concentrations and extents of Eu depletions in lunar fines tend to decrease with increasing distance away from the Apollo 14 landing site.

Two fragments of plagioclase (2.88 mg) were picked by hand, and an additional sample of plagioclase (2.14 mg) was separated magnetically from 14063. The concentrations of REE in those samples fall within the range reported for lunar feldspars (6). If these plagioclase crystals are considered to have formed in equilibrium with a liquid whose composition matches that of the igneous fragments of 14063, solid/liquid distribution coefficients of 0.16 (La), 0.155 (Sm), 1.10 (Eu), and 0.15-0.18 (heavy REE) can be calculated. These values are high, but not outside the range reported in the literature (7). Some reported anorthosites must have come from liquids whose REE concentrations exceeded those of chondrites and which were enriched in REE relative to chondrites.

References:

(2) PET, Science 173 681 (1971).

The specific heats of Apollo 14 samples 14163,186 and 14321,153 have been measured between 90 and 350 K using an adiabatic calorimeter. The specific heats of both the soil and breccia increase monotonically from about 0.06 cal g⁻¹ K⁻¹ at 100 K to about 0.20 cal g⁻¹ K⁻¹ at 350 K. Smoothed values for the specific heat at 50 K intervals are as follows:

<table>
<thead>
<tr>
<th>Temperature K</th>
<th>Specific heat cal g⁻¹ K⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Fines 14163,186</td>
</tr>
<tr>
<td>100</td>
<td>0.065</td>
</tr>
<tr>
<td>150</td>
<td>0.106</td>
</tr>
<tr>
<td>200</td>
<td>0.139</td>
</tr>
<tr>
<td>250</td>
<td>0.163</td>
</tr>
<tr>
<td>300</td>
<td>0.184</td>
</tr>
<tr>
<td>350</td>
<td>(0.204)</td>
</tr>
</tbody>
</table>

The specific heats of the lunar soils from the Sea of Tranquillity, the Ocean of Storms, and Fra Mauro are quite similar being within 10 percent of one another at all temperatures between 100 and 400 K.
Isotopic Anomalies in Lunar Rhenium.

U. Herpers, W. Herr, H. Kulus and R. Michel
Institut fuer Kernchemie der Universitaet zu Koeln

In the course of an earlier attempt to prove the existence of "solar wind" Rhenium in lunar surface material and to search for a possibly "deficient" Re-187, which made have suffered a faster "bound state" \( \beta^- \)-decay (1), we observed in contrast a slight enhancement in Re-187 (2). To prove this and by reason of the extremely low Rhenium concentrations (igneous rocks < 0.1 ppb) neutron activation and counting techniques had to be improved. Suitable samples were subjected to an intense neutron bombardment, the Re was chemically separated and analyzed by \( \gamma \)-spectroscopy by a 25 cc Ge(Li)-detector. The Re-187/Re-185 ratio was checked by the 137 keV \( \gamma \)-line of Re-186 and the 155 keV line of Re-188. Lunar basaltic rocks gave the lowest Re-contents, ranging from 0.02 to 0.09 ppb, which roughly corresponds to achondrites. Whereas the soils were found to be richer in Re, 0.6 ppb for Apollo-12 and 0.95 - 1.4 ppb for 14-regolith, indicating a larger meteoritic contribution. The 14-breccias proved to range between rocks and soils.

The determination of the isotopic composition of Re showed, that Re-187 is remarkably enriched. An excess of 1.4 to 1.8% was measured in the 14-regoliths and was even higher in breccias up to 29%. The observed anomaly is explained by the reaction W-186(n,\( \gamma \))W-187 \( \beta^- \)Re-187. Consequently the W content were determined (table 1). The W/Re ratios are between 1100 (14-soils) and 41000 (breccia 14.321).

However, because the neutron-capture of W-186 does not only occur at the lunar surface, but also during pile activation, the measured excess of Re-187 had to be corrected for the reactor irradiation. This was done by calculation and by comparison of respective standards. The lunar share was found to be between 20 and 60% of the brutto excess of Re-187 (table 2). It is concluded that the lunar neutron capture of W-186 is mainly due to epithermal neutrons. The cross section has a pronounced maximum at 18.8 eV and is relatively large up to \( \sim \)100 eV. A contribution of W-184 (n,\( \gamma \)) to Re-185 is in the first approximation negligible. The possibility of deriving time integrated neutron fluxes and neutron spectra is discussed.
Additional analytical data on some noble elements, especially on Os/Re ratios are presented. These ratios were normally found near 1. They rise up to about 40 in the 14-soils. Extraordinary high Os-contents up to about 50 ppb were determined.

(1) Clayton, Nature 224, 56(1969)
Isotopic-Anomalies in Lunar Rhenium.
U. Herpers

<table>
<thead>
<tr>
<th>sample</th>
<th>type</th>
<th>W [ppm]</th>
<th>Re* [ppb]</th>
<th>ratio W:Re [x 10^3]</th>
</tr>
</thead>
<tbody>
<tr>
<td>14.163</td>
<td>a</td>
<td>1.32 ± 0.02</td>
<td>1.316 ± 0.004</td>
<td>1.1 ± 1.0</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>1.94 ± 0.03</td>
<td>0.957 ± 0.007</td>
<td>2.0 ± 2.0</td>
</tr>
<tr>
<td>259</td>
<td>a</td>
<td>1.55 ± 0.02</td>
<td>1.16 ± 0.04</td>
<td>1.6 ± 1.6</td>
</tr>
<tr>
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<td>b</td>
<td>1.85 ± 0.03</td>
<td>0.957 ± 0.007</td>
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</tr>
<tr>
<td>321</td>
<td>a</td>
<td>0.82 ± 0.01</td>
<td>0.957 ± 0.007</td>
<td>2.09 ± 2.09</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>0.95 ± 0.01</td>
<td>0.957 ± 0.007</td>
<td>2.09 ± 2.09</td>
</tr>
<tr>
<td></td>
<td>c</td>
<td>0.97 ± 0.01</td>
<td>0.957 ± 0.007</td>
<td>2.09 ± 2.09</td>
</tr>
<tr>
<td>.305</td>
<td>Breccia</td>
<td>1.74 ± 0.03</td>
<td>0.020 ± 0.003</td>
<td>2.0 ± 2.0</td>
</tr>
<tr>
<td>.305</td>
<td>Breccia</td>
<td>2.39 ± 0.03</td>
<td>0.083 ± 0.003</td>
<td>2.9 ± 2.9</td>
</tr>
<tr>
<td>12 053</td>
<td>rock</td>
<td>0.28 ± 0.01</td>
<td>0.078 ± 0.007</td>
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</tr>
<tr>
<td>021</td>
<td>rock</td>
<td>0.12 ± 0.01</td>
<td>0.032 ± 0.002</td>
<td>0.4 ± 0.4</td>
</tr>
</tbody>
</table>

* Re contents based on $^{185}$Re ($n,γ$) $^{186}$Re (T=906 h)
** given by Wanke et al (1971)

Table 1

Contribution of the Reactor Irradiation to the $^{187}$Re - Anomaly

<table>
<thead>
<tr>
<th>Apollo-14 samples</th>
<th>$Δ^{187}$Re brutto [10^{-15}g]</th>
<th>$Δ^{187}$Re reactor [10^{-15}g]</th>
<th>$Δ^{187}$Re lunar [10^{-12}g]</th>
<th>$^{186}$W</th>
<th>$Δ^{187}$Re (Lun) *</th>
<th>$^{186}$W</th>
</tr>
</thead>
<tbody>
<tr>
<td>14.163</td>
<td>3.5</td>
<td>0.94</td>
<td>2.6 ± 0.3</td>
<td>18.8</td>
<td>0.14 ± 0.01</td>
<td>0.027 ± 0.01</td>
</tr>
<tr>
<td></td>
<td>1.9</td>
<td>1.21</td>
<td>0.7 ± 0.3</td>
<td>27.6</td>
<td>0.12 ± 0.01</td>
<td>0.027 ± 0.01</td>
</tr>
<tr>
<td>259</td>
<td>3.8</td>
<td>1.13</td>
<td>2.6 ± 1.4</td>
<td>22.2</td>
<td>0.12 ± 0.01</td>
<td>0.027 ± 0.01</td>
</tr>
<tr>
<td>321</td>
<td>3.9</td>
<td>3.18</td>
<td>0.7 ± 0.09</td>
<td>10.7</td>
<td>0.071 ± 0.007</td>
<td>0.001 ± 0.006</td>
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<td>5.1</td>
<td>4.13</td>
<td>1.0 ± 0.09</td>
<td>12.4</td>
<td>0.091 ± 0.006</td>
<td>0.001 ± 0.006</td>
</tr>
<tr>
<td>.305</td>
<td>Breccia</td>
<td>11.9</td>
<td>7.39</td>
<td>4.5 ± 0.3</td>
<td>24.8</td>
<td>0.18 ± 0.01</td>
</tr>
<tr>
<td>.305</td>
<td>AEI soil</td>
<td>8.9</td>
<td>6.21</td>
<td>2.7 ± 0.2</td>
<td>33.9</td>
<td>0.081 ± 0.005</td>
</tr>
</tbody>
</table>

all data normalized to 50mg weight (indiv. weights 49 - 104mg)

* $Δ^{187}$Re (Lun) = $\int \frac{Δ^{187}}{185} \int dE$ * $\sigma(E) \cdot \Phi(E,t)$ dE dt
Study on the Cosmic Ray Produced Longlived Mn-53 in Apollo-14 Samples.

W. Herr, U. Herpers and R. Woelfle *
Institut fuer Kernchemie der Universitaet zu Koeln

For exposure ages, activity profiles of the lunar surface and variations of the cosmic flux, the comparison of short and longlived spallation radionuclides is exceedingly helpful. Especially one of the latter group nuclide Mn-53 deserved our interest. The K-emitting Mn-53, having a relatively high spallation yield, can be transformed by neutron capture to $\gamma$-emitting Mn-54. We first applied this technique to a larger number of iron meteorites, proving the usefulness and advantages of this technique even for mg quantities (1-3). Depth profiles of Mn-53 in lunar rocks were recently presented by SHRELLDALFF (4) and also by our group (5).

In the last decade, however, one difficulty in dating-applications etc. on the Mn-53 basis, remained, because of an uncertainty in the respective half life. Earlier figures range from $T = 1.9 \times 10^6$ y. Quite recently, new attempts were made in establishing the half value time, using different methods. Obviously, weight has the work of Honda et al. (6) succeeding by massspectrometric techniques on a man made sample, coming to a value $T = (3.7 \pm 0.4) \times 10^6$ y.

Three years ago, in view of the importance of this constant (2), we started independantly an experiment in order to determine the Mn-53 halflife via activation. The principle is: MnO$_2$ was extracted from the iron meteorite "Duchesne" and aliquots were neutron irradiated:

1. for a "short" period of 24 days in a $10^{14}$ n cm$^{-2}$ s$^{-1}$ flux
2. for a "long" period of 345 days, with a suitable Mn-54 monitor.

From the two describing activation equations the number of Mn-53 atoms are eliminated, thus allowing the evaluation of the cross section $\sigma_{53} = 70 \pm 10$ b.

The determination of the halflife was then possible by relating to the "Peace River" decay rate (best known value by J. R. Arnold; priv. comm.). So we resulted in $T_{53} = (3.8 \pm 0.7) \times 10^6$ y. Evidently this figure is in best agreement with Honda's value. The present work is based on our own figure.

*KFA Juelich.
The counting was very much improved by a 75 cc Ge(Li)-Well-type detector, made by J. Eberth, Institut of Nuclear Physics, Cologne (resolution: 3.8 keV at 1.332 MeV; efficiency: 1.4% at the Mn-54 835 KeV line; background: 0.30 cpm under the total line).

The Mn-53 values related to iron are given in the table. The 11-14 soil samples are all in the range of 340-380 dpm/kg Fe, with the only exception of the "comprehensive soil" 14.259 (~1 cm from surface), being higher by a factor of nearly 2. Evidently this is due to a contribution of solar protons. Remarkable is the relatively low Mn-53 content of the mikrobreccia 14.305 BDI (interior part). This could mean that the exposure age is surprisingly short only about 6 x 10^6 y. Further we redetermined sections from rock 10.017. The Mn-53 depth-dependance is demonstrated by the diagram. We were also able to check an aliquot of the "Tokyo - Standard", kindly provided by Prof. M. Honda. Evidently best agreement is reached.

1.) Herpers, U., W. Herr and R. Woelfle, Radioactive Dating and Methods of Low-level Counting, 199, IAEA, Vienna (1967)
**53Mn-Content of Lunar Material**

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Type</th>
<th>Sample Weight</th>
<th>$^{54}$Mn (total)</th>
<th>Contrib of $(n,2n)$ plus $(n,p)$</th>
<th>$^{53}$Mn</th>
<th>Fe</th>
<th>$^{53}$Mn</th>
</tr>
</thead>
<tbody>
<tr>
<td>14,163</td>
<td>soil</td>
<td>248</td>
<td>383 ± 0.016</td>
<td>27.8</td>
<td>26.9 ± 1.1</td>
<td>7.5</td>
<td>359 ± 54</td>
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<td>14,250</td>
<td></td>
<td>231</td>
<td>662 ± 0.019</td>
<td>17.6</td>
<td>49 ± 7.5</td>
<td>7.7</td>
<td>630 ± 96</td>
</tr>
<tr>
<td>12,070</td>
<td></td>
<td>501</td>
<td>14.1 ± 0.2</td>
<td>25.8</td>
<td>49 ± 3.5</td>
<td>13.2</td>
<td>380 ± 57</td>
</tr>
<tr>
<td>10,084</td>
<td></td>
<td>990</td>
<td>17.7 ± 3.2</td>
<td>30.2</td>
<td>41 ± 7</td>
<td>12.0</td>
<td>341 ± 58</td>
</tr>
<tr>
<td>14,305 BD1 (interior)</td>
<td>brecc.</td>
<td>504</td>
<td>375 ± 0.015</td>
<td>48.4</td>
<td>12.9 ± 1.9</td>
<td>7.9</td>
<td>163 ± 24</td>
</tr>
<tr>
<td>12,021</td>
<td>rock</td>
<td>842</td>
<td>25.4 ± 0.3</td>
<td>38.8</td>
<td>33 ± 4</td>
<td>14.2</td>
<td>232 ± 20</td>
</tr>
<tr>
<td>0.022</td>
<td></td>
<td>1038</td>
<td>357 ± 0.4</td>
<td>36.6</td>
<td>38 ± 4</td>
<td>16.4</td>
<td>232 ± 24</td>
</tr>
<tr>
<td>0.053</td>
<td></td>
<td>1070</td>
<td>41.2 ± 0.4</td>
<td>34.1</td>
<td>41 ± 3</td>
<td>15.3</td>
<td>268 ± 20</td>
</tr>
<tr>
<td></td>
<td>0.5nHCl solut.</td>
<td>1000</td>
<td>16.89 ± 0.39</td>
<td>0.26</td>
<td>0.39 ± 0.059</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*) $^{53}$Mn = "Standard" - Solution supplied by courtesy of Prof. M. Honda with a value of 0.395 dpm/g solution (priv. comm.)

**53Mn-Values in Relation to the Depth**

![Diagram showing Mn values in relation to depth](image-url)
Inert gas contents of size fractions of 14259 and 14163 can be summarized in equations of the type $X = A/d + B$, where $X$ = gas content in $10^{-8}$ cm$^3$ STP/g; $d$ is mean particle diameter of size fraction in microns.

<table>
<thead>
<tr>
<th>Sample</th>
<th>He$^4$</th>
<th>Ne$^{20}$</th>
<th>Ar$^{36}$</th>
<th>Kr$^{84}$</th>
<th>Xe$^{132}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>14259</td>
<td>4.02</td>
<td>6.05</td>
<td>2.35</td>
<td>9.0</td>
<td>1.7</td>
</tr>
<tr>
<td>14163</td>
<td>0.74</td>
<td>1.70</td>
<td>0.67</td>
<td>2.9</td>
<td>0.70</td>
</tr>
<tr>
<td></td>
<td>$(10^6)$</td>
<td>$(10^4)$</td>
<td>$(10^2)$</td>
<td>$(10^2)$</td>
<td>$(10^2)$</td>
</tr>
</tbody>
</table>

The gas contents of bulk samples of 14003, 14259, and 14163 are very similar: He$^4 \approx 3000$; He$^4 \approx 8,000,000$; Ne$^{20} \approx 150,000$; Ar$^{36} \approx 40,000$; Kr$^{84} \approx 20$; Xe$^{132} \approx 4$; all in $10^{-8}$ cm$^3$ STP/g. Single fragments from 14166 and 14167 have gas contents from one to two orders of magnitude less. From a plot of He$^4$ vs Ne$^{20}$ we conclude that nearly all of the samples contain $(3 \pm 1) \times 10^{-3}$ cm$^3$ STP/g of radiogenic He$^4$. The average He$^3$ content of the 14259 size fractions is $260 \times 10^{-8}$ cm$^3$ STP/g, that of 14163 is $270 \times 10^{-8}$ cm$^3$ STP/g. The average Ne$^{21}$ contents are $47 \times 10^{-8}$ cm$^3$ STP/g (14259) and $64 \times 10^{-8}$ cm$^3$ STP/g. The following equations represent Ar$^{40}$ vs Ar$^{36}$ in size fractions:

14259: $\text{Ar}^{40} = 0.81 \text{Ar}^{36} + 1.87 \times 10^{-4}$ cm$^3$ STP/g
14163: $\text{Ar}^{40} = 1.87 \text{Ar}^{36} + 2.21 \times 10^{-4}$ cm$^3$ STP/g

It is interesting to note that, although the slopes of the two lines differ considerably, the Ar$^{40}$-intercepts agree within $\pm 7\%$. Also, the intercept values of about $2 \times 10^{-4}$ cm$^3$ STP/g are nearly three times larger than that of 10084 ($0.71 \times 10^{-4}$ cm$^3$ STP/g). According to the LSPET Apollo 14 report, the K contents of Apollo 14 fines, 4000-5000 ppm are about three to four times larger than 10084 (1200 ppm). Hence we conclude that all, or nearly all of the Ar$^{40}$ on the intercept represents in situ decay of K$^{40}$.

Now that Ar data from fines in four locations on the moon
are before us, we have learned that the $\text{Ar}_{40}/\text{Ar}_{36}$ ratio in bulk fines as well as from internally determined $40/36$ slopes (from size fractions) vary not only from site to site, but within a given landing area itself. We have learned that many fines, but not all, contain parentless $\text{Ar}_{40}$ (defined here as $\text{Ar}_{40}$ that is correlated with solar wind $\text{Ar}_{36}$). The ratio total $\text{Ar}_{40}$/total $\text{Ar}_{36}$ as determined in bulk samples is nearly always greater than the $40/36$ slope from size fractions because of the positive $\text{Ar}_{40}$ intercept. However, it appears reasonable to correct the bulk $\text{Ar}_{40}$ measurements reported in the literature by $-0.6 \times 10^{-4}$ cm$^3$ STP/g per 1000 ppm of K. When this is done (inert gas data from Funkhouser et al, Heymann and Yaniv, Hintenberger et al, Hohenberg et al, Marti et al Apollo 11 Lunar Science Conference Proceedings; also Hintenberger et al, Funkhouser et al, Vinogradov Apollo 12 Lunar Science Conference Proceedings) we find:

<table>
<thead>
<tr>
<th>Samples</th>
<th>Range</th>
<th>Average (Measurement)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Apollo 11</td>
<td>0.7 - 1.4</td>
<td>1.1 (9)</td>
</tr>
<tr>
<td>Apollo 12</td>
<td>0.3 - 0.5</td>
<td>0.35 (8)</td>
</tr>
<tr>
<td>Apollo 14</td>
<td>0.8 - 1.9</td>
<td>1.3 (3)</td>
</tr>
<tr>
<td>Luna 16</td>
<td></td>
<td>0.65 (1)</td>
</tr>
</tbody>
</table>

(Note: the 12032 and 12033 samples which we have measured do not contain parentless $\text{Ar}_{40}$). There is an interesting correlation between average $\text{Ar}_{40}/\text{Ar}_{36}$ slopes and the ages of the rocks: the Apollo 11 site has the youngest ages (about 3.3 b.y.) and the smallest average (0.35); the Apollo 14 site has the oldest ages (about 3.9 - 4.0 b.y.) and the greatest average (1.3). The Apollo 12 values fall in between. At first sight the correlation appears to be purely accidental; however, if one assumes that the rock ages represent the earliest time at which atmospheric $\text{Ar}_{40}$ could become implanted in the regolith the correlation suggests that the average concentration of neutral $\text{Ar}_{40}$ in the lunar atmosphere decreased rapidly from 4-3 b.y. ago; in fact more rapidly than the decay of $\text{K}_{40}$ in the Moon itself. We suggest the following working hypothesis. Following an igneous extrusion on the surface, a substantial regolith evolved in each location on a time scale of a few hundred million years or less. A record of the average $\text{Ar}_{40}$ concentration in the lunar atmosphere is preserved in the fines for different periods in the 3-4 b.y. past era. The regolith need not be well mixed vertically, but the present top of the regolith must be substantially "contaminated" with particles that were exposed to solar wind early in the history of the regolith. Mixing of the regolith from different
regions such as Tranquillity, Procellarum, and Fra Mauro has always been poor. The content of neutral atmospheric Ar\textsuperscript{40} decreased because of K\textsuperscript{40} decay, cooling of the Moon's near-surface regions and perhaps because of a decrease in frequency and intensity of near-surface igneous activity. The significance of 12032 and 12033, which contain little if any parentless Ar\textsuperscript{40} is that these materials appear to have arrived in their present location less than 1 b.y. ago, judging from their apparent K-Ar ages (both samples are slightly contaminated with Ar\textsuperscript{40} rich fines of the 12070 type, but corrections can be made). The variations of 40/36 ratios within a given landing site (e.g. 14259 vs 14163) could be due to variations of local slopes of the terrain relative to the path of the Ar\textsuperscript{40} ions at impact, to differences in average trapping efficiencies of two regolith materials, to the degree of vertical mixing (contamination), secular variations in the energy distribution of Ar\textsuperscript{40} ions impinging on the regolith. We have attempted to establish correlations between 40/36 values and slopes of the terrain as taken from contour maps. No clear-cut correlation is apparent.
GRAIN-SIZE AND SHAPE CHARACTERISTICS OF LUNAR FINES AND SOME TERRESTRIAL COMPARISONS.

H. Heywood  Chemical Engineering Department, University of Technology, Loughborough, Leics., England.

The grain size distribution and shape characteristics have been determined on samples 10 084, 171 and 12 057, 72 returned by Apollo 11 and 12 missions respectively. Some of the characteristics of the latter sample have been described in a previous report, which also explains the procedures adopted for the sizing analyses (1). The only difference in the present procedure is that sieve plates with round hole apertures made by electro-forming were used over the complete size range of 20 to 1000 μm.

### Abbreviated sieving analysis

<table>
<thead>
<tr>
<th>Round hole diam.: μm</th>
<th>Per cent weight passing hole</th>
</tr>
</thead>
<tbody>
<tr>
<td>10 084</td>
<td>12 057</td>
</tr>
<tr>
<td>500</td>
<td>94</td>
</tr>
<tr>
<td>200</td>
<td>82</td>
</tr>
<tr>
<td>100</td>
<td>67</td>
</tr>
<tr>
<td>50</td>
<td>45</td>
</tr>
<tr>
<td>20</td>
<td>17</td>
</tr>
</tbody>
</table>

Arith. mean. aperture, μm 127 140

### Shape characteristics

The theory for defining numerical shape factors has been described in the previous report and may be summarised as follows:-

The orthogonal limiting dimensions of a particle are L, B and T in decreasing order of magnitude. The elongation ratio of the profile of a particle is \( n = L/B \). If \( a \) is the projected area of the profile in the plane of L and B, then a two-dimensional shape factor is the Projected area ratio \( \sigma_a = a/BL \) as
Grain-size and shape characteristics
H. Heywood

shown in the adjacent figure. A three dimensional shape factor is the volume coefficient \( a_v,a = \frac{\text{volume of particles}}{(\text{projected area diameter})^3} \). The projected area diameter is the diameter of the circle which encloses the same area as the profile of the particle, namely \( d_a = \frac{(4 a/n)^{1/2}}{2} \).

The sieving procedure used for the determination of grain-size composition produced about 16 closely graded fractions of particles. Determinations of the shape factors \( n \) and \( a_a \) were made on optical microscope photographs of each of these fractions. A summary of these measurements is given below:

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Mean value for all fractions</th>
<th>Range of mean values for individual fractions</th>
</tr>
</thead>
<tbody>
<tr>
<td>10 084,171</td>
<td>( n = 1.386 )</td>
<td>1.33 to 1.42</td>
</tr>
<tr>
<td>12 057,72</td>
<td>( n = 1.376 )</td>
<td>1.32 to 1.41</td>
</tr>
<tr>
<td>10 084,171</td>
<td>( a_a = 0.78 )</td>
<td>0.74 to 0.83</td>
</tr>
<tr>
<td>12 057,72</td>
<td>( a_a = 0.80 )</td>
<td>0.75 to 0.84</td>
</tr>
</tbody>
</table>

In general, the finer particles below 100 \( \mu \text{m} \) are the most elongated and tend to have a more nearly prismoidal shape, but this variation does not occur uniformly over the complete size range. There are anomalies that indicate variations in the mineralogical composition of the different size fractions but further work is necessary to show whether these variations are random or are significant features in the composition of the lunar fines.

The mean particle volume was measured for the size fractions exceeding 100 \( \mu \text{m} \) particle diameter and the following were the mean values of the volume coefficient.

Sample 10 084,171 \( a_v,a = 0.25 \) Sample 12 057,72 \( a_v,a = 0.22 \)

The general conclusions are that these two samples of lunar fines are very similar, both as regards grain-size composition and particle shape. These shape measurements are being extended by electron microscope photography to particles smaller than 5 microns.

The relative number of glass spherules in the samples is a matter of interest: a precise count has not yet been made but they are present to the number of approximately 10 to 20 per 1000 other particles of the same size in the material larger than 50 \( \mu \text{m} \). However, in the size fractions smaller than 2 \( \mu \text{m} \) there appears to be an increase in the relative number of spheres, and the adjacent photograph shows a remarkable cluster of
Grain-size and shape characteristics  
H. Heywood

spherules at 60 K magnification
in the size range 0.1 to 0.2 μm.

Comparative studies on terrestrial particles

Measurements of shape factors are being made on crushed particles of terrestrial minerals similar to those found in the lunar fines. Preliminary examination has shown that sharply angular fragments are much more common in the terrestrial minerals than in the lunar fines. A study of the mode of formation of glass agglomerates and glass spherules may be promoted by examination of the fused ash particles produced by the combustion of powdered coal. There is an extensive literature on the break-down of liquid drops and a study of these may elucidate the mechanism of spherule and 'dumb-bell' formation in the lunar fines.

Reference

LUNAR ORBITAL MASS SPECTROMETER EXPERIMENT   J. H. Hoffman,  
R. R. Hodges, The University of Texas at Dallas, Dallas, Texas,  
75230, and D. E. Evans, Manned Spacecraft Center, Houston, Texas, 77058

A magnetic sector field mass spectrometer was flown on Apollo 15 as part of the Scientific Instrument Module (SIM) bay experiment complement to measure the composition and number density of the neutral lunar atmosphere. The instrument was mounted on a 7.3 meter bi-stem boom extended from the SIM bay and operated by a crew member from the Command Module during lunar orbit and trans-earth coast (TEC). Data from the experiment are applicable to several areas of lunar studies, namely, the understanding of the origins, transport mechanisms and sinks of the atmosphere. Light gases, such as hydrogen, helium and neon probably originate from neutralization of solar wind ions, whereas molecular gases, such as carbon dioxide, carbon monoxide, hydrogen sulfide, ammonia and water vapor may be produced by lunar volcanism.

Preliminary results from the Apollo 15 experiment indicate a large number of gas molecules of many different species exist in the vicinity of the spacecraft in lunar orbit. Most of these gases appear to be of spacecraft origin with levels of several (H$_2$O, N$_2$ and CO$_2$) exceeding the upper bound on native lunar gases by one to two orders of magnitude. This gas cloud seems to be moving with the CSM, since its measured density is largely independent of the angle of attack of the mass spectrometer plenum. Many interesting variations of this gas cloud are observed as a function of orbital parameters. However, during TEC the amplitudes of all peaks in the spectra decreased by a factor of five to ten and a boom retraction test showed no increase in gas densities down to a distance of 1.25 meters from the CSM surface. The outgassed molecules appear to be a collisionless, outwardly streaming gas beyond this point. The clear implication of this result is that the higher levels of contaminant gases in lunar orbit was due to orbital mechanics. However the mechanism of production of a stable gas cloud about an orbiting body is obscure, because of the low gas temperature required to maintain the detected level.
CARBON COMPOUNDS IN APOLLO 12, 14 AND 15 SAMPLES

P. T. Holland, B. R. Simonait, P. G. Mszolek, W. H. McFadden and A. L. Burlingame, Space Sciences Laboratory, University of California, Berkeley, CA. 94720

The gases released on the desorption of a variety of samples have been analyzed by gas chromatography and high resolution mass spectrometry. Initially, the Apollo 12 samples were analyzed for various 14C isotopes, 20Ne and 36Ar are released as well as the normal C, H, and higher deuterated hydrocarbons. The yields correlate with the solar carbon content of the samples and the C, H, and 14C released may be regarded as indigenous while the deuterated products result from hydrolysis of carbide material.

3.2 grams of the 14240 SESC sample were sieved and repackaged in the UCSSL clean room dry-box facility. A size distribution was obtained very similar to that reported for the 14149 bottom-trench sample confirming that these two samples are very similar. Dissolutions were performed on the five size fractions in the range 120 to 320. The yields of CH4, C2H4, 20Ne and 36Ar are plotted in Fig. 1 against the inverse mean radius. It is seen that the yields of these products correlate with the surface area and provide a direct confirmation that the CH4, like the 20Ne and 36Ar is surface correlated and therefore probably arises from interaction of solar wind implanted carbon and hydrogen (5). The C2H4 yields roughly correlate with the square of the surface area supporting this model for bond formation in the finer fractions. The observed correlation of the CH4 yields with grain size also indicates that the solar wind is promoting carbide formation on the surface of the lunar fines. Other deuterocarbons released include C2D4, C2D6, C3D8, C3D10.

Evaporation of these size differentiated samples confirmed the common site and surface correlation of the CH4, 20Ne and 36Ar. These gases show very similar evolution curves with most of the gas being released in the 400-900°C range.

Dissolution and pyrolysis of the 14148, 14156, and 14149 trench samples showed that their carbon chemistry and solar wind exposure were very similar to that of the 14240 SESC and Apollo 11 and 12 fines-of-high-carbon-content. CH4 and 36Ar yields of around 100 ng/g and 200 ng/g respectively were observed. There was some decrease in solar wind derived products with depth in the trench. Preliminary examination of the breccias 14067 and 14069 shows that these samples are high in carbon and that their carbon chemistry and solar wind exposure are similar to the fines samples. This is in agreement with
Carbon Compounds in Apollo 12, 14 and 15 Samples

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fusion of fines at moderate temperatures as a model for breccia formation.

In contrast the complex fragmental rock 14066 releases no CH$_4$, $^{20}$Ne or $^{36}$Ar on pyrolysis or dissolution although the total carbon content is 90 ppm (2). This indicates that during its formation, shock and heating have released all the solar wind rare gases but leaving at least some of the carbon or that the rock was formed from material of low solar wind content but higher in carbon than the basaltic lunar material thus far examined. The carbon present shows similar release patterns on pyrolysis to fines material with over 80% being released as CO above 900°C.

Other interesting components released from the soil samples by DF include D$_2$S, DCN and CS$_2$. The D$_2$S yields correspond to sulphur contents of 150-300 µg/g and indicate that the principal form of sulphur is as metal sulphides, probably Troilite (FeS$_2$). The carbon released as protobiologically important DCN and CS$_2$ amounts to about 0.3 µg/g and 0.1 µg/g respectively and are probably also formed by solar wind interactions.

Pyrolysis experiments have also shown that HCN and CS$_2$ are released. These experiments also confirmed that the bulk of the carbon is released as CO when temperatures above 1000°C are attained and that the bulk of the CO$_2$ and H$_2$O released, particularly below 400°C is contamination. Nitrogen is mainly released above 800°C and indigenous contents are about 100 µg/g. The origin of benzene and other hydrocarbons observed during pyrolysis will be discussed in reference to labelling experiments and preliminary results on Apollo 15 soil samples will be presented.


(2) Preliminary Examination of Lunar Samples from Apollo 14 LSPET, Science, 173, 681 (1971).
Carbon Compounds in Apollo 12, 14 and 15 Samples

P. T. Holland

Figure 1

$\text{CD}_4$, $\text{CH}_4$, $^{20}\text{Ne}$ and $^{36}\text{Ar}$ yields from sample 14240 as a function of inverse grain radius.
Apollo 14 sample 14310,21 is a fine-grained intersertal basalt containing 54.3 modal% plagioclase, 36.8% pyroxene (approximately 23.8% augite, 11.4% hypersthene, and 1.6% pigeonite), 3.4% opaque (ilmenite and ulvospinel), and 5.5% residual phases (mostly glass).

It appears to be the first basaltic sample retrieved from the moon which preserves mineral phases, or even chemical characteristics of a given mineral phase, indicative of a prior, pre-melt history. We have identified mineral fragments which have come from two or three distinctly different rock types, possibly indicating a heterogeneous source region for the magma represented by rock 14310. The identification of these pre-melt mineral fragments is based on the distinction between normal and anomalous zoning profiles of pyroxenes.

In grossly simplified terms, the pyroxene crystallization sequence is from magnesian hypersthene (Wo4En79Fs17) to pigeonite to subcalcic augite. This sequence is more "primitive" (earliest formed pyroxene lower in Fe) than the hypersthene to pigeonite to augite to pyroxferroite sequence described by us in the only Apollo 12 sample with hypersthene cores, sample 12065; it is useful, therefore, to describe sample 14310 in comparison to 12065. A major difference between the two samples is that pigeonite occurs between hypersthene and augite in only a few grains in 14310, the hypersthene being overgrown directly by augite, or pigeonite only, in most grains. In 14310, the first augite to grow on hypersthene is more calcic (Wo37En42Fs21) than that on pigeonite (Wo30En45Fs25), the latter being similar to the augite in contact with pigeonite cores in pyroxene phenocrysts of sample 12065. The hypersthene in 14310 zones continuously from Wo4En79Fs17 to the unusually calcic composition of Wo11En58Fs31, adjacent to the augite; that in contact with the pigeonite is less calcic (Wo4En69Fs26.5). The composition of the hypersthene cores in the pigeonite phenocrysts of sample 12065 are almost identical (Wo5En68Fs27), but the pigeonite is
more iron-rich in 14310: Wo7En61Fs32 compared to Wo7En66Fs27. The most Fe-rich typical augites and pigeonites in 14310 have the compositions Wo34En28Fs38 and Wo11En41Fs48, respectively. The zoning and overgrowth effects are interpreted to be the result of rapid growth of pyroxenes in a closed system which was enriched in Mg relative to Ca, Al, and Fe as compared with most of the Apollo 11 and 12 basalts.

An anomalous zoning feature of the pyroxenes is the occurrence of a polysynthetically twinned pigeonite grain which zones from a more iron-rich core (Wo8En54Fs38) to a magnesian rim (Wo6En68Fs26) and is overgrown by magnesian hypersthene (Wo4En75Fs21). This grain and others like it definitely reflect a pre-melt history because no crystals which have grown stably in a melt have ever been reported which increase in Mg/Fe ratio from core to rim.

A second anomalous feature, and one of major petrogenetic significance, is a relatively more homogeneous core of hypersthene within the hypersthene described above which has a slightly more iron-rich composition (Wo4En74Fs22) than the adjacent part (Wo4En79Fs17) of the zoned hypersthene which has overgrown it. When together, the two hypersthenes are distinguishable, in some cases, by different optic orientations; others are in perfect optical continuity. The boundary between the two is irregular and is suggestive of resorption. In one core-hypersthene grain there are blebs of coexisting high-Al, high-Ti, moderately high-Na calcic pyroxene (approximately Wo40En47Fs13, with the total composition: Na2O, 0.14; MgO, 15.8; Al2O3, 7.9; SiO2, 46.7; CaO, 15.7; TiO2, 3.4; Cr2O3, 0.6; MnO, 0.16; FeO, 7.6). A second hypersthene core, similar in composition to the other, has contained in it only plagioclase blebs, rather than high-Al clinopyroxene.

Because (1) the core-hypersthenes are not always in optical continuity with the rim-hypersthenes, (2) the contact between them is irregular, and (3) the core-hypersthene is comparatively homogeneous and can have coexisting blebs of augite, we suggest that there were, during crystallization of rock 14310, at least two different structural varieties of hypersthenes.

The compositions of the coexisting hypersthene and augite in the core-hypersthene are similar to discrete grains of a magnesian hypersthene (Wo4.5En77Fs18.5) and average clinopyroxene (approx. Wo39En49Fs13) from breccia sample 14063. Both pairs may belong to the same family of tie-lines as found in Apollo 15
rock 15415 ("genesis rock") which contains hypersthene with the composition of Wo$_{2.5}$En$_{58}$Fs$_{39.5}$ and augite with the composition Wo$_{46}$En$_{39}$Fs$_{16}$. This latter tie-line is of genetic significance in that it crosses the hypersthene-augite tie-line defined by the overgrowth relationship in 14310, and it defines a wider solvus, implying lower temperatures of equilibration for rock 15415. The three hypersthene-augite pairs define an unusually wide (for the moon, so far) solvus but are remarkably similar in position to hypersthene-augite tie-lines described for plutonic rocks associated with terrestrial anorthosites.

We suggest that rock 15415 and the hypersthene and coexisting augite in the core of the pyroxenes of rock 14310 and in the mineral fragments of breccia 14063 are samples of plutonic rocks which form a part of the lunar crust. There are important differences, however, between the blebs of augite in the cores of the pyroxenes of sample 14310, the augite in 14063, and the augite in 15415. The augite blebs in 14310 are extremely enriched in Al$_2$O$_3$ and TiO$_2$; they may be the most enriched in Al$_2$O$_3$ yet reported from the moon. The augite in 15415 is unusually low in these components, and the augite in 14063 has a distinctly different relative proportion of Al$_2$O$_3$, TiO$_2$, and Cr$_2$O$_3$. These results suggest the plutonic terrain of the moon is chemically heterogeneous.

We continue to find permissive evidence that the mare basalts were generated near the lunar surface; we have not found any unambiguous evidence of crystallization in the lunar interior, or incorporation of definite xenocrysts from the lunar interior in the Apollo 11, 12, 14, 15 and Luna 16 samples examined by us, with the possible exception of the core-hypersthenes in sample 14310. A model consistent with all geochemical, age, and petrologic information, including comparison with terrestrial equivalents to the lunar rocks, is that the mare lavas are partial melts from the near-surface (<10km) of the moon and that the source material consists of layered intrusions with the bulk composition of gabbroic anorthosite, with possibly some additional, related ferrogranites. The most plausible mechanism for generating magmas from such a crust is by impact melting.

Heavy lines, 14310 trends
Light lines, 12065 trends
Dashes imply chemical discontinuities

Relative Concentrations of Al, Ti and Cr in Lunar Pyroxenes
The variations in relative content of Al, Ti, and Cr in the lunar pyroxenes can reflect independent variations in the concentrations of these components in the melt during crystallization. We have shown for the Apollo 12 pyroxene phenocrysts that an increase in the proportion of octahedrally coordinated Al from core to near the rim (Figure) is related to an increase of the concentration of Al in the melt prior to nucleation of plagioclase. With rapid growth of plagioclase, the content of Al in the melt dropped rapidly compared to Ti, and the new composition of the pyroxene (12065 Edges) reflected this change in melt composition. The relative content of the three components in the rims of the Apollo 12 phenocrysts is close to those of all analyses of Apollo 11 clinopyroxenes, including both rim and core analyses, although absolute amounts of each element vary widely within and between samples. This similarity is consistent with a conclusion that plagioclase and ilmenite were crystallizing simultaneously throughout all pyroxene growth in most Apollo 11 samples, whereas plagioclase, and, based on textural evidence, ilmenite, crystallized only during the last stages of pyroxene growth in the porphyritic Apollo 12 basalts.

The Al-Ti-Cr trend for most of the growth of the Apollo 12 pyroxenes lies nearly along a line of constant relative Ti content and is towards increasing relative Al content. In marked contrast, the trend of the pyroxenes in Apollo 14 sample 14310 (Figure) is away from the Al corner, towards the Ti-Cr edge. Most of the hypersthene trend lies along a line of nearly constant relative Cr content.

It is reasonable to suggest that plagioclase crystallized with most of the pyroxene and that ilmenite crystallized late in sample 14310 because (1) plagioclase is abundant (54.3 modal per cent compared to 22% in 12065 and 27% in 10058), and (2) plagioclase laths cut into most areas of the pyroxene whereas ilmenite appears texturally as an interstitial phase. We therefore conclude that the trend for sample 14310 shown in the Figure is that for pyroxene growing during an increasing concentration of Ti relative to Al in the melt, similar to the last stages of pyroxene growth in the Apollo 12 samples. When the concentration of Ti in the melt was sufficiently high to result in the crystallization of ilmenite, leading to an abrupt change in melt composition, the change in composition trend of the pyroxene was towards the Ti-Al edge of the diagram.
Al, Ti AND Cr IN LUNAR PYROXENES

L.S. Hollister

The third component in the figure, Cr, can also be an independent variable. Hence, different magmas with the same relative order of crystallization of the phases ilmenite, plagioclase, and pyroxene may give different trends or plot in a different field, depending on the initial relative concentrations of the three components in the melt. Thus, all analyses of pyroxene which crystallized with ilmenite and plagioclase would not be expected to lie in one small region of the Figure, and, indeed, there are systematic differences. In order of increasing Cr with respect to Ti are Luna 16 sample G36, Apollo 11 samples 10058 and 10062, Apollo 12 sample 12065 Edges, Apollo 11 sample 10069 and Apollo 14 sample 14063, and finally Apollo 15 sample 15415. This should not be taken to imply a genetic line of liquid ascent, but rather to emphasize that melts may have different relative concentrations of Ti and Cr even if they are in equilibrium with pyroxene, plagioclase, and ilmenite. However, the initial differences in relative concentrations of the three components do reflect important differences in the initial melt compositions for the Apollo 11, 12, and 14 basalts, and it is probable that the composition of the pyroxene in sample 15415 as presently constituted was not in equilibrium with the melts from which the Apollo 11, 12, and 14 basalts crystallized.

Independent of the trends for most of the pyroxenes in sample 14310 is an analysis of high magnesian augite blebs in hypersthene (Wo4En74Fs22), called "primitive augite" in the Figure. This augite has the approximate formula

\[
(\text{Na}_{0.01}\text{Ca}_{0.74}\text{Fe}_{0.23}\text{Mn}_{0.01})(\text{Mg}_{0.87}\text{Ti}_{0.09}\text{Cr}_{0.02}\text{Al}_{0.05})(\text{Al}_{0.29}\text{Si}_{1.71})_6.
\]

An interpretation of this unusual grain of coexisting augite and hypersthene is that it crystallized at the beginning of crystallization of a primitive Mg, Al, Ti-rich magma; it could also be interpreted as a xenocryst from the lunar interior.

The most important, immediate use of diagrams similar to that of the Figure is that data plotted on them emphasize differences in the Apollo basalts and give information on the relative order of crystallization of the three phases plagioclase, ilmenite, and pyroxene.
In our previous work (Housley et al., 1970; 1971) we have shown, 1) that the dark gray fines typical of the Apollo 11 and 12 sites contain about 5 to 10 times as much Fe metal as the igneous rocks characteristic of the sites, 2) that almost all this excess Fe metal is associated with the glassy material in the fines, and 3) that much of the excess Fe metal is present as extremely small particles. Continuing this study we found that the fines 14003 and 14163 contain amounts of Fe metal similar to those in the Apollo 11 and 12 fines and that again much of it is present as very small particles. The igneous rock 14310 also contains about a factor of 5 less Fe metal.

Motivated initially by a desire to refine our knowledge concerning the association of this Fe metal we devised a simple technique for making quantitative magnetic separations of the fines. The sample is spread over the bottom of a flat bottomed Al container which in turn sets on a flat supporting plate between the poles of a strong permanent horseshoe magnet providing a field of about 1 kOe in the gap. A thin flat soft Fe plate in contact with the upper magnet pole can be lowered to any desired height above the bottom of the sample container. In order to reduce the tendency of grains to stick to each other or the container, the latter is filled with pure alcohol to a level above the bottom of the soft Fe plate. A separation is made by gently sliding the sample dish around until all parts of the sample have been directly under the soft Fe plate. The plate is then raised and the fraction collected removed. Another fraction can then be obtained by repeating the procedure with the soft Fe plate lower. Repeat collections at a given separation after stirring the sample collect very little material. Conversely a magnetic fraction after being stirred around in a clean sample dish is almost entirely collected again at the same separation at which it was initially collected. Therefore the technique results in magnetically clean fractions. It works well for particles at least spanning the size range 0.045-1.0 mm and can be used with little danger of sample loss or contamination.

A significant amount (~5%) of material is collected from the lunar fines at a separation of 6 mm allowing them to be separated into at least 8 distinct fractions. Paramagnetic material was not collected from a synthetic lunar rock powder at separations greater than 0.5 mm proving that the fraction in which a grain from the fines falls is determined largely by its metallic Fe content.

We have so far separated the light (sp. g. < 3.3) < 75 μm fraction of the fines 10084,85 and the 45-75 μm fraction of the fines 15101,92. One immediate observation from these separations was that a large fraction of the total material including many particles having the appearance of mineral grains was picked up at separations much larger than could be accounted for by
Study of Excess Fe Metal in the Lunar Fines

R. M. Housley

Paramagnetism. For example, 67% of the 15101.92 sample was collected at separations greater than 2.0 mm. This indicates that significant Fe metal is adhering to these grains.

The magnetic material from 10084.85 was examined by a variety of microscopic techniques and has been used in a series of Mössbauer runs at applied magnetic fields up to 55 kOe. These data show 1) that at least half and perhaps considerably more of the Fe metal is present as single domain particles and 2) that most of the metal has a hyperfine field corresponding to pure Fe metal rather than an Fe-Ni alloy.

The most magnetic (collected at 6 mm separation) and least magnetic (not collected at 2 mm separation) fractions of the 15101.92 fines have been microscopically examined in immersion oil grain mounts and in polished grain mounts using both reflected and transmitted light. They have also been studied with a scanning electron microscope (SCM) and x-ray powder photos have been taken.

The most magnetic particles form a remarkably uniform group. They are dark irregular, largely vesicular glass containing abundant mineral fragments. Under the SCM they are seen to have rounded glassy surfaces largely coated with welded smaller fragments. In the polished section at high magnification considerably less than half of the grains show any recognizable metallic Fe but by far the majority contain regions of inhomogeneous milky appearing glass. It seems likely that fine grained Fe is the cause of this appearance and perhaps of the overall darkness of the grains. In turn, we believe that it is likely that grains such as these are responsible for the low albedo of the lunar surface.

The least magnetic fraction does not contain any glass welded aggregates similar to those in the most magnetic fraction. It however contains in addition to mineral grains a considerable amount of fairly homogeneous transparent glass and some rock fragments.

Preliminary study of the x-ray diffraction data suggests that except for the strong enrichment in Fe metal the crystalline phases are present in about the same relative proportions in the glass welded aggregates of the most magnetic fraction as in the mineral grains of the least magnetic fraction.

The low Ni content and association with glass welded aggregates whose overall appearance suggests formation in a hot gas and dust cloud leads us to conclude that most of the Fe metal in the lunar fines probably results from the reduction of Fe rich silicate grains in such a cloud.

It seems likely that the magnetic separation technique described could be used to advantage in almost any type of study of the lunar fines since it clearly and easily separates fractions that have different physical characteristics and have had different physical histories.

We wish to thank L. H. Hackett, Jr. for operating the SCM.

References

Study of Excess Fe Metal in the Lunar Fines

R. M. Housley

LUNAR SOIL POROSITY AND ITS VARIATION AS ESTIMATED FROM FOOTPRINTS AND BOULDER TRACKS

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A number of estimates of lunar soil porosity have been made based on core tube samples obtained during the Apollo 11-15 Missions. The self-recording penetrometer, used during Apollo 15, provided data from which an additional porosity estimate could be made. While some adjustment of core tube porosity values is required to account for disturbance during sampling, the wide range in observed porosities (about 32% to 58%) suggests that lunar soil porosity may best be described in terms of statistical parameters. Unfortunately, however the number of values obtained using core tube samples alone is somewhat too small to yield a statistical variation in porosity—especially at any one landing site.

A statistical study of porosity variations has been made using two sources of lunar data: (1) Astronaut footprints - Apollo Missions 11, 12, 14 and 15 and (2) Boulder tracks - Lunar Orbiter Missions II, III, and V.

Photographic coverage of a large number of footprints is available for each landing site and these footprints have been analyzed as indicators of soil porosity within the top 10 to 20 cm of lunar soil. Through an extensive model testing program using a lunar soil simulant and theoretical analysis of footprints as plateload tests (including an account of the effect of reduced gravity), the correlation shown in Figure 1 was developed between astronaut footprint depth and soil porosity. The analysis of the footprint data was accomplished by comparing the behavior of the actual lunar soil to that of a lunar soil simulant prepared at the University of California, Berkeley. Model tests on other simulants have provided a basis for adjusting the curve in Fig. 1 as needed to account for variation in soil gradation. The correlation in Fig. 1 has been used to estimate the variation in soil porosity at each of the Apollo sites and at different locations within each of the landing areas.

Additional data on the variation of porosity of lunar soil has been obtained from the study of tracks made by rolling boulders. Theoretical analyses of the deformation mechanism associated with rolling boulders have led to the development of a relationship between the boulder track geometry and the mechanical properties of the soil. Sixty-nine lunar boulder tracks from 19 different locations on the moon have been examined using Lunar
LUNAR SOIL POROSITY AND ITS VARIATION AS ESTIMATED FROM FOOTPRINTS AND BOULDER TRAKCS

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FIG. I. PREDICTED VARIATION OF FOOTPRINT DEPTH WITH POROSITY.
Orbiter photography. Measurements of the track widths indicate that some of the boulders sank considerably deeper than others. It is suggested that the lunar surface materials vary from place to place and that the state of compaction as reflected by porosity is probably one of the controlling variables. Using bearing capacity theory, the friction angle of the lunar soil was estimated to range from 19° to 53° with an arithmetic average of 37°. Most of the values were between 24° and 47° and this range suggests a corresponding porosity range of 60 to 35 percent for a specific gravity of 3.1

Porosities from footprints and boulder tracks have been compared with core tube porosities to develop a comprehensive statistical description of the lunar soil porosity.

A direct correlation between the existing porosity and the most probable mode of soil formation (deposition) is still being sought. However, available data showing the wide range in soil porosity indicate that several mechanisms of soil formation may have been operative—often within a relatively small region. The magnitude of local variability in soil porosity suggests that a loosening process (such as impacting meteors) may have often been super-imposed on a soil formation process which produced a soil profile which was initially very dense.
BISTATIC-RADAR OBSERVATIONS OF THE LUNAR SURFACE WITH APOLLOS 14 & 15

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Transmissions from the orbiting Apollos 14 and 15 command-service modules have been used to observe oblique scattering of radio waves from the lunar surface. Signals at 0.13 and 1.16 m. wavelengths originate on the spacecraft, are reflected by the lunar surface, and received on the earth. Quantitative measures of surface slope statistics (r.m.s. slope and slope frequency distributions) on scales of approximately 200 wavelengths, effective surface reflectivity to depths of about 20 wavelengths, and bulk dielectric constants are obtained simultaneously. In addition, the polarization state of the received signal may be used to qualitatively assess variations in small scale roughness.

Root-mean-square slopes on the scale of 20 to 200 m. in length vary between 1 and 8 deg. In the mare areas observed thus far, the 200 m. r.m.s. slopes are roughly one-half those obtained at 20 m. In highland areas, the 200 m. and 20 m. slopes are nearly equal, albeit they may have markedly different slope frequency distributions.

From the polarization data the lunar surface on the scale of 0.13 m. is highly uniform in structure and shows little correlation with large scale features. On the scale of 1.16 m. marked variations appear over large areas.

Surface reflectivity shows marked wavelength dependences. At 0.13 m. the reflectivity data are consistent with scattering from a uniform dielectric half space with dielectric constant near 3. At 1.16 m. no such simple relationship holds and the data must be interpreted in terms of local variations in surface reflectivity, most likely associated with the depth of the regolith. Analysis of data from the Cavalerius formation indicate that this feature consists of reduced bulk density to a depth of many meters.
New data support the pre-mission suggestions (1,2) that Hadley Rille is a collapsed lava tube. Hadley Rille generally trends in directions that are controlled by pre-mare structures. The southern half of the rille, south of the Apollo 15 site, follows a mare-filled graben that is concentric to the Imbrium Basin; however, the zigs and zags in the course are mainly north and east-northeast, directions that do not reflect pre-mare structures in adjacent highlands. The northern half of the rille follows a general direction that is parallel to pre-mare Imbrium-radial faults, is less sinuous than farther south, and intersects several fault-like features, all of which suggest some structural control along this part of the course. At the Apollo 15 site, the east rim is 30-40 m higher than the west rim as if the rille were a normal fault. The rille cannot be a simple fault, however, because the two sides do not match; instead, at bends, the outside has a greater radius than the inside, which is partly attributed to recession of the rims by mass-wasting.

The rille owes at least part of its present profile to collapse, as if into a buried conduit or deep, narrow trough. Northwest of the Apollo 15 site, interruptions in the rille are interspersed with rimless, elongate bowl-shaped depressions that appear to be formed by collapse. Incipient slumps are clearly recognizable, and slumping is presumably responsible for downbowed of the mare surface immediately adjacent to the interrupted portions of the rille. South of this area the rille is continuous, but still the deepest parts are the widest. This is a different relationship than that shown by river channels in which depth and width vary inversely so that the cross-sectional area tends to remain constant. The geometry of the rille can, however, be explained by collapse with more extensive foundering at the widest points. Most of the rille has a V-shape profile apparently formed by recession of the rims and coalescence of talus aprons from either side. Depth-to-width ratios increase toward the south, showing that the slopes of the rille walls steepen southward.

The possibility that the mare surface subsided differentially approximately 100 m after it crusted over is suggested by an unevenness of the surface, a cap of mare-like rocks on North Complex, a possible high-lava mark on the base of Mount Hadley, and grabens at the margin of the mare. Such a subsidence could have resulted from drainage into and along a lava tube that became the rille.
Regolith and underlying outcrops of mare basalt are exposed along the upper walls at the rille. Blocky talus derived from the outcrops occupies the lower parts of the rille walls. The rille is mainly a depository but not a source of local impact ejecta, so regolith is eroded off the rim into the rille and tends to collect at the bottom of the rille.

Details of the outcrops of mare basalt are well displayed in telephoto stereophotography of the southwest wall of the rille near the Apollo 15 site. Exposures there are mostly restricted to the upper 35 m of the wall, but locally extend as far as 60 m below the rim. Differences in aspects of the outcropping rocks suggest that more than one flow unit of basalt is present. Most of the outcrops are massive units, averaging at least 10 m thick. Some of these massive outcrops are pitted and break irregularly, others break cleanly along joints, and others show discontinuous horizontal planar parting. They vary also in albedo. Crude, subvertical columnar joints are present locally. One series of massive outcrops is characterized by inclined joints. Below this series is a well layered outcrop 8 m thick; several vertical fractures cut across more than one of the layers, suggesting the layers may be within the same cooling unit. Locally, above the massive outcrops, is a poorly exposed, dark, hackly-surfaced and crudely layered unit no more than a few meters thick. One light-toned slabby outcrop near the mare surface resembles shelly pahoehoe. Numerous talus blocks derived from the massive outcrops are 10 m across or more; blocks of basalt this large are rare on Earth, demonstrating that the lunar basalt flows are both thick and remarkably unjointed compared to terrestrial counterparts.

The outcrop ledge is absent where the rille abuts Apennine massifs, showing that the rille cuts through the mare basalt and into the massif material. Southward along the rille, the outcrop ledge in the upper walls of the rille thickens and is more continuous. This suggests a regional variation in the mare strata. Soft material above the massive outcrops also thickens southward, from 5 m of regolith at the Apollo 15 site to 15-25 m south of Hadley C; possibly pyroclastics are present in the south, in addition to a regolith produced by impact gardening.

Talus slopes below the outcrops show gentle subhorizontal benches, inflections, and concentrations of blocks which may be indicative of subtalus stratigraphy.

References and Notes

(3) Prepared under NASA contracts T-65253-6, R-66, and NASW-417.
We are studying the origin and nature of the thermoluminescence (TL) of lunar fines by measuring the TL of individual grains. Figure 1 shows the sixteen brightest natural glow curves obtained from several hundred grains of the 200-400 mesh fraction of 12033. Also shown is the sum of the glow curves of these sixteen individual grains and the glow curve obtained from a 0.5 mg bulk sample (~1500 grains). It can be seen that only a few dozen of the brightest grains will account for the majority of the TL from the bulk sample; i.e. most of the TL is emitted by a small minority of the grains. The bright grains are transparent and have a specific gravity $<2.89$. The most common type of TL glow curve, shown in Fig. 1 with solid lines, has a low temperature peak around 225-275°C. A microprobe analysis of several of these grains has shown just an AlSiCa phase (we are unable to detect Na) and they are evidently some form of plagioclase, although the majority of the plagioclase grains give very little natural TL (<50 cps). Other grains, shown by long dashed lines, have higher temperature peaks and are usually found to contain an AlSiK phase and sometimes a SiCaP phase. We have not yet determined which phase gives the high temperature TL. Still other grains (short dashed lines) have both the low and high temperature peaks and are presumably "composite" grains. Similar results have been found for Apollo 12 double core samples and for the Apollo 14 core 14230.

The association of phosphates (which are high in radioactivity) with some of the TL grains indicates that the TL, particularly at glow curve temperatures above 300°C, may be induced partly by internal radiation as well as by cosmic rays. At low glow curve temperatures the TL comes partly from plagioclase grains, in which the radiation should be mainly from cosmic rays, and partly from the "composite" grains in which the TL internal radiation dose rate could be important. Thus the radiation dose rates appropriate to the TL emitted by a bulk sample are very complex, and could change from one sample to another with variation in radioactivity content or distribution. Measurements on bulk samples are further complicated by the presence of other mineral grains which give very little natural TL but which do contribute to the artificially induced TL, especially at temperatures $<250°C$. Despite these limitations, bulk sample measurements give useful information, e.g. distinguish material from within a few centimeters of the surface from deeper material. However we have been unable, from measurements on bulk samples, to get results of the accuracy required to measure temperature gradients in core samples (1). Efforts to achieve greater accuracy by measuring individual grains at various core depths are in progress and will be reported.
A survey using 1 mg bulk samples was made of the Apollo 14 and 15 fines 14141, 14163, 14230, 14259, 15231, 15471, 15531 and of the Apollo 14 trench. The glow curve shapes induced by artificial radiation of the Apollo 14 fines are very similar to those of Apollo 11 and 12 fines, indicating similar TL minerals. Relatively more low temperature TL (<300°C) is induced in the Apollo 15 fines. The samples are about as sensitive as the Apollo 12 core fines except for 14141 which is about 4 times brighter. The samples 14141, 14163, 14259, 15471 and 15531 all show the effect of diurnal heating near the lunar surface, having relatively low natural TL at glow curve temperatures <250°C. They are similar to Apollo 12 core samples of <3 cm, indicating a similar depth if the diurnal heat wave penetrated the same distance. Sample 15231, from underneath a large boulder, is similar to Apollo 12 core samples of 3–4 cm. Samples from all depths of the core 14230 have natural glow curve shapes which are similar to the Apollo 12 core from about 7 cm downward; none look like surface material. This implies that several centimeters of material from the top of the core were lost when the follower fell out. An anomaly is that samples from the top, middle and bottom of the Apollo 14 trench do not show any depth dependence; all have low TL like the surface samples.

The effect of solar flare radiation has been seen in the natural TL of rocks 14310 and 12063. The TL decreases steadily with depth by a factor of ~2.5 in the first centimeter and is then nearly constant to 5 cm, the maximum depth available.

The stored TL is affected by white roomlight, being bleached at glow curve temperatures of 200 to 250°C and induced at 150°C (Fig. 2). All our operations are now carried out in red light which has much less optical effect. The effects of optical bleaching by the sun have also been seen in several lunar rocks where the TL at the very surface (<1 mm) is low. Some of these optical effects may be produced by plagioclase as shown by optical measurements on the plagioclase rock 15415 (Fig. 3).

Fig. 1. Natural TL glow curves of fourteen grains of 12033. Grain size: 200–400 mesh. Blackbody has been subtracted. Also shown is the light sum of the grains and the TL emitted by a bulk sample.
REFERENCE


Fig. 2. Dependence of stored natural TL upon total time of exposure to white light (o) and red light (x), for 23-cm material from the Apollo 12 double core. For each sample, the natural TL measured after exposure to light has been normalized by the subsequent TL output (measured at 325°C) induced by a beta dose of 1.5 x 10^5 rads.

Fig. 3. Glow curves of a 1 mg sample of rock 15415; natural (N), artificial (A) produced by a beta dose of 20 krad, and the TL induced by a 12 h exposure to white fluorescent bulb roomlight. Curve A has been decreased by x10. Blackbody has been subtracted.
Two chemically distinct types of lunar anorthosites and related high Al basaltic rocks have been found in the lunar samples. The most common type of anorthosite has low K concentrations, is exemplified by the Apollo 15 specimen 15415, and is chemically similar to the type described by Wood et al. (1970) in the Apollo 11 soil. There is evidence for a second type of lunar anorthosite characterized by high K concentrations (Fig. 1) and containing minor minerals such as zircon, apatite, ilmenite and K-feldspar (Table 1 and Fig. 1). The mineral assemblages of the high K groups suggest that they were derived from KREEP basalts.

Our isotope dilution data (Table 2 and Fig. 2) show: 1) that Eu and Sr values are relatively constant except for one high K sample (12033,97,7) which has much higher Eu and Sr contents and 2) that abundances of Ba and the trivalent rare earths vary markedly. The second characteristic is related to mineralogical purity, as illustrated by plotting Ce vs Mg (Fig. 3).

Samples 15415 and 12033,97,7 are nearly pure plagioclase and we assume that their chemical composition is that of liquidus plagioclases. We can calculate some chemical characteristics of their parent liquids using liquid/plagioclase distribution coefficients measured on coexisting plagioclase phenocrysts and basalt glass or fine grained matrix from oceanic ridge basalts (last two columns Table 2) except for Eu (Green et al., 1971). The calculated rare earth, Ba and Sr patterns are shown in Fig. 4 along with the plagioclase data from which they were calculated and the pattern for KREEP basalt glass from 12033 soil. The calculated pattern for the low K anorthosites is unlike that of any lunar basalt observed to date and is essentially chondritic in relative abundances although much higher in concentrations. In particular the calculated Eu anomaly is very small and may really be non-existent. Conversely, the KREEP anorthosite fragment appears to have crystallized from a liquid with substantial Eu and Sr anomalies and with much higher rare earth and Ba concentrations, similar to that observed for KREEP basalt glass.

Hubbard and Gast (1971) have suggested that the KREEP type of basalt was produced by 1-3% partial melting of a shallow primitive layer (<100 km deep) in the moon that was substantially enriched, relative to chondrites, in refractory trace elements, Ca and Al. They further suggested the plagioclase is a major component of that region of the moon. We suggest that the parent liquid calculated for the low K anorthosites was produced by much more extensive melting of a similar feldspar rich region.

Chemical Composition of Lunar Anorthosites and Their Parent Liquids
N. J. Hubbard


<table>
<thead>
<tr>
<th>Table 1. Mineral Assemblages and K2O Contents of Plagioclases from Anorthositic Coarse Fines:</th>
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<tr>
<td><strong>High K Type</strong></td>
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<tr>
<td>14321,17,17 98% Plagioclase An85, 1% Zircon, 0.5% Whitlockite, 0.5% Pyroxene. K2O = 0.15%.</td>
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<td>12033,85c 99% Plagioclase An80, 1% Olivine Fa30, 1% Ilmenite. K2O = 0.20%.</td>
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<td><strong>Low K Type</strong></td>
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<tr>
<td>*14161,35,2 90% Plagioclase An95, 9% Olivine Fa40, 1% Orthopyroxene. K2O = 0.04%.</td>
</tr>
<tr>
<td>*10085,11,146 White fragment donated by Elbert King. Not analyzed by microprobe. See Table 2 for partial chemical analysis.</td>
</tr>
<tr>
<td>*10085,11,47 White fragment donated by Elbert King. Not analyzed by microprobe. See Table 2 for partial chemical analysis.</td>
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<td>10085,87,1 100% Plagioclase. K2O = 0.01%.</td>
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<td>*Samples that were analyzed by isotope dilution. See Table 2.</td>
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**Fig. 1**

**Fig. 2**

**Fig. 3**
Chemical Composition of Lunar Anorthosites and Their Parent Liquids

N. J. Hubbard

Fig. 2

Fig. 4

TABLE 2

<table>
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<th>Element</th>
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<th>15415</th>
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*Values bracket those of Green et al. (10).
The major chemical characteristics of the Apollo 14 samples analyzed by us are essentially identical with those of the KREEP basalts from the Apollo 12 soil samples. Basalt sample 14053 is chemically intermediate between the KREEP basalts and the Apollo 12 mare basalts. The U, Ba and REE abundance patterns of Apollo 14 soils, breccias and basalt 14310 are identical with those of Apollo 12 KREEP basalts and extend to lower concentrations, reaching a minimum in 14310 (Fig. 1). The lower U, Ba and REE concentrations of soils are in part due to admixture of mare basalts and other rock types. However, KREEP basalts with lower U, Ba and REE concentrations are to be expected if this basalt type formed by partial melting. KREEP basalts are unique among lunar materials in their high Ce/Yb ratios, large negative Eu and Sr anomalies and high U, Ba and REE concentrations.

During basaltic igneous processes Ce, U, Ba and K are essentially coherent in their chemical behavior. Plots of these elements relative to each other show that Ce and U are coherent for Apollo 11, 12, 14, 15 and Luna 16 soil, breccia and basalt samples and that KREEP basalts have a U/Ce ratio ~50% greater than other lunar materials. Relative to the Ce-U coherence and the general Ba, Ce, U and K coherence in lunar samples Ba, K and Ce (or U) are rather incoherent in KREEP basalts. This disruption in trace elements normally coherent in basaltic igneous processes suggests that non-igneous processes have been active. The events that produced this disruption are thought to occur as follows: (1) formation of basaltic liquids of KREEP composition by small percentages of partial melting, (2) crystallization of this liquid to produce basalts with K, Rb, Ba, U, REE, etc. rich groundmass, (3) a poorly defined metamorphic event that brecciated and annealed these basalts at temperatures high enough to melt and differentially segregate the high K, Ba components of the groundmass as granophyre patches and thus disrupt the igneous coherence of K, Ba and Ce(U) on a scale similar to our sample sizes (1-10 mm).

Several of our samples have been extensively annealed (metamorphosed) and contain granophyre patches. It is known from laboratory experiments (Gibson and Hubbard, 1971) that K and Rb can be differentially volatilized at temperatures similar to those required to effect the metamorphism. Two samples (14006 and 14161,35,3) have strikingly low K and Rb concentrations and high K/Rb ratios (Table 1, Fig. 2) but normal Ba, U, REE, etc. and major element composition. The percentages of K and Rb assumed lost from these samples and the accompanying changes in K/Rb ratios are very comparable with the experimental results, thus indicating that thermal volatilization depleted the K and Rb concentrations and increased the K/Rb ratios.

REFERENCES:
E. K. Gibson, Jr. and N. J. Hubbard, this volume.

ACKNOWLEDGEMENTS:
We thank Brij Bansal, Stan Church and Ian Ridley for assistance in analytical work.
CHEMICAL COMPOSITION OF APOLLO 14 MATERIALS AND EVIDENCE FOR ALKALI VOLATILIZATION

N. J. Hubbard

P. 409

**TABLE 1. ISOTROPIC DILUTION DATA FOR APOLLO 14 SAMPLES**

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**TABLE 2. X-RAY FLUORESCENCE DATA FOR MAJOR, MINOR AND TRACE ELEMENTS IN APOLLO 14 SAMPLES**

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Phase relations of lunar and terrestrial pyroxenes at one atm.

J. Stephen Huebner and Malcolm Ross, USGS, Washington, D. C. 20242

Phase equilibrium experiments delineating the augite-pigeonite solvus (12,13) and orthopyroxene-clinopyroxene reaction relationship (13) have been combined with previous experimental studies of the bounding systems to yield a phase diagram for pyroxene quadrilateral compositions. This diagram helps explain crystallization, reaction relationships, inversions, and exsolutions of pyroxenes in lunar and terrestrial basic volcanic and intrusive rocks.

The pyroxene quadrilateral En-Di-Hd-Fs (MgSiO₃-Ca₀.₅Mg₀.₅SiO₃-
Ca₀.₅Fe₀.₅SiO₃-FeSiO₃) is a part of a planar section through the system MgO-
CaO-FeO-SiO₂. During pyroxene crystallization the system is only pseudoternary because the molecular ratio (Mg+Ca+Fe):Si=1, if precipitation of olivine (Mg,Fe,Ca)₂SiO₄ or silica occurs. We have ignored the presence of other components; in the lunar pyroxenes, Al₂O₃+Cr₂O₃+TiO₂, the most important additional components, total about 3.5 wt.%. We adopted the following nomenclature for pyroxenes. Calcium-rich, C2/c pyroxenes (Wo>24) are augite (A); orthorhombic (Pbcm) pyroxenes (Wo<4) are orthopyroxene (O); pigeonite (P) is a calcium-poor clinopyroxene (Wo<20) with P2₁/c symmetry at low temperatures that inverts to C2/c symmetry at high temperature (10). We have not observed evidence of protoenstatite (over the composition range En₁₄-87) and have excluded it from this discussion.

The solidus diagram is shown in Fig. 1. Isotherms are established across the augite-pigeonite field (our work), and along the MgSiO₃-CaO-Mg₀.₅SiO₃ (5) and Ca₀.₅Fe₀.₅SiO₃-FeSiO₃ (3) joins. We experimentally observed melting during the reaction of Mg-rich orthopyroxene (En>61) to pigeonite, a finding in disagreement with Bowen and Schairer (4). The solidus intersects the immiscibility regions formed by the three pairs of pyroxenes A+O, O+P, A+P. Olivine (F) + silica (S) is stable for compositions near the Fs corner; high-temperature-wollastonite solid solution (W) is the stable phase for very Hd-rich compositions above solidus temperatures, but inverts to augite on cooling.

The subsolidus (Fig. 3) is characterized by two essentially colinear miscibility gaps, A+0 and A+F. The boundary between the compositional gap O+P and the pigeonite field slopes, with decreasing temperature, toward "Hd" and intersects the A+P immiscibility "tunnel" so that the calcium-poor side of the three-phase triangle O+A+P moves toward iron-rich compositions. With decreasing temperatures the A+P tie lines move so that P becomes increasingly more iron-rich than A (1).

Pigeonite is not stable at low temperatures, but inverts to orthopyroxene on cooling. The subsolidus diagram accounts for the inverted pigeonites of Hess (8): pigeonite host exsolves (001) augite before and (or) during inversion of the host to orthopyroxene. Continued cooling of orthopyroxene host results in exsolution of (100) augite. Where pigeonite is found at room temperature, as in the Apollo 12 rocks, it appears to be "locked in" by the presence of finely intergrown augite lamellae.
PHASE RELATIONS

A tentative liquidus diagram is presented in Fig. 2. Although there is a primary field of olivine (15,11), the stable existence of pyroxene with the same Mg:Ca:Fe proportions as the olivine at low temperature, commonly resorbed olivine phenocrystals, and reaction to form pyroxene suggests olivine becomes unstable at a peritectic point (P1). Consumption of olivine by reaction with liquid has been observed experimentally (7,2).

We observed experimentally the reaction of $P+A \rightarrow O+L$ on heating. Hollister et al. (9) observed hypersthene cores in pigeonite from lunar rock 12065, and Biggar et al. (2) found that early precipitated, Ca-poor pyroxene reacted out during crystallization of pigeonite from melt. We interpret the indicated reaction as peritectic, P2. We do not yet know the compositions of the liquids in the assemblages OPL and OAL as they descend in temperature to the 4-phase equilibrium, or PAL as it descends from the equilibrium, but analysis of our experimentally produced liquids is in progress. The possibly resorbed pigeonite cores with augite rims observed by Gay et al. (6) and the pyroxene crystallization trend observed by many lunar investigators suggest that $L \rightarrow P+A$ over the early part of the trend, later changing to $L+P \rightarrow A$.

Chemical analysis and behavior on heating define two groups of pigeonites in Apollo 12 rocks, specifically 12021. We associate the more calcium-poor pigeonite ($\sim$Wo9) with the region O+P, and the more calcium rich pigeonite ($\sim$Wo3) with the region P+A.

Other reaction points must be present, but they have not been studied experimentally. Point P3 is A+W+S+F+L and is probably peritectic. Wollastonite solid solutions are rarely found in nature. Perhaps wollastonite crystallization can serve as a model for pyroxferroite crystallization in the lunar rocks. Point P4 is arbitrarily drawn as an eutectic, yielding the subsolidus assemblage augite + olivine + silica + orthopyroxene or pigeonite.

Experiments in progress suggest the surprising result that the solidus is not lowered appreciably in temperature by the addition of plagioclase to the system, simplifying the application of this diagram to the interpretation of lunar rocks. However, even slight changes in the solidus position may change the character of points P3 and P4, greatly influencing the trend of highly differentiated liquids.

References

Fig. 1, Solidus diagram, pyroxene quadrilateral. Topology of fields near Hd-Fs join tentative.

Fig. 2, Tentative liquidus diagram, pyroxene quadrilateral.

Fig. 3, Isothermal section, approx. 900°C, through pyroxene quadrilateral. Pigeonite stable; wollastonite unstable.
The 40Ar-39Ar age dating technique has been applied to separated minerals (plagioclase, pyroxene, quintessence, and "ilmenite") as well as to unseparated whole rock samples from Apollo 14 and Apollo 15. These investigations make it clear that the age resolution possible by the 40Ar-39Ar method is limited not by experimental errors but rather by the interpretation of variations in the K-Ar age spectrum which exceed the experimental errors.

A monotonic increase of apparent age with release temperature is frequently observed in 40Ar-39Ar experiments. It is a well-understood consequence of prior gas loss and does not affect the validity of an age determined from a high-temperature plateau. However, seemingly unambiguous ages differing by 0.1 AE have been determined by well-defined plateaux in the apparent age spectra of whole rock and plagioclase samples of rock 15555.

Prior low-temperature losses of Ar observed in 14073 and 14310 are reflected principally in the quintessence phase (high K interstitial glass), confirming earlier suggestions. The quintessence separates for both 14073 and 14310 nevertheless define high-temperature apparent age plateaux which are marginally lower than the ages of the corresponding unseparated samples or mineral separates. The effect of low ages in the non-retentive quintessence is a lowering of the apparent age of the unseparated whole rock and is the presumed explanation for the apparent age difference between the plagioclase and unseparated samples of 15555. The degree to which this effect is operative in other rocks has not been established.

A well-defined pattern of apparent ages has emerged from the study of a number of Apollo 14 samples. The apparent ages increase as a function of gas release to a reasonably well-defined plateau followed by a significant decrease in apparent age in the final 10-20% of gas release. This anomalous behavior is not observed in plagioclase and quintessence separates of these rocks, but is particularly strong in pyroxene separates, and less so in an "ilmenite" separate.


Since plagioclase shows the best gas retention characteristics of all the phases studied and has not been observed to show the anomalous behavior of the pyroxene, it must be regarded as a more reliable and more easily interpreted age indicator than the other minerals and even unseparated whole rock samples.

A second type of anomalous apparent age spectrum is observed in several samples of 14083 and 14321, both fragmental rocks from the edge of Cone crater, and in a shocked igneous fragment from soil 14001. Both light and dark fractions of 14083 were studied; the 14321 samples were a group of light-colored igneous fragments showing some evidence of shock, a well-defined basaltic clast, two samples of dark, fine-grained, partly vesicular material, and a finely-crushed split of one of these dark fractions. The K concentrations (and \(^{40}\)Ar) vary by a factor of 4 among these samples. These samples show a peculiar characteristic release pattern in which the apparent ages rise to very high values in the early stages of release (~ 10%) and then decline steadily to very low values at higher extraction temperatures. This release pattern is found to varying degrees in all of these samples with the extreme case being the crushed fragment of 14321, which varied from 4.3 AE to 2.4 AE.

The average age of each of these samples, computed from the ratio of total \(^{40}\)Ar to total \(^{39}\)Ar, is remarkably constant at 3.92-3.93 AE, excepting only the 14321 basaltic clast, whose lower 3.86AE age is reasonably attributed to gas loss in the first release fraction. This conservation suggests that the pattern is due to a redistribution, without significant loss, of \(^{40}\)Ar (or K). The relatively simple age spectrum provides a boundary condition on suggested redistribution processes. The crushing experiment has shown that little of the \(^{40}\)Ar is located in positions such as large vesicles and interstitial voids easily accessible by crushing. \(^{39}\)Ar/\(^{37}\)Ar release patterns in several of the samples reflect relatively small variation in the apparent K/Ca ratio as compared to rocks in which large amounts of \(^{40}\)Ar are contained in K-rich glass.

These large and variable differences in the age spectra notwithstanding, the agreement of the total gas ages also suggests that relatively precise ages can be extracted from data on samples where this pattern is observed, if low temperature losses have had no serious effect on the second release fraction. If the \(^{40}\)Ar or K has been redistributed in a single event, it must have occurred less than 3.5 AE ago. In particular, if redistribution occurred during brecciation, then this event must be separated from the Imbrium impact event, which occurred more than 3.75 AE ago.
HISTORY OF LUNAR BRECCIAS, I.D. Hutcheon, P.P. Phakey, P.B. Price and R.S. Rajan, Dept. of Physics, Univ. of California, Berkeley, CA 94720

With optical and high-voltage electron microscopy we are studying the complicated history of Apollo 14 and 15 breccias.

1. **Uranium distributions using Lexan plastic prints.**

Fission track studies of the distribution of uranium in Apollo 14 breccias have demonstrated striking differences between Apollo 14 samples and basalts from Apollo 11 and 12. Tracks produced by thermal neutron-induced fission of $^{235}\text{U}$ are recorded in Lexan plastic overlays, providing a quantitative mapping of uranium locations in the rock sections. The whole-rock uranium concentrations in all seven Apollo 14 breccias studied are comparable (several ppm) but exhibit substantial variation in their distribution. In the interior of 14311 the uranium distribution is extremely uniform. No stars or centers of high U concentration are present, in contrast to the situation in 14301, 14315, and 14318, which are characterized by highly irregular uranium distributions.

Illustrating the extreme disequilibrium between components of a breccia is the great variation in the concentration of uranium within a breccia. In 14066 two feldspar crystals within 100μ of each other differ in U-content by a factor of 10². Similarly, feldspar crystals in 14318 with a uranium level as low as 10 ppb are embedded in fine-grained material containing 3 - 4 ppm of uranium. Fine-grained material in the form of a microbreccia firmly shock-welded to feldspar crystals has been observed in 14321 and 14311. The uranium is primarily located in the breccia portion, but the presence of spontaneous fission tracks in the adjacent feldspar portion suggests limited diffusion of uranium through the interface between the breccia and crystal. Characteristic of many feldspars are internal surfaces ("cleavages") strongly enriched in U (determined by Lexan mapping) and other heavy elements such as Ba, Sr, K, and Zr (determined by ion microprobe). These observations must be borne in mind when searching for extinct elements and when doing leaching experiments in geochronology.

2. **$^{244}\text{Pu}$ fission tracks in Apollo 14 breccias.**

a) Uniformly distributed U and Pu in whitlockite within breccia 14321: A 300μ whitlockite crystal from the matrix of 14321 contains $8.6 \times 10^7$ spontaneous fission tracks per cm². From the measured $^{238}\text{U}$ concentration we find that the number of $^{244}\text{Pu}$ fission tracks is about half that due to fission of $^{238}\text{U}$ ($\rho_{\text{Pu}}/\rho_{\text{U}} \approx 0.5$). To discriminate against cosmic-ray induced spallation tracks, we counted only long tracks in a plastic replica observed in an SEM. Assuming condensation of the solar system 4.6 $\times 10^9$ years ago and an initial Pu/U ratio of .016, we calculate that the whitlockite crystal began storing tracks 3.95 $\times 10^9$ years ago. The fission track age of the whitlockite grain is consistent with the Rb-Sr age of a basaltic clast from 14321, reported to
be $3.95 \pm 0.04 \times 10^9$y (Papanastassiou and Wasserburg, 1971) and with a lower limit on the whole-rock K-Ar age of 14321, reported to be $3.92 \times 10^9$yr (Turner et al., 1971). The consistency of these results implies that Pu and U have geochemically similar behavior in whitlockite. Because of the low fractional yield of fissionogenic Xe isotopes (-6%) compared with fission tracks (100%), it will be very difficult to see excess Xe isotopes from $^{244}$Pu fission in a 3.95 $\times 10^9$ year old sample. Fission tracks provide a very sensitive means of determining small differences in age for samples older than 4.0 $\times 10^9$ years.

### Table 1. $^{244}$Pu Fission Track Analysis

<table>
<thead>
<tr>
<th>Breakdown of tracks</th>
<th>Track density $(\text{cm}^{-2} \times 10^{-7})$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total tracks in whitlockite</td>
<td>$9.36 \pm 0.72$</td>
</tr>
<tr>
<td>Fe-group cosmic-ray tracks</td>
<td>$0.15 \pm 0.05$</td>
</tr>
<tr>
<td>Reactor neutron-induced fission tracks</td>
<td>$0.60 \pm 0.08$</td>
</tr>
<tr>
<td>Cosmic-ray neutron-induced fission tracks</td>
<td>$\leq 0.02$</td>
</tr>
<tr>
<td>High-energy cosmic-ray-induced tracks</td>
<td>$\leq 0.03$</td>
</tr>
<tr>
<td>Spontaneous fission tracks</td>
<td>$8.56 \pm 0.73$</td>
</tr>
<tr>
<td>Tracks from $^{238}$U fission in 3.95 Gy</td>
<td>$5.64 \pm 0.42$</td>
</tr>
<tr>
<td>Tracks from $^{244}$Pu fission in 3.95 Gy</td>
<td>$2.92 \pm 0.84$</td>
</tr>
</tbody>
</table>

Observed $\rho_{Pu}/\rho_{U} = 0.52 \pm 0.16$

b) Heterogeneously distributed U and Pu in feldspars within breccia 14311: We have described in a separate Abstract how background tracks from solar and cosmic ray particles can be preferentially erased without disturbing fission tracks - by annealing each mineral at a pre-determined temperature. This technique has enabled us to study spontaneous fission tracks in feldspar crystals from 14311. Lexan prints show that although the whole-crystal uranium level is low ($<100$ppb), uranium at the several ppm level is localized in cleavages and inclusions. Once the actual uranium concentration in a cleavage is determined from a Lexan print, we find that the observed number of spontaneous fission tracks emanating from the cleavage (TINCLES) is consistent with a 3.95 Gy age for the crystal and a $^{244}$Pu fission contribution similar to that in whitlockite of 14321. We find no evidence for an excess of fission tracks in cleavages that could be attributed to superheavy transuranic elements.

3. Shock and radiation-damage effects in matrix and clasts.

Much of the area of a thin-section of a breccia turns opaque or frosty when etched under conditions that would normally reveal tracks. We have used a 650 kev electron microscope to examine sections that have been ion-beam-thinned. Strong shock features are common, particularly in the fine-grained matrix. Highly localized shock waves have produced extensive twinning and glass laths on a submicroscopic scale and account for the odd etching behavior. Generally the grains showed no evidence of prolonged thermal meta-
History of Lunar Breccias

I.D. Hutcheon

We are studying correlations between shock intensity and survival of tracks produced in soil grains before they were incorporated into breccias. As an example, 14315 contains some grains with track densities of $10^8$ to $10^9$/$\text{cm}^2$ that survived the process of breccia formation. Two breccias (14315 and 14318) have submicroscopic features in common: sub-parallel to parallel sets of microfractures are present; some glass veins and planar features are observed in plagioclase; lamellar twins are common in pyroxenes; intergranular porous space is absent. These features are indicative of a low level of shock intensity, which is demanded if pre-existing tracks are to survive breccia formation.

4. Depth distribution of heavy cosmic-ray tracks.

In an igneous rock such as 14310, the density of heavy cosmic-ray tracks decreases very smoothly with depth, varying from nearly $10^7$/cm$^2$ at the surface to $\sim 2 \times 10^6$ at the center. No significant variations in density are observed in neighboring crystals. By contrast, in the large breccia 14321, though the trend is for cosmic-ray track density to decrease with depth, there are variations in adjacent crystals by almost a factor 10. We infer that some tracks produced in the grains before their assemblage into 14321 have survived and are superimposed on the more recent radiation. Both the shape and magnitude of the track density gradient in 14311, though incomplete, are consistent with the results obtained for 12022 by Barber et al., 1971. The buried rock 14301 contains $\sim 10^7$ Fe tracks per cm$^2$ in its interior. This high cosmic ray track density for an interior sample of a rock suggests a surface residence time $>>10^7$ years.

References


We have examined more than 100 of the Apollo 14 rocks in hand specimen and have divided them into two broad groups: relatively homogeneous crystalline rocks and fragmental rocks (breccias). The crystalline rocks are mostly metaclastic, but a few, including the second largest sample returned, are aphyric basalts. All of these are considered to be clasts dislodged from fragmental rocks. Megascopic properties of the fragmental rocks suggest a four-fold subdivision as follows: (1) friable breccias (F1) with light clasts dominant over dark ones; clasts larger than one millimeter are scarce; fragmental glass is comparatively abundant; (2) coherent breccias (F2) with light clasts dominant, which contain comparatively abundant fragmental glass; (3) friable to moderately coherent breccias (F3) with dark clasts dominant, which lack fragmental glass; and (4) coherent breccias (F4) with dark clasts dominant, which lack fragmental glass. Table 1 gives the sample numbers of rocks larger than 50 grams so classified, and the total number of rocks larger than 1g in each of the four groups.

<table>
<thead>
<tr>
<th></th>
<th>HOMOGENEOUS CRYSTALLINE ROCKS</th>
<th>METACLASTIC ROCKS</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>B (BASALTIC)</td>
<td>C (METACLASTIC)</td>
</tr>
<tr>
<td>&gt;50 g:</td>
<td>053, 310</td>
<td>&gt;50 g: 053, 310</td>
</tr>
<tr>
<td>&gt;1 g:</td>
<td>9 ROCKS</td>
<td>&gt;1 g: 9 ROCKS</td>
</tr>
<tr>
<td></td>
<td>F2</td>
<td>F4</td>
</tr>
<tr>
<td>&gt;50 g:</td>
<td>051, 265</td>
<td>&gt;50 g: 066, 169, 264, 302, 305</td>
</tr>
<tr>
<td></td>
<td>267, 271</td>
<td>267, 271</td>
</tr>
<tr>
<td></td>
<td>301, 307</td>
<td>301, 307</td>
</tr>
<tr>
<td></td>
<td>313, 315</td>
<td>313, 315</td>
</tr>
<tr>
<td>ME</td>
<td>&gt;1 g: 30 ROCKS</td>
<td>&gt;1 g: 35 ROCKS</td>
</tr>
<tr>
<td></td>
<td>F1</td>
<td>F3</td>
</tr>
<tr>
<td>&gt;50 g:</td>
<td>041 - 042 - 045 - 047</td>
<td>&gt;50 g: 063, 064, 082</td>
</tr>
<tr>
<td>&gt;1 g:</td>
<td>25 ROCKS</td>
<td>&gt;1 g: 6 ROCKS</td>
</tr>
<tr>
<td>LIGHT</td>
<td>DARK</td>
<td></td>
</tr>
</tbody>
</table>

We emphasize that the classification is descriptive. Without intending to require a genetic connotation, we note that the F1 group of fragmental rocks correspond to the "porous, unshocked microbreccias," the F2 rocks to the "shock compressed microbreccias," and the F3 and F4 rocks to the "thermally metamorphosed microbreccias" of Chao and others (1).

We also studied the character and distribution of more than 3,000 clasts in 200 thin sections of 19 rocks, and were able to divide them into 22 varieties (Table 2).
TABLE 2. PRINCIPAL TYPES OF CLASTS AND MICROCLASTS

<table>
<thead>
<tr>
<th>SYMBOL</th>
<th>CLAST-MICROCLAST TYPE</th>
<th>SYMBOL</th>
<th>CLAST-MICROCLAST TYPE</th>
</tr>
</thead>
<tbody>
<tr>
<td>GLASSES</td>
<td></td>
<td>MINERALS</td>
<td></td>
</tr>
<tr>
<td>A1</td>
<td>Pale yellow to brown glass</td>
<td>B1</td>
<td>Plagioclase</td>
</tr>
<tr>
<td>A2</td>
<td>Colorless glass</td>
<td>B2</td>
<td>Clinopyroxene</td>
</tr>
<tr>
<td>IGNEOUS ROCKS</td>
<td></td>
<td>B3</td>
<td>Olivine</td>
</tr>
<tr>
<td>C1</td>
<td>Intersertal</td>
<td>B4</td>
<td>Orthopyroxene</td>
</tr>
<tr>
<td>C2</td>
<td>Ophitic basalt</td>
<td>B5</td>
<td>Opaque minerals</td>
</tr>
<tr>
<td>C3</td>
<td>Intergranular basalt</td>
<td></td>
<td>METAMORPHIC ROCKS &amp; MINERALS</td>
</tr>
<tr>
<td>C4</td>
<td>Variolitic basalt</td>
<td></td>
<td>D1 Recrystallized plagioclase</td>
</tr>
<tr>
<td>C5</td>
<td>Vitrophyric basalt</td>
<td>D2</td>
<td>Light-colored metaclastic rock</td>
</tr>
<tr>
<td>C6</td>
<td>Graphic quartz-Alkali feldspar</td>
<td>D3</td>
<td>Metabasalt</td>
</tr>
<tr>
<td>C7</td>
<td>Plagioclase, orthopyroxene cumulates</td>
<td>D4</td>
<td>Dark-colored metaclastic rock</td>
</tr>
<tr>
<td>C8</td>
<td>Hypautomorphic gabbro, norite</td>
<td>D5</td>
<td>Recrystallized olivine</td>
</tr>
</tbody>
</table>

We had difficulty in counting clasts larger than 1 mm because of the small size of the sections and because of the scarcity of such fragments in the F1 and F3 groups. We therefore gathered much of our data in the 0.1 to 1.0 mm microclast range. Where clast and microclast distribution data are compared (see Fig. 1A vs. 1B), it is apparent that the smaller size group contains a larger proportion of glass and mineral fragments than the larger size group but is otherwise quite similar. Fig. 1A shows a triangular plot comparing the proportion of glass clasts to those of light colored rocks (the total of D1, D2, and D3 types) and dark colored rocks (D4) in thin sections available to us. A clear separation is found between F1, F2, and F4 types in terms of light and dark clast ratios. It is also apparent that F2 rocks are more apt to contain glass than F4 types. Much better statistical reliability can be obtained from the microclast data (Fig. 1B). The same clear separation of light and dark clasts of F2 and F4 rocks is apparent. Furthermore, the proportion of D4 clasts in F1 rocks is, on the average, even less than that of F2, although some overlap is seen. The proportion of glass microclasts is also significantly different between the groups: only two rocks classified as F4 types contain glass microclasts, whereas all of the F2 types contain significant amounts, and F1 rocks contain more than 50 percent glass in this size range. Only two F3 rocks are available for counting, and one of these proved to be layered. The counts suggest its darker layered component is like that of F4 material, whereas its lighter component is unlike any of the other rocks. A plot of microclasts of igneous rocks (all of Group C in Table 2) against dark (D4) and light (D1, 2 and 3) metamorphic rocks is given in Fig. 1C. The light and dark metamorphic separation is again apparent between F4 and F1-F2 groups. The proportion of igneous rocks in the F1 rocks is clearly very low, and there is an apparent tendency for F2 rocks to contain fewer igneous rock clasts than F4 rocks, although the areas overlap somewhat. Finally, all lithic microclasts were plotted against glass and mineral fragments in Fig. 1D. While the clear separation of types F1, F2, and F3 is largely due to glass content, it is apparent that type F2 contains considerably higher proportions of lithic than mineral fragments in this size range, whereas the reverse is generally true for the other groups.
FIGURE 1. Triangular plots of clast and microclast distributions. Symbols same as those of Tables 1 and 2. (1A at upper left, 1D at lower right). The clast and microclast populations clearly support the hand specimen classifications of groups F1, F2, and F4. Furthermore, the lack of overlap in the fields of F1, F2, and F4 rocks in Fig. 1 shows that at least three and probably four distinct stratigraphic breccia units are present in the Fra Mauro area, each distinguished by a characteristic clast population.

REFERENCES
The isotope ratios of elements in minerals of terrestrial origin rarely exhibit deviations from the accepted values (isotope abundance ratios) and then only in cases where the disturbance can be directly attributed to the nuclear genesis or decay of one or more of the isotopes in the natural system. Fractionation by chemical processes, although theoretically possible, is seldom substantiated in naturally occurring minerals. Fractionation might be expected to be more pronounced in elements whose isotopes are very different in mass i.e. $\text{H}_1$ and $\text{H}_2$ and various attempts have been made to measure the $\text{H}/\text{D}$ ratio in nature.

Of similar importance in recent years has been the ratio $\text{Li}^6/\text{Li}^7$ which has been the subject of considerable controversy. Whilst considering an analytical program for the rock-systems of South Africa checks were made on the isotopic ratios in ten of the more abundant systems and differences were found which were apparently too large for experimental error. Certain inconsistencies remained and these have been chronologically interpreted. In the case of basalts a relationship with age appears feasible with the lunar samples falling reasonably into agreement.

**Method**

A spark source mass spectrometer (Varian - Mat SMIBF) fitted with a laser-source was used in the investigation. The samples were introduced as pressed pills ($\phi = 1 \text{ cm}$) and Figure I shows a pill which has been exposed to the laser radiation. Use was made of the good sensitivity of the system for lithium (Fig. II). The ten samples had lithium concentrations in the range $0.5 - 30 \text{ ppm}$ and spectra were recorded for standard samples as well as the ten mentioned. A section of a mass spectrum can be seen in Fig. III. In this case the spark spectrum was run alternately so that the difference in signal for the same exposure can be appreciated. The lines are broad and rectilinear in profile so that a precise measurement can be made of the intensity of each line. Thus calibration functions can be constructed for the isotopes where the precision approaches the value of $\pm 1\%$ (the statistical limit). A computer program has been written to check and eliminate points which lie more than $3\sigma$ from the line arriving at a calibration function for the lines which has an overall precision of approximately $\pm 2\%$. As can be seen in Fig. IV the scatter of the points about the line is small.

**Results on materials analysed**

Our value for the $\text{Li}^7/\text{Li}^6$ ratio for W1 is $11.1 \pm 0.3$, agreeing well with the accepted value of 11.15. A recent study of lithium isotope ratios
LITHIUM ISOTOPE RATIOS

P.F.S. Jackson

(normal quoted as \((\text{Li}^7)/(\text{Li}^6)\)) in terrestrial material has been reported by Kaviladze, Abaludze and Melashivili\(^2\) who indicate that all previous determinations showing discrepancies in this ratio, have been subject to "systematic" or experimental error and there is no indication of lithium isotope fractionation in terrestrial material. Other authors have found values varying from 10.8 - 13.0 and Bernas, Gradsztajn, Reeves and Schatzman\(^2\) adopt a value of 12.5 in their paper.

Results on the ten selected samples are seen in Fig. V. It can be seen that the results are inconsistent. However, results on samples from extraterrestrial sources appear more consistent (Fig. VI). Some of the values were repeated using the spark source. In all cases reasonable agreement was established between spark source/laser source results. A chronological relationship was plotted and the results are shown in Fig. VII.

As the results on these samples have been obtained by two separate methods, the values for the lithium isotope ratios that have been obtained may be considered reasonable estimates of the mean values for the samples concerned. The precision of our methods although poorer than that of others studying lithium isotopic ratios is sufficient to suggest a distinct variation in the mean value of each sample. Physical differences between salts of the two isotopes are so small that such a difference in isotopic composition cannot be wholly interpreted by this means and another explanation involving nuclear particle bombardment is being sought. However, in the latter case certain isotopic ratios of other elements have to be measured even where the element is present only at ultratrace level (< 1 ppm) and the work on separating these is in hand.

References

Fig. II - Detection limits (ppm) in the laser source

Fig. III - Section of mass spectrum comparing mass 7 for spark and laser source
LITHIUM ISOTOPE RATIOS

P.F.S. Jackson

Fig. I - Sample pill and laser track

<table>
<thead>
<tr>
<th>BASALTS</th>
<th>AGE</th>
<th>$\text{Li}^7/\text{Li}^6$</th>
</tr>
</thead>
<tbody>
<tr>
<td>V.C.</td>
<td>2.3</td>
<td>8.6 ± 0.5</td>
</tr>
<tr>
<td>D.R.</td>
<td>2.8</td>
<td>10.0 ± 0.5</td>
</tr>
<tr>
<td>J.W.P.</td>
<td>3.1</td>
<td>13.1 ± 0.6</td>
</tr>
<tr>
<td>OK-4</td>
<td>3.5</td>
<td>14.6 ± 0.8</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>GRANITES</th>
<th>AGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>CA-1</td>
<td>0.5</td>
</tr>
<tr>
<td>NIM-G</td>
<td>2.0</td>
</tr>
<tr>
<td>F.F.</td>
<td>2.0</td>
</tr>
<tr>
<td>K.3</td>
<td>2.3</td>
</tr>
<tr>
<td>H.H.</td>
<td>3.2</td>
</tr>
<tr>
<td>DM</td>
<td>3.3</td>
</tr>
<tr>
<td>GC.48A</td>
<td>4.0</td>
</tr>
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*EXTREMELY LOW LITHIUM CONTENT

Fig. V - Results on terrestrial rocks

Fig. IV - Calibration curves for mass 6 and 7

EXTRA TERRESTRIAL SAMPLES

<table>
<thead>
<tr>
<th>LUNAR</th>
<th>AGE</th>
</tr>
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<tbody>
<tr>
<td>1416-151</td>
<td>3.3</td>
</tr>
<tr>
<td>14259-77</td>
<td>3.3</td>
</tr>
<tr>
<td>14321-222</td>
<td>3.3</td>
</tr>
<tr>
<td>14321-222/F</td>
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METEORITIC

<table>
<thead>
<tr>
<th>AGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1936 MACIBINI</td>
</tr>
<tr>
<td>1925 QUEEN MERCY</td>
</tr>
<tr>
<td>1912 LEEUWONTEIN</td>
</tr>
</tbody>
</table>

Fig. VI - Results on lunar and meteoritic material.

Fig. VII - Age of rock plotted against lithium isotopic ratio
THE ANALYSIS OF LUNAR MATERIAL RETURNED BY APOLLO 14


Ten rocks representative of the South African systems with ages varying from 2.0 - > 4.0 x 10^9 y. together with the three lunar samples (14163-151, 14259-77, and 14321-222) were analysed comprehensively by several analytical methods. The sample 14321 was found to contain heterogeneously distributed Fosterite, the analysis of which has been included in the results. These methods included wet chemical techniques and various instrumental methods of analysis. The results from all methods were compared and the best average value selected to represent the mean. Comparisons between the terrestrial basalts and the lunar samples showed few distinctive differences in major composition and one particular terrestrial sample, a lava from Ndikwe in the North Eastern Natal region of Southern Africa bears very similar analytical composition and age to the lunar samples returned by Apollo 14 and its analysis has been included for convenience.

Methods

The more precise of the chemico-analytical methods made use of a discrete elemental separation using combinations of ion exchange systems. Each element, having been totally isolated, was subsequently determined under the most ideal conditions for the anticipated concentration range. The samples were originally dissolved by heating the pulverized material with a mixture of hydrofluoric, hydrochloric and sulphuric acids in a controlled temperature environment. This solution technique was also made use of in atomic absorption spectrophotometry. In this case, however, the elements were determined under non-ideal conditions and buffers were used, together with other agents, to suppress interference effects present in multi element solutions. Standards (U.S.G.S.) were run simultaneously so that the calibration functions were obtained under working conditions.

Of the instrumental techniques available most use was made of x-ray spectrography. The various approaches used were:

1. Pressed powder method
2. Pellet method
3. Double melt method
4. Melt and Lucas Tooth regression
5. Direct Electron Excitation.

All these methods have been reported or are shortly to be so. One interesting feature is the use of a multiple regression analysis. The results
THE ANALYSIS OF LUNAR MATERIAL
P.F.S. Jackson

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obtained from this regression series are extremely good. Complementary to the fluorescence techniques used for the determination of the heavier elements (Z > 11) direct electron excitation was used for the determination of the lighter elements.

Finally a spark source mass spectrometer with a low voltage discharge as ion-source has been used to determine elements in the concentration ranges 1% - 1 ppm to supplement those obtained by the other methods.

Results

The results are compared in three sections:

MAJOR CONSTITUENTS (%)

Of the four samples, three lunar and the Ndikwe lava sample, the major constituents (> 5%) were SiO₂, Al₂O₃, CaO, MgO and FeO. Figure I presents the values obtained on each (the best average value).

<table>
<thead>
<tr>
<th>Element</th>
<th>14163-151</th>
<th>14259-77</th>
<th>14321-222</th>
<th>Ndikwe</th>
<th>Fosterite</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>-</td>
<td>49,0</td>
<td>47,2</td>
<td>54,2</td>
<td>Major</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>17,10</td>
<td>16,99</td>
<td>14,02</td>
<td>15,0</td>
<td>-</td>
</tr>
<tr>
<td>CaO</td>
<td>10,65</td>
<td>10,71</td>
<td>10,00</td>
<td>9,24</td>
<td>0,35</td>
</tr>
<tr>
<td>MgO</td>
<td>9,00</td>
<td>9,32</td>
<td>10,91</td>
<td>5,59</td>
<td>Major</td>
</tr>
<tr>
<td>FeO</td>
<td>10,22</td>
<td>10,56</td>
<td>13,11</td>
<td>9,23</td>
<td>7,5</td>
</tr>
</tbody>
</table>

Fig. I - Major constituents of lunar sample received from Apollo 14 compared to Ndikwe lava

MINOR CONSTITUENTS (%)

<table>
<thead>
<tr>
<th>Element</th>
<th>14163-151</th>
<th>14259-77</th>
<th>14321-222</th>
<th>Ndikwe</th>
<th>Fosterite</th>
</tr>
</thead>
<tbody>
<tr>
<td>K₂O</td>
<td>0,53</td>
<td>0,47</td>
<td>0,44</td>
<td>0,48</td>
<td>0,44</td>
</tr>
<tr>
<td>TiO₂</td>
<td>1,84</td>
<td>1,82</td>
<td>2,01</td>
<td>0,73</td>
<td>0,085</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0,50</td>
<td>0,42</td>
<td>0,36</td>
<td>0,132</td>
<td>0,01</td>
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<tr>
<td>Na₂O</td>
<td>0,65</td>
<td>0,65</td>
<td>0,71</td>
<td>1,17</td>
<td>-</td>
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<tr>
<td>MnO</td>
<td>0,127</td>
<td>0,136</td>
<td>0,17</td>
<td>0,145</td>
<td>0,08</td>
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<tr>
<td>Cr₂O₃</td>
<td>0,18</td>
<td>0,165</td>
<td>0,25</td>
<td>0,033</td>
<td>0,08</td>
</tr>
<tr>
<td>Ba</td>
<td>0,106</td>
<td>0,097</td>
<td>0,063</td>
<td>0,0185</td>
<td>0,0005</td>
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</tbody>
</table>

Fig. II - Minor constituents of materials
THE ANALYSIS OF LUNAR MATERIAL

P.F.S. Jackson

P.426

TRACE CONSTITUENTS (ppm)

<table>
<thead>
<tr>
<th>Element</th>
<th>14163-151</th>
<th>14159-77</th>
<th>14321-222</th>
<th>Ndikwe</th>
<th>Fosterite</th>
</tr>
</thead>
<tbody>
<tr>
<td>V</td>
<td>43</td>
<td>34</td>
<td>85</td>
<td>139</td>
<td>20</td>
</tr>
<tr>
<td>Ni</td>
<td>280</td>
<td>330</td>
<td>130</td>
<td>103</td>
<td>3</td>
</tr>
<tr>
<td>Nb</td>
<td>62</td>
<td>54</td>
<td>63</td>
<td>11</td>
<td>&gt; 2</td>
</tr>
<tr>
<td>Zr</td>
<td>760</td>
<td>720</td>
<td>710</td>
<td>89</td>
<td>&gt; 5</td>
</tr>
<tr>
<td>Y</td>
<td>240</td>
<td>210</td>
<td>190</td>
<td>16</td>
<td>&gt; 3</td>
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<tr>
<td>Rb</td>
<td>13</td>
<td>14</td>
<td>12</td>
<td>13</td>
<td>&gt; 1</td>
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<td>Sr</td>
<td>235</td>
<td>248</td>
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<tr>
<td>Co</td>
<td>33</td>
<td>33</td>
<td>33</td>
<td>43</td>
<td>25</td>
</tr>
<tr>
<td>Cu</td>
<td>14</td>
<td>16</td>
<td>10</td>
<td>45</td>
<td>&gt; 1</td>
</tr>
</tbody>
</table>

Fig. III - Trace elements in returned material.

Precision

The precisions of methods of analysis are naturally dependent on several factors. To demonstrate the precision of some of the methods we have compared the results of several methods at different concentrations.

<table>
<thead>
<tr>
<th>Chemical</th>
<th>At. Abs.</th>
<th>Mass spec.</th>
<th>X-ray powder</th>
<th>X-ray pellet</th>
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<tr>
<td>Al₂O₃ (%)</td>
<td>13,98</td>
<td>14,24</td>
<td>15,1</td>
<td>14,6</td>
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<tr>
<td></td>
<td>14,03</td>
<td>14,53</td>
<td>15,1</td>
<td>14,7</td>
</tr>
<tr>
<td></td>
<td>14,03</td>
<td>13,84</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>TiO₂ (%)</td>
<td>2,00</td>
<td>2,0</td>
<td>2,05</td>
<td>2,01</td>
</tr>
<tr>
<td></td>
<td>1,99</td>
<td>1,7</td>
<td>2,04</td>
<td>2,00</td>
</tr>
<tr>
<td></td>
<td>2,00</td>
<td>2,2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>MnO (%)</td>
<td>0,177</td>
<td>0,192</td>
<td>0,15</td>
<td>0,15</td>
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<tr>
<td></td>
<td>0,174</td>
<td>0,192</td>
<td>0,155</td>
<td>0,15</td>
</tr>
<tr>
<td></td>
<td>0,175</td>
<td>0,192</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Fig. IV - Precisions of some of the methods employed.

The precision of the methods can be seen to be better than 2%. The intermethod comparison gives an indication of the order of the accuracy anticipated. Analytically the lunar sample has been compared to ten local samples and in the case of the major constituents it would seem that one of the locally obtained basalts would make a reasonable standard for comparison purposes.

The analysis of 14321-222 was complicated by its heterogeneity and the basic inclusion on the otherwise homogeneous rock phase was identified as Fosterite (the analysis of which was obtained by mass spectrographic technique or 3 mg of the separated inclusion which was isolated under a microscope). The actual inclusion will be shown on a photograph and the analysis of the Fosterite is included in the abstract for interest.

A more complete analysis of the three lunar samples will be available for presentation at the conference.
X-Ray-Studies of Plagioclases and Pyroxenes
H.Jagodzinski and M.Korekawa, Institut für Kristallographie und Mineralogie der Universität München, 8 München 2, Theresienstr. 41, B.R.D.

Plagioclase

16 crystals from Apollo 14 lunar rocks 14310.106 were studied by single crystal X-ray diffraction methods. Besides Weissenberg and Buerger precession cameras a monochromatically focussing single crystal camera with high resolution power (1), (2) was used. Crystal No. 2 shows that it consists practically of only two individuals described by a complex "albite-Karlsbad" twin law (mirror plane (010) and rotation axis [001]) and crystals No. 5 and No. 15 show also only a pair of twins after the albite law. With these exceptions the other crystals investigated show three or four individuals after albite and "Karlsbad" (rotated around [001]) twin laws.

On precession and Weissenberg photographs 'a' and 'b' or 'a', 'b' and 'c' (diffuse in direction b*) were observed. Crystal No. 5 shows additionally 'e' and 'f' reflections. Weissenberg photographs of the crystal (0.1 x 0.1 x 0.1 mm) were carefully taken with better collimated incident X-rays. On extremely long exposed photographs (200 hours) the following characteristics on the 'a', 'b' and 'e' reflections were observed:

On 0kl photographs the position of 'e' reflections is symmetric with respect to 'b' reflections (like satellites). On 4k1 photographs the same is not true, additionally 'a' reflections are split into two reflections. This is characteristic for an "incoherent" two-phase pattern of submicroscopically intergrown phases with different An-contents.

The analogous position of reflections were observed on the photographs of some corresponding terrestrial specimens An(70)~An(77), in which approximately An(65) and An(80) were determined as exsolution end members (3). An(65) has intermediate structure and An(80) anorthite-like structure with P-lattice. a and c axes of both phases are parallel to each other and the boundary between these two phases is most probably (010). Whether the formation of twins and the exsolution are correlated or not is not yet known.

Single crystal photographs of high resolution power of crystal No. 6
X-Ray-studies of Plagioclases and Pyroxenes

H. Jagodzinski and M. Korokawa

were taken in a focusing camera. 'a' reflections observed were more diffuse than the corresponding reflections of terrestrial specimens. No super-satellite reflections were observed.

Pyroxenes

X-ray studies of 6 crystals of clinopyroxene (pigeonite) and two single crystals of orthopyroxene of lunar sample 14310.106 have been investigated by optical and X-ray methods, applying various single crystal techniques including focusing or semi-focusing devices. The following lattice constants have been determined for the pigeonites:

\[
a = 9.66, \quad b = 8.38, \quad c = 5.219, \quad \beta = 108.8^\circ; \quad \text{space group } \frac{2}{1}.
\]

The lattice constants differ slightly from previous investigations of lunar pyroxenes, indicating a lower Ca- and higher Fe-contents. Most of the crystals showed a more or less pronounced diffuseness of the reflections (hkl) with h+k=2n, which is complementary to the diffuse reflections h+k=2n+1 reported elsewhere (4), (5). On the other hand they were also different from the streaks connecting the pigeonite and augite reflections (6). Diffuse streaks originating from the pigeonite reflections point into the direction where augite reflections are expected, sometimes a very diffuse reflection has approximately the position of the expected augite reflection.

Orthopyroxenes are very rare in sample 14310.106. Although the reflections are generally less diffuse than the corresponding reflections of the pigeonites described above. It could be shown, that the diffuseness of reflections is much more pronounced for reflections with large k indices when compared with reflections with big values of h. The dependence of diffuseness as a function of l has not yet been determined. A marked heterogeneity of reflection intensities is correlated with the broadening of reflections.

X-ray photographs with focusing monochromators show the typical effects of "polygonization" in extremely small parts of the crystal. The splitting of Bragg-reflections observed for special values of (hkl) is apparently anisotropic. The polygonization pattern shows very sharp maxima, which seems to be due to curvatures in certain areas of the crystal. All maxima are connected by diffuse intensities. Thus is seems to be improbable, that the effects observed are due to different orientations of mosaic blocks Additional reflections of lower intensity are also observed, the interpretation of which can not yet been given.

The theory of X-ray diffraction of lamellar phase separation (2 phases) has been developed in case of congruent and incongruent planes of intergrowth. If the averaged thickness of lamellae exceeds approximately 100 Å, the calculated diffraction pattern is satisfactorily described by an incoherent superposition of both diffraction pictures.
Thin lamellae cause diffuse scattering, which differs appreciably from the incoherent model: Asymmetric streaks accompany the Bragg reflection, and in case of an averaged thickness of a few lattice constants, the diffraction pattern of the thin lamellae is replaced by diffuse bands, the asymmetry of which is described by the imaginary part of intensity constants. Diffuse streaks connect the reflections of the two phases, they are straight in case of congruent planes of intergrowth, but they may be curved if incongruent planes of intergrowth occur.

Lunar An-rich feldspars show exsolution lamellae with incongruent planes and lunar pyroxenes congruent and possibly incongruent planes of intergrowth. We thank Prof. Laves and his coworkers for stimulating discussions.

Literature

(1) S.R.Clark, M.Ross and D.E.Appleman Amer. Mineral. 56 (1971) 888-908
(2) H.Jagodzinski and M.Korekawa Naturwissenschaften (1965) 52, 640
to be published
(3) M.Korekawa
Amer. Mineral. 54 (1969) 725
PETROLOGY OF MARE FECUNDITATIS. Petr Jakes, Lunar Science Institute; W. I. Ridley*, Arch M. Reid, Jeff Warner, Russell S. Harmon, Robin Brett, NASA Manned Spacecraft Center; and Roy W. Brown, Lockheed Electronics, Houston Texas

Two 0.025 gm samples of soil returned by the Luna 16 probe from Mare Fecunditatis have been investigated by optical and electron microprobe techniques. Analyses for major elements were made on 1020 minerals and 266 glasses. The two soils are from the near surface (horizon A) and from a depth of approximately 30 cm (horizon D). No significant differences between the soils were noted.

Particles in the soil are of four types: 1) mineral fragments, 2) glass, some containing mineral fragments and/or crystallites, 3) agglutinates and microbreccias, and 4) crystalline lithic fragments largely with ophitic to subophitic texture and consisting dominantly of plagioclase, clinopyroxene, ilmenite, olivine, rare ulvospinel, Ni-Fe, troilite. These fragments apparently represent the mare basalts indigenous to the area.

The major minerals in the soil in order of decreasing abundance are, pyroxene 40%, feldspar 29%, oxides 19%, olivine 11%, metal + sulfide 1%.

Glasses. Average compositions of the major glass components are shown in Table 1. 23% of the glasses are Al-rich, Fe, Cr-poor basaltic glasses with lower Ca/Al ratio and higher Mg/Mg+Fe (.69) compared to the remaining glasses. These glasses have Ca/Al ratios almost identical to high-Al basaltic glasses from the Fra Mauro soil, and are considered to have a non-mare origin.

Two major non-mare types are recognized: 1) Basaltic glasses with 21-30 wt.% Al2O3. These comprise 22% of all analyzed glasses, and have the major element composition of feldspathic basalt or anorthositic gabbro. 2) Glasses with over 30 wt. % Al2O3. These comprise 1.3% of analyzed glasses and include gabbroic anorthosite and anorthosite compositions. Glasses with KREEP compositions are virtually absent in our sample of the Luna 16 soil. Two glasses have a granitic composition similar to the rare granitic glasses in the Fra Mauro, Apollo 11, and 12 soils. Remarkably low Na2O is a characteristic feature of the granitic glasses.

Fe, Ti, Cr are higher, and Ca, Al lower in the majority of glasses, compared to those described above, and they are considered to be mare-derived. The mare glasses are divided into two groups (Table 1), a major group named Fecunditatis type A (70% of all glasses) with less than 5% TiO2, and a minor group named Fecunditatis type B (7% of all glasses) with greater than 5% TiO2. This latter group has up to 22 wt. % FeO and most closely resembles high-titania glasses from the Apollo 11 soil.

Type A glasses have a preferred composition (Table 1) that is distinctly different from other mare basaltic glass compositions. In particular they have less FeO, TiO2, more Al2O3, and lower Ca/Al ratio (.76?) than other mare basalts. In most elemental abundances and ratios it is intermediate between other mare basalts and non-mare basalts.

Pyroxenes. Augite, sub-calcic augite and pigeonite are most abundant, and only 2 low-Ca pyroxenes may be orthopyroxenes. Ferro-augite and ferropigeonite are less
abundant, pyroxferroite and ferrohedenbergite rare. Two pyroxene trends are observed in lithic fragments, although more may exist. One trend from augite cores, through low Ca ferro-augite to pyroxferroite rims, and a second trend from pigeonite cores through ferropigeonite to pyroxferroite rims. Pigeonites and augites contain significant Ti, Al', Cr as in Apollo 11 and 12 clinopyroxenes. Most pyroxenes plot along an Al:Ti = 2:1 line similar to Apollo 11 clinopyroxenes, although a few grains with low Al, Ti plot near to the Al:Ti = 4:1 line indicating some octahedral alumina.

Plagioclase. Plagioclase compositions vary from An99 to An68, but 60% are in the range An90-An99, and show less than 2 mole % Al variation in individual grains. Fe varies from 0.18-1.8 wt. % and generally the most calcic plagioclases contain the lowest Fe. The plagioclases are quite distinct from the more sodic plagioclases in Apollo 11 high-K basalts but are undistinguishable from plagioclases in Apollo 11 low-K basalts and Apollo 12 basalts.

Opaque Minerals. The following minerals occur in the soil: ilmenite, Cr-Al ulvospinel, Ti-Al-Cr spinel, Fe-Ni and troilite. The majority of ilmenites lie within the compositional range of other mare ilmenites, but a few with greater than 2 wt. % MgO may indicate a higher Mg/Mg+Fe ratio in the melt on crystallization. Cr-Al ulvospinels range from Ulv57 Cr30 Hc13 to Ulv94 Cr1 Hc5 and generally fall in the range of Apollo 12 ulvospinels. The range in Ti-Al-Cr spinels is FeCr2O4 40-34 mole %, FeAl2O4 32-15 mole %, Fe2TiO4 29-47 mole %, and most resemble Apollo 11 spinels. The high Al2O3 contents of these spinels may reflect the high Al2O3 content of the Fecunditatis type A basalts.

Olivine. 75% of all analyzed olivines fall in the range Fo57-Fo68. Rare grains have Fo75 compositions. Only 8% of the grains are more fayalitic than Fo31, the most iron-rich grain being of Fo11 composition. Zoning is not extensive, rarely exceeding 2 mole % Fo. Mn varies from 0.1-1.4 mole % tephroite and is lowest in Mg-rich olivines. Overall the olivines are more iron-rich than those in Apollo 11 and 12 basalts.

Discussion. If we assume that a preferred glass composition approaches the composition of a rock type, then Fecunditatis type A basalts are unlike Apollo 11, 12 basalts, although the Mg/Mg+Fe ratios are similar. If the latter reflects degree of partial melting from a uniform source composition, then the higher Al2O3, and lower FeO, MgO, TiO2 and Cr2O3 in Fecunditatis type A basalts suggests their source region is unlike that of other mare basalts. Notably, some basaltic fragments with igneous textures indicate the earlier crystallization of plagioclase in Fecunditatis basalts, compared to Apollo 11 and 12 basalts. The intermediate chemistry of these basaltic glasses, between other mare basalts and non-mare glasses, indicates a source region that is more aluminous than that of the Apollo 11 and 12 basalts, but less aluminous than the source of non-mare rocks.
Luna 16: Composition of Glasses

<table>
<thead>
<tr>
<th></th>
<th>Granite</th>
<th>Anorthositic</th>
<th>Highland Basalt</th>
<th>Fecunditatis B Basalt</th>
<th>Fecunditatis A Basalt</th>
<th>Weighted average</th>
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</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>77.57</td>
<td>43.02 (2.79)</td>
<td>45.20 (1.68)</td>
<td>41.53 (2.36)</td>
<td>44.57 (2.64)</td>
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<tr>
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<td>n.d.</td>
<td>.08 ( .08)</td>
<td>.54 ( .51)</td>
<td>7.00 (2.44)</td>
<td>2.62 (1.21)</td>
<td>2.45</td>
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<tr>
<td>Al₂O₃</td>
<td>10.76</td>
<td>35.34 (1.79)</td>
<td>25.28 (2.33)</td>
<td>11.94 (2.21)</td>
<td>15.92 (3.10)</td>
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</tr>
<tr>
<td>Cr₂O₃</td>
<td>.02 ( .03)</td>
<td>.11 ( .06)</td>
<td>.22 ( .07)</td>
<td>.23 ( .07)</td>
<td>.20</td>
<td></td>
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<tr>
<td>FeO</td>
<td>1.42</td>
<td>1.06 (1.03)</td>
<td>6.49 (2.71)</td>
<td>18.15 (2.19)</td>
<td>14.43 (2.57)</td>
<td>12.81</td>
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<td>1.82 (2.03)</td>
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<td>9.01 (1.87)</td>
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<td>CaO</td>
<td>1.01</td>
<td>18.86 (1.54)</td>
<td>14.28 (1.49)</td>
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<td>11.86 (1.54)</td>
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<td>Na₂O</td>
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<td>.38 ( .16)</td>
<td>.32 ( .16)</td>
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<td>K₂O</td>
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<td>.05 ( .15)</td>
<td>.07 ( .11)</td>
<td>.14 ( .08)</td>
<td>.09 ( .08)</td>
<td>.09</td>
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<tbody>
<tr>
<td>CaO/Al₂O₃</td>
<td>.534</td>
<td>.565</td>
<td>.94</td>
<td>.745</td>
<td></td>
</tr>
<tr>
<td>MgO/MgO+FeO</td>
<td>.493</td>
<td>.560</td>
<td>.308</td>
<td>.385</td>
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</tbody>
</table>

<p>| | | | | | |</p>
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<th></th>
<th></th>
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<tr>
<td>No. analyzed</td>
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<td>16</td>
<td>64</td>
<td>14</td>
<td>170</td>
</tr>
<tr>
<td>Percentage of glass grains</td>
<td>0.4</td>
<td>1.3</td>
<td>21.6</td>
<td>7.2</td>
<td>69.9</td>
</tr>
</tbody>
</table>

n.d. = not determined
A PARTICLE OF NEARLY PURE NICKEL IN SOIL SAMPLE 12003.17

J.Jedwab, G.Naessens and R.Wollast
Brussels Free University - 50 Av.Roosevelt - Brussels 1050
BELGIUM

During the microscopical investigation of a polished mount of soil sample nr.12003.17 a particle of about 0.2 x 0.1 mm was noticed, owing to its complexity: it was composed of an aggregate of mineral and glassy fragments incompletely wrapped in a metallic foil of a few microns thickness (Fig.1). As the latter had a yellowish hue and a reflectivity in oil higher than that of the iron grains commonly found in the lunar soils (R=52.5% at 548 mm, as compared to 50.0% for iron) we performed EMP analyses of the metal. This proved to be almost pure nickel (Fig.2) with 2 to 5 % iron and 0.5 % cobalt (1). With the possibility of a perryite occurrence in mind we sought especially Si and P but with negative results.

Under the reflecting microscope, the metal presents the following characters. The outline exposed by the polishing is not continuous, suggesting that the foil has some holes. Transverse breakings are sometimes observed at foldings and are clearly mechanical. The walls are either straight and parallel, or show a swollen mossy structure towards the exterior. The metal appears also between some of the silicate grains, which is especially visible with the help of SEM and EMP.

Micro-hardness is dissymmetrically distributed, with the softer direction parallel to the foil elongation. Approximate value of Hv (for 2 and 5 g loads) = 234 kg mm-2. The metal contains no inclusions visible without etching. Owing to the transparency of the silicates, one could observe the inner surface of the metal in its untouched stage: cracks, scratches and imprints may be observed.

EMP-analyses of the enclosed silicates have shown that they matched compositions of lunar pyroxenes and felspars. This fact and the tightness of contact with the metal makes a glaring contamination during mount preparation unlikely.

After removal of a part of the enclosed minerals by means of the hardness tester; the innermost surface of the metal foil has been studied with the SEM. (Fig.3). The places left empty bear imprints, scratches and irregular radial cracks. The overall surface is a succession of holes and saddles or sharp crests. Similar morphologies have been described in impacted steels (2) and in the metallic particles of the Ramsdorf chondrite (3). A torn edge of the metal foil shows a fractured pattern perpendicular to the surface (Fig.4).
Nearly pure Nickel particle
J.Jedwab

P. 434

From the morphological point of view, the observations thus favor an impact origin for the described particle. But the authors are not aware of such a high nickel content in metallic particles from lunar soils or rocks(4). Among meteoritic materials found on Earth, the Oktibbeha find, with his 62% Ni may be quoted here, but this value is known to be a limiting one for cooling Fe-Ni systems(5). Very high nickeliferous particles are known especially among shock-metamorphosed materials (up to 91.9% Ni in spherules from the Henbury glass, with iron and cobalt amounts matching those of the lunar particle(6)) or among heat ablated ones(up to 77.5% Ni in oceanic spherules(7)). The mechanisms of Ni-enrichment advocated in such cases, i.e. differential oxidation or dissolution of the iron (8) are difficult to transfer to the lunar conditions. Accordingly, no natural mechanism for producing free large particles of nearly pure nickel can be proposed by the authors.

A study of more particles of similar compositions, purposely searched, is evidently needed before any sound hypothesis of origin could be enunciated.

NOTES AND REFERENCES

(1) We thank Professor K.Keil (Univ.of New Mexico, Albuquerque) for checking this unusual Ni-analysis and also for confirming the compositions of the enclosed particles.
(8) R.Brett (1967) Amer.Min. 52. 721.

(Figures on third page).

2. EMP view of the same. Ni distribution.

3. SEM, secondary mode. Lower half after repolishing and removal of silicates. (a): area enlarged in fig. 4. (b): Crack and imprint. (c): Scratches.

4. SEM-view of 3a area, rotated 180° and tilted. (a): Polished and (b): broken metal. (c): imprinted areas left by Pyl and Py2 in fig. 1. (d): same crack as in 3b. Inner surface of the metal is rough and finely pitted.
Cold cathode ionization gauges were included in the ALSEPs to measure the amount of gas on the moon. The gauge is a conventional magnetron gauge in a stainless steel envelope, with an axial magnetic field of about 900 gauss and an applied voltage of 4500 volts. The aperture of the gauge was covered but not sealed, and the cover was removed by remote command after the gauge was deployed.

The solar wind constitutes a steady source of lunar atmosphere as it impacts on the lunar surface and is neutralized. The particles should become imbedded in the surface on impact, but once the surface is saturated with a given constituent, that constituent ought to be released at the same rate as it is brought to the moon by the solar wind. After release from the surface, light particles - hydrogen and helium - escape within a few hours by virtue of their high thermal velocities, even at the temperature that prevails on the night side of the moon. Thermal escape for neon and heavier particles is slower than escape by a competing process - ionization followed by acceleration due to solar-wind fields. The lifetime for this latter process is the ionization lifetime, which is a few months. Neon ought to be the most prominent constituent of that part of the lunar atmosphere that is of solar wind origin, with a daytime concentration of about $6 \times 10^4$ cm$^{-3}$. Heavier constituents should be less prevalent because of their lesser abundance in the solar wind and lighter constituents less prevalent because of their more rapid thermal escape. Gases that do not escape rapidly by thermal escape and that do not condense or adsorb on the surface at lunar nighttime temperatures should distribute themselves in concentration according to a $T^{-5/2}$ law. This should apply to neon and give rise to a nighttime concentration about 26 times the daytime concentration.

A second possible source of lunar atmosphere is release from internal sources - the means by which the earth and presumably other terrestrial planets acquired their atmospheres. A release rate at the same average rate per unit mass as on earth would constitute a much more important source of lunar atmosphere than is the solar wind. Further, such release would probably be intermittent with occasional bursts of correspondingly greater amplitude, making them stand out above the background more than would be the case if the source were steady.

The most important result so far obtained from the cold cathode ionization gauge is that the nighttime concentration is very low, about $2 \times 10^5$ cm$^{-3}$ - almost an order of magnitude below the value expected for neon from the solar wind. This indicates that the lunar surface is not saturated with neon and that less neon is being released from the lunar surface than impinges on it as the solar wind impacts the surface. The observed low nighttime
concentration also indicates that contaminant gases from the lunar module, and possible other constituents of the daytime lunar atmosphere as well, are very efficiently adsorbed on the cold nighttime lunar surface.

Several transient gas events have been observed at both the Apollo 14 and 15 sites, and these form a pattern. A large gas release lasting a day or more occurred after the first lunar sunset in each case, with two smaller and shorter releases either superimposed on or following the big release. A few other events have also been observed, but the frequency of occurrence for Apollo 14 has fallen off markedly with time suggesting that they were caused by lunar hardware, not release from the lunar interior. Too few data are available from Apollo 15 to permit any such generalization in that case.

The gas concentration during the daytime has been greater than at night by about two orders of magnitude, undoubtedly due to release of adsorbed contaminant gases. The observed time constant for release is about two months. It is not yet clear whether steady-state daytime conditions have been reached at the Apollo 14 site and no data are yet available for the decay rate at the Apollo 15 site. The concentration is temperature dependent, the maximum occurring near midday.

The highest degree of contamination in the landing site after departure of the astronauts occurred at the first lunar sunrise. The sporadic release of gas from lunar hardware was especially prominent after the first lunar sunset and some of this gas became adsorbed on the lunar surface in the immediate vicinity. As the site became sunlit, the initial release of adsorbed gas was very rapid. On Apollo 14, the gas concentration rose two orders of magnitude in two hours at sunrise and then fell gradually for a few hours before leveling off. There is no evidence of any release of gas from the lunar interior during this time, although the record is somewhat noisy in a way that could obscure short bursts of a few minutes to an hours duration.

The release of adsorbed gases from the landing site as the sun came up must be considered as a possible source of the "water vapor" seen by the SIDE at this time. The SIDE is sensitive only to ions; its mass analyzer reportedly identified particles with a mass-to-charge ratio near 18 with 49.6 eV energy from about 19.5 to 33 hours after sunrise. The total ion detector noted a nearly monoenergetic group of ions with energies near 70 eV/unit charge; particularly notable is the absence of lower energy ions, as this indicates that the source of the ions cannot be photoionization of contaminant gas released from the landing site.

The observed gas concentration at the landing site was about $10^7$ cm$^{-3}$. Assuming that the concentration varied according to an inverse square law of distance from the LM, the gas in a unit column was about $10^{14}$ cm$^{-2}$. The photoionization rate for most gases is near $10^{-7}$ sec$^{-1}$, giving rise to a source strength of about $10^4$ ions cm$^{-2}$ sec$^{-1}$ from the contaminant cloud. Once formed, an ion is accelerated by the electric field associated with the solar wind and its embedded magnetic field; a typical value for this field is about 1 volt km$^{-1}$ and the usual scale size is much larger than the dimensions that we are concerned with here. Until an ion has accelerated over a distance
equal to a large fraction of its gyro radius, which is large compared to a lunar radius for a heavy ion, its motion is mainly dominated by the electric field rather than by its interaction with the magnetic field. Therefore, newly formed ions can be seen by the SIDE only if the electric field is accurately aligned with its field of view and in such a sense as to accelerate the ions towards the SIDE. The spectrum for contaminant gas should have most particles at low energy, because the gas cloud is densest near the landing site, permitting most ions to be accelerated only over short distances and thus acquiring only a few eV energy (90% of the gas is within the first kilometer, 99% within the first 10 km). The observed monoenergetic spectrum could not be formed by photoionization of such a cloud centered on the landing site; it could be formed by photoionization only if there were a localized gas cloud, not more than a few kilometers across, located of the order of fifty to one hundred kilometers above the landing site. Thus it does not seem possible to explain the SIDE observations in terms of contaminant gas from the landing site. The arguments also indicate that other gases of lunar origin are unlikely to explain the observations.

We are determining the low frequency velocities and damping factors of acoustic waves in rock powders in vacuo with a view to understanding the lunar seismic signals, which are clearly so different from the earth's.

Many features of the lunar seismic signals can be understood in terms of the Gold-Soter model,\(^1,^2\) in which the outer layer of the moon is largely composed of rock powder the compaction of which increases in some manner with depth. The long rise and fall times of the lunar seismic signals would result from the diffusion of acoustical energy through the powder where the scattering mechanism would arise from inhomogeneities such as the undulating lunar surface and powder density fluctuations. The energy would be confined to such a powder layer by what is essentially total internal reflection, the increasing compaction with depth providing the necessary increase in velocity with depth.

The diffusion process results in long path lengths which can only result in the observed large signals at long delays if the acoustical damping factor \(n(=1/2Q)\) is small. This factor thus emerges as the crucial rock powder parameter, though the severity of the constraint that it be small is eased by the low wave propagation velocities that occur in rock powders. It would seem that to satisfactorily account for the lunar seismic signals \(Q\) need not exceed 2000, and it could be less.

Measurements by Hunter et. al.\(^3\) at frequencies above 7 kHz give values of \(Q \sim 5\). We did not consider this a discouraging result because most of the energy in the lunar seismic signals is near 1 Hz, at which frequency, for a given wave amplitude, the particle strain is about a factor of \(10^7\) less than that at \(10^4\) Hz, which could result in much lower damping at the lower frequency. A result by Schmidt at 20 Hz\(^4\) also gave a \(Q\) value \(\sim 5\) but in this case adequate precautions against container damping had not been taken. None of this earlier work had been done in vacuo.
We are therefore measuring $Q$ in vacuo and as close to 1 Hz as is convenient, which at present is at about 10 Hz in well settled powder. The lowest limit is imposed upon us by the size of the laboratory and is about 6 Hz. We set up longitudinal standing waves in a horizontal trough of rock powder and observe the free decay. This trough is very compliant horizontally, is of low mass, and is very lightly supported by fine wires on low loss supports. Any residual support system effects can only decrease $Q$, therefore any result we obtain may be a lower limit on $Q$ of the powder. From measurements performed on the system we estimate that the support system probably limits values of $Q$ observed in vacuo to an absolute minimum of 1000. To test for support system effects we have worked with 2 different masses of powder in the trough and the results were essentially the same in each case. We are therefore not yet being limited by damping in the support system. Our results are summarized overleaf.

The best value of $Q$ obtained so far, 120 in as yet imperfect conditions, gives support to the model in which the observed seismic signals result from propagation through variously compacted layers of rock powder. We think it possible that when fluids are removed more thoroughly from the powder then values of $Q$ in excess of 1000 can occur, especially when the powder has been pre-compacted.

Low frequency acoustic compressional waves in the rock powder at the lunar surface travel at about 50 m/s and measurements exist that indicate that the overburden weight alone can increase this velocity to 3000 m/s at a depth of about 6 km, at which depth the material would still be essentially a powder. Little energy put in at the surface could penetrate such a layer. Date from first arrival times of seismic energy on the moon in fact correspond to propagation velocities of about 3000 m/s.

MEASUREMENTS OF THE ACOUSTICAL PARAMETERS OF ROCK
POWERS, AND THE GOLD-SOTER LUNAR MODEL.

Initial Author B. W. Jones

Summary table of results on acoustic wave
propagation in powdered basalt

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Effect on Q</th>
<th>Effect on c</th>
</tr>
</thead>
<tbody>
<tr>
<td>Powder support system.</td>
<td>Negligible at intrinsic powder Q values below 1000.</td>
<td>Negligible.</td>
</tr>
<tr>
<td>Ageing: settling and partial removal of fluid films</td>
<td>6x increase, at least.</td>
<td>50% increase at least.</td>
</tr>
<tr>
<td>Frequency</td>
<td>&lt; 5% change over first 3 modes at least.</td>
<td>&lt; 1% change over first 4 modes at least.</td>
</tr>
<tr>
<td>Amplitude</td>
<td>2x increase between 100μm and 10μm, thereafter &lt; 5% down to at most 0.01μm.</td>
<td>Negligible.</td>
</tr>
<tr>
<td>Ambient air pressure.</td>
<td>2x increase between 760 Torr and 0.1 Torr, thereafter &lt; 5% increase down to 0.03 Torr.</td>
<td>Negligible.</td>
</tr>
</tbody>
</table>

Result at low amplitudes on aged powder at < 0.1 Torr pressure:

density = 1.34±0.01 gm/cm³, (c.f. 1.9-1.9 gm/cm³ for the lunar surface).

longitudinal wave velocity = 47.3±0.5 m/s, (c.f. 45 m/s for the surface layer at Surveyor sites).

Q = 118±5
RARE GAS MEASUREMENTS IN THREE APOLLO 14 SAMPLES, W.A. Kaiser, Univ. of California at Berkeley, Physics Dept., Berkeley, Calif., 94720.

Ne, Kr, and Xe were measured in the breccias 14066 and 14318 and in igneous rock 14310. The breccia 14066—the consortium sample under the supervision of John H. Reynolds—had an extremely low solar wind contribution, thus we measured essentially pure cosmogenic produced Ne, Kr, and Xe. Cosmogenic spectra as well as a $^{81}$Kr-$^{83}$Xe exposure age were obtained. The cosmogenic $^{131}$Xe/$^{126}$Xe ratio determined is the lowest yet found in a lunar sample, probably resulting from a high contribution of REE elements (cosmogenic $^{130}$Xe/$^{126}$Xe ratio < 0.9). The breccia 14318—consortium leader, M. Tatsumoto—contained predominantly a solar wind contribution. The igneous rock 14310 contained essentially pure cosmogenic produced rare gases. The $^{131}$Xe/$^{126}$Xe ratio is ca. 7. The rock displayed essentially Ba produced cosmogenic Xe. The $^{81}$Kr-$^{83}$Kr exposure age found is $270 \times 10^{6}$y.

COSMOGENIC Kr AND Xe VERSUS DEPTH IN THE ESTHERVILLE MESOSIDERITE; FURTHER PROOF THAT THE LUNAR $^{131}$Xe ANOMALY IS A DEPTH EFFECT, W.A. Kaiser, Univ. of California at Berkeley, Physics Dept., Berkeley, Calif., 94720.

Previous studies using lunar rocks to determine cosmogenic Kr and Xe ratios in depth studies have had two critical disadvantages. First, depth effect experiments are limited by the small size of the lunar samples. Second, because of the "gardening" effect on the lunar surface, whole layers of material shift easily. As a consequence, the lunar samples have been irregularly exposed to cosmic rays. These two disadvantages were eliminated by using three core samples from different depths of one of the bigger bodies of the Estherville samples. A correlation between the cosmogenic $^{131}$Xe/$^{126}$Xe ratio and the $^{78}$Kr/$^{83}$Kr ratio was found. The cosmogenic $^{131}$Xe/$^{126}$Xe increased with depth and the $^{78}$Kr/$^{83}$Kr ratio decreased with depth confirming the observations of EBERHARDT et al. [1] and BOCHSLER et al. [2]. The cosmogenic Kr and Xe spectra as a function of depth are included in this paper.


RARE GAS STUDIES IN LUNA-16-G-7 FINES BY STEPWISE HEATING TECHNIQUE: A LOW FISSION SOLAR WIND Xe, W. A. Kaiser, Univ. of California at Berkeley, Physics Dept., Berkeley, Calif., 94720.

He, Ne, Ar, Kr, and Xe were examined in a dust sample ($<125 \mu$m) of Luna-16 in twelve temperature steps with especially small intervals in the low temperature range (80°C steps). The gas concentrations, as well as their relative abundances, are in general agreement with values reported by VINogradov [1] for Luna-16 and values found in Apollo 11 fines [2-8] except for Ne.

The $^{3}$He/$^{4}$He ratios decrease with increasing release temperatures as observed in Apollo 11 fines studies [2] [3]. The initial $^{3}$He/$^{4}$He ratio is 17.6% higher than the total $^{3}$He/$^{4}$He ratio, and thus similar to the results obtained by PEPE et al. (20.7%) [3] but lower than the 36% found by the Berkeley group in Apollo 11 fines [2].
The $^{20}$Ne/$^{22}$Ne ratios are generally lower than found by Hohenberg et al. [2] and Pepin et al. [3] in Apollo 11 fines, but oscillate in the same manner as observed in those studies. The $^{21}$Ne/$^{22}$Ne ratio increased up to an outgassing temperature of $160^\circ$C and then decreased, reaching its lowest value at $400^\circ$C.

The $^{38}$Ar/$^{36}$Ar ratio followed generally the same pattern but was of less statistical significance. The $^{40}$Ar/$^{36}$Ar ratio had its minimum at $550^\circ$C of $0.710 \pm 0.003$. The K-Ar age, based on the "excess" $^{40}$Ar and a K-value of 880 ppm [9], is unreasonably high ($6.73 \times 10^9$y). The solar wind Kr composition is within the error bars, in agreement with the results obtained by Eberhardt et al. [4]. No $^{129}$I was seen. The "trapped" Xe is lower in fission-affected isotopes than the Xe from Apollo 11 fines [4] and also lower than the $1000^\circ$C fraction of Marti's Pesyanoe measurements [10]. On the other hand, the non-fission-affected isotopes are in good agreement with values obtained by Eberhardt et al. [4]. AIR-Xe [11] is interpreted as a fractionated solar wind Xe with the composition found in this study. The percentage of the fractionation is $4.12 \pm 0.5$ percent per atomic mass unit. An exposure age based on $^{83}$Kr and $^{128}$Xe and the chemical abundances of Sr, Y, Zr, Ba, and Ce (REE) [9] was estimated which is definitely higher than that for the Apollo 11 landing site.

REFERENCES:
THE AVERAGE $^{130}$Ba(n,γ) CROSS SECTION AND THE ORIGIN OF $^{131}$Xe ON THE MOON, W.A. Kaiser, Univ. of California at Berkeley, Physics Dept., Berkeley, Calif., 94720, and B. L. Berman, Univ. of California, Lawrence Livermore Laboratory, Livermore, Calif., 94550.

The average $^{130}$Ba(n,γ) cross section has been measured for a neutron spectrum similar to the one at the lunar surface. The pulsed-neutron facility at the Livermore 100-MeV Electron Linear Accelerator served as the source of radiation. The spectrum and flux of neutrons were measured simultaneously with the sample irradiation by the neutron time-of-flight technique. Chemically pure natural BaCl$_2$ samples were irradiated with a total integrated neutron flux of $1.6 \times 10^{13}$ neutron/cm$^2$. The $^{131}$Xe resulting from neutron capture on $^{130}$Ba was measured mass spectrometrically. The results show that the integrated product of the neutron flux and the neutron-capture cross section is equal to $1.8 \times 10^{-10}$, whence the average cross section $\bar{\sigma} = 12$ b. If, however, the energy range is restricted to the most likely region for large capture resonances (1 - 1000 eV), the resulting $\bar{\sigma} = 220$ b. This large value enables us to conclude that the anomalously high concentration of $^{131}$Xe in lunar rocks probably has been produced via this reaction.

Work performed in part under the auspices of the U.S. Atomic Energy Commission.
ANALYSIS OF LUNAR LASER ALTIMETRY. W.M. Kaula, G. Schubert, Univ. of California, Los Angeles, 90024; W.L. Sjogren, Jet Propulsion Lab., Pasadena, California, 91103; H.D. Wollenhaupt, Manned Spacecraft Center, Houston, Texas, 77058

The Apollo 15 lunar altimeter obtained a total of about 5 revolutions of good data. The principal indications of this limited sample are as follows.

- The mean radius of the moon is 1737.0 km.
- The center-of-mass is displaced from the center-of-volume of the moon 2 km in a direction 37° east of the direction of the earth.
- The ellipticity of the moon's surface has about 700 m amplitude, with the long axis at about 90° longitude with respect to the earth.
- The mean height of the terrae above the maria is about 3 km with respect to the center-of-mass.
- A depression 1400 km wide and 2 km deep with respect to the 1737 km sphere (or 5 km deep with respect to the surrounding highlands) is centered at 180° longitude.
- The depths of the ringed maria Serenitatis, Crisium, and Smythii are inversely correlated with their widths.
- The maria are extremely level, ranging not more than ±150 meters over some stretches of 200 to 600 kilometers.
- The terrae are quite rough, with a median change of about 600 meters for the measurement interval of about 40 km, and frequent changes of more than 1500 m.
GAMMA RAY MEASUREMENTS OF APOLLO 12, 14, & 15 LUNAR SAMPLES

J. E. Keith, R. S. Clark and K. A. Richardson*
NASA MSC

The results of the gamma ray spectroscopy (see O'Kelley, et. al., 1970) of two Apollo 12 samples, 18 Apollo 14 samples and 24 Apollo 15 samples are presented in Tables 1 and 2. These measurements made during preliminary examinations were counted about 2500 minutes; those made afterward were counted about 5000 minutes. However, all counting times were adjusted according to the activity of the sample and to operational requirements within the LRL.

The sample spectra were resolved into their components by comparison with spectra obtained from various distributed sources, by a computer program, STRIP, which subtracts successively proportionate amounts of the various standard spectra from the sample spectrum. Comparable results are obtained by least-squares methods. The errors listed in the tables are one standard deviation and include the estimated error in the standards and an estimate of the error due to inappropriateness of the standard, as well as the counting statistics error. In many cases, the inappropriateness of the standard is the most important source of error. In the twelve cases where the sample was measured more than once, the weighted average of the results and its standard deviation are listed. An upper limit is listed in those cases where no peak can be distinguished and is reported as three standard deviations.

Apollo 14 samples tended to be higher in potassium, uranium, and thorium than samples from either Apollo 12 or Apollo 15, and this made the determination of cosmogenic radionuclides, notably Mn\textsuperscript{54}, difficult. The ratios of primordial radionuclides remain roughly what has been seen previously. For instance, the K/U ratio of Apollo 14 fines is about 1200, and that of Apollo 15 fines about 1400, and the Th/U ratio is about 3.7 for both Apollo 14 and 15.

The levels of cosmogenic activity reflect the chemical composition, the degree of exposure to solar and cosmic radiation, and the surface to volume ratio of the sample. In addition, the Co\textsuperscript{56} and Mn\textsuperscript{54} are somewhat higher in Apollo 14 samples, due to the large proton flare which immediately preceded that mission. The variation of Al\textsuperscript{26}, Na\textsuperscript{22} and Co\textsuperscript{56} with depth in the Apollo 14 trench fines suggest that the bottom and middle of the trench were contaminated by surface materials in contrast to the Apollo 15 trench samples. Three Apollo 15 breccias have Al\textsuperscript{26}/Na\textsuperscript{22} ratios less than or equal to one, and are probably unsaturated with respect to Al\textsuperscript{26}.

*Now with the Geological Survey of Canada, Ottawa, Canada.
### TABLE 1 - APOLLO 12 AND 14 RESULTS

<table>
<thead>
<tr>
<th>Sample</th>
<th>Wt. (gms)</th>
<th>Th (ppm)</th>
<th>U (ppm)</th>
<th>K (percent)</th>
<th>$^{26}$Al (dpm/Kg)</th>
<th>$^{22}$Na (dpm/Kg)</th>
<th>$^{54}$Mn (dpm/Kg)</th>
<th>$^{56}$Co (dpm/Kg)</th>
<th>$^{46}$Sc (dpm/Kg)</th>
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<tbody>
<tr>
<td>14066</td>
<td>497.5</td>
<td>15.3±1.3</td>
<td>4.2±.2</td>
<td>.72±.02</td>
<td>103±6</td>
<td>43±6</td>
<td>5±12</td>
<td>51±7</td>
<td>6±3</td>
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<tr>
<td>14070</td>
<td>1370.0</td>
<td>13.2±1.0</td>
<td>3.6±.5</td>
<td>.604±.006</td>
<td>62±18</td>
<td>27±6</td>
<td>&lt;33</td>
<td>8±2</td>
<td>.4±.5</td>
</tr>
<tr>
<td>14081</td>
<td>380.6</td>
<td>13.9±1.7</td>
<td>3.8±.2</td>
<td>.533±.010</td>
<td>74±13</td>
<td>46±11</td>
<td>4±13</td>
<td>3±6</td>
<td>.6±1.6</td>
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<td>14082</td>
<td>600.2</td>
<td>12.0±.5</td>
<td>3.27±.14</td>
<td>.49±.03</td>
<td>117±7</td>
<td>41±3</td>
<td>10±11</td>
<td>28±10</td>
<td>4±3</td>
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<td>14083</td>
<td>1100.0</td>
<td>12.7±.8</td>
<td>3.6±.2</td>
<td>.402±.015</td>
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<td>38±7</td>
<td>16±6</td>
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<td>&lt;4</td>
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<td>14085</td>
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<td>13.8±1.3</td>
<td>3.6±.4</td>
<td>.39±.03</td>
<td>139±19</td>
<td>84±9</td>
<td>&lt;70</td>
<td>80±20</td>
<td>5±3</td>
</tr>
<tr>
<td>14086</td>
<td>115.0</td>
<td>8.8±.7</td>
<td>2.1±.08</td>
<td>.328±.007</td>
<td>146±16</td>
<td>58±3</td>
<td>&lt;28</td>
<td>52±10</td>
<td>4±3</td>
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<tr>
<td>14087</td>
<td>63.0</td>
<td>4.2±.3</td>
<td>1.24±.11</td>
<td>.206±.009</td>
<td>120±13</td>
<td>55±4</td>
<td>6±11</td>
<td>34±10</td>
<td>1.6±1.6</td>
</tr>
</tbody>
</table>

#### CLASTIC ROCKS

#### CRYSSTALINE ROCKS

#### FINES

#### TWO APOLLO 12 ROCKS
### Table 2 - Apollo 15 Results

<table>
<thead>
<tr>
<th>Sample</th>
<th>Weight (grams)</th>
<th>Th (ppm)</th>
<th>K (percent)</th>
<th>U (ppm)</th>
<th>26Al (dpm/kg)</th>
<th>54Mn (dpm/kg)</th>
<th>46Sc (dpm/kg)</th>
<th>56Co (dpm/kg)</th>
<th>22Na (dpm/kg)</th>
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<tbody>
<tr>
<td>Clastic Rocks</td>
<td></td>
<td></td>
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<tr>
<td>15206</td>
<td>85.5</td>
<td>12.0±1.3</td>
<td>3.2±4</td>
<td>0.487±0.018</td>
<td>40±5</td>
<td>49±5</td>
<td>7±8</td>
<td>&lt;30</td>
<td>&lt;8</td>
</tr>
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<td>15205</td>
<td>334.4</td>
<td>12.0±4</td>
<td>2.9±5</td>
<td>0.440±0.008</td>
<td>48±6</td>
<td>48±4</td>
<td>50±30</td>
<td>&lt;13</td>
<td>&lt;7</td>
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<tr>
<td>15465</td>
<td>364.9</td>
<td>5.9±11</td>
<td>1.46±13</td>
<td>0.234±0.004</td>
<td>120±30</td>
<td>56±14</td>
<td>31±18</td>
<td>&lt;19</td>
<td>&lt;5</td>
</tr>
<tr>
<td>15265</td>
<td>314.2</td>
<td>5.05±12</td>
<td>1.27±07</td>
<td>0.211±0.08</td>
<td>72±8</td>
<td>37±3</td>
<td>12±15</td>
<td>8 ± 6</td>
<td>&lt;15</td>
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<tr>
<td>15558</td>
<td>1333.3</td>
<td>3.42±18</td>
<td>1.01±04</td>
<td>0.170±0.006</td>
<td>84±5</td>
<td>36±5</td>
<td>23±5</td>
<td>9 ± 3</td>
<td>3.0 ± 0.7</td>
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<tr>
<td>15255</td>
<td>240.4</td>
<td>3.5±3</td>
<td>0.92±0.07</td>
<td>0.156±0.019</td>
<td>111±7</td>
<td>43±4</td>
<td>26±3</td>
<td>11±8</td>
<td>4 ± 4</td>
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<td>15456</td>
<td>117.8</td>
<td>3.5±2</td>
<td>0.86±0.08</td>
<td>0.156±0.004</td>
<td>79±8</td>
<td>36±4</td>
<td>4±5</td>
<td>5 ± 4</td>
<td>0.5 ± 1.3</td>
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<td>15086</td>
<td>172.1</td>
<td>3.2±2</td>
<td>0.76±0.03</td>
<td>0.143±0.003</td>
<td>39±6</td>
<td>50±6</td>
<td>22±6</td>
<td>11±3</td>
<td>2.5 ± 0.7</td>
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<tr>
<td>15459</td>
<td>92.0</td>
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<td>9±6</td>
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<td>0.143±0.002</td>
<td>73±13</td>
<td>58±12</td>
<td>29±17</td>
<td>12 ± 4</td>
<td>4.1 ± 1.3</td>
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<td>15431</td>
<td>145.4</td>
<td>4.86±0.15</td>
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<td>0.186±0.005</td>
<td>66±7</td>
<td>36±4</td>
<td>&lt;12</td>
<td>12 ± 12</td>
<td>3 ± 4</td>
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CHONDRULES IN APOLLO 14 BRECCIAS AND ESTIMATION OF LUNAR SURFACE EXPOSURE AGES FROM GRAIN SIZE ANALYSES; Elbert A. King, John C. Butler and Max F. Carman, Dept. of Geology, Univ. of Houston, Houston, Texas, 77004.

Numerous spherical, subspherical and rounded bodies occur in at least three Apollo 14 breccia samples. Many of these objects are identical in texture, size and general mineralogy to common varieties of meteoritic chondrules. These lunar chondrules are not the relatively simple glass spherules that have been observed and described by many investigators in lunar samples from previous Apollo Missions, although some of the chondrules are partly glass and others are devitrified glass in part. The textures observed in lunar chondrules include spherical aggregates of mineral grains with indistinct margins; spherical brown glass chondrules enclosing euhedral olivine and/or pyroxene; spherical chondrules with pyroxene and plagioclase crystals that nucleate on the surface of the sphere and radiate into the interior of the chondrule with interstitial brown and turbid glass; and less perfectly spherical chondrules with basaltic, microbreccia and other fine-grained rock textures. The chondrules range in size from less than 0.1 mm. to more than 1.0 mm. Chondrules are abundant in samples 14313, 14318 and 14301 occupying more than 10% by volume of the observed portions of these samples. Rare to moderately abundant chondrules have been observed in samples 14305, 14306, 14311 and 14162. All of these samples contain numerous highly shocked mineral grains and rock fragments.

Why are chondrules present in Apollo 14 samples and are not recorded in the Apollo 11 or Apollo 12 samples? The major difference between the Apollo 14 and the Apollo 11 and 12 landing sites probably is that the Apollo 14 landing site is on the Fra Mauro Formation, believed to be mostly crater ejecta from the Imbrian event, and the Apollo 11 and 12 landing sites are on relatively undisturbed extrusive rocks that have not been involved in a large impact. Thus, we believe that the chondrules in the Apollo 14 samples may have been produced by the huge impact and accompanying phenomena that formed the Imbrian Basin. It does not appear that all of the chondrules have formed by the same mechanism. At least three different mechanisms may have formed chondrules in the Apollo 14 breccias: (1) Crystallization of shock-melted silicate droplets. This mechanism requires that pre-existing lunar rock, or a mixture of surface silicate materials, be melted by the energy of an impact, that the fluid silicate forms a roughly spherical shape due to surface tension, and that the silicate body subsequently crystallizes. The crystallization may have occurred after supercooling, as has been experimentally demonstrated by Keil, Blander and co-workers, or possibly more slowly as devitrification of a predominantly glass chondrule in fall back or base surge deposits that were at elevated temperatures; (2) Rounding of rock clasts in base surge deposits. The base surge deposits from the Imbrian event must have traveled across the lunar surface for as much as several hundred km. before final deposition. Abrasion and interaction of clasts and particles in the base surge may have produced consid-
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erable rounding of some rock fragments prior to deposition. This mechanism may have produced many of the multimineralic chondrules that seem to have igneous and microbreccia rock textures; (3) The inclusion of rock fragments and clasts in base surge and fall back deposits of greatly different bulk chemical composition may have caused rapid diffusion of cations between the surrounding deposit and the clasts, particularly if the deposits were initially hot. This mechanism is less certain and more speculative than the first two, but is the only process that occurs to the authors to explain the chondrules that appear to be clasts of dunite surrounded by fine-grained matrix in which there appear to be concentric diffusion halos around the clasts. Regardless of the exact mechanisms that form chondrules in Apollo 14 samples, the fact remains that at least three major types of chondrules are present in lunar samples and that these same types of chondrules are common in chondritic meteorites.

The suggestion that chondrules in the lunar samples are the result of processes accompanying large impacts leads to the possibility that at least some, perhaps many, meteoritic chondrules may have been formed by the same impact-related mechanisms. If some meteoritic chondrules are produced in large impacts, this requires that at least one of the bodies involved in the event be a fair-sized planetary object, especially for mechanisms (2) and (3) to be valid. Where could these impacts have occurred? Probably the largest iron and silicate planetary body that ever existed in the Solar System was the proto-sun, the initial accumulation of matter at the center of the Solar System that was still accreting, but had not yet gained sufficient mass to become a star. Very large volumes of chondritic rocks may have been produced on the surface of the proto-sun by impacts of accreting objects, but the energy required to get pieces of this chondritic material away from the proto-sun into outer regions of the Solar System, distances of two or three AU and more, is extremely great. This mechanism not only requires very great energy, but also requires the movement of material away from the proto-sun early in Solar System history. This direction of movement probably would be contrary to the direction of movement of most other Solar System debris at that time. However, at the time that stellar activity was initiated in the Sun, perhaps some of the chondritic outermost surface could be propelled far away from the center of the Solar System in the same general direction of movement as much other Solar System debris. If the large impact hypothesis for the origin of chondrules is correct, chondrules and chondritic rocks may be an inescapable result of the terminal stages of accretion of all of the silicate-iron planets and large moons.

A total of 26 lunar fines samples (less than 1 mm.) has been sieved, and the grain size frequency distributions of the samples have been determined. These samples include three surface samples from Apollo 11, ten surface and trench samples from Apollo 12, and thirteen surface, trench and core tube samples from Apollo 14. In general, the grain size frequency distributions are characterized by very poor sorting, platykurtosis, and low values of skewness. Samples from Apollo 14 contain both the coarsest and the finest fines samples yet analyzed. A fines sample collected on the flank of Gone Crater is the coarsest sample with a graphic mean size of 112 microns. If only relatively undisturbed samples that were collected away from crater rims and coarse ejecta blankets are considered, the average graphic means of the Apollo
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11, 12 and 14 fines samples are 63 microns, 69 microns and 49 microns respectively. Therefore, it seems probable that the graphic mean size (and other size distribution indices) correlate with the accepted ages of the assumed lunar surface material at different landing sites. Specifically, the graphic mean size of the surface regolith fines decreases with increasing meteoroid bombardment exposure age, as we have previously suggested. This relationship can be used to estimate the meteoroid bombardment exposure age of a surface site from size analysis information alone. This estimation requires that a number of samples are collected from relatively undisturbed sites away from coarse ejecta deposits. Furthermore, we find in the samples that we have analyzed that there is a tendency, in general, for surface samples to be finer than those from depth within the regolith. As we have previously reported previously, the size frequency distributions of many lunar fines samples are bimodal. The broad mode in the 1μ to 3μ size interval is due to the sticking together of finer particles by glassy spatter from nearby impacts, and the more narrowly defined mode near 5μ is due to the addition of finer particles, probably as secondary ejecta from the presumably older regolith surfaces of the lunar highlands. Using an approximately linear relationship between sample graphic mean size and age (which may not be justified) we estimate that there are regoliths in the southern lunar highlands with meteoroid bombardment exposure ages of approximately 4 billion years. Our estimated age for the Apollo 14 site is 3.7 billion years, and the youngest well-dated Apollo 14 rocks cluster around a value of 3.88 billion years (Wasserburg and co-workers). Our estimated age should approximate the Imbrian event, and the strontium-rubidium age should approximate the cooling of the last igneous rock prior to the Imbrian event, assuming that sampling is adequate.

1. He, Ne, Ar, Kr, and Xe isotopes were determined by mass spectrometry in bulk samples, in unetched and etched sieve fractions and in separated structural components of Apollo 14 and 15 fines and rocks. Results on the elemental and isotopic abundances of trapped gases in fines are similar to previous findings, except for slightly higher $^{36}\text{Ar}/^{38}\text{Ar}$ ratios (Table 1 and Fig. 1). A large $^{40}\text{Ar}/^{36}\text{Ar}$ excess in coarse fines 14161 is peculiar (Table 2, bottom). Fragmental rock 14303 is almost free of solar wind gases. Apollo 14 and 15 soil exposure ages, improved by etching experiments, are similar to those of Apollo 11 and 12 soils, but comprehensive coarse fines 14257 have a much lower exposure age (Table 2). Conventional K-Ar ages for Apollo 14 and 15 fines (model ages 2.9 b.y.) and for rock fragments are given in Tables 2 and 3. K and U were determined by neutron activation. Table 3 also contains exposure ages for different rock fragments and the U-He age of 14303. In addition, the $^{39}\text{Ar}/^{40}\text{Ar}$ method was applied to crystalline fragments from 14303 and 14161. Mare-type basalts of 14161 have $^{39}\text{Ar}/^{40}\text{Ar}$ ages near 3.95 b.y. in agreement with results by other authors, but 2 non-mare type basalt of the same sample constitute a new age group for Apollo 14 material at ~3.55 b.y. (Table 3).

2. Mainly to study the location and variation of trapped solar wind, weighed single grains (10-300 μg) of 14003 soil were analyzed for He, Ne, and Ar isotopes by a single grain technique (50 specimen, Fig. 2) and by He-micro probe analysis (153 specimen). The handpicked minerals were identified by optical and X-ray methods. Within each mineral type, trapped gas concentrations vary by 1-2 orders of magnitude with ilmenites 10-100 times enriched (max. 1.6 cc $^{4}\text{He}$/g). The distributions for shocked plagioclases do not differ from unshocked samples. There is a clear correlation between color of glass spherules and trapped gas. Mean concentrations increase in the order brownish, yellow, green, black by about a factor of 10. The least fractionated gases are found in a dark spherule ($^{4}\text{He}/^{20}\text{Ne}=400$) and in ilmenites (e.g., $^{4}\text{He}/^{20}\text{Ne}=240, ^{20}\text{Ne}/^{36}\text{Ar}=43$).- For 15 grains, individual exposure ages up to ~2 b.y. have been found (Figure 3).

3. To gain information about concentration profiles of trapped gases, linear heating (5°C/min) was applied to measure gas release patterns for He and Ne in single glass spherules (6-110 μg) and minerals of 14003 and 10084 soils. Two out of nine gas release patterns are shown in Fig. 4. It turns out that even chemically similar spherules behave distinctly different. This underlines the necessity to study systems as simple as possible, that is, single particles. The differential Ne release displays single peaks between 760°C and 940°C with reasonable half widths (ca 150°C). The peak temperature seems to increase with the total Ne/cm² concentration. These patterns can be explained by a surface concentration profile and an activation energy of 50±10 kcal/mol. Such high activation energies
for glasses may be caused by traps due to the heavy radiation damage in the outer 400 Å-layer. Contrary to Ne, the He release is quite complex. The temperature of the maximum differential He release is sometimes higher and sometimes lower than for Ne (Fig. 5). Also, in addition to a first peak between 7900 K and 8600 K (ca. 2000 K halfwidths) often a second broad maximum is observed. If this is not caused by a radiogenic component—which would require very high U-concentration—this must be due to He diffusion from the original implantation sites (~400 Å) into the interior of the grains on the moon, since the activation energy for He cannot be higher than for Ne. Quantitative model calculations to obtain the corresponding concentration profiles as well as implantation experiments are under way. Extrapolation of the data for 4000 K shows that the He diffusion losses on the moon amount to 1% in only a few weeks, as compared to a saturation time of ~100 y. Therefore, diffusion largely affects the He distribution and explains the low total He/Ne ratios observed. Contrary, sputtering only shifts the surface but does not change the relative profile. Preliminary activation energies for He were calculated to be 45 ± 10 kcal/mol. — 4. The He content of unpolished pitted surfaces and polished cross sections of fragmental rock 14303 were analyzed in situ using the He-microprobe. Surprisingly, the heavily pitted “Top”-surface contains trapped He up to 10^{-2} cc/cm^2 only in relatively small spots. Other areas are free of trapped He inside of microcraters. Radiogenic He is seen in regions of U-enrichments. The less frequently pitted “North”-surface displays trapped He much more frequently. — 5. Charge assignments made to heavy ion tracks in lunar minerals are assumptions based on little evidence. To provide such evidence, lunar pyroxenes were bombarded by artificially accelerated Ca-ions of ~2 MeV/amu in the Heidelberg Tandem accelerators. The preexisting cosmic ray tracks were used to develop the Ca tracks (track in track method). It was found that such tracks were generated in the interior of the pyroxenes only. Thus the total etchable range of Ca-ions could be determined. From thermal annealing experiments applied to artificial 1 MeV/amu Fe-tracks in pyroxenes, a length decrease of ~1 μ during 1 b.y. at lunar temperatures was extrapolated. It is concluded that thermal effects on the track length distributions are negligible. The comparison of the lengths distributions of cosmic ray tracks and of 2 MeV/amu Ca ion tracks in the same grain made it possible to ascribe the peak between 4 and 6 μ to Ca (Fig. 6). Provided the peaks of the lengths distributions are due to even atomic number ions, an etchable range of Fe-ions between 15 and 18 μ results. However, often the abundance maximum is not situated in this region. A detailed study of the relative track abundance ratios (Co+Ni)/(all 20<Z<28 tracks) and (Cr+Mn)/Fe shows an increase of (Cr+Mn) relative to Fe with decreasing (Co+Ni). This correlation clearly indicates that fragmentation of the incident ions plays an important role. This has consequences for the proper charge assignment as well as for the interpretation of track abundance ratios in terms of cosmic ray elemental abundances. The relative (Co+Ni) abundance decreases with increasing penetration depth of the cosmic radiation while the fragmentation of Fe nuclei seems to populate preferentially Cr and Mn. This correlation can also be used to estimate the burial depths of individual grains. — 6. Steep gradients of track densities were observed in the outer 50μ of 14303 feldspar crystals from the pitted “North” surface but not in grains from the interior of this fragmental rock. — Samples from 6 mm depth yield an estimated solar flare exposure time of ~0.5 m.y.
Table 1. Rare Gas Concentrations \(^{10^{-6}}\) in Gases

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<th>Air</th>
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<th>CO(_2)</th>
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<th>1.00% Ne</th>
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Table 2. Exposure Ages of Single Soil Particles

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<tr>
<td>Mica Type A</td>
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<td>Volcanic Rock</td>
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Table 3. Glass Spheres

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<th>Radius (μm)</th>
<th>Density (g/cm³)</th>
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<td>2.5</td>
</tr>
<tr>
<td>1600</td>
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Figure 1. Trapped Gases in Various Soils

Figure 2. Maxwellian Error

Figure 3. Exposure Age of Soil Particles

Figure 4. Number of Tracks

Figure 5. Track Length (μm) vs. Exposure Age (ka)

Figure 6. Track Length (μm) vs. Exposure Age (ka)
Microbreccia 14305-4 and rock fragments from a one-gram sample of coarse fines (14257) were studied in detail by optical and electron microprobe techniques.

Microbreccia 14305-4 and several fragments of breccia in the coarse fines show completely recrystallized matrices and their clasts exhibit the following shock and heat effects:
   a) brecciation of single-crystal and rock fragments
   b) partial melting and reaction zones along outer edges of and fractures in single-crystal fragments
   c) microfaulting as shown by displaced twin lamellae
   d) devitrification and recrystallization of spheroidal clasts originally probably glass spherules.

Almost totally gradational contacts between some rock clasts and the matrix were also noted. Other microbreccia fragments show only partially devitrified matrices and some are completely unrecrystallized with glass clasts and glass matrix. Pyroxene compositions of various breccia types are shown in fig. 1 and compositions of glass fragments and unrecrystallized matrix (m) in fig. 2.

Five mineralogically and texturally distinct igneous rock fragments in the coarse fines were selected for study. These are (see fig. 1 for pyroxene and olivine compositions):
   A-microgabbro. Very similar to the relatively coarse grained Apollo 12 material, with pyroxene ranging from pigeonite to subcalcic augite, showing relatively little chemical zoning in single pyroxene grains; plagioclase (An 88) and ilmenite and traces of troilite and iron are the remaining constituents.
   B-olivine-ilmenite basalt. The olivine (Fo 57-60) is coarse-grained and set in a finer grained, finely intergrown matrix of augite to ferroaugite, plagioclase (An 76-84) and ilmenite. TiO₂ and Al₂O₃ contents, in these pyroxenes range from 1.70 to 4.55 and 2.5 to 6.1 respectively.

*at Department of Geology, University of Vermont, Burlington, Vt.
Fig. 1. Pyroxene and olivine compositions in microbreccias and igneous rock fragments A-E. • = clinopyroxene; o. = orthopyroxene; x = olivine. Dashed lines connect coexisting phases and curved arrows indicate compositional zonation from core to rim in a single grain.

Fig. 2. Compositions of glass fragments in and matrix (m) of unrecrystallized breccia. Dashed line outlines compositional range of Apollo 11 glasses in fines. No. 11 is scoriaceous glass average for Apollo 11 (after Frondel et al., 1970 and Winchell et al., 1970).
C-ophitic basalt. Almost identical in texture to rock 14310 (LSPET), consisting of subhedral plagioclase laths (An 86-90), some light colored magnesian pigeonite, minor dark brown augite, and traces of ilmenite and troilite. Small amounts of K-rich glass occur interstitially.

D-anorthosite. Highly granulated, fine-grained plagioclase (An 90-96) with about ten percent olivine (Fo 60) and magnesian pigeonite.

E-norite. Strongly sheared and granulated plagioclase (An 86-93) enclosing larger, anhedral bronzite grains. The basaltic rocks (A,B,C) show considerably less Fe-enrichment and zonation of single crystal pyroxene grains than was observed in the pyroxenes of the Apollo 12 basalts.

The outer surfaces of several of the rock fragments and many of the glass fragments and spheroidal bodies in the coarse fines show fine-scale spallation features and microcratering due to the impact of hypervelocity particles. Similar craters occur on some iron-nickel fragments in the coarse fines. Most of the iron-nickel occurrences are spheroidal or globular in form and their Ni and Co contents range from 2.0 to 9.1 and 0.34 to 0.68 weight percent respectively.

Three size fractions of the fines (14259-23) were analyzed chemically:

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<th></th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>TiO₂</th>
<th>Cr₂O₃</th>
<th>FeO</th>
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<td>1)</td>
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<td>17.5</td>
<td>1.74</td>
<td>0.25</td>
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<tr>
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<td>0.27</td>
<td>10.36</td>
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<td>1.82</td>
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<td>10.25</td>
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<th>CoO</th>
<th>MgO</th>
<th>CaO</th>
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<th>K₂O</th>
<th>P₂O₅</th>
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<td>0.56</td>
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<td>0.06</td>
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<td>9.50</td>
<td>11.10</td>
<td>0.72</td>
<td>0.59</td>
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<td>100.29</td>
</tr>
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<td>3)</td>
<td>0.14</td>
<td>0.06</td>
<td>0.01</td>
<td>8.30</td>
<td>11.65</td>
<td>0.70</td>
<td>0.55</td>
<td>0.66</td>
<td>100.39</td>
</tr>
</tbody>
</table>

*) S is present in each analysis.

The FeO and TiO₂ contents are distinctly lower than in the average analyses of Apollo 11 and 12 fines, whereas the Al₂O₃ and SiO₂ contents are much higher. These differences reflect the higher plagioclase and lower ilmenite contents of the rocks and resulting fines and the relatively magnesian clinopyroxene and orthopyroxene occurrences.
NUCLEAR MAGNETIC RESONANCE PROPERTIES OF RETURNED LUNAR SAMPLES:

Nuclear magnetic resonance (NMR) measurements have been made in the dispersion mode at room temperature on samples of the following fines: 10084 and 14163-168; and crystalline rocks: 12021-55 and 14321-166. Spectra from the following nuclei have been studied in one or more of the lunar samples and compared to those observed from selected terrestrial samples: $^{31}$P($I=1/2$), $^{27}$Al($I=5/2$), and $^{23}$Na($I=3/2$).

The dispersion derivative of $^{31}$P from 14163-168 recorded at 30 MHz is very weak and rather broad and appears to be shifted by a slight amount, approximately 5 gauss, to the high field side of the unperturbed resonant field, $H_0$. This shift was measured relative to a sharp, narrow absorption derivative of $^{31}$P from a sample of chlorapatite, used to locate the position of $H_0$. For comparison the $^{31}$P spectrum from a terrestrial sample of GSP-1 granodiorite (weight percent $P_2O_5 = 0.28\%$) was also very weak and broad in shape similar to that of $^{31}$P in 14163-168 but with apparently little or no shift measurable to within $\pm 2$ gauss. The $^{31}$P shift in 14163-168 may be associated with small internal magnetic fields similar in magnitude to those which oppose the laboratory field at the aluminum and sodium sites reported in earlier NMR studies of lunar material. In contrast to the somewhat indirect evidence for internal magnetic fields that can be cited on the basis of the field dependence of the $^{23}$Na and $^{27}$Al nuclear electric quadrupole interactions in lunar materials, the $^{31}$P nucleus can be considered to provide more direct and unambiguous evidence for the presence of these effects since its spectrum contains no quadrupole perturbations.

The $^{23}$Na and $^{27}$Al nuclear magnetic resonance spectra from all lunar samples studied thus far, as well as those from several terrestrial samples such as BCR-1 basalt, exhibit an inverse dependence of their line widths on $\nu_0$, the unperturbed resonant frequency, and a broad, asymmetric frequency dependent line shape characteristic of second-order nuclear electric quadrupole perturbations. Only their "central" ($m=1/2 \leftrightarrow m=-1/2$) transitions, arising from slightly differing distributions of quadrupole interactions from sample to sample, appear in the lunar spectra. The satellite transitions, $m=\pm 3/2 \leftrightarrow m=\pm 1/2$ and $m=\pm 5/2 \leftrightarrow m=\pm 3/2$, have been split out over such a broad frequency range that they are unobservable. The distributions of quadrupole interactions in the Apollo 14 samples resemble qualitatively those found previously for Apollo 11 and Apollo 12 specimens and are taken to indicate, principally, the degree of atomic disorder at the aluminum and sodium sites and the presence of mineral components which contain strains. The relation between this disorder and strain and the factors which produced it such as the rate of quenching and the crystallization history of the sample has not as
NMR of Lunar Samples
D. Kline

yet been clearly established.

Despite interference from the extraneous Knight-shifted aluminum response arising from the NMR probe, the shapes of the 27Al high-field (low-frequency) peaks and their splittings from $H_0$, is easily resolved. The interference from the background probe line is less serious at the highest frequency used, 20 MHz, than at the lower frequencies used previously due to the relatively stronger sample intensities and the greater separation of the probe line from $H_0$ at high frequencies. In addition use of an "aluminum free" NMR probe which operates in the range 6-15 MHz and shows no 27Al spectrum, has permitted more precise studies of both 27Al and 23Na spectra at lower frequencies than was possible in earlier studies.1

For sample 14163-168 the width of the high-field side of the 27Al spectrum, measured from $H_0$ at half-maximum intensity, was found to be 39(+1), 46(+2), 76(+5) gauss at spectrometer frequencies of 20, 13, and 8 MHz, respectively. Accompanying this was a decrease in intensity and an increase in the splitting of the 27Al high-field peak from $H_0$, found to be 10(+1), 12(+1), and 17(+2) gauss, respectively. This inverse behavior with frequency indicates the dominant presence of second-order quadrupole perturbations in the 27Al spectrum of 14163-168. Similar results have been found for the other lunar samples as well with the fines samples in general tending to show somewhat larger values of widths and peak splittings, and thus atomic disorder relative to those found in samples of crystalline rocks. Specifically, for samples 12021-55, 14321-166 and 14163-168 at a spectrometer frequency of 20 MHz the high-field widths at half maximum are 28(+1), 33(+1), and 39(+1) gauss, respectively, while the peak splittings are, respectively, 6(+1), 7(+1), and 10(+1) gauss. The comparable figures for a terrestrial sample of BCR-1 basalt are 26(+1) and 8(+1) gauss. In addition, the experimentally measured ratios of the 27Al high-field widths at half maximum, at 13 MHz and 20 MHz, which are 33(+1)/28(+1)=1.18(+0.07) for 12021-55, 37(+1)/33(+1)=1.12(+0.06) for 14321-166, and 46(+2)/39(+1)=1.18(+0.08) for 14163-168 are possible indications that small internal magnetic fields exists at the aluminum sites in the lunar samples, opposing the laboratory field. Such a conclusion may be made by comparing the above ratios with those obtained from computer simulations of 27Al lunar spectra using the quadrupole parameters of the eight sites of anorthite ($e^2qQ/h=8.42, 7.25, 6.81, 6.30, 5.54, 4.90, 4.30$, and 2.66 MHz and $\eta=0.66, 0.76, 0.85, 0.88, 0.42, 0.42, 0.53, 0.66$) with varying amounts of isotropic magnetic fields shifts.1 These simulations show that the computed ratio of high-field widths at half maximum at 13 MHz and 20 MHz decrease somewhat when magnetic field shift parameters are included relative to the value of approximately 1.32 obtained from simulations when no magnetic field shifts are included. For comparison, the comparable ratio for BCR-1 basalt is 32(+1)/26(+1)=1.23(+0.09). Taken together the experimental ratios quoted above for the lunar samples are considered to deviate sufficiently from calculated value of 1.32 to suggest, at least indirectly, the presence of internal magnetic fields of the order of 5-10 gauss. The magnitudes of these internal magnetic fields cannot be quoted precisely because of the relatively large experimental uncertainties involved and, also, because it is not clear
if these internal fields remain constant or vary with varying external magnetic fields.

Previous NMR studies of $^{23}$Na in lunar samples have been hampered by very weak signal intensity.\textsuperscript{1} More recently, improved signal-to-noise for Na, particularly in two samples of fines, 10084 and 14163-168, has been obtained through use of larger samples than previously, 7-8.5 grams, and study at higher frequencies, up to 20 MHz. Examination of the $^{23}$Na spectra from 14163-168 and 10084 over a range of frequencies from 20 MHz to 16 MHz showed that the spectral widths were somewhat large in the Apollo 11 sample as compared to the Apollo 14 sample, indicating greater apparent site disorder in the former. Computer simulations indicate that the $^{23}$Na spectra of 10084 and 14163-168 are not inconsistent with distributions of quadrupole interactions whose coupling constants lie mainly in the range 2-4 MHz and whose asymmetry parameters lie largely in the range 0.25-0.75. No special significance may be attached to such distributions. The criteria for choosing the limits of quadrupole parameters noted above are based mainly on the magnitude and ratio of the high-field peak splitting and width at half-maximum as calculated from a given distribution. In addition a comparison of the $^{23}$Na high-field widths at half-maximum, at 16 MHz and 20 MHz, in a manner similar to that used for the case of the aluminum spectra, suggests the possibility of small internal magnetic fields at the sodium sites in the lunar samples.


*Deceased
THE NEAR-SURFACE VELOCITY STRUCTURE OF THE MOON; R. L. Kovach, Stanford University, Stanford, Calif., 94035 and J. S. Watkins, University of North Carolina, Chapel Hill, North Carolina, 27514

During the Apollo 14 mission to the Fra Mauro site a shallow seismic refraction experiment was carried out by the astronauts to aid in answering some questions concerning lunar geology and near-surface internal structure. What is the nature and thickness of the lunar regolith? What is the seismic velocity of the lunar near-surface material? Are there major velocity discontinuities recondite in the upper layers of the Moon?

Two P-wave velocities are evident in the observed travel-time data. A direct arrival was observed with a P-wave velocity of 104 m/sec together with a faster refracted arrival possessing a velocity of 299 m/sec. Seismic signals were also generated by the LM ascent and recorded by the Apollo 14 passive seismometer at a distance of 178m (584 ft). A clear first and second arrival can be recognized. The observed travel time for the second arrival is nearly identical to that for a direct seismic wave propagating with a velocity of 104 m/sec. A first arrival was observed with a travel time somewhat faster than that predicted by a refraction from the top of a 299-m/sec horizon, suggesting that a material with a faster intrinsic compressional wave velocity lies beneath the 299-m/sec material. Estimates of the thickness of the underlying material range from 19 to 76 meters.

A large amplitude later arrival was also observed on many of the thumper generated seismograms. We have tentatively identified this arrival as a direct shear wave arrival traveling with a velocity of 62 m/sec (202 ft/sec). Because of the lack of three component information it cannot be proved that this is a shear wave arrival, but the measured value is in fairly good agreement with the value of 23 m/sec deduced from Surveyor touchdown dynamics. The measured value of \( V_s = 0.6 \ V_p \) is also in good agreement with velocity measurements on Apollo 11 returned lunar samples. The value for Poisson's ratio is 0.23. Spectral analyses of the lunar seismograms and subsequent analyses of the amplitude decay with distance lead to an estimate for \( Q \) of the lunar near-surface material in the range from 50-100.

From the seismic signals produced by the LM ascent at the Apollo 12 site the near-surface compressional wave velocity was measured at 108 m/sec. This value is in exceedingly close agreement with that measured for the surface material at the Apollo 14 site. It can be convincingly argued that the fragmental and comminuted layer that covers much of the lunar surface, the regolith, although locally variable in thickness, possesses remarkably similar acoustic properties.
Near Surface Lunar Structure
R. L. Kovach

We have therefore concluded that the 104 m/sec layer is the lunar regolith. At the Apollo 14 site, the layer is 8.5 m thick. The lunar regolith (i.e., 104 m/sec layer) appears from photogeologic evidence to be comprised mainly of overlapping ejecta blankets from post-Fra Mauro craters.

The 104 m/sec layer is underlain by a layer whose thickness is estimated to range from 19 to 76 m and whose velocity is 299 m/sec. The abrupt change in seismic velocity and, by inference, of other physical properties to the 299 m/sec layer, indicates that the lunar regolith is indeed a well-defined unit in lunar stratigraphy.

The nature of the 299 m/sec layer is somewhat less clear than that of the 104 m/sec layer. Photogeologic mapping of the area around the Apollo 14 site had tentatively suggested that the 299 m/sec layer might consist of Fra Mauro material or possibly of volcanics. Both explanations are not inconsistent with the observed velocities. However, if the 299 m/sec layer is volcanic, it is difficult to explain why so few igneous rocks were returned by Apollo 14 astronauts since a number of young craters in the vicinity appear to have penetrated into and through the 299 m/sec layer. The velocity of the 299 m/sec layer is comparable to that observed on Earth in the upper part of Meteor Crater ejecta and, thus, is quite reasonable for ejecta from Mare Imbrium. We, therefore, have concluded that the 299 m/sec layer is the Fra Mauro Formation.

The relatively low compressional wave velocities that were measured at the Apollo 14 site argues against the presence of substantial amounts of permafrost in the lunar near-surface at this particular site. Permafrost had been speculated to be present in the lunar near-surface. Measured velocities in permafrost vary greatly -- depending on such factors as lithology, porosity, and degree of interstitial freezing -- but typically range from 2438 to 4572 m/sec (8000 to 15,000 ft/sec).

The near-surface velocity-depth function for the Moon increases from about 100 m/sec at the surface to ~ 6 km/sec at a depth of 15-20 km or so. This is an enormous velocity change compared to terrestrial standards. The lunar seismic data do not yet preclude the presence of major velocity discontinuities in the upper 10 km of the Moon.

Rock 14318 is a densely welded microbreccia consisting of fragments of shock-modified and recrystallized anorthosite, norite, troctolite, KREEP norite, basalt, dunite, as well as glass fragments, glass spherules, and chondrules, embedded into a partly glassy, fine-grained matrix. Chondrules, glasses, and lithic fragments are welded to the matrix, and fractures usually cut across particle-matrix boundaries. Glass spherules (and fragments) are sometimes recrystallized and partly devitrified but can readily be distinguished from chondrules: the latter often have igneous textures (i.e. large crystals embedded in a fine-grained matrix)[1], and resemble in all textural properties meteoritic chondrules and artificial chondrule-like spherules that were formed by spontaneous crystallization of highly supercooled, molten silicate droplets [2]. The results of this study may be summarized as follows: i) Lithic fragments in 14318 range widely in composition and are divided into an anorthositic-noritic-troctolitic [ANT] group, KREEP norite group, basalt group, and dunite. Although most fragments have shock-modified and recrystallized rather than igneous textures, they are considered to be derivatives of igneous rocks (i.e. they are metaigneous rocks). ii) ANT lithic fragments are similar to those reported from Apollo 11 [3], Apollo 12 [4] and Luna 16 [5]. Norites are low in K and P and are not to be confused with KREEP norites. However, 14318 does, in addition, contain KREEP norite fragments that are similar to previously described KREEP norite lithic fragments from Apollo 12 [4-6]. Separation of KREEP norites from ANT norites is not always sharp but gradational (Fig. 1). iii) The basaltic lithic fragment suite is represented by only one specimen, but similar fragments were found in low abundance in other Apollo 14 samples. The basaltic fragment is similar to rock 14053 but lower in K and P than the lithic fragments of a second basaltic suite, the K-rich basalt group, found in other Apollo 14 rocks [7]. However, in 14318, this group is represented by K-rich basalt glasses and chondrules (Fig. 1; Table I). iv) A single lithic fragment of feldspathic dunite (magnesian) was found, lending additional support to earlier suggestions that ultramafic rocks occur on the moon [3-5]. v) Glasses are the compositional analogs to the ANT and KREEP norite lithic fragments. The third glass group, the K-rich basalt glass, has no lithic fragment counterparts in rock 14318 but does in other Apollo 14 sections [7]. K-rich basalt glasses are similar to mare basalts in major elements but are enriched in K and P, indicating they may not be mare-derived (Table I). Although REE were not determined it is possible that the K-rich basalt glasses represent a suite of rocks enriched in KREEP in comparison to KREEP norite. Glass cementing microbreccia 14318(matrix glass) is

*Speaker
A Chondrite of Lunar Origin
Kurat, Keil, Prinz, and Bunch

Table I: Bulk compositions of lithic fragment, glass, and chondrule groups from 4318.4 (broad electron beam techniques; wt.\%)

<table>
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<th>&quot;Igneous&quot; lithic fragments</th>
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<th>Chondrules</th>
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<tr>
<td></td>
<td>ANT KREEP norite</td>
<td>Basalt</td>
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<tr>
<td>SiO₂</td>
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<td>K₂O</td>
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<tr>
<td>P₂O₅</td>
<td>0.10</td>
<td>0.57</td>
</tr>
<tr>
<td>Total</td>
<td>98.68</td>
<td>98.38</td>
</tr>
</tbody>
</table>

Specimens | 4 | 7 | 1 | 1 | 13 | 12 | 9 | 22 | 3 | 2 |

predominantly of KREEP norite composition, with compositional similarities to ANT norite (Fig. I). vi) Chondrules are the compositional analogs to the ANT, KREEP norite, and K-rich basaltic lithic fragments. Most chondrules, however, are of ANT norite composition (Fig. I). vii) Apollo 11 and 12 and Luna 16 mare-type basaltic lithic fragments and glasses are rare. iix) On the basis of its texture, it is suggested that rock 14318 and its chondrules[1] originated in a large-scale impact event. The similarity in texture between 14318 and polymict-brecciated chondrites suggests that meteoritic chondrites may have formed in similar large-scale impact events on parent meteorite bodies. However, the question of the origin of meteoritic chondrules by either impact or condensation from the solar nebula cannot be resolved on the basis of the lunar evidence.

"Igneous" Lithic Fragments

- ANT Group
- KREEP Norite
- Dunite
- Basalt
- I4310
- I4053
- Fines (Average)

Glasses

- ANT Group
- Spherules
- Fragments
- Matrix Glass
- KREEP Norite
- Spherules
- Fragments
- Matrix Glass
- K-Rich Basalt
- Spherules
- Fragments

Chondrules

- ANT Group
- KREEP Norite
- K-Rich Basalt

"Igneous" lithic fragments, glasses, and chondrules from rock 14318

Figure 1

Weight Percent FeO

Weight Percent Al₂O₃
Melting experiments on the Apollo 14 high-alumina basalt 14310 (new chemical analysis given in Table 1) have been made in a pressure range 5 to 30 kilobars (kb) and in a temperature range 1100° to 1470°C with graphite capsules. Synthetic glass of this rock composition was also used as a starting material. Below about 10 kb, calcic plagioclase is the liquidus phase and is followed by chromian spinel (~20 wt.% Cr₂O₃), aluminian orthopyroxene and clinopyroxene with lowering temperature. Between 10 and 20 kb, spinel is the liquidus phase and above 20 kb garnet (~Ca₂0Mg₅₈Fe₂₂) is the liquidus phase. Aluminian clinopyroxene is the second phase to crystallize above 15 kb. The melting interval is 120 – 180°C between 5 and 25 kb. In the 1 atm quenching runs made on the synthetic glass, olivine crystallizes at least near 1220°C. The liquid of the 14310 rock composition shows cotectic nature at about 10 and 20 kb; however, the coexisting phases are plagioclase+spinel and spinel+garnet, respectively, at these pressures. In the system anorthite-olivine-silica, the composition of rock 14310 (and 14310 – 10% plagioclase) as well as feldspathic basalt (or gabbroic anorthosite) which Reid et al. (1971) have suggested to be a lunar highland material lies on the plagioclase side of the olivine-plagioclase or pyroxene-plagioclase cotectic boundary at least up to 10 kb (~300 km). On the other hand, the possible lunar 'mantle' material (peridotite or pyroxenite with or without a small amount of plagioclase) lies on the olivine or pyroxene side of the cotectic boundary. If rock 14310 is not a plagioclase cumulate rock or a plagioclase cumulate rock with addition of plagioclase less than 10%, the rock cannot be a product of direct partial melting of the possible materials of the lunar interior under anhydrous conditions. It is also not a product of fractional crystallization of the magma which was formed by direct partial melting of the lunar interior. Therefore, a double-stage model must be considered for the origin of the rock 14310. One possibility is that the magma was generated by bulk melting or partial melting of a plagioclase cumulate rock which had been formed by a large-scale differentiation of magma in the shallow part of the lunar interior. It is mentioned that the magma of the 14310 rock composition may be produced by partial melting of gabbroic anorthosite.

---

**Table 1**

| SiO₂  | 46.88, TiO₂  | 1.19, Al₂O₃  | 21.68, FeO  | 8.22, MgO  | 7.42, CaO  | 12.55, Na₂O 0.72, K₂O 0.50, Cr₂O₃ 0.25, MnO 0.13, P₂O₅ 0.17, Total 99.71 (wt.%) |

(Analyzed by H. Haramura with conventional wet chemical analysis method.)
The above discussion is based on the assumption that there was no loss or gain of alkalies and other elements during the magmatic stage. Recently, however, Brown and Peckett (1971) have suggested a significant (~70 wt.% or more) alkali loss for rock 14310. If this is true, the above discussion may be subject to change. To examine this, melting experiments have been made on the synthetic glass of the 14310 rock composition with 3.2 wt.% Na₂O and about 1% K₂O. The plagioclase liquidus drops drastically and olivine crystallizes to higher pressures, and in consequence, olivine, plagioclase, spinel and orthopyroxene appear within 20° at about 3 kb (~100 km). This evidence suggests that the liquid of this composition could be formed by direct partial melting of plagioclase and spinel bearing peridotite at depths of about 100 km. Therefore, if significant alkali loss occurred, as suggested by Brown and Peckett, the origin of rock 14310 could be explained by a single-stage model.

Orthopyroxene in rock 14310, which occurs as the core of pigeonite and subcalcic augite, contains 1.5 - 3.1 wt.% Al₂O₃, whereas that crystallized at 10 kb at 1250°C and at 5 kb at 1200°C contains 5.7 - 7.7% Al₂O₃ and 4.3%, respectively, presumably as a Tschermak's component, supporting the low-pressure crystallization of the 14310 orthopyroxene. Upon cooling of the magma, small-scale fractional crystallization took place and the following compositional variations in pyroxenes and plagioclase are observed: orthopyroxene Ca₄Mg₇₉Fe₁₇-Ca₅Mg₆₂Fe₃₂, pigeonite Ca₉₀Mg₃₈Fe₃₂-Ca₁₁Mg₁₅Fe₇₄, Ca-rich clinopyroxene Ca₃₄Mg₄₆Fe₂₀-Ca₁₁₇Mg₁₇Fe₆₆, plagioclase An₉₄Ab₆-An₇₃Ab₂₇. In the late stage of crystallization, minor minerals such as Ba-bearing K-feldspar, phosphate minerals, Zr-rich and rare earth-bearing minerals crystallized and the residual liquids that were concentrated in the interstices of early crystallized plagioclase and pyroxene became potassic rhyolite in composition. In some interstices, liquid immiscibility occurred, having produced rhyolitic and Ti-rich, pyroxenitic liquids.

Another crystalline rock 14053, which has been also analysed with the microprobe, contains plagioclase (An₂-An₇₄), pigeonite (Ca₁₂Mg₆₀Fe₂₈-Ca₁₃Mg₴₄Fe₄₃), subcalcic augite and ferroaugite (e.g., Ca₅₅Mg₄₅Fe₇₀-Ca₂₃Mg₶₆Fe₇₁), hedenbergitic clinopyroxene (Ca₄₂Mg₂₈Fe₅₆), ilmenite, spinel (both Ti-bearing chromite and Cr-bearing ulvospinel), olivine (Fa₇ and Fa₉₇), cristobalite, K-feldspar with mesostasis of high K, rhyolitic composition. The bulk composition of rock 14053 is significantly different from that of rock 14310 and rather similar to those of the Apollo 12 crystalline rocks with intermediate FeO/(FeO+MgO) ratio, which is higher than that of rock 14310.

References:
Lunar high-alumina basalt (21.68% Al₂O₃)

* synthetic glass

Graph showing phase transitions under varying temperature and pressure conditions.
ELECTRON PETROGRAPHY OF APOLLO 14 AND 15 SAMPLES

by

J.S.Lally, R.M.Fisher (U.S.Steel Research Center, Monroeville, Pennsylvania), J.M.Christie, D.T.Griggs (Geology Dept., University of California, Los Angeles), A.H.Heuer, G.L.Nord, Jr. and S.V. Radcliffe (Division of Metallurgy and Materials Science, Case Western Reserve University, Cleveland, Ohio)

Comparative optical and high-voltage (up to 1 mev) electron petrographic analyses have been conducted on basalt and breccia samples from Apollo 14 and anorthosite from Apollo 15. The analyses were complemented by localised chemical analysis on the same specimen areas by electron microprobe and scanning microscopy techniques. We report here the preliminary results for four samples - including each of the above types.

Sample 14310/159 (basalt):

The optical structure in thin sections of this crystalline rock exhibited principally plagioclase and pyroxene with an intersertal texture. Only minor amounts of opaque phases are present. The pyroxene occurs as both ortho and clino-pyroxene and zoning from the former to the latter is apparent. In the electron microscope, the density of dislocations in both the plagioclase and pyroxene is comparable to that we observed earlier in the igneous rocks from Apollo 11 and 12 [1,2]. Fine-scale exsolution is present in the pyroxene. Under appropriate electron diffraction conditions, antiphase domain boundaries (APB's) can be imaged in the same areas and are found to be coarser in scale than the exsolution. In the plagioclase also, a fine-scale exsolution has been identified. These substructural features indicate that 14310/159 resembles the mare basalt samples examined from the earlier landings in that it exhibits little or no evidence of mechanical deformation. The presence of exsolution in the plagioclase and its absence in Apollo 11 and 12 basaltic plagioclases indicates a slower cooling history, or re-heating, for rock 14310.

Sample 14321/42 (breccia):

This sample is seen in optical thin-sections to contain fragments of igneous rocks, previously consolidated breccia and individual mineral crystals ranging from centimetre dimensions to the limit of optical resolution. The finest material in the matrix surrounding these features is optically isotropic, denoting that it is either amorphous or composed of submicroscopic crystallites. Shock-induced deformation features (twinning and bending in plagioclase and pyroxene) are apparent,
with considerable evidence of local annealing or recovery, notably near the outer surface of large mineral fragments, but also in the interior as indicated by partially resorbed twins. In electron transmission, the optically isotropic matrix could be resolved and was found to consist of a heterogeneous mixture of pyroxene and plagioclase crystallites containing a small amount (of the order of 10%) of glass. Virtually no porosity was present. The sub-structural features indicative of deformation are likewise distributed extremely heterogeneously, with regions of very heavy deformation containing recrystallization nuclei existing adjacent to largely recovered structures. Deformation/recrystallization processes appear to have been particularly active at grain contacts both in the matrix and at the matrix-clast interfaces. Mechanical twins, indicative of shock deformation, are also present. In addition, plagioclase in both the clasts and the matrix exhibits more extensive fine-scale exsolution than in 14310. The pyroxenes showed a well-developed APB structure. Both of these features are indicative of reheating following deformation. We conclude that in this breccia sample cohesion was achieved by deformation and subsequent or simultaneous recrystallization caused by the high pressures and temperatures associated with impact events on the lunar surface. This process we termed "shock sintering" [3]. The glass present in the matrix was probably also formed by the shock events, but is not thought to play a major role in the consolidation.

Sample 14161/38 (coarse fines):

Two specimens from this sample were examined in detail as they differed significantly from the breccia 14321. Sample 14161/38/4 is a vesicular brown glass containing angular mineral fragments. In the electron microscope tabular, micron-size plagioclase crystals produced by devitrification were common. Fine metallic particles observed throughout the glass and mineral fragments were shown by electron diffraction to be iron-nickel alloys in the f.c.c. phase field. Sample 14161/38/7 consists of angular mineral fragments with about 10% of vesicular glass. The glass, which contains abundant fine (~100Å) metal spherules, wetted all the mineral fragments and the glass-mineral boundaries appeared gradational rather than sharp. A puzzling feature of this sample is the high density of particle tracks in many mineral fragments at depths up to 1 mm. from the surface of the sample. The track densities are in the range of 3 to 7x10^10 per cm^2. We consider that the tracks must have formed in the fragments before consolidation. It is difficult to understand how this glass welded breccia was formed without annealing the particle tracks.

Sample 15415 (anorthite):

Only plagioclase fragments (less than 1 mm) from this rock have been available to us to date. Using a new technique, the fragments have been mounted and thin-sectioned.
Optically the fragments were variably twinned and appeared to be homogeneous chemically and lacking in evidence of deformation. Electron probe analysis indicated a composition of $\text{An}_{93} - \text{An}_{95}$. The electron microscopy showed the dislocation density to be distinctly less than igneous rocks of Apollo 11, 12, and 14. No other evidence of deformation was present, although abundant submicroscopic growth twins were evident in some fragments. The most striking substructural features were easily resolved "c"-type domain boundaries, much larger than "c"-type domains in other more albitic lunar plagioclase. The large size of these domains is ascribed mainly to the high An content. Inclusions of calcic augite are present in the plagioclase on a microscopic scale and submicroscopic precipitates of clinopyroxene have been observed on plagioclase twin-boundaries.

REFERENCES

ALPHA SPECTROMETRY OF A SURFACE EXPOSED LUNAR ROCK

Alpha spectrometry was carried out on the top surface of lunar rock in view to detect the Lead $^{210}$ and Polonium $^{210}$ which could be deposited on the surface of the Moon as decay products of outgassed Radon from the lunar soil.

This activity was proposed as 4 dps/cm$^2$ by Kraner et al. (1) and as from 1 to 0.1 by Lambert and Grjebine (2).

The Polonium $^{210}$ activity should give rise to an $\alpha$ emission of definite energy, 5.30 MeV, as opposed to the activity of $\alpha$ emitters produced by disintegration of Uranium and Thorium included in the rock which will produce a continuous spectrum.

Lunar rock No. 14321, piece No. 212 provided by the NASA was measured on an $\alpha$ semiconductor spectrometer. The sample was cut off from the upper area of a main stone.

Two faces have been measured: the first one (about 1.5 cm$^2$) was facing the outside but presents only one distinct pit. For control, an inside face was also measured.

The $\alpha$ spectrometer consisted of a surface barrier detector (ORTEC 2 cm$^2$) followed by a linear amplifier and a 200 channel analyser. It was located in an air-conditioned room and fed with a battery stabilized electric power.

The measurements were carried out for one month and a half (i.e. 1176 hours of effective counting) without any change of the position of the sample. The stability of the apparatus was checked every week through a pulse calibrated generator: no shift was noted. The background was of the order of 2 counts per 100 keV for 1176 hours.

Obtained data leave no significant activity for Polonium $^{210}$ in excess. With the statistical errors involved, the upper limits permissible for an activity of Polonium $^{210}$ (in the range of 5.1 to 5.3 MeV) would be 22.3 counts per 1176 hours. A yield estimated at 13%, therefore brings this figure to approximately 2 $10^{-5}$ dps/cm$^2$. 
This result is in agreement with the measurements performed by Economou and Turkevich (3) on the filter of Surveyor III television camera who found an upper limit of 5 $10^{-3}$ dps/cm$^2$, and Lindstrom et al. (4) with 4.5 $10^{-4}$ dps/cm$^2$ in Apollo XI fines.

The lowering of this limit by a factor of 20 by our measurement suggests that either there is no outgassing of Radon from the Lunar soil, or some other complicated process must be involved.

1) - Lindstrom et al. (4) have suggested that, in the Lunar soil, the recoiling atom, after the $\alpha$ emission of Radon 222 is not slowed down in the vacuum space between grains and therefore is imbedded in the next crystal. However, some desintegrations certainly occur inside the grains and provide some Radon atoms with a very low kinetic energy, able to diffuse between the grains: an order of 10% is in agreement with our present experiments.

2) - The fines have shown a very low thermal conductivity. Using the superficial temperature of the Moon surface measured by infrared emission (5), we have calculated the variations of temperature with depth in the Lunar soil. According to Birkebak et al. (6), Bastin et al. (7), Warren et al. (8), a thermal conductivity of $2 \times 10^{-3}$ watt m$^{-1}$ K$^{-1}$ gives at 13cm of depth a temperature of $210^\circ$ K ± 10$^\circ$K. These very low temperatures mean that the adsorption possibilities of the Lunar soil could be very large for Radon.

Nevertheless, experiments simulating Lunar conditions performed in the laboratory show that the most important parameter is not the low temperature but the highly degassed state and the size of the fines. In a certain condition, less than 0.1% of the Radon may cross through a layer of 1cm.

3) - The erosion by solar wind is as well considered as a possible effect for removing Lead 210 from the surface of the rocks.

References
3. T.E. Economou and A.L. Turkevich, Houston Conference (January 1971)
ALPHA SPECTRUM OF LUNAR ROCK

G. Lambert


The Heat Flow Experiment is one of the instruments in the Apollo Lunar Surface Experiments Package (ALSEP) emplaced on the moon during Apollo 15. This instrument is designed to measure the heat flow from the lunar interior by determining the vertical gradient of mean temperature and the thermal conductivity in the upper few meters of the regolith.

The essential measurements are made by slender, meter-long temperature sensing probes that are placed in hollow, thin-walled fiberglass bores. These bores are drilled into the regolith by the Apollo Lunar Surface Drill. The instrument is designed to make two independent measurements of heat flow with two probes buried about 10 m apart.

The relatively high density of the regolith at Hadley Rille prevented full penetration of either of the two fiberglass bores to the planned 3 m depth. Both bores were driven about 1.6 m into the subsurface; however, one of the probes, designated #2, could not pass to the bottom of the borestem because of an obstruction. The bottom of this probe is only one meter below the surface.

Data received during the first 1 1/2 months after the probes were emplaced has been analyzed. During this period the initially warmer probe and bores stem cooled slowly toward the initial undisturbed temperature distribution in the regolith. The mean temperature at 100 cm below the surface is 252.3° K at probe #1 and 250.7° K at probe #2. These values are nearly 30° K higher than the average surface temperature over a lunation cycle, which is about 217° K. Most of this increase in mean temperature is due to the high dependence of heat transfer on ambient temperature in the upper several cm of the lunar soil.

Eight temperature sensors below 70 cm were unaffected by the large variations of surface temperature during the first 30 days after emplacement. The cooling histories of these thermometers were extrapolated to their equilibrium values using the theory of cooling of a cylinder in an infinite medium. After applying appropriate corrections for the shunting effect of the probe and bores stem, these extrapolated temperatures yield an accurate determination of the temperature field that existed in the subsurface before the probes were inserted. This analysis indicated that the undisturbed vertical temperature
gradient below 70 cm is $1.75 \pm 0.09^\circ$ C/M.

Each of the heat-flow probes has heaters surrounding four of the thermometers. These heaters can be turned on by command from earth to conduct in situ experiments to determine the thermal properties of the soil surrounding the boresite. Six such experiments were conducted in August and September. The analysis of these experiments indicated that the conductivity of the regolith is not uniform with depth. The values obtained from three experiments are tabulated below:

<table>
<thead>
<tr>
<th>Depth, cm</th>
<th>Conductivity, W/cm·°K</th>
</tr>
</thead>
<tbody>
<tr>
<td>49</td>
<td>$1.4 \times 10^{-4}$</td>
</tr>
<tr>
<td>91</td>
<td>$1.7 \times 10^{-4}$</td>
</tr>
<tr>
<td>138</td>
<td>$2.5 \times 10^{-4}$</td>
</tr>
</tbody>
</table>

*Estimated error of these measurements is ±10%.

An independent estimate of thermal conductivity can be made from the cool down history of the thermometers. Such estimates are dependent on knowledge of the initial temperature and consequently are not very accurate (±50%). Analysis of the first 100 hours of data indicate that the conductivity increases with depth at a rate similar to that shown by in situ measurements, the value at 91 cm is $1.5 \times 10^{-4} \pm 50\%$ and at 138 cm is $2.7 \times 10^{-4} \pm 50\%$. These values are in good agreement with the in situ determinations.

The conductivity data combined with the gradient determination indicate that the average heat flow through the regolith is $3.3 \times 10^{-6}$ watts/cm² (0.79 x $10^{-6}$ cal/cm²sec) at the probe #1 site. For comparison this value is about 1/2 the average heat flow from the earth.

The shallow emplacement of probe #2 prevents an accurate estimate of the mean gradient with only 1000 hours of data. When several lunations of data become available, the heat flow at this site can be calculated with suitable accuracy.

Because of the proximity of the heat flow measurement to the surface, it is susceptible to a number of disturbances. The principal ones are caused by topography and refraction of heat flow associated with sloping interfaces between materials of differing thermal conductivity. The effects of the most important topographic features, Hadley Rille and the Apennine Front, have been estimated by modelling them with very simple geometries. These preliminary estimates indicate that positive corrections on the order of 10 to 20% should be applied to the observed heat flow value.

Realizing that future measurements may produce major changes in our conclusions, we will here consider briefly the implications of the observed heat flow taking it at face value. For a planetary body the size of the moon very little heat flow results from initial temperatures even if they were at the solidus. Most of the present heat flow must result from the decay of long
lived radioisotopes. Isotopic abundances appropriate for ordinary chondrites would result in present rates of heat loss of 1.0 to 2.0 \times 10^{-6} \text{ watt/cm}^2. A moonwide heat flow of 3.3 \times 10^{-6} \text{ watts/cm}^2 would require much higher rates of heat production at present than "chondritic" bulk composition for the moon can provide.
With the successful installation of a geophysical station at Hadley Rille, on July 31, 1971, on the Apollo 15 mission, and the continued operation of stations 12 and 14 approximately 1100 km SW, the Apollo program for the first time achieved a network of seismic stations on the lunar surface. A network of at least three stations is essential for the location of natural events on the moon. Thus, the establishment of this network was one of the most important milestones in the geophysical exploration of the moon.

The major discoveries that have resulted to date from the analysis of seismic data from this network can be summarized as follows:

1. Lunar seismic signals differ greatly from typical terrestrial seismic signals. It now appears that this can be explained almost entirely by the presence of a dry, heterogeneous layer which blankets the moon to a probable depth of a few km with a maximum possible depth of about 20 km. Seismic waves are highly scattered in this zone. Seismic wave propagation within the lunar interior, below the scattering zone, is highly efficient. As a result, it is probable that meteoroid impact signals are being received from the entire lunar surface.

2. The moon possesses a crust and a mantle, at least in the region of the Apollo 12 and 14 stations. The thickness of the crust is approximately 65 km in the Fra Mauro region of Oceanus Procellarum and may consist of two layers. The contrast in elastic properties of the rocks which comprise these major structural units is at least as great as that which exists between the crust and mantle of the earth. (See Toksoz et al., for further discussion of seismic evidence of a lunar crust).

3. Natural lunar events detected by the Apollo seismic network are moonquakes and meteoroid impacts. Moonquakes are recorded at an average rate of 600 per year at station 12 and 1800 per year at station 14. The greater number of moonquakes detected at station 14 is a consequence of the thick layer of unconsolidated material at the Fra Mauro site which makes this station more sensitive. Sufficient data from station 15 have not yet been obtained to make a meaningful estimate of average rate of detection. All of the moonquakes are small (maximum Richter magnitudes between 2 and 3).
MOONQUAKES AND LUNAR TECTONISM

Gary Latham

4. Moonquakes can be divided into two categories: (1) repeating moonquakes with fixed source locations, and (2) moonquake swarms with variable source locations. With few exceptions, the repeating moonquakes occur at monthly intervals near times of perigee and also show correlations with longer-term (7-month) lunar gravity variations. They originate at not less than 10 different locations. However, a single focal zone accounts for 80 percent of the total seismic energy detected. The epicenter of the active zone has been tentatively located at a point 600 km SSW from the Apollo 12 and 14 stations. The focus is approximately 800 km deep. Each focal zone must be small (less than 10 km in linear dimension). Changes in record character that would imply migration of the focal zone or changes in focal mechanism have not been observed in the records over a period of 20 months. Cumulative strain at each location is inferred. Thus, the moonquakes appear to be releasing internal strain of unknown origin, the release being triggered by tidal stresses. The occurrence of moonquakes at great depths implies that (1) the lunar interior at these depths is rigid enough to support appreciable stress, and (2) that maximum stress differences occur at these depths. If the strain released as seismic energy is of thermal origin, this places strong constraints on acceptable thermal models for the deep lunar interior.

5. Episodes of frequent small moonquakes, called moonquake swarms, have been observed. During periods of swarm activity, events occur as frequently as one event every 2 hours during intervals lasting several days. The occurrence of moonquake swarms shows a semi-monthly periodicity; hence, they also appear to be induced by lunar tides. The sources of moonquake swarms have not been determined, but they appear to be widely distributed. Moonquakes in one swarm show a gradual increase in magnitude over a period of 2½ days, ending abruptly with the largest moonquake yet recorded. This moonquake is distinguished not only by its size, but by the high frequency of the signal it generated. The epicenter has been located in the vicinity of Mare Crisium. This pattern of activity is similar to that observed from terrestrial volcanic eruptions. Thus, while the main episode of volcanism ended on the moon 3 billion years ago, it may not be completely extinct.

6. The average rate of seismic energy release within the moon is far below that of the earth. Thus, internal convective motion leading to significant lunar tectonism appears to be absent. Presently, the outer shell of the moon appears to be relatively cold, rigid, and tectonically stable compared to the earth. Moonquake swarms may be generated within the outer shell of the moon as a result of continuing minor adjustments to crustal stresses.
BULK,REE AND OTHER ELEMENTAL ABUNDANCES IN FOUR APOLLO 14 CLASTIC ROCKS AND THREE CORE SAMPLES, TWO LUNA 16 BRECCIAS AND FOUR APOLLO 15 SOILS.

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Abundances of eight bulk elements (TiO\textsubscript{2}, Al\textsubscript{2}O\textsubscript{3}, FeO, CaO, Na\textsubscript{2}O, K\textsubscript{2}O, MnO and Cr\textsubscript{2}O\textsubscript{3}) and ten trace elements (Sc, V, Co, U, Ba and five REE, La, Sm, Eu, Yb and Lu) have been measured by sequential INAA. Results are given in Table 1. Abundances of five additional trace elements, (Zr, Hf, Tb, Ta and Th) are presently in data reduction. Abundances of Dy, given in our preliminary abstract, are not reliable REE abundance indicators.

Apollo Core Tube Samples 14230 (STA G, 14, 19 and 22cm depths): The core tube was drilled ~100m E of North Crater and presumably in North and Center Crater ejecta. Abundances of all 18 bulk and trace elements in the three core tube samples, 56-74mg, agree with those obtained in three Ap. 14 soils gathered at the LM and at STA G. This suggests a uniform soil composition at least to ~22cm depth within ~200m distance from the LM landing site.

Clastic rock 14063, STA Cl (Tatsumoto consortium): This rock was collected from the continuous Cone Crater ejecta blanket. Samples of 59-81mg were analyzed by us before they were analyzed by E. Anders and coworkers. The compositions of five Al\textsubscript{2}O\textsubscript{3} rich (>20%) clasts generally are rather similar in composition; however, detailed inspection shows some differences in their REE contents. Clasts BS and B11-3 are quite similar. All have low KREEP or noritic component.

Clastic rock 14083 STA Cl (Wasserburg consortium): This rock was a piece of 14082 rock, which was chipped from the top surface of the "White" rock, ~20m near Cone Crater rim. Abundances in a black and white fragment and in a pure dark interior sample are quite similar to the composition of KREEP or noritic matter for all elements, except for the high Cr\textsubscript{2}O\textsubscript{3}abundance in the pure dark matter. The higher CaO and Al\textsubscript{2}O\textsubscript{3} contents and lower FeO and TiO\textsubscript{2} suggest higher plagioclase and lower pyroxene and ilmenite contents in the pure white matter. Since it appears probable that clastic rocks 14063 and 14083 are Cone Crater ejecta, the chemical dissimilarity between these two rocks suggests a very complex and heterogeneous stratigraphy for the Frau Mauro formation. The same conclusion was reached by LSPET (1), who analyzed two rocks, 14321 and 14065, both of which were on the Cone Crater ejecta blanket and which also yielded quite dissimilar chemical compositions.

Clastic rock 14318 STA H (Tatsumoto consortium): Although picked up ~100m NW of the LM, this rock was in the Cone Crater ejecta blanket (2). Based on the analysis of a 31mg mixed sawdust sample, the overall rock composition is slightly lower in Al\textsubscript{2}O\textsubscript{3} content and slightly higher in FeO compared to igneous rock 14073. The close similarity in the abundances of all the other sixteen
bulk and trace elements in the clastic 14318 and igneous 14073 rocks suggests a genetic relationship between them. Also the similarity in composition between 14318 and the average Ap. 14 soil suggest that either the soil is pulverized from breccia rocks such as 318 or that 318 was compacted from the soil in a meteoritic impact event. If ejected from Cone Crater, the chemical composition of 14318 underscores the complexity of the Frau Mauro formation as noted in the 14083 discussion. The composition of clasts 26A and 26B are largely KREEP matter. Clasts 26C and 27B have much higher K₂O and Ba abundances compared to KREEP; however, normal KREEP abundances of U were found in these two clasts.

Clastic rock 14066 STA F (Reynolds consortium): This rock was picked up from smooth terrain, ~100m SSE of Weird Crater. The compositions of the sawdust and four clasts are all different. The LSPET (1) analysis for 14066 is in fair agreement with this work. The overall rock (sawdust) composition is slightly lower in Al₂O₃ and higher in K₂O and REE compared to the average in Ap. 14 soils. The KREEP component in clastic rock 14066 is higher than that observed in clastic rock 14318; e.g. the REE contents in 14318 are 220% higher compared to rock 14066. Both the "Matrix small clasts" (21,1.01 sample) and the "White igneous clast" (21,2.04,2 sample) have high K₂O abundances of 1.3 and 1.44%, respectively. This high K₂O abundance is not correlated with the REE abundances, which are about half the average REE abundances observed in KREEP. Many fragments in rock 12013 also had high K₂O abundances and low REE abundances compared to average KREEP composition (3). REE abundances in all samples from the four clastic rocks studied in this work exhibit chondritic normalized REE patterns that are parallel to the REE pattern for KREEP; these patterns differ from the concave REE patterns observed in six fragments of 12013 (3).

Luna 16 breccias A36 (~7cm) and G41 (~30cm depth): In general, the chemical compositions of these small breccia samples are similar to the Luna 16 soil composition (4) with the exceptions of the alkalis; i.e., Na₂O and K₂O abundances are higher in these breccias by factors of two or greater.

Apollo 15 soils, 15021,24 (LM), 15471,49 (STA 4), 15501,40 (STA 9) and 15531,52 (STA 9a): In general, the chemical compositions of all four soils are different; e.g. lower alkali and Al₂O₃ abundances were found soil 15531 (the near Rille soil) and the highest Cr₂O₃ and V contents in any lunar soils were found in 15531 soil. Abundances of the REE in these four Ap. 15 soils vary within a wide range; e.g. La varies from 10.9 to 26ppm. These REE abundances are lower than those measured in Ap. 11, 12 and 14 soils. The REE abundances in 15531 soil are less than the REE abundances in Luna 16 soil (4). The Chondritic normalized REE patterns of the Apollo 15 soils parallel the observed distribution in 12070 soil.

References and Notes
5. This work was supported by NASA grants NGL 38-002-039 and 38-002-020.
Heating experiments at 950–990°C confirm the hypothesis that lunar pyroxferroite is metastable at low pressures with respect to the three-phase assemblage Ca-rich clinopyroxene plus fayalite plus tridymite. Rate studies of the breakdown reaction may provide information on the cooling rates of lunar igneous rocks containing pyroxferroite.

The discovery of the new mineral pyroxferroite in a number of lunar igneous rocks has posed an enigma, for phase equilibrium studies had shown that synthetic Mg-free pyroxferroite is stable only at pressures greater than 10 kb (Lindsley & Burnham, Science, 1970). The lunar occurrences—late-stage minerals associated with tridymite, cristobalite, and glass in volcanic or near-surface rocks—strongly suggested, on the other hand, that the natural material crystallized at very low pressure. One possible explanation was that lunar pyroxferroite might be stabilized by minor constituents that are missing in the synthetic material. However, the metastable crystallization from synthetic glasses of Ca-Fe pyroxenoids approaching pyroxferroite in composition (Bowen, Schairer, and Posnjak, Am. J. Sci., 1933; see also Lindsley and Munoz, Am. J. Sci. 267A, 1969) suggested that pyroxferroite may have crystallized metastably and was then cooled so rapidly that a more stable assemblage was unable to form. This hypothesis received support from observations of an apparent breakdown symplectic—fayalite, a silica mineral, and a Ca-enriched metasilicate—after pyroxferroite (Ware and Lovering, Science, 1970) which indicates that at least some pyroxferroites may have cooled sufficiently slowly to permit transformation to a more stable assemblage.

Single crystals of pyroxferroite were separated from samples 12021, 38 and 12021, 109 by E. Dowty for Mössbauer studies. Selected crystals, from 0.1 to 0.2 mm, were mounted in epoxy, polished, and analyzed by electron microprobe for Fe, Ca, Mg, Si, Al, Ti, and Mn; the range of compositions Wo12En13Fs75-Wo14En4Fs82 is closely similar to that reported by Boyd et al. for another sample of rock 12021. Single-crystal precession x-ray diffraction photographs confirmed that the crystals are pyroxferroite; some showed minor amounts of oriented clinopyroxene (Table I). It is not yet clear whether this clinopyroxene occurs as fine lamellae or as epitaxial discrete crystals. Next the pyroxferroite crystals were wrapped in envelopes of Ag70Pd30 foil, sealed in evacuated silica-glass tubes, heated to temperatures ranging from 950–990°C, and quenched in water.

In initial experiments (14163, 14168, Table I), the crystals were converted to Fe-depleted clinopyroxene + tridymite, apparently by incongruent vaporization of FeO-rich components (although the silica-glass capsules were made as small as convenient, their capacity was approximately 10⁵ times the volume of the pyroxferroite crystals). To counteract this effect, the free
Pyroxferroite Breakdown
D. H. Lindsley

Pyroxferroite with several hundred times the mass of the lunar crystal was added. In experiments utilizing these precautions (14167 and 14166, Table I), one pyroxferroite crystal (14166) heated for a total of 5 days, 23 hours, at 960±3°C did not break down but did develop weak reflections in the x-ray diffraction photographs attributable to the appearance of oriented clinopyroxene. Another (14167) was transformed completely in 68.5 hours at 990±3°C to an intimate intergrowth of clinopyroxene + fayalite + tridymite (Fig. 1); the grain size of the clinopyroxene and fayalite was just coarse enough to be resolvable by electron microprobe techniques (analyses given in Table I). Precession photographs and optical examination reveal that the fayalite and tridymite occur as minute, randomly oriented inclusions within a single crystal of clinopyroxene that presumably formed topotaxially with respect to the parent pyroxferroite.

These preliminary findings may indicate that lunar basalts containing pyroxferroite have cooled to temperatures below 990°C in less than 3 days.

Fig. 1. Portion of pyroxene quadrilateral showing compositions of pyroxferroite crystal 14167, and of the breakdown products clinopyroxene and fayalite + tridymite (projected).
Pyroxferroite Breakdown
D. H. Lindsley

Table I. Data for heating experiments on pyroxferroite.

<table>
<thead>
<tr>
<th>Starting Materials Crystal No.</th>
<th>14163</th>
<th>14168</th>
<th>14167</th>
<th>14166</th>
</tr>
</thead>
<tbody>
<tr>
<td>Probe Composition</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Pyroxferroite</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>a(Å)</td>
<td>6.62</td>
<td>6.62</td>
<td>6.64</td>
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<tr>
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<tr>
<td>Tridymite</td>
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<td>YES</td>
<td>YES</td>
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</tr>
<tr>
<td>Fayalite</td>
<td>NO</td>
<td>NO</td>
<td>FO7</td>
<td>---</td>
</tr>
</tbody>
</table>

Mainly pyroxferroite (minor pyroxene)
VANADIUM ISOTOPIC COMPOSITION AND THE CONTENTS OF CHROMIUM, IRON MAGNESIUM, TITANIUM AND VANADIUM IN APOLLO 11 AND 14 LUNAR FINES AND AN APOLLO 14 ROCK; M.E. Lipschutz (Chem. and Geosciences Depts., Purdue Univ., Lafayette, Ind. 47907 and Visiting Associate Prof., Dept. of Physics and Astronomy, Tel-Aviv Univ., Ramat-Aviv, Tel-Aviv, Israel), H. Balsiger (Physik. Inst., Univ. Bern, Switzerland) and P. Rey (Chem. Dept., Purdue)

A number of noble gas nuclides constitute very sensitive monitors for the irradiation of solid material by energetic charged particles and, based upon study of such nuclides, it is clear that the lunar surface has been thus irradiated during the last several hundred million years. Suggestions have also been advanced that lunar material (as part of the solar system) was irradiated in a similar manner early in its history, perhaps under conditions such that gaseous nuclides (and other temperature- or time-dependent) radiation monitors would not be retained. In order to examine this possibility, it is necessary to utilize a radiation monitor that is insensitive to these other processes and one of the most sensitive monitors satisfying this requirement is the vanadium isotopic composition.

In previous studies (Balsiger et al., 1969; Pelly et al., 1970) we established that the vanadium isotopic composition in terrestrial samples was the same as that in eight chondrites (principally primitive ones of petrologic grades 1-4) to within 1%. Albee et al. (1970) noted that the vanadium isotopic composition in a sample of fines and a rock from Apollo 11 was the same as that in terrestrial samples to within 2% and 3%, respectively, and we obtained similar results (to within 1%) from study of fines and three rocks from Apollo 12 (Lipschutz et al., 1971). However, the $^{50}$V/$^{51}$V ratios in Apollo 12 lunar material tended to be slightly higher than those in meteorites and we felt it worthwhile to determine whether this trend was apparent or real by studying additional lunar and meteoritic samples. Here we report $^{50}$V/$^{51}$V results for fines from Apollo 11 and 14 and for two different portions of a fragmental rock from Apollo 14 and the concentrations of V, Cr, Fe, Mg and Ti in aliquots of the same samples.

The lunar specimens investigated included a sample of the Apollo 11 fines (10084) studied by Albee et al. (1970) and a sample of Apollo 14 fines (14163). The fragmental Apollo 14 rock (14321,184) appears quite complex petrographically; it mainly consists of dark crystalline lithic fragments (lunar basalt types with sub-ophitic structures) and fragmental lithic fragments (dark, consolidated microbreccia) embedded in a lighter colored fragmental matrix (D. Weill, personal communication). Inasmuch as we were particularly interested in determining the $^{50}$V/$^{51}$V ratios in primitive material we obtained samples of the two inclusion types (sub-samples 1c and 20/22, respectively) but not the matrix.

All specimens (1.5-2 grams) were treated chemically as described by Pelly et al. (1970) except that about 10-15% of each original sample solution was reserved for determination of Cr, Fe, Mg and Ti by atomic absorption spectro-
VANADIUM ISOTOPIC COMPOSITION

M. E. Lipschutz

Vanadium was determined by isotope dilution and the 50\textsubscript{V}/51\textsubscript{V} isotopic ratios in the unspiked samples were determined mass-spectrometrically using techniques described by Balsiger et al. (1969) and Lipschutz et al. (1971). The contents of Fe, Mg and Ti in standard rock BCR-1 and Cr in standard rock DTS-1 are similar to those of lunar material and we measured these to test the specific atomic absorption techniques used for analysis of lunar samples.

The 50\textsubscript{V}/51\textsubscript{V} ratios in these lunar samples (Table 1) are similar to those in four Apollo 12 samples (Lipschutz et al., 1971). The maximum difference of the group means is 1%, corresponding to the lunar-meteoritic difference plus two weighted standard deviations, (0.012±2x0.005)x10\textsuperscript{-3}. This difference may arise from long-term instrumental drift or from an actual difference in the integrated proton fluxes (Δ\phi) received by these two sorts of solar-system material. In the latter case the maximum limits are 1.6 x 10\textsuperscript{18} "hard" protons/cm\textsuperscript{2} (cosmic-ray spectrum, E>30 MeV), and 8 x 10\textsuperscript{19} "soft" protons/cm\textsuperscript{2} (solar-flare spectrum, E>10 MeV), respectively (cf. Balsiger et al., 1969).

The conflicting trends exhibited by the ferromagnesian element contents in the basaltic and microbreccia portions of rock 14321 (Table 3) are in accord with its complex petrographic characteristics. The contents of Fe, Cr and V are lower in the microbreccia than in the basaltic portion of this rock whereas Mg exhibits the opposite behavior. The contents of Fe, Cr and V in the microbreccia and of Mg in the basaltic portion are similar to those in lunar fines 14163. Titanium is more abundant in the basaltic portion than in the microbreccia and both portions contain more Ti than does fines sample 14163. Our data are generally similar to those of LSPET (1971) in that Fe, Ti, Cr and V are less abundant in Apollo 14 than in Apollo 12 samples whereas the Mg contents are generally similar. In our limited sampling of Apollo 14 material V tends to follow Cr rather than Fe as in Apollo 12 igneous rocks and fines (Lipschutz et al., 1971).

References

### Table 1. Vanadium isotopic ratios and concentrations in lunar samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>$^{50}V/^{51}V^*$</th>
<th>V (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>10084,247</td>
<td>fines</td>
<td>$(2.470\pm0.064)\times10^{-3}$</td>
<td>$\dagger$</td>
</tr>
<tr>
<td>14163,129</td>
<td>fines</td>
<td>$(2.477\pm0.065)\times10^{-3}$</td>
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</tr>
<tr>
<td>14321,184-1c</td>
<td>lunar basaltic</td>
<td>$(2.456\pm0.015)\times10^{-3}$</td>
<td>133</td>
</tr>
<tr>
<td>14321,184-20/22</td>
<td>microbreccia</td>
<td>$(2.483\pm0.047)\times10^{-3}$</td>
<td>44</td>
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</tbody>
</table>

* The uncertainty listed for each sample includes a statistical error (three estimated standard errors) as well as those errors arising from correction of mass-50 peak (cf. Lipschutz et al., 1971).

† This portion of the sample was lost in a laboratory mishap.

(a) LSPET (1971)

### Table 2. Mean $^{50}V/^{51}V$ ratios in lunar, terrestrial and meteoritic samples

<table>
<thead>
<tr>
<th>Material</th>
<th>No. investigated</th>
<th>Weighted $^{50}V/^{51}V$ and standard deviation</th>
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</thead>
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<td>$(2.461\pm0.003)\times10^{-3}$</td>
</tr>
<tr>
<td>Terrestrial</td>
<td>4</td>
<td>$(2.458\pm0.004)\times10^{-3}$</td>
</tr>
<tr>
<td>Meteoritic</td>
<td>10</td>
<td>$(2.449\pm0.004)\times10^{-3}$</td>
</tr>
<tr>
<td>All</td>
<td>22</td>
<td>$(2.455\pm0.002)\times10^{-3}$</td>
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### Table 3. Fe, Mg, Ti and Cr in standard rocks and lunar samples

<table>
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<th>Mg (percent)</th>
<th>Ti (percent)</th>
<th>Cr (percent)</th>
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<td>-</td>
<td>0.014(a)</td>
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<td>BCR-1</td>
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<td>13.98</td>
<td>12.3(c)</td>
<td>4.96</td>
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<td>8.22</td>
<td>8.06(b)</td>
<td>5.77</td>
<td>5.9(b)</td>
</tr>
<tr>
<td>14321(1c)</td>
<td>12.07</td>
<td>10.1(b)</td>
<td>5.68</td>
<td>7.2(b)</td>
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<tr>
<td>14321</td>
<td>7.75</td>
<td>7.0(b)</td>
<td>7.03</td>
<td>8.6(b)</td>
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</tbody>
</table>

* References: (a) Flanagan (1969) - mean values; (b) LSPET (1971); (c) Wänke et al. (1970) - mean values of all investigations of Apollo 11 fines.
Earthshine and Near Terminator Photography

D. D. Lloyd


Earthshine Photography During Apollo 15, fifteen photographs of the moon were taken under Earthshine illumination. These were the first Earthshine photographs taken from lunar orbit.

The photographs are of photometric interest, particularly as they involve double reflection of sunlight--by the earth, then the moon--prior to photographic exposure. Certain published data on the mean illumination of the moon by the crescent earth predicted lower exposure values than obtained for each measured area. The apparent albedo values obtained for the floor of the crater Aristarchus were anomalously higher than those obtained of the surrounding maria.

Preliminary geological observations suggest that the special Earthshine illumination conditions make possible some new analyses. For example, new brightness and contrast patterns not seen before include: (1) a bright area highlighted in the southern portion of the floor of the Cobra Head; and (2) bright and dark bands located on the walls and rim of the crater Aristarchus that are oriented radially.

Near Terminator Photography During Apollo 15 ten sequences of photographs were taken starting a few minutes prior to crossing of the terminator and continuing past the terminator.

The preliminary scientific results apply to three general areas:

1. A sequence of photographs obtained in Mare Vaporum showing maria structure in detail.
2. Four sequences in the Aristarchus plateau area and in Oceanus Procellarum area west of the Aristarchus plateau.
3. Five sequences in highland areas.

Selected pictures are reproduced and discussed in the accompanying figures.

Many photographs were obtained that showed lunar surface areas within a few degrees of the terminator and are of significant geologic interest. Many geological features stand out in a manner not normal in conventional lunar photography, providing additional data on the surface morphology and configuration of a large number of lunar surface structures.

Slight variations (less than 1/2°) in slopes near horizontal produced significant variations in scene contrast. This special effect enables the eye to recognize unusual geological features not recognized before—for example, a dome at least 25 km in diameter in Mare Vaporum.

In principle, a mid-infrared emission spectrum of a planetary surface offers information diagnostic of bulk mineralogical composition, or rock type. This is because features which occur in mid-infrared spectra are caused by vibrations within the basic repeating units of the material. Thus, the number and location of the bands depends directly upon the molecular structure and, hence, on mineralogical composition.

There has, however, been serious disagreement concerning the amount of such information available in lunar infrared emission. Early laboratory work showed that the molecular vibration bands (reststrahlen bands) of silicate minerals become more difficult to detect as the particle size of the minerals becomes small (Lyon, NASA Tech. Note TND-1871, 1963). Because the lunar surface layer has long been known to be composed primarily of fine particulate material, it has been generally assumed that it radiated essentially as a black body in the mid-infrared.

Yet, early observations of lunar emission did detect differences in spectral behavior from place to place on the surface (Hunt & Salisbury, Science, 146, 641, 1964), and one observer determined that lunar emission departed significantly from black body behavior (Murcray, J.G.R., 70, 4959, 1965). On the other hand, later observations indicated that, although the emissivity of some areas differed from that of their surroundings, the emissivity of most areas did not (Goetz, J.G.R., 73, 1455, 1968). The most recent measurements, which have the advantage of being made with a balloon-borne telescope above most of the infrared-absorbing constituents of the atmosphere, showed that the moon did, indeed, depart from black body behavior (Murcray et al, J.G.R., 75, 2662, 1970). They also showed that the spectra display a pronounced peak near 8.0 microns, the position of which varies from place to place on the surface (see Figure 1).

Although the presence of an emissivity peak was explained by the work of Conel (J.G.R., 74, 1614, 1969) as a spectral feature associated with the principal Christiansen frequency, this work predicted a 2 or 3% feature at most, rather than the 10 to 20% effect shown in Figure 1. Logan and Hunt (Science, 169, 865, 1970) have recently shown that this large departure from black body behavior is due to the thermal environment of the lunar surface. That is, heating of the surface by illumination in vacuum, while it radiates to cold space, will result in a very steep thermal gradient near its surface.
MID-INFRARED SPECTRA
L. M. Logan

Figure 1. Spectral emissivity of six different lunar regions obtained with a balloon-borne telescope. The strong band near 9.6 microns is due to absorption by residual ozone in the earth's atmosphere. (From Murcray et al., J.G.R., 75, 2662, 1970).

Such a thermal gradient is accompanied by a strong emissivity peak.

Laboratory study of terrestrial rocks in a simulated lunar environment has demonstrated that the position of the emissivity peak, together with the average departure of a spectrum from black body behavior, are two parameters that serve to clearly distinguish among different rock types (Logan et al, in preparation). To apply the technique to the moon, however, it must be demonstrated that lunar rocks follow the same rules of peak location and average emissivity as do terrestrial rocks.

To this end the emission spectrum of a sample of Apollo 14 soil (14259) has been obtained in a simulated lunar environment (see Figure 2). We find that the peak location and average emissivity of this sample are typical of silica-poor basaltic rocks, which is entirely consistent with the composition of the soil. The departure from black body behavior displayed by the Apollo
MID-INFRARED SPECTRA

L. M. Logan

14 soil spectrum is generally not so great as that determined by the balloon-borne telescope system for some areas on the moon. This may be due to a true difference in emissivity for these areas, or to inadequate calibration of the balloon spectrometer. The position of the Apollo 14 soil emission peak (8.24 microns) is identical with that previously determined for the Central Highlands with the balloon system. That the two peaks are identical is reasonable, if the regolith in the Central Highlands and the debris ejected from the Imbrian basin to form the Frau Mauro formation are both derived from the lunar crust. It is noteworthy that an emissivity peak at this long a wavelength implies that the Central Highlands regolith cannot be derived from "anorthosite", if by that term is meant a rock composed almost entirely of calcic plagioclase. On the contrary, the regolith must contain abundant pyroxene and/or olivine, as does the Apollo 14 soil.

Figure 2. Emission spectrum of Apollo 14 soil obtained in a simulated lunar environment.

We conclude, therefore, that 1) the moon deviates significantly from black body behavior in its infrared emission; 2) spectra already available of different areas on the lunar surface indicate that the Central Highlands material, although less basic than the maria materials, is gabbroic rather than anorthositic; and 3) mid-infrared spectroscopy is a valid lunar remote sensing tool, which could be most useful in indirect exploration.
URANIUM GEOCHEMISTRY AND LATE-STAGE (MESOSTASIS) MINERALOGY OF APOLLO 14 LUNAR ROCKS.

J.F. Lovering, D.A. Wark, D. Sewell and C. Frick, School of Geology, University of Melbourne, Parkville, Victoria 3052, Australia.

I. BASALTIC VOLCANIC ROCKS: Fission track "Lexan" print studies indicate an overall 0.56 ± 0.06 ppm U in an Apollo 14 fine to medium grained porphyritic olivine basalt 14072-4. The uranium is generally concentrated in micron-size phases within the mesostasis areas with chlor-apatite (100-600 ppm U) the most abundant source (~75% of highly enriched sources), followed by whitlockite (~50 ppm U, ~20% of sources) and baddeleyite (~170 ppm U, ~5% of sources). A zircon grain has been recognized in this rock and is the first unequivocal zircon occurrence in a lunar rock of basaltic composition. Uranium abundances of 3 to 4 ppm are also observed in minor K-rich "glass" areas up to 100 μ across in the mesostasis. The zirconium enriched mineral assemblage in the mesostasis of this Apollo 14 basalt (i.e. baddeleyite-zircon) is quite distinct from the baddeleyite tranquilityite-phase B assemblages previously observed by Lovering and Wark (1971) in the Apollo 11 and 12 basaltic rocks. The Apollo 14 basalts are apparently enriched in zirconium relative to the Apollo 11 and 12 basalts and the late-stage zirconium-rich phases forming probably reflect these and other subtle differences in composition.

II. BRECCIAS: The Apollo 14 breccias are enriched in uranium relative to the Apollo 14 basalt 14072-4 but very similar to 3.7 ± 0.6 ppm uranium abundances reported in 14310. Breccia 14313-14 appears to fall into Group 2 Breccias as defined by LSPET (1971a). Many clasts consist of medium to highly shocked fragments of basaltic mineralogy with overall uranium contents of 3 to 6 ppm U (cf. basalt fragment 14310) in which the uranium-enriched phases are observed to occur in the mesostasis. The most abundant uranium-enriched sources are whitlockite and/or chlor-apatite (less than 10 microns across) with baddeleyite and zircon coating ilmenite grains and rare tranquilityite sources. K-rich "glass" is also observed with a uranium content of ~1.1 ppm U. Other clasts consist of buff-coloured glasses with uranium contents ranging from 2 to 5 ppm U. These glasses have average compositions around ~53% SiO₂, ~1.5% TiO₂, ~13% Al₂O₃, ~10% FeO, ~10% CaO, ~13% MgO, ~1% K₂O. They commonly contain relics of high-uranium mineral grains (i.e. zircon ~45 ppm U; whitlockite, ~65 ppm U), up to 50 microns across. These glasses and their relic high-U phase, closely resemble KREEP glasses observed in Apollo 12 breccias. There are also other clasts of rounded brown glass with < 0.4 ppm U and individual mineral fragments with very low U abundances (e.g. plagioclase ~2.5 ppm U). Individual mineral clasts of uranium-enriched phases zircon (up to 40 microns
across with \( \sim 92 \) ppm U) and baddeleyite occur. A few grains of a high-relief, red-brown mineral with high uranium abundances (\( \sim 1000 \) ppm U) have been observed. Preliminary analyses indicate a composition enriched in Zr-Ti-Fe-Ca and very closely related to phase B described in Apollo 11 and 12 rocks by Lovering and Wark (1971).

Breccia 14305-77 is probably a Group 3 type (LSPET, 1971a) in which both the matrix and the abundant clasts show evidence of variable recrystallisation. One breccia clast (containing \( \sim 3.7 \) ppm U) within breccia 14313-14 was texturally identical to this recrystallized breccia 14305-77 suggesting that the Group 2 breccia 14313 was formed after some recrystallized Group 3 breccias had formed. Breccia 14305-77 contains numerous mineral fragments of uranium-poor phases (e.g. plagioclase, pyroxene, olivine, etc.) and uranium-enriched phases (e.g. whitlockite \( \sim 64 \) ppm U) and phase B (\( >1000 \) ppm U). Rock clasts of partially recrystallized plagioclase-rich igneous rocks and partially devitrified glasses are common. A characteristic recrystallized volcanic rock clast composed of anorthitic plagioclase laths, irregular outlined clinopyroxene grains and minor K-feldspar grains set in an abundant matrix of pale-brown "glass" high in SiO\(_2\) (\( \sim 90\% \)) and K\(_2\)O (\( \sim 7\% \)) with minor FeO, TiO\(_2\), and CaO shows a high overall U content of \( \sim 8.2 \) ppm U. Individual U-enriched minerals (e.g. zircon, \( \sim 160 \) ppm U, coating on ilmenite; whitlockite, 60-100 ppm U) occur but most of the uranium in the clast is contained within the interstitial Si-,K-rich "glass" areas which have variable U contents ranging from 10 to 13 ppm U. Another highly shocked and partly crystallized volcanic clast composed of plagioclase, clinopyroxene and ilmenite contains \( \sim 3.6 \) ppm U overall with uranium concentrated in an interstitial K-rich material and in zircon and Ca-phosphate grains. The partly recrystallized glasses typically show relict uranium-enriched Ca-phosphate and other uranium-poor phases set in a matrix of red-brown and rare colourless glass. The red-brown glasses are variable in composition but are enriched in SiO\(_2\) (\( \sim 87\% \)) and K\(_2\)O (\( \sim 12\% \)) with minor amounts (\(< 2\% \)) of FeO, TiO\(_2\), CaO, MgO. Uranium concentrations range from 5 to 17 ppm U.

Clearly the relatively uranium-enriched breccias contain clasts of igneous rocks themselves enriched in uranium (and probably also potassium) with the uranium highly concentrated in individual whitlockite, apatite, baddeleyite and zircon crystals and with significant concentrations of uranium (\( \sim 17\% \)) in K-rich "glass" areas in the mesostasis. These rock clasts are most likely the source of the high-U glasses also common in the breccias. The source of the individual mineral clasts of the Zr-Ti-Fe-Ca phase B in the breccias cannot be immediately traced to any volcanic clasts observed in this study.
NEUTRON AND SPALLATION EFFECTS IN FRA MAURO REGOLITH. G. W. Lugmair and K. Marti, Chemistry Dept., University of California, San Diego, La Jolla, California 92037

Cosmic rays, both galactic and solar, have bombarded lunar surface materials for hundreds of millions of years. Induced nuclear transformations can be used to unravel the fate of individual rocks and soil particles, also turnover and erosion rates can be determined. In previous publications we have attempted to combine and correlate spallation data with anomalies produced by neutron capture in Gd to determine average irradiation depths for several Apollo 12 rocks (Marti and Lugmair, 1971; Lugmair and Marti, 1971). The present paper reports on an extension of this work to Apollo 14 material including a number of soil fragments ("peanuts", 4-10mm). Particle track densities were determined by Arrhenius et al. (1972), and these data can be correlated with our own. Our Apollo 12 rock data showed that the relative spallation mass yields of Kr can be used as a shielding or depth parameter. Also, Lingenfelter et al. (1971) have shown that slow neutron flux data could, in principle, be used as depth parameter whenever it is possible to decide between two possible solutions. These parameters, however, only define relative depth scales as long as no calibration data are available for both the spallation Kr yields and the neutron flux below the lunar surface.

Neutron capture anomalies in $^{158}$Gd/$^{157}$Gd up to 0.7% are observed in Apollo 14 materials (Table 2). On the other hand, Gd in rock 14321 and terrestrial Gd are very similar. In addition to anomalies in Gd due to capture of <0.3eV neutrons, we report anomalies in Kr of Apollo 14 samples due to capture in Br of epithermal resonance (30 - 300eV) neutrons. Kr in Apollo 11A rocks and in rock 14321 is considered to represent pure spallation Kr. Since Zr/Sr abundance ratios appear to be similar in 11A and 14 rocks, the $^{80}$Kr and $^{82}$Kr spallation yields of all other rocks are obtained by interpolation. Computed relative $^{80}$Kr and $^{82}$Kr excesses are plotted in Figure 1. The correlation of the two relative excesses is very good, and the slope agrees with calculated Br resonance integrals. We consider this as strong evidence for resonance neutron capture in Br. Figure 2 shows that the $^{80}$Kr excess also correlates with the neutron dosage $\phi$ obtained from the Gd isotopic anomalies. The ratios $^{131}$Xe/$^{126}$Xe in our samples correlate with both the $^{78}$Kr/$^{83}$Kr spallation ratio and the neutron dosage $\phi$.

In order to get more information on the history of Apollo 14 soils and their components, we have determined $^{81}$Kr-Kr ages of a number of individual fragments and have investigated their spallation and neutron capture records (Tables 1 and 2). In cases where anomalies from Br are found, the production ratio $P_{81}/P_{83}$, which is used in Kr dating, cannot be obtained by interpolation from measured $^{80}$Kr and $^{82}$Kr, but can be inferred from the $P_{81}, P_{83}$-correlation of Apollo 11A rocks and rock 14321 since the major target elements, Zr and Sr, have similar abundance ratios. The $^{81}$Kr-Kr ages obtained range from 260 to 700 m.y. (Table 2). Most of the fragments
Neutron and Spallation Effects in Fra Mauro Regolith

G. W. Lugmair

have been shielded to a variable degree, but at least one indicates a close-
to-surface irradiation. Different pieces of information, including a track
record, seem to give a consistent picture.

A grouping of exposure ages around 27 m.y. for materials from near Cone
 crater appears to date this impact event. All other samples indicate much
larger exposure ages.

The search for a large lunar rock with a simple irradiation history con-
tinues. Results we have obtained from the largest Apollo 15 rock 15555 indi-
cate a well shielded location. However, the orientation of our small chip
relative to the rock and the lunar surface are unfortunately not known, and a
gradient, therefore, cannot be ruled out. The Kr data are given in Table 1. An
$^{81}$Kr-Kr exposure age of $81 \pm 18$ m.y. is calculated from these data. The
spallation Xe data, in contrast to Kr, are different from those in Apollo 12
rocks and indicate a higher abundance ratio of the REE and Ba.

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flux revisited; manuscript to be submitted to Earth Planet. Sci.

Lugmair, G. W. and Marti, K.; Neutron capture effects in lunar gadolinium
and the irradiation histories of some lunar rocks; Earth Planet. Sci.

Marti, D. and Lugmair, G. W.; $Kr^{81}$-Kr and K-Ar$^{40}$ ages, cosmic-ray spal-
lation products, and neutron effects in lunar samples from Oceanus Pro-
Suppl. 2, 2, 1591 (1971).

Table 1: Isotopic composition and concentration of Kr, normalized to $^{83}$Kr = 100,
in Apollo 14 and 15 samples.

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<thead>
<tr>
<th>14160-4</th>
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</table>
Neutron and Spallation Effects in Fra Mauro

G. W. Lugmair

Table 2. Neutron Fluxes (10^3/n/cm^2) and Irradiation Ages in Apollo 14 Rocks and Rock Fragments

<table>
<thead>
<tr>
<th>SAMPLE</th>
<th>%K, COBALT SAT</th>
<th>BF</th>
<th>SF</th>
<th>T Neutron</th>
<th>SF Neutron</th>
<th>SF COBALT SAT</th>
<th>ERR</th>
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<tr>
<td>14604-1</td>
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<td>0.27</td>
<td>299</td>
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<td>1.93 ± 0.29</td>
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<td>697</td>
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<td>299</td>
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<tr>
<td>14604-23</td>
<td>1.508 37 ± 0.000 15</td>
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<td>299</td>
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<tr>
<td>Terr 60X</td>
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<td>1.93 ± 0.29</td>
<td>697</td>
<td>0.27</td>
<td>299</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*1* For the calculation of neutron fluxes, 2 x 10^19 n/cm^2 was used.

Figure 1: 8Keuexcess / 8Kspall, versus 8Keuexcess / 8Kspall.

The slope of the correlation line is 2.69 ± 0.28

This shows that the Keuexcess values are due to spallation neutron capture on 76.8Kr in the energy range of 0.1 - 200 MeV.

Figure 2: Correlation of the absolute Keuexcess due to spallation neutron capture on

Kr with the neutron dosage of energies > 0.04 MeV calculated from the full integral spectrum. The spread of Keuexcess indicated by the shaded sector reflects a variable spall in Apollo 14 samples.
REGOLITH DYNAMICS
D. Macdougall, B. Martinek and G. Arrhenius
University of California, La Jolla, California

This paper is an interim report of studies designed to elucidate dynamic processes occurring in the lunar regolith. Our data consist of information about cosmic ray particle tracks in silicate crystals, in particular track densities, presence or absence of track density gradients, and angular distribution of tracks within individual crystals.

There has been a great deal of discussion about regolith mixing, and a number of authors have stated that the regolith is well mixed at least to the penetration depth of the Apollo 12 double core (1,2). Track data from the two Apollo 11 cores are ambiguous in this respect (3), but the track density distributions observed in closely spaced samples of the Apollo 12 double core clearly indicate that mixing is poor over depths of 15 to 20 cm (4). Although we have not yet been able to study core samples from Apollo 14, particle track features observed in surface scoop samples indicate that the same is true at this site.

A technique not used before for lunar samples, consisting of angular distribution measurements and in particular measurements of azimuth angles, has provided information about orientation changes of small grains, rocklets and rocks. Details will be published elsewhere but one essential feature is that a sharp, singly peaked azimuthal angle distribution indicates no change in orientation during irradiation. Measurements on grains from several layers of the Apollo 12 double core show that many crystals now buried were not extensively moved about while lying within a few centimeters of the surface. Similarly measurements of track density as a function of grain size may indicate lack of mixing. Figure 1 shows a typical case. The frequency of occurrence is drawn in percent of total for ease of comparison, and all samples with densities $> 90 \times 10^6$ and/or having steep track density gradients are included in the last column to the right. In spite of poor statistics it is evident that there is no striking difference in track densities from one size class to another, except that the $< 1$ mm sample has the largest proportion of grains with high track densities and gradients, i.e. grains exposed on the surface. Similar features are observed in other samples where a variety of grain sizes are available, excepting that absolute track densities differ considerably from one sample to another; put another way the "exposure time" as measured by the integrated flux varies from place to place on the surface. These observations are in keeping with the throw out model proposed earlier (4). The large proportion of high density and gradient grains in the smallest size class indicates frequent moving of these particles compared to larger grains, as might be expected from the steep rise toward
small diameters in the size spectrum of impacting material (5). Angular distribution measurements such as those described above suggest that several of the centimeter size rocklets from the bulk sample may have spent relatively long periods of time (50 to a few hundred my) within half a meter of the surface but without any change in orientation. A simple model of layer by layer deposition by throw out from discrete cratering events with continual turning and mixing of grains in the top several hundred microns is no doubt greatly oversimplified but fits the present data well. Of course large fragments and rocks are tumbled occasionally. Five of sixteen rocks examined so far by track techniques are thought to have tumbled at least once, but since all were picked up from the surface they constitute a biased sample.

As members of a consortium studying wallnut sized rocks we have analyzed nine rocks of centimeter size from sample 14160. Only one is texturally an igneous rock (14160, 9); others are fragmental rocks or breccias. At least two orthogonal slices of each were examined for fossil tracks. The results (Table 1) show that only two exhibit evidence for surface irradiation while the others have no track density gradient over distances of 0.5 to 1 cm. As described earlier, extensive angular distribution measurements suggest that several of these rocks received their entire track compliment without any shift in orientation. If all tracks were formed in a continuous irradiation spanning the entire Kr exposure age (time spent within 1 m of the surface) then the depth of burial was 14 to 15 cm. (Table 1, column 6). This is an over simplification since they were actually scooped from the top 6-8 cm of soil.

Another aspect of our studies has been the search for irradiation records acquired before consolidation by individual components of rocks and
Table 1. Track densities and $\text{Kr}^{31} - \text{Kr}^{33}$ exposure ages (K. Marti, Pers. Comm.) for sample 14160

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Size (mm)</th>
<th>Olivine</th>
<th>Feldspar</th>
<th>$\text{Kr}^{31} - \text{Kr}^{33}$ (my)</th>
<th>depth (cm)</th>
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<tr>
<td>14160.3a</td>
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<td>14160.3b</td>
<td>12x9x4</td>
<td>$\leq 100$</td>
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<td>20</td>
<td>15</td>
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<td>14160.4</td>
<td>12x9x3</td>
<td>13</td>
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<td>28</td>
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<td>18</td>
<td>700</td>
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<tr>
<td>14160.6</td>
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<td></td>
<td></td>
<td>$\leq 100$</td>
<td>14</td>
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<td>16</td>
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</table>

* All samples show constant density throughout except 14160, 3b and 3.

rocklets in the class known as fragmental rocks or breccias. The rocks studied exhibit a wide range of textural features suggesting varying degree of lithification or recrystallization, and many resemble texturally the type of meteorite material known as polymict breccia. At least one of the rocks (10018, 29) is known to be rich in solar wind gases, but in contrast to the gas rich meteorites, none of the samples examined to date has been found to contain irradiated grains. This and other observations concerning the geometry of surface (unshielded) particle irradiation illustrates the considerable differences between the irradiation and consolidation conditions in the lunar regolith on the one hand and the corresponding environment of formation of gas rich meteorites on the other.


The Planetology Branch, Ames Research Center, received the mirror of the Surveyor III Television camera from the Jet Propulsion Laboratory to search for micrometeoroid impact sites. Previous investigators had already studied this mirror but we had the first opportunity to observe it with a Scanning Electron Microscope (SEM). We were also allowed to remove the two trunnions (aluminum pieces on each side bearing the mirror) for SEM analysis. In this paper we show the results of this examination.

The Surveyor III mirror was manufactured by Electrodepositing a nickel plating of 80 μm thick on a beryllium block. This plating was polished, and then vacuum-coated with a 0.2 μm thin film of aluminum. Finally, a 0.1 μm thin coating of silicon oxide was deposited on top of the aluminum film. The Surveyor III mirror was exposed to the lunar environment for 944 days following an abnormal landing which caused a heavy deposition of dust on its surface.

When the mirror was first removed from the Surveyor III television camera, it was totally covered with dust except for the portion wiped off by the astronaut's finger. During analysis by other investigators later, a 2 cm wide strip of dust along its main axis was taken off by acetate-film stripping. Also an area was rubbed off by a cotton tipped swab (Q-tip). In our investigation we examined three areas. The first was an untouched area covered with debris; second, the area where the dust had been stripped off; and third, the area where dust had been rubbed off with the Q-tip.

Prior to SEM examination we scanned the surface with an optical microscope, mainly along the main axis of the mirror. From this examination we found that the untouched areas of the mirror were covered with a homogeneous layer of dust. The surface was pitted and the concentration of pits increased from the bottom to the top of the mirror. Two transition areas were observed along two oblique shadow lines in the lower part of the mirror. The bottom part of the mirror contained only a very small number of pits compared to the zone above the shadow lines. This bottom portion seems to have been shielded from impacting particles and can be used as a control surface.

The optical scan showed, further, that the mirror was covered with a contaminant which may have been deposited from combustion of the vernier engines. This contaminant cannot be found in the lower part of the mirror.

*On leave from the DERTS, Department de l'Office National d'Etudes et de Recherches Aerospatiales, Toulouse, France, under a grant from the European Space Research Organization.

**On contract from Trapelo/West Division of LFE Corporation.
below its rotation axis. This material causes the interference bands and the brown coloration of the mirror and seems to contribute to the glare of the surface. This contaminant was more adherent than the dust and was only removed by Q-tip rubbing. In the optical microscope it looks like a deposit of irregular-shaped, micron-sized blisters. We could not determine its composition. However, a closer view with the SEM showed that these features were mainly circular with diameters ranging from 8 to 40 \( \mu \text{m} \) and had cracked rims raised above the surface of the mirror. They were probably contamination which got on the mirror before the Surveyor launching. These spots were covered by lunar dust and could not have been made after the recovery. The energy-dispersive spectrometer attached to the SEM showed that these spots were compounded mainly of potassium, aluminum and silicon which is the composition of the thermal paint used on the spacecraft.

The small sample chamber on the SEM confined our examination to the central portion of the Surveyor mirror. The three areas of interest (untouched region, stripped area, and Q-tip rubbed zone) were visible in this central 3 x 3 cm area. In the untouched area the search for impact sites was difficult due to the great amount of dust on the surface. This dust consists mainly of angular particulates with 1% to 2% of spherical particles. The concentration of the debris is about 3 x 10\(^7\) particles/cm\(^2\). Few particles are larger than 4 \( \mu \text{m} \) and 60% are smaller than 2 \( \mu \text{m} \). The size and concentration of these particles are in agreement with similar measurements on the Surveyor III optical filters. The composition of the particles \((\text{Mg, Al, Si, Ca, Ti, Fe})\) as given by the X-ray spectrometer is consistent with a lunar origin.

In the stripped area and the Q-tip zone of the mirror where lunar dust did not obstruct the view, impact features were found at a concentration of 10\(^6\)/cm\(^2\). The largest of the impact features are elongated gouges; the smallest are nearly circular dents. Their sizes range from 25 \( \mu \text{m} \) length and 8 \( \mu \text{m} \) width, down to 1 \( \mu \text{m} \) diameter. Nearly 75% are smaller than 4 \( \mu \text{m} \). The shape and the orientation of most pits seem to show that they were caused by rather low velocity particles (some hundred meters per second) striking the mirror with an oblique or grazing angle and coming from some source below the mirror. Imbedded in some pits we found material whose composition is consistent with lunar origin.

When the trunnions were removed from the mirror, these pieces showed a brown coloration on the part exposed to the lunar environment. The left trunnion was colored more than the right one. A portion of each trunnion was fastened tightly against the mirror by screws and was completely protected from the lunar environment. This served as an excellent control area free of any features caused by the lunar exposure. Under SEM examination, all surfaces of the aluminum trunnions were found to be rather rough, with parallel polishing grooves. Careful scanning of the trunnion areas which had been exposed to the moon revealed numerous impact pits. An average pit concentration of 3 x 10\(^4\)/cm\(^2\) was found on the left trunnion and 1 x 10\(^5\)/cm\(^2\) on the right. Most of these pits were hemispherical or elliptical gouges crossing the polishing grooves with a lip raised downrange from the direction of impact.
Their sizes ranged from 0.5 μm to 10 μm. Spherical lunar particles were found imbedded in several pits. Examination of similar areas on the shielded portions of each trunnion failed to reveal any impact pits.

**DISCUSSION**

No hypervelocity impacts were found on the 5 mm² scanned area on the mirror or on the 6 mm² scanned area on the trunnions. Because of changes in elevation of the mirror during its operation on the moon, it is difficult to compute its actual exposure to direct impacting micrometeoroids. However, with some assumptions we can calculate a flux limit of $1 \times 10^{-2} \text{m}^{-2} \text{sec}^{-1} (2\pi \text{ steradians})^{-1}$ for particles with masses larger than $1 \times 10^{-13} \text{g}$.

On the trunnions as well as on the mirror, the numerous low velocity impacts seem to be caused by lunar soil particles. Most likely, this pitting was due to lunar material being blasted toward the camera by the Surveyor's vernier engines. This is apparent because the mirror was facing the No. 3 vernier engine upon landing; the lower part of the mirror was shielded by the housing and is relatively free of pits; the direction of the pitting material was from below the mirror; and the velocity of the lunar soil particles blown from the surface by the vernier engines would be fast enough to cause the observed damage.
The dynamics of ions in the lunar atmosphere are related to phenomena such as: gas clouds released into the lunar atmosphere by lunar venting or man-made impacts, lunar ions detected by surface ion detectors, and the apparent presence of trapped lunar atmosphere in the lunar soil samples. A recent model for the dynamics of lunar ions describes their energy and flux as a result of acceleration by fields in the solar wind and at the lunar surface (Manka and Michel, 1970). The purpose of this paper is to discuss some further characteristics of this model and to present, using correlated data from lunar surface and orbital magnetometers and surface ion detectors, an analysis of several ion events which support the theory.

The model has already been applied to studies of noble gases found in lunar samples and to results of ALSEP ion detectors. At present it appears that, in addition to the solar wind gases trapped in lunar surface samples, there is a significant amount of lunar atmosphere which is being ionized and accelerated into the moon with sufficient energy to be trapped; furthermore, some isotopic fractionation of the trapped atmosphere is likely and the magnitudes of these effects have previously been estimated. The lunar atmosphere can also be detected in ionized form; the Suprathermal Ion Detector (SIDE) sees sporadic bursts of lunar ions (Freeman, et al., 1971). Different types of ion events are seen, each being characteristic flux, peak energy, and phase of the lunar orbit in which they occur. Many of the types of ion events, including low energy events (tens to hundreds of electron volts), occur preferentially around local sunrise and sunset, as is predicted by this acceleration model. Special gas clouds could be generated by spacecraft impacts or by sudden lunar venting (which could produce an event such as the one recently attributed to vented water vapor (Freeman et al., 1972). Such
clouds may be detected by an ion detector, provided that field conditions are such that ions formed in the cloud are accelerated into the detector look angle. Such an acceleration mechanism is, then, crucial to the connection between the atmosphere or gas event itself, and its detection in the lunar surface samples or by charged particle experiments.

The proposed mechanism is based on photo- and charge exchange ionization of the gas and subsequent acceleration into the moon by electric (E_{SW}) and magnetic (B_{SW}) fields associated with the solar wind. This acceleration has several unique characteristics. One expected feature is that the fluxes of heavier ions should be quite directional, approximately along E_{SW}, and predominately at right angles to both B_{SW} and the solar wind velocity (V_{SW}). This is because the ion is initially accelerated along the interplanetary electric field, given by E_{SW} = - V_{SW} x B_{SW}. Another feature is that the ions should be energetic, compared to the terrestrial ionosphere, with energies of tens of ev to a few kev. These directions and energies could be modified by lunar surface electric and magnetic fields; and the ions might even be used to probe these surface fields. We discuss the application of this model to gas clouds from sudden, localized events; the sporadic detection of ions from the event may be due to the orientation and fluctuation of E_{SW}.

The expected ion energy and flux spectra are given and estimates are made of the effects of surface electric and magnetic fields. Lunar ion trajectories are studied by comparing flux measurements from the SIDE, which looks in the ecliptic plane, with magnetic measurements from the Lunar Surface Magnetometer and the Ames Magnetometer on lunar orbiting Explorer 35 (Dyal and Parkin 1971; Colburn et al., 1971). The model predicts that near local lunar sunrise and sunset, ions from the lunar atmosphere could enter the detector when the interplanetary magnetic field is out of the ecliptic in such a way that the V x B electric field drives ions into the detector; in addition, a source of lunar nighttime particles could be ions from the earth's bow shock traveling up the interplanetary magnetic field. Thus a condition for lunar ions to enter a detector would be that E_{SW} is anti-parallel to the detector look direction, while a condition that bow shock ions enter a detector would be that B_{SW} is approximately parallel or anti-parallel to the detector look direction. Comparison of sunrise - sunset ion events, detected by the SIDE,

...
LUNAR IONS

R. H. Manka

with the corresponding magnetometer data shows that in most cases B_{sw} was strongly out of the ecliptic and had the proper orientation for E_{sw} to drive lunar ions toward the detector. Analyses of several low energy events are presented for which detailed time correlations between ion flux and B_{sw} are found.

Implications of these findings are discussed with respect to other detected ion events, the assumptions used in calculations of ion trapping, and the interaction of the solar wind with the lunar atmosphere. In particular, some properties of the ion dynamics during the possible "water vapor" event of March 7 are discussed. Using preliminary data from the Ames Explorer 35 Magnetometer and the lunar surface Solar Wind Spectrometer, the strength of the interplanetary electric field is calculated. If the assumption is made that the observed ion energy is derived entirely from E_{sw}, then the ions originate at heights of 16 to 32 km, consistent with the scale height for water vapor (~ 25 km).

We thank Drs. H. K. Hills and Conway Snyder for making available preliminary data.

References


Our general survey of coarse fine particles from the Apollo 14 soil samples is summarized in the abstract we originally submitted to the Lunar Science Conference and will be published in the Proceedings Volumes. In our talk and this amended abstract, we wish to dwell on one particularly interesting class of soil particle.

Fig. 1 shows the normative mineralogy of all particles of non-mare lunar material, from lunar soil samples, analyzed by our group to date. The distinction between mare and non-mare materials is not always easy to make; we assumed particles containing > 2% TiO$_2$ were mare-derived, and excluded them from the plot. Three classes of non-mare material can be distinguished: plagioclase-rich anorthositic particles, norites rich in K and P (normative orthoclase, apatite), and a small group of pale green glasses (arrow) consisting of about one-third normative plagioclase and two-thirds normative mafic silicate minerals.

We have found these pale green mafic glasses in < 1-mm fines samples from the Luna 16 and Apollo 14 core tubes. These glasses are relatively rare in the lunar soils, but not so rare as Fig. 1 indicates. They are sufficiently rare that there is little likelihood of finding them among the small numbers of > 1-mm soil particles allocated for study; and they have escaped notice among < 1-mm soil particles because few investigators make detailed chemical studies of individual particles in "fine fines" samples.

The lunar mafic green glasses are compositionally similar to howardites, one of the Ca-rich classes of meteorites. The comparison is drawn in columns three and four of Table I. Averaged values in these two columns sometimes do not match very closely, but both howardites and lunar green glasses are variable in composition, and for all oxides but CaO the compositional ranges overlap. A comparison between the

![Fig. 1. Normative compositions (microprobe analyses) of non-mare particles analyzed by us in lunar soil samples. Arrow: mafic green glasses.](image)
normative mineralogy of Ca-rich meteorites and howardites is drawn in Figs. 2 and 3. Mineral compositions and abundances do not match perfectly for howardites and lunar green glasses, but the difference is small.

Four interpretations of the significance of howarditic glass in the lunar soil can be considered:

(1) The glass is derived from howarditic meteorites that have bombarded the Moon. This is unlikely; ordinary chondrites, ~30X more abundant among terrestrial meteorite falls than howardites, are not found in the lunar soil, either intact or impact melted; evidently stony meteorites are, with rare exceptions, melted on impact and mingled beyond recognition with lunar materials.

(2) Howarditic meteorites are derived from the Moon (a possibility suggested by Duke and Silver, Geochim. Cosmochim. Acta 31, 1637, 1967). Also unlikely: if impacts on the Moon deliver howardites to us as meteorites, they should also deliver anorthositic rocks, KREEP-rich norite, and Ti-rich basalt.

(3) Howarditic material is developed by a complex process of igneous differentiation, brecciation, and mixing (as indicated by textures) in at least two places: in the Moon and in one or more parent meteorite planets elsewhere in the solar system. We regard it as too much of a coincidence that this chain of events would produce such similar end products on two planets in the solar system; textures notwithstanding, it seems to us that the howarditic composition has some fundamental geochemical significance that carries over from one planet to another.

(4) The special importance of this composition does not have to do with melting relationships in silicate systems; howardites do not lie near the eutectic composition in the system anorthite-forsterite-silica. We suggest that howardites are a fair approximation to the raw material from which some of the inner planets or pre-planetary masses accreted. These included our Moon; and while it seems that almost the entirety of the Moon has been partially melted and differentiated, some small proportion of its original substance must have survived more or less unfractonated, to serve later as a source for the lunar green glasses.

Fig. 2. Comparison of proportions of major minerals in the norms of Ca-rich meteorites, green glasses, and other lunar materials: 1, 15415 (the Genesis Rock); 2, gabbroic anorthosite 15418; 3, igneous norite 14310; 4, "green clod" 15923.

Fig. 3. Comparison of composition of normative pyroxenes in Ca-rich meteorite and lunar materials. Symbols and number key as in Fig. 2.
Table I. Comparison of average compositions (and compositional ranges) of Ca-rich meteorite classes and lunar mafic green glasses. Analyses recalculated to sum 100%.

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<th>Stony Phase, Meteorites</th>
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<th>19993</th>
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<tr>
<td></td>
<td>Eucrites b</td>
<td>Mesosiderites c</td>
<td>Howardites d</td>
<td>Genesis Bean e</td>
<td>MSC Glass f</td>
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<tr>
<td>SiO₂</td>
<td>48.67 (47.04-49.96)</td>
<td>55.61 (49.27-60.59)</td>
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<td>46.68 (44.64-50.74)</td>
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<td>TiO₂</td>
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<td>11.88 (10.33-11.15)</td>
<td>8.87 (5.10-11.15)</td>
<td>11.38 (9.26-17.45)</td>
<td>8.98 (5.11)</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.39 (0.06-0.83)</td>
<td>0.53 (0.04-1.15)</td>
<td>0.62 (0.04-1.15)</td>
<td>0.34 (0.07-0.70)</td>
<td>0.56 (0.47)</td>
</tr>
<tr>
<td>FeO</td>
<td>18.42 (15.02-20.41)</td>
<td>17.77 (15.29-18.43)</td>
<td>17.03 (15.29-18.43)</td>
<td>17.03 (15.29-18.43)</td>
<td>16.79 (15.29-18.43)</td>
</tr>
<tr>
<td>MgO</td>
<td>7.42 (6.47-8.67)</td>
<td>15.18 (12.22-20.61)</td>
<td>15.27 (12.22-20.61)</td>
<td>15.27 (12.22-20.61)</td>
<td>15.27 (12.22-20.61)</td>
</tr>
<tr>
<td>MnO</td>
<td>0.42 (0.21-0.78)</td>
<td>0.63 (0.40-0.62)</td>
<td>0.81 (0.22-0.72)</td>
<td>0.28 (0.03-0.34)</td>
<td>0.29 (0.03-0.34)</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.05 (0.02-0.05)</td>
<td>0.08 (0.01-0.02)</td>
<td>0.10 (0.01-0.02)</td>
<td>0.06 (0.00-0.01)</td>
<td>0.15 (0.00-0.01)</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.14 (0.09-0.27)</td>
<td>0.78 (0.26-1.47)</td>
<td>0.10 (0.26-1.47)</td>
<td>0.14 (0.09-0.27)</td>
<td>0.10 (0.09-0.27)</td>
</tr>
</tbody>
</table>

a Eucrites and howardites contain <1% mesosiderites 49 to 99% of metal and sulfide phases.
b Average of Bertha, Bouchard, Jossea, Juvinas, Lubolola, Nunsini, Moore County, Passamonte, Perambol, Petersberg, Stannem, Chervony Kut, and Sioux County; analyses referenced in Mason (1967) and Urey and Craig (1953).
c Average of stony phases of Patwar (Jarosevich and Mason, 1969), Crab Orchard, Vass Maorte, Estherville, Hainholz, Veramnb, Lowicz, Morristown, and Bondoc (Powell, 1971).
d Average of Frankfort, Binds, Pavlovka, Chaves, Gunn, Yurtuk, Le Tellier, Kapoua, and Brent; analyses referenced in Mason (1967).
e Average of Genesis Bean and demi-Genesis Bean, from 14233, 2; pale-green glasses (200-214), (200-75), and (200-145) from level G, Luna 10; and pale-green glasses (257-21) and (260-67) from 14236, 30 and 14230, 2, respectively.
f Our microprobe analyses.
g Average of ten "mare derived glasses" containing <1.0% TiO₂ from <1 mm fines sample 14259 (Brown et al., 1971).
h Only post-1950 values for meteorites are used; supplemented by Edward and Urey (1956) and Schmitt (1966).
COMPOSITION OF LUNA 16-G52, AND APOLLO 14 FINES


Luna 16-G52 was a 0.69 mg rock fragment, a small pellet approximately 0.7 mm diameter, with D = 2.88 g/cc. Scanning electron microscope photographs showed a rough surface extensively coated with small irregular particles, probably largely glass. A polished thin section showed that the fragment consisted almost entirely of devitrified glass, enclosing rare laths of anorthite. Microprobe analyses, using a defocussed beam, gave somewhat variable results, the following being fairly representative: SiO$_2$ 42.3, TiO$_2$ 0.16, Al$_2$O$_3$ 24.4, FeO 6.4, MnO 0.00, MgO 7.5, CaO 15.3, Na$_2$O 0.02, K$_2$O 0.01; this composition corresponds to a gabbroic anorthosite, with a norm of 70% anorthite, 14% olivine, 8% diopside, 7% hypersthene, and 0.3% ilmenite. The fragment probably represents a shock-melted anorthositic rock.

We have examined the following samples of fines (<1 mm) from the Apollo 14 collections: 14003,26 (0.25 g); 14003,79 (2.01 g); 14163,42 (0.26 g); and 14165,5 (0.25 g). Samples 14003 and 14163 were collected near the landing site, whereas 14165 was the fine residue in one of the collecting bags of the second EVA and probably represents material from a wider area. The samples were dry-sieved; the grain-size distribution in 14003 and 14163 was essentially identical, as follows (in weight percent): > 0.50 mm, 7; 0.25-0.50 mm, 11; 0.15-0.25 mm, 8; 0.074-0.15 mm, 15; 0.043-0.074 mm, 12; < 0.043 mm, 47. Sample 14165 contained relatively more coarser material (42% > 0.15 mm) and less extreme fines (33% < 0.043 mm).

The analyses were made of 14003,79, one of an "as received" split, and the other of the <0.043 mm fraction. The results are as follows (weight percent):
LUNA 16-052, AND APOLLO 14 FINES

Brian Mason

The compositions are essentially identical except for a slightly higher \( \text{Al}_2\text{O}_3 \) content in the <0.043 fraction; this corresponds to a slightly larger feldspathic component in the extreme fines. Normative calculation of the analysis of the <0.043 fraction gives (weight percent): \( \text{Or} 0.4, \text{Ap} 1.0, \text{Il} 3.5, \text{Or} 3.3, \text{Ab} 5.3, \text{An} 43.4, \text{Wo} 3.6, \text{En} 22.5, \text{Fs} 16.9, \text{Q} 0.2 \). Compared to our analyses of the Apollo 11 and 12 fines (10084 and 12001), this Apollo 14 material is notably higher in \( \text{Al}_2\text{O}_3, \text{K}_2\text{O}, \text{and P}_2\text{O}_5 \), somewhat higher in \( \text{SiO}_2 \), and lower in \( \text{FeO} \) and \( \text{TiO}_2 \). This corresponds to a source richer in feldspar and phosphate and poorer in ilmenite and ferromagnesian minerals. Whereas the Apollo 11 and 12 fines are olivine-normative, the Apollo 14 fines are slightly quartz-normative.

These Apollo 14 fines consist largely of glass-bonded agglutinates and breccia fragments. Mineral grains make up a minor component of the finer fractions (<0.15 mm). Calcium-rich plagioclase is the most abundant mineral. Next most abundant are pyroxenes with a wide range of composition, from magnesium-rich pigeonites to iron-rich augites. A little olivine, ranging in composition from \( \text{Fa}_{11} \) to \( \text{Fa}_{46} \), was observed, and some opaque material, largely ilmenite and a few particles of nickel-iron.

This research was supported by NASA grant NGR 09-015-146.
SPINEL AND HORNBLENDE IN APOLLO 14 FINES


LSPET (1971) reported a pink, high-relief, isotropic mineral in several of the Apollo 14 fragmental rocks, and tentatively identified it as garnet or spinel. We found a grain of this mineral, 100 microns across, in the heavy fraction separated from 14163.42 with methylene iodide (D = 3.32). Microprobe analysis gave (weight percent) Al₂O₃, 61.2; MgO, 22.1; FeO, 8.6; Cr₂O₃, 7.4; TiO₂, 0.3. This analysis (omitting the small amount of titanium) calculates to (Mg₀.₆₆Fe₁.₅) (Al₁₄.₈Cr₁.₂) O₃₂, in excellent agreement with the spinel formula. An isolated grain of this spinel, 0.4 x 0.1mm, was also observed in a thin section of breccia 14319.13, and another in Apollo 15 fines (15471,27).

Spinels of this composition are rare in the lunar material so far examined. Lunar spinels usually fall in the chromite - ulvospinel series. Keil et al. (1970) described an anorthositic fragment in Apollo 11 sample 10059 consisting largely of plagioclase (An₉₅), with minor amounts of olivine (Fo₈₁), and a spinel with comparable composition: Al₂O₃, 68.0; Cr₂O₃, 2.16; FeO, 3.33; MgO, 25.9.

Spinels of this composition are also rare or non-existent in terrestrial rocks, judging from a survey of published analyses. The color of the transparent red spinels of gem quality is evidently due to the presence of chromium, but these spinels are almost pure MgAl₂O₄ with a maximum Cr₂O₃ content of about 1%. Most terrestrial chromium spinels are very dark in color, evidently because of the presence of ferric iron, with is essentially absent from the lunar rocks. The closest analogy known to us is the pink spinel in aluminous graphitic xenoliths in the native iron-bearing Disko basalts, West Greenland (Melson and Switzer, 1966). Here, as in the lunar samples, very low oxygen fugacities effectively eliminated ferric iron.

However, the presence of chromian spinel in the lunar rocks is explicable in terms of synthetic studies. Spinel appears on the liquidus in many systems involving CaAl₂Si₂O₈-Mg₂SiO₄. Osborn and Tait (1952) have investigated the system diopside-forsterite-anorthite, and find an extensive spinel field extending from the anorthite-forsterite join towards the diopside apex. On the anorthite-forsterite join spinel appears on the liquidus at about 1450° for compositions between about 55% and 85% anorthite.
The solidus temperature is 1320°, at which point, if equilibrium is maintained, the spinel will be completely resorbed and the solid phases will be anorthite and forsterite. Under non-equilibrium crystallization spinel may survive, or it may be effectively removed from the system by gravitational settling from the melt. The association anorthite-olivine-spinel recorded by Keil et al. (1970) in Apollo sample 10059 suggests non-equilibrium crystallization of an anorthositic magma undersaturated in silica and with a high Mg/Fe ratio. Most of the lunar anorthosites are anorthite-pyroxene rocks, and MgAl_2O_4 is not compatible with pyroxene at magmatic temperatures, hence these rocks do not contain spinel (sensu strictu); however, they may contain chromite.

The same polished thin section (14163,42A) in which the spinel was discovered also contains a few grains of a dark brown, non-pleochroic, anisotropic mineral with relief and birefringence similar to the abundant pyroxene (which is colorless to pale brown in section). The refractive indices were determined by the immersion method (with some difficulty, due to paucity of material and the high absorption), and found to be α = 1.700, γ = 1.725. Microprobe analysis gave (weight percent): SiO_2, 39.9; TiO_2, 4.0; Al_2O_3, 12.3; FeO, 19.6; MnO, 0.4; MgO, 8.2; CaO, 11.6; Na_2O, 2.7; K_2O, 0.8; sum 100.5. When this analysis is recalculated in terms of 23 O it gives the formula \((Ca_{1.88}Na_{0.79}K_{0.35})(Mg_{1.84}Mn_{0.05}Fe_{2.47}Al_{0.18}Ti_{0.45})Al_{2}Si_{6}O_{23}\), in excellent agreement with the amphibole formula. It has not yet been possible to obtain X-ray diffraction data on the mineral, but both the optical properties and the chemical composition support the identification as an amphibole. Specifically it may be described as a kaersutite or titanian pargasite.

The identification of amphibole in lunar samples, especially those that have been subjected to extensive laboratory processing, naturally raises the question as to whether it might be a terrestrial contaminant. This does not seem to be the case. Although we have not found this mineral in any sections of Apollo 14 rock fragments, we have found it in another fines sample (14003,79). Terrestrial amphiboles of comparable composition are rare; a survey of kaersutite analyses shows that few if any of them have Fe/Mg ratios as high as in this lunar specimen.

As far as we know, amphibole has been recorded twice before in lunar samples. Gay et al. (1970) found crystals of amphibole in a vug in rock 10058, a Type B (coarse-grained) microgabbro; this amphibole is, however, quite distinct in composition from our material, being high in SiO_2 (54.51%) and low in TiO_2 (0.16%) and Al_2O_3 (0.75%). Dence et al. (1971) have described a gray-blue amphibole which they found as isolated grains in a sample of 12021, a coarse-grained dolerite. Its composition is rather similar to our material except for TiO_2, which is quite low (0.1%).
SPINEL AND HORNBLende IN APOLLO 14 FINES

B. Mason

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We have examined the two basalts returned by the Apollo 14 mission for this amphibole, without success. However, this was not unexpected, since practically all the rocks returned by that mission were breccias. In view of the very heterogeneous nature of these breccias, an extensive examination of them might well turn up this amphibole. However, detection may well be difficult, since the color of the amphibole is similar to much of the glass in the breccia material.

References

Two samples of lunar fines from Apollo 14 (14003,27 and 14163,70) were examined by trimethylsilylation and gas chromatographic separation of the anionic constituents. The dominant anion in each sample was SiO$_4$$^4^-$. Minor constituents were the chain anions Si$_2$O$_7$$^6^-$ and Si$_3$O$_{10}$$^8^-$. The derivative of the cyclic ion Si$_4$O$_{12}$$^8^-$ was barely detectable, in contrast with the results for Apollo 11 and 12 fines. The yield of SiO$_4$$^4^-\text{derivative}$ was somewhat less than could be accounted for by the presence of 6-8% olivine, as found in the modal analyses. However, the complexity of the composition of the fines makes quantitative estimation of constituents very difficult. Overall, the chromatographic pattern was consistent with a small contribution due to glass and/or other minerals superimposed on the pattern due to olivine alone.

A synthetic lunar glass was prepared at 1200°C, P$_2$O$_5$ = 10$^{-13}$ atm. Microscopic examination of coarsely ground fragments showed orange to pale-straw coloured glass fragments, many containing small regularly shaped (cubic, star shaped, octahedral) isotropic crystals. Silylation and chromatography revealed the anion SiO$_4$$^4^-$ along with a trace of dimeric species, Si$_2$O$_7$$^6^-$.  

Volume percentages of the main constituents as determined by point counting were:

<table>
<thead>
<tr>
<th></th>
<th>14003,27A (6300/cts)</th>
<th>14163,70A (1600 cts)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Olivine</td>
<td>8 ± 2</td>
<td>6 ± 2</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>12 ± 2</td>
<td>16 ± 2</td>
</tr>
<tr>
<td>Pyroxene</td>
<td>10 ± 2</td>
<td>13 ± 2</td>
</tr>
<tr>
<td>Glass</td>
<td>51 ± 2</td>
<td>51 ± 2</td>
</tr>
<tr>
<td>Aggregates</td>
<td>12 ± 2</td>
<td>9 ± 2</td>
</tr>
<tr>
<td>Opaques</td>
<td>2 ± 1</td>
<td>2 ± 1</td>
</tr>
<tr>
<td>Unknowns</td>
<td>5 ± 3</td>
<td>3 ± 2</td>
</tr>
<tr>
<td><strong>Totals</strong></td>
<td><strong>100%</strong></td>
<td><strong>100%</strong></td>
</tr>
<tr>
<td>Glass beads, dumbbells, etc.</td>
<td>1473/mg</td>
<td>700/mg</td>
</tr>
<tr>
<td>Opaque beads</td>
<td>88/mg</td>
<td>-</td>
</tr>
</tbody>
</table>
The dust appears more heterogeneous than the Apollo 11 and 12 samples, and contains a considerable percentage of minerals which could not be positively identified by purely optical means during counting.

Mineralogically of interest is the discovery in sample 14003,27 of 6 optically uniaxial-negative grains with low birefringence of the order of 0.010 and $n_\omega \approx 1.67$, three of which had two cleavages at 90° angle, as well as 5 optically uniaxial-positive grains, with lower refractive index, $n_\omega \approx 1.64$, also with birefringence of about 0.010. The small size of these grains prevented more exact optical characterization, but the suggestion is made that these may be melilitelite-group minerals, the possible presence of which was indicated in our previous studies.

Apollo 14 samples, 14310 (igneous rock), 14321 (breccia) and 14163 (fines), were analysed for rare earth elements. To ensure the high accuracies in the determination by stable isotope dilution, the standard RE solutions were calibrated gravimetrically and titrimetrically. The differences between gravimetric and titrimetrical results were less than 0.6%, except Eu (1.3%).

Besides the lunar samples, ordinary chondrites were also investigated for RE abundances. As a result, it was found that most of the chondrites have more or less individualities in fine aspects of relative abundances of RE elements. Of the chondrites studied by us, the Leedey chondrite has been judged to be the safest choice as regards the relative abundances of RE elements. Therefore this chondrite has been used for normalization in Masuda-Coryell plotting.

As seen in Fig. 1, the chondrite-normalized RE patterns for three samples studied are very similar to each other, although there are differences in absolute concentrations. Apart from the cerium and europium anomalies, the pattern appears to be composed of two rectilinear segments, one from La through Sm and another from Gd through Lu; there is a small gap between Sm and Gd. The cerium anomalies for samples 14163, 14321 and 14310 are 10.9, 12.2 and 13.0%, respectively; the mean value is 12.0%. In spite of some uncertainties of these numerical values, the difference in cerium anomaly between 14163 and 14310 might be regarded significant, but a further study will be requested to conclude the variability of cerium anomaly.

The rectilinear patterns as seen in Fig. 1 indicate that these materials are primary liquid-type materials which were left as residual melts after the fractional solidification. This is not only in line with the fact that the Fra Mauro materials are the highland materials, but also suggests that the Moon encountered the large-scale total melting. (A few low-rubidium samples from Tranquility Base are primary solid-type materials.)

It would be worth mentioning that the slope for the span from Gd through Lu is similar, though not identical, to that for the sample 12035 which we infer to be the closest to the
"original" magma of the moon. This in accord with our consequence that the bulk partition coefficients of RE for lunar fractionation do not change greatly for the span from Gd through Lu.

As seen in Fig. 1, the points for La, Nd and Sm fall on or very close to a rectilinear line, whereas in terrestrial igneous rocks, lanthanum deviates often from a smooth curve. This would suggest that the water content in lunar magmas was much lower than in terrestrial magmas.

Independently of the work mentioned above, A. Masuda, S. Yanagita, H. Mabuchi, K. Notsu and M. Ojima are working with the measurement of lanthanum isotopic composition. The results obtained thus far appear to indicate the slightly higher La-138/139 ratio for lunar samples, but it is premature to draw any conclusion about this matter.
RARE-EARTH PATTERNS OF THE APOLLO 14 SAMPLES

A. Masuda

Sample/Chondrite

La Ce Nd Sm Eu Gd Dy Er Yb Lu

△ 14321,165
● 14163,161
○ 14310,126
Introduction.
The possible use of lanthanum isotopic composition for detecting the effect of proton irradiation on planetary matter during the whole history of the solar system was proposed by Mabuchi and Masuda\textsuperscript{1}). The isotope abundance of \(^{13}\text{La}\) is very low (0.08\%) and may be easily affected by the nuclear reaction \(^{138}\text{Ba}(p, n)\), which has the cross section peak at 20 MeV.

We have studied and are studying the abundance ratios of \(\text{La-138}\) to \(\text{La-139}\) in the Apollo 14 samples for two purposes; first to search for the effect of primordial irradiation and secondly to see if there is any isotope anomaly at the surface due to irradiation by solar flare particles.
**Experiment**

After sample was dissolved, lanthanum was separated from other rare earth elements by ion exchange. The isotope ratio measurement was performed by using the LaO\(^{+}\) ion produced by Ta single filament surface ionization.

**Results**

Table 1 gives the \(^{138}\text{La}/^{139}\text{La}\) ratios of the samples studied. These data are not corrected for the errors arising from mass discrimination effect at ion source and multiplier. The interfering ions for the measurement of \(^{133}\text{La}^{16}\text{O}^{+}\) are \(^{138}\text{Ce}^{16}\text{O}^{+}\), \(^{154}\text{Sm}^{+}\) and \(^{138}\text{Ba}^{16}\text{O}^{+}\). The corrections for \(^{138}\text{La}^{16}\text{O}^{+}\) are shown in Table 1 compared with \(^{138}\text{La}^{16}\text{O}^{+}\). The correction for \(^{138}\text{Ba}^{16}\text{O}^{+}\) was negligibly small (less than 0.05% of \(^{138}\text{La}^{16}\text{O}^{+}\)). The statistical errors shown in Table 1 are standard deviations (±2σ).

Thus far, it appears that the ratio for the lunar soil LL163 is slightly higher than those for the terrestrial basaltic sample JB1 and the lanthanum reagent which were also studied by us for lanthanum isotopic composition. However, it is too premature for us to give any conclusion about the reality of the difference. A further cautious examination and multiplicate repetition of measurement are going on now.
Table 1. Lanthanum isotope abundance ratios

<table>
<thead>
<tr>
<th>Sample</th>
<th>$^{138}$La/$^{139}$La($\times 10^{-3}$)</th>
<th>$^{138}$Ce$^{16+}$</th>
<th>$^{154}$Sm$^{+}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>14163</td>
<td>0.8931 ± 0.0084</td>
<td>0.19%*</td>
<td>1.3%*</td>
</tr>
<tr>
<td>JB1**</td>
<td>0.8885 ± 0.0054</td>
<td>0.20%</td>
<td>2.8%</td>
</tr>
<tr>
<td>La reagent</td>
<td>0.8873 ± 0.0024</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Inghram et al. 2) 0.8888
White et al. 3) 0.8933

*Relative to $^{138}$La$^{16+}$.
**Japanese basaltic rock, prepared as a reference sample.

References

Photointerpretation of the superb images acquired by the panoramic camera on Apollo 15, combined with preliminary measurements from the metric camera and laser altimeter, results from S-band tracking and X-ray spectrometry, as well as first looks at analysis of returned samples from Apollos 14 and 15, allow more rigorous testing of many hypotheses on the origin and development of the Moon.

The high-resolution photographs and returned samples seem to confirm the hypothesis that the Imbrium basin formed as the result of the hypervelocity impact of a cosmic body. The Fra Mauro Formation at the Apollo 14 site may be the ejecta blanket deposited by the base surge generated by the impact. The fragmental debris apparently retained enough heat to weld on coming to rest. The fact that the matrix recrystallized seems to confirm the hypothesis of an impact-generated nuée ardente (hot debris cloud). Debris from the upper levels (0-10 km) of the crust was carried farthest out, and is present, for example, at the Apollo 14 site; closer to the original crater rim, as at the Apollo 15 site, materials from deeper crustal levels (20-30 km) were deposited.

This hypothesis, derived from study of terrestrial natural and experimental craters, may be confirmed by the observation that at the Fra Mauro (Apollo 14) site many of the samples consisted of breccias within breccias such as might be expected if previously brecciated near-surface materials had been re-brecciated during the Imbrium event. The breccias at Apennine/Hadley are simple and derived from greater depth by one impact—that which formed the Imbrium basin.

The linear ridges arranged radially to the Imbrium basin at the Fra Mauro site are part of the ejecta blanket whose surface form ranges from equidimensional hummocks closest to the original crater rim, through braided ridges, and finally transverse ridges at the outer margin of the ejecta sheet. The varied forms reflect flow velocities and turbulence in the mobile base surge. According to this hypothesis the highest velocity and turbulence produces the equidimensional hummocks, and the lowest produces the transverse dunes at the outer edge of the deposits. This mobilized fluidized deposit, i.e., the material carried in the base surge, is the upper portion of the overturned flap generated by the shock wave. The observed depositional patterns at Imbrium and Mare Orientale are similar to those formed at the base of turbidites in terrestrial marine landslides.
AN APOLLO 14-15 VIEW OF LUNAR GEOLOGY

H. Masursky

The laser tracking data confirm the tracking data from Apollos 8-14 and the Ranger, Orbiter, and Surveyor data and indicate that the earth-based hemisphere is low with respect to the nominal lunar radius of 1738 km and that the far side is high. The laser data indicate that the front side is 2 to 5 km low and that the average level is 2 km low on the west limb in Oceanus Procellarum, descending to 5 km low at Mare Smythii on the east limb. The far side is 2 to 5 km high with a 9.5 km difference between the floor of Smythii and the rim of the crater Gagarin on the far side. There appears to be a central depression along the 180° longitude line as well as a central high in the region of Descartes on the front side. Preliminary analysis by Sjogren and Wallenfautl indicates that the center of figure is displaced by more than 1.5 km from the center of mass.

The distribution of the low density crustal material indicates that it was affected by a gravity field similar to the present one. This distribution indicates that the differentiation of the lunar crust took place after the Earth and Moon were in positions at least somewhat like the present ones.

The large positive gravity anomaly (mascon) in the Imbrium basin is confined to the center of the original crater and results from the central uplift analogous to the central peaks of smaller craters. Mantle material, rich in olivine and pyroxene, is lifted up into the differentiated felsic crust. The whole inner basin is covered by later mare basalt derived by partial melting of the mantle induced by impact unloading of the crust. The Apennine front appears to be a tilted crustal block covered by layered Fra Mauro Formation formed by minor surges within the major base surge like the "white sands" in the deposits around Meteor Crater, Arizona. Rings of positive and negative gravity anomalies surround the inner basin except where they are masked by the adjacent Mare Serenitatis structure. These rings reflect warps and dislocations in the crust-mantle interface developed by the outward spread of the shock front following the impact. The location of the anomalies is determined by the shock mechanics that produces lateral flow in the crust and underlying mantle and is only slightly modified by the emplacement of the post-impact basalts.

Variation in regional elevation of terrain and the morphology of the geologic units correlate closely with regional variations in the silicon/aluminum ratio as determined by the orbital X-ray fluorescence experiment of Adler and Trombka.

Hadley Rille, like other sinuous rilles photographed, has an extremely irregular floor apparently caused by intermittent eruptions along a fault controlled course. The thinly layered basalt flows in the rille walls photographed by the astronauts on the surface indicate that the rille formed largely in response to the channelized flow of mare material. However the shape of the channel has been appreciably modified by the transportation of fragmental debris by gaseous eruptions.
The major trends of many rilles appear to be determined by the local tectonic pattern. In many places they cross the regional slopes diagonally. The inner meandering valley on the floor of some rilles indicates that there may have been extensive modification of the original tectonic control by the flow mechanics of the extruded volcanics. The irregular longitudinal profile of the floors of many rilles indicates extrusion at many sites along the channel in the fashion suggested by Schumm.

Near-terminator panoramic and metric photographs indicate that the lobate basaltic lava flows imaged by earth-based telescopic photographs and shown in Lunar Orbiter V photographs in the center of the Imbrium basin are ubiquitous in the western part of the basin and the north end of Oceanus Procellarum. In places there are as many as six superimposed highly digitate flows whose patterns suggest eruption from fissures along mare ridges.
MICRO-SCALE EROSION ON THE LUNAR SURFACE BY HYPERVELOCITY IMPACT, SOLAR WIND SPUTTERING AND THERMAL CYCLING


The programme simulates experimentally three of the most dominant aspects of the lunar environment affecting surface morphology on the sub-centimetre scale, to establish criteria and parameters which will lead to a better understanding of the lunar surface and its environment. Hypervelocity impact parameters and crater profiles are measured by the impact of electrostatically accelerated micron spheres. Thermal cycling of the lunar sample zone containing these impact craters and the associated shock damaged areas is then achieved by temperature cycling under radiative heat transfer. Stereoscan electron microscopy of selected impact zones permits accurate evaluation of any induced cycling damage.

Preliminary results from hypervelocity impacts have established features of penetration on sample 14310.158 (a medium grained basaltic crystalline rock) over the velocity range 2 km sec$^{-1}$ to 10 km sec$^{-1}$, and on a comparison surface of quartz. Spallation zones, with radial fractures and the formation of platelets, are extensive at low velocities, but a progressive decrease of this damage relative to the primary crater is noted with increasing velocity. At high velocities plastic flow conditions predominate in the formation of the crater, which is of hemispherical profile and has raised lips. An extended damage area beyond this is still observed at the highest velocities observed to date, but in the form of a shallow conical area slightly raised from the surface. Often a single radial fracture is observed in this cone. The transition from shock damage to plastic flow occurs at a lower velocity for the lunar sample than for quartz, and the retention of platelets in the spalled areas is noticeably strong compared to quartz, which readily sheds conchoidal fragments at low velocities. Observed hypervelocity features show general similarity to the class of lunar micro-craters which exhibit hemispherical form and to simulated micro-impacts on terrestrial rocks (Bloch et al. 1971), but there are
distinct differences from centimetre-scale impacts, where a.
reduction of the effective material strength leads to extended
spall damage (Hörz, 1969)...

Thermal cycling of impact zones on sample 14310 has
failed to detect any surface changes at or near 90 impact
areas after 10 cycles of 210°K excursion. The thermal cycles
had a rise time of approximately 15 minutes and a periodicity
of 24 hours. Since on a microscale the temperature differences
across impact zones during hear flow are minute, it is likely
that the temperature extremes are the major features of thermal
cycling on the lunar surface, and it is likely that this
preliminary test is fairly representative of the damage which
could be induced on the lunar surface. The failure to detect
induced displacement or fracture of even the spalled zones in
which platelets appear quite loosely bonded, indicates the
strength of this type of sample on a micro-scale and suggests
that thermal cycling is not a major force. Nevertheless the
clearly crystalline and fractured nature of the surface of this
sample (identified as top surface) defies explanation by hyper-
velocity impact alone. We are left without explanation of the
generation of this surface condition unless (i) sample
orientation deductions are false or, less likely, (ii) the
sample has been recently emplaced on the surface or (iii) the
sample has suffered recent spall damage from secondary impact.
It appears from the results of our experimental observations
to date that spall change and thermal cycling are not respon-
sible for the development of this crystalline surface. Other
rock surfaces have been accounted for very reasonably by
primary hypervelocity impact (R. Bloch, 1971(2)). For
centimetre scale rocks the surface impact erosion is dominated
by particles in the radar and faint visual range, and is
calculated to be 10⁻⁷ cm yr⁻¹ (McDonnell & Ashworth, 1971). It
must be remembered that as a consequence of the dominance of
large (and thereby infrequent) impacts that local erosion rates
between impacts may be very much lower, and the observed
surface erosion deduced by solar flare tracks of 10⁻⁸ cm yr⁻¹
(Fleischer et al. 1971 etc) may be quite compatible with this
higher impact erosion rate.

Sputtering by solar wind simulation using hydrogen and
helium at 400 km. second velocity will be initiated shortly,
but a review of satellite data has indicated a much lower
Helium flux than used in early sputter estimates. It is
suggested that the average sputter rate on the lunar surface
may be as low as 0.02 cm yr⁻¹ (McDonnell & Ashworth, 1971)
compared to previously accepted figures in the region of 10⁻１ cm yr⁻¹
(Wehner, 1963). The implications of this have significance for
pico-gram particle flux rates of interplanetary dust deduced
Micro-scale erosion on the Lunar Surface

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from sub-micron surface crater counts (Neukum et al 1972) where equilibrium can be established between particle crater formation and erasure by solar wind sputtering.

References


VAPOR PHASE CRYSTALLIZATION IN APOLLO 14 BRECCIAS AND SIZE ANALYSIS OF APOLLO 14 SOILS. D.S. McKay, U.S. Clanton, G.H. Heiken, D.A. Morrison, R.M. Taylor, NASA Manned Spacecraft Center, Houston, TX 77058; and G. Ladle, Lockheed Electronics, Houston, TX 77058.

Vapor Phase Crystallization. Highly recrystallized or annealed breccias make up 48% of Apollo 14 coarse fines (1-4 mm) which we examined. The presence of vugs is an important feature of these fragments and of some of the larger breccias. Sixty percent of the 1-4 mm well recrystallized breccia fragments in our allocation have vugs and eighty percent of these fragments at Cone Crater have vugs. Some of the vugs are self contained and isolated but others are partially interconnected.

Many of the vugs contain well developed crystals growing from the walls and bridging across the open spaces. We have identified by scanning electron microscope and energy dispersive x-ray analysis the following euhedral crystals: plagioclase, low Ca pyroxene (Figure 1), ilmenite, apatite (Figure 2), whitlockite, iron (Figure 3), nickel-iron and troilite. Some of the nickel-iron crystals have a partial crust of iron sulfide. The nickel content of the nickel-iron crystals is variable and ranges up to about 12%. Approximately 2% phosphorus is present in some of the nickel-iron crystals. Some of the apatite crystals (Figure 2) contain about 2% chlorine.

The growth of crystals in interior vugs, the development of detailed and complete crystal faces, the abundance of growth steps, and the delicate network of crystals separated by open spaces are strongly suggestive of growth from a vapor phase.

The growth of crystals in the vuggy recrystallized breccias indicates that a process exists in cooling lunar ejecta blankets for the transfer and redistribution of materials in a vapor phase.

An analogy may be made with the vapor phase crystallization of cooling terrestrial ash flows. Vapor phase crystallization in the cooling Fra Mauro base surge deposit may have produced many of the vug crystals found in the Apollo 14 breccias.

Terrestrial ash flows typically display vapor phase alkali exchange, loss of volatile gases, and the development of fumaroles and sublimes. A similar vapor phase transport of material (perhaps including parentless radiogenic components) during the cooling of hot lunar ejecta blankets may have been a particularly important lunar process when the most extensive ejecta blankets were formed early in lunar history.

Size Analysis. Eleven soil samples from Apollo 14 have been analyzed by sieving and optical methods for sizes down to about 1 micron. The Folk and Ward size parameters from this data are presented in Table 1 which incorporates
Curator data on all material coarser than 1 mm with our own data on material finer than 1 mm. The contingency sample 14003, is notably coarser than other surface soil samples taken in the smooth flat area near the LM; it may have been disturbed by the LM exhaust and the finer fractions partially blown away. The graphic mean size was also calculated using only material finer than 1 mm. This reduces the extreme differences shown in Table 1 and slightly changes the relative order of sample coarseness. The graphic mean sizes for material less than 1 mm are 14141: 125 micrometers; 14003: 100 micrometers, 14230,130: 107, micrometers; 14149: 94 micrometers; 14230,113: 87 micrometers; 14260: 86 micrometers; 14230,121: 72 micrometers, 14156: 66 micrometers; 14259: 65 micrometers 14148: 62 micrometers; and 14163: 57 micrometers. The average of the Apollo 14 soils and cores is coarser than either Apollo 11 or 12 but this results from the extremely coarse Cone Crater soil and trench bottom soil. If these two are removed, Apollo 14 becomes slightly finer than Apollo 11 or 12. It is not as well sorted, however.

TABLE 1

<table>
<thead>
<tr>
<th>Sample</th>
<th>Graphic Mean (Mz)</th>
<th>Inclusive Graphic Mean</th>
<th>Inclusive Graphic Standard Deviation</th>
<th>Graphic Skewness</th>
<th>Graphic Kurtosis</th>
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<td>2.85φ(139μ)</td>
<td>2.33φ</td>
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<td>14141,30</td>
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</table>
VAPOR PHASE CRYSTALLIZATION IN APOLLO 14...

D.S. McKay

Figure 1, left. Long prismatic low calcium pyroxene crystals growing outward from a vug wall. Smaller equant crystals are mainly plagioclase. An iron crystal is partially visible behind the tip of the largest pyroxene crystal. Note the growth steps on some of the crystals. These crystals have apparently grown from a vapor phase in the open pore spaces in the recrystallized breccias.

Figure 2, right. Doubly terminated euhedral apatite crystals suspended from the wall of a vug. Most of the apatite crystals contain about two percent chlorine.

Figure 3, left. An iron crystal growing from the wall of a vug. Most of the equant crystals on the vug wall are plagioclase.
CLASSIFICATION AND SOURCE OF LUNAR SOILS; CLASTIC ROCKS; AND INDIVIDUAL MINERAL, ROCK, AND GLASS FRAGMENTS FROM APOLLO 12 AND 14 SAMPLES AS DETERMINED BY THE CONCENTRATION GRADIENTS OF THE HELIUM, NEON, AND ARGON ISOTOPES. G. H. Megrue and F. Steinbrunn
Smithsonian Institution, Astrophysical Observatory, Cambridge, Mass. 02138

The spatial distribution of the helium, neon, and argon isotopes, which have been produced on the lunar surface by solar-wind irradiation, cosmic-ray spallation reactions, and in situ decay of the radioactive isotopes of uranium, thorium, and potassium, has been determined by laser probe mass spectrometry for the purpose of (1) classifying lunar soils, fragments, and clastic rocks (2) determining ultimately the source and temporal frequency of impact events on the lunar surface and (3) studying the temporal variation of solar-wind irradiation.

The concentration gradient of the solar helium, neon, and argon isotopes within an individual glass cylinder has been measured by selective volatilization in intervals of 50 microns depth from the outer surface to the center of the grain and then to the opposite outer surface. The highest solar $^{4}\text{He}/^{20}\text{Ne}=155$ was found on the exterior of the grain. Decreasing $^{4}\text{He}/^{20}\text{Ne}<100$ were measured towards the center of the grain. The results indicate that the solar gases are found within some glasses at a depth greater than their expected ranges. Consequently, these glasses have formed from soil material that was previously irradiated by the solar wind and incompletely outgassed during impact melting. Subsequent to formation of the glass cylinder, a second solar irradiation produced the surface enrichment of the solar gases.

The concentration gradients of the radiogenic gases of $^{4}\text{He}$ and $^{40}\text{Ar}$ were measured in a KREEP grain (figure 1). The results show a constant $(^{4}\text{He})_I/(^{40}\text{Ar})_I$ as a function of depth within the grain and indicate that thermal diffusion of radioactive gases from this grain has been negligible. A radiogenic $^{4}\text{He}/^{3}\text{He}$ cosmogenic=1800 is observed to be constant throughout the interior of the grain, as the $^{36}\text{Ar}/^{38}\text{Ar}$ varies from a solar value of 5.2 at the surface to a cosmogenic value of .65 in the interior.

A summary of the solar $^{4}\text{He}/^{36}\text{Ar}$ and $^{20}\text{Ne}/^{36}\text{Ar}$ (figure 2) from Apollo 12 soils and breccia, and Apollo 14 soil, breccia, and glass fragments show unique variations for each sample with a factor of 10 variation in $^{4}\text{He}/^{36}\text{Ar}$ and a variation in $^{20}\text{Ne}/^{36}\text{Ar}=5$. The lowest $^{4}\text{He}/^{36}\text{Ar}$ and $^{20}\text{Ne}/^{36}\text{Ar}$ from glass and a welded soil fragment are attributed to heating and diffusion of the solar wind gases. Breccia 14047 has consistently lower $^{4}\text{He}/^{36}\text{Ar}$ and $^{20}\text{Ne}/^{36}\text{Ar}$ than the corresponding ratios for soil 14163 and breccia 14301. Again this could be understood as resulting from diffusion
Table 1. Distribution of Radiogenic $^{40}$Ar and Exposure Ages of Fragments Selected from Apollo 14 Samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>$^{40}$Ar/Ar</th>
<th>$^{40}$Ar, AD</th>
<th>E-4, Ar</th>
<th>$^{36}$Ar</th>
<th>$^{36}$Ar</th>
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<td>Rocks</td>
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<tr>
<td>(a)</td>
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<td>(c)</td>
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<tr>
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<td>53</td>
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<td>14161.31</td>
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<td>97</td>
<td>570</td>
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</table>

Figure 1

Figure 2
of solar gases from different grain-size materials during the impact brecciation process. A soil agglomerate from 14161 has distinctly different solar 4He/36Ar and 20Ne/36Ar than the corresponding ratios from the soil at the Apollo 14 site; however, the similarity of these ratios with the ratios found for soil 12070 indicate that soil of the Apollo 12 type has been transported to the Apollo 14 site.

Cosmogenic exposure ages which have been calculated either from the cosmogenic 3He or 21Ne abundances are summarized in Table 1. The exposure ages for lithic fragments within breccias indicate that breccia formation took place relatively recently (14047<33 m.y.; 14301<20 m.y.; and 14321<20 m.y.) and probably is associated with the formation of cone crater. Exposure ages>100 m.y. are found in 2 KREEP grains and one ropy glass.

The radiogenic 40Ar/36Ar solar and absolute abundances of 40Ar for individual KREEP, glass, and plagioclase fragments (Table 1) are extremely variable. K-Ar ages are calculated for average potassium abundances as determined by Wood et al. The highest radiogenic 40Ar abundance is found in light fine-grained material from breccia 14301. This material must have a potassium concentration>4.7 per cent for a K-Ar age<4 x 10⁴ years; furthermore, this breccia, which because of its young exposure age is most likely from Cone crater and the Fra Mauro formation, is either the source of "parentless" radiogenic 40Ar or indicates that a high potassium material within the Fra Mauro formation underlies the basaltic mare and therefore is the source of the salted 'magic' component into the lunar soils.

References:


The basaltic rocks from Apollo 14 fall into at least two major groups which are believed to reflect extensive differentiation of pre-Imbrium crustal rocks. These groups are a plagioclase-rich one (14310-type) and pyroxene-rich one (14053-type). Modal analyses on the two 'type specimens' (sections 14053,5, and 14310,6) give, respectively, in volume percent pyroxene = 51.6, 32.0; plagioclase=41.7, 66.3; opaques, mainly ilmenite = 2.9, 0.5; cristobalite 1.7, n.n.; mesostasis = 1.4, 0.5; metal 0.3, 0.1; pore space 0.4, 0.6. Representatives of these two groups also occur in some coarse-fine samples (e.g. 14258), as clasts in breccias, and among the walnut-size samples. The Fe/Fe + Mg ratio in liquidus pyroxenes differ in the two magma series, and reflect fundamental magma compositional differences. The Fe/Fe + Mg ratio in mafic phases and the An content of first-crystallizing plagioclase are as follows:

<table>
<thead>
<tr>
<th></th>
<th>Plagioclase-rich</th>
<th>Mafic-rich</th>
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<tr>
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<td>14258,24D</td>
<td>14310</td>
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<tr>
<td>Pyroxene</td>
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<td></td>
</tr>
<tr>
<td>Low calcium</td>
<td>15</td>
<td>18</td>
</tr>
<tr>
<td>High calcium</td>
<td>31</td>
<td>36</td>
</tr>
<tr>
<td>Olivine</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Plagioclase (An)</td>
<td>95</td>
<td>95</td>
</tr>
</tbody>
</table>

Both these groups are reported to have similar argon-argon ages. This may mean that these ages have been 'reset' by melting in the Imbrium impact event, or that extensive generation and differentiation of lunar crust occurred in a very short time just prior to Imbrium impact. Texturally, anorthite and orthopyroxene appear to be the low-pressure liquidus phases in the plagioclase-rich series, whereas olivine and pigeonite are the liquidus phases in the mafic group. The plagioclase-rich group cannot be a derivative of low pressure crystal fractionation of the mafic series because the former has considerably lower Fe/Fe + Mg liquidus pyroxene compositions than the latter, whereas fractional crystallization would produce the reverse relationship. The plagioclase-rich group appears to be a member of the "KREEP" materials from Apollo 12 and in clasts and fines in Apollo 14 materials.
Mafic 'basalt' 14053 has similarities to some Apollo 11 and 12 basalts, such as high mafics/plagioclase ratio, low alkalis, and contents of 'incompatible' elements. Magnesian pigeonite (most Mg-rich composition found is En64Wo8Fs28) is the most abundant early crystallizing phase. It forms large prisms, reaching up 2 mm long, which poikilitically enclose anorthite (most calcic composition is An96) and olivine (most magnesian is Fo66). The large pigeonite crystals are bordered by rims of zoned augite and ferroaugite. A discontinuity in zoning (miscibility gap?) occurs between subcalcic augite (Fs42Wo16En4) and augite (Fs31Wo33Fs31) which is similar to the gap shown by some Apollo 12 pyroxenes. During the final stages of crystallization, small amounts of pyroxferroite and ferrohedenbergite crystallized, mainly as discrete small highly colored granules in the matrix. Electron probe analysis of early olivine, magnesian pigeonite, ferroaugite, ferrohedenbergite, pyroxferroite, and fayalite give, respectively, in weight percent, SiO2 = 36.0, 50.3, 46.5, 44.7, 43.4, 29.8; Al2O3 = 0.20, 1.86, 1.21, 1.13, 1.07, 0.23; TiO2 = 0.10, 0.90, 1.04, 0.90, 0.72, 0.17; FeO = 32.0, 19.6, 31.2, 36.1, 49.5, 66.0; MnO = 0.19, 0.23, 0.32, 0.35, 0.41, 1.27; MgO = 31.2, 20.7, 5.2, 0.75, 0.18, 2.5; CaO = 0.4, 6.5, 14.3, 14.1, 2.4, 0.05; and Cr2O3 = 0.35, 0.58, 0.19, 0.01, 0.01, 0.13. Grains of fayalite are common in the interstitial areas and are associated with K-rich glass, metal, and other phases. Other late stage minerals include tranquillityite, fluorapatite, and baddelyite (?). Ilmenite appears to have crystallized mainly late. Fe metal occurs in small amounts, some crystallizing evidently early (high Ni, to Ni = 3.8, Co = 0.8) from a metallic immiscible liquid. Ti- and Cr-bearing metallic iron (Ti = 1.4, Cr = 0.5) are associated with reduction of late Fe-Ti-Cr-oxide grains. Nearly pure iron forms spongy intergrowths with fayalite and other late phases. These late metallic phases were mostly related to late stage reduction reactions. The Ti and Cr contents of late metal may be real, although smearing of adjacent grains or secondary x-ray fluorescence may be involved.

At least two immiscible glasses formed during the final stages of solidification. Like those reported from some other Apollo basaltic rocks, one is an FeO-rich low silica, low alkali glass, the other a high K2O, high silica glass. Electron probe analysis of these give, respectively, SiO2 = 78.3, 48.8; Al2O3 = 11.9, 1.78; TiO2 = 0.25, 1.45; FeO = 0.48, 29.3; MgO = 0.05, 1.72; CaO = 2.56, 15.1; Na2O = 0.53, 0.05; K2O = 5.4, 0.05; BaO = 1.27, 0.08; F = 0.08, 0.20; P2O5 = 0.05, = 0.09; Cr2O3 = 0.05, = 0.09; Cr2O3 = 0.05, = 0.10.
DEUTERIUM ANALYSIS OF HYDROGEN EXTRACTED FROM LUNAR MATERIAL.

L. MERLIVAT - G. NIEF and E. ROTH - Centre d'Etudes Nucléaires de Saclay - D.R.A. - Bte Ple n° 2 - 91-GIF-sur-YVETTE

I - EQUIPMENT AND PROCEDURE.

1 - About 200 mg aliquots of a sample are transferred from the shipment container to a gold foil wrapping, in a glove box, under dry nitrogen. They are inserted in a silica tube, connected to a ground joint and a stopcock, then attached to a vacuum line where they can be heated to 1100°C by a D.C. current furnace.

A liquid air trap removes condensable vapors from desorbed gases. A vanadium furnace absorbs non rare gases. Pressure is followed by a Pirani gauge.

Non absorbed gases are pumped out. Hydrogen desorbed from the vanadium is transferred to a charcoal trap at -195°C, then taken to the mass spectrometer line.

2 - On the mass spectrometer the sample, attached to the double inlet system, is compared with a protonium standard (D/H ratio=1.0 x 10^-6) at the same pressure. Precision (2σ) of ratio measurements is ± 0.1 x 10^-6 for 20 mm³ samples, ± 0.5 x 10^-6 for 2 mm³, and ± 2 x 10^-6 for 0.5 mm³ samples.

II - PRELIMINARY CHECKS.

Terrestrial basalt in contact with protonium gas is used to investigate operating conditions (T, P, treatment time) and contamination.

a) Under 600°C recovery yield of hydrogen is higher than 95%, and protonium in contact with basalt and the line under extraction conditions, is not contaminated by deuterium by more than 0.4 to 0.6 parts per million.

b) Above 600°C the recovery yield drops. At 800°C protonium in contact with the basalt is completely absorbed in twenty minutes.

III - RESULTS.

The gases of a first 235.1 mg aliquot from the sample 14.003,78 (fines) have been desorbed after pumping 1/2 hour at 275°C. Volumes of fractions collected in each temperature interval, and corresponding D/H ratios are given in Table I part 1. The results can be compared to those of [1] and [2]. To investigate further whether the deuterium found can originate from water adsorbed on the fines, by exchange with desorbing hydrogen, two experiments have been carried out.

a) Ordinary water vapor saturated at room temperature was kept in contact 72 hours, with the sample previously outgassed at 800°C.
Protonium is then introduced on the extraction line, the sample heated and the gas sampled under the same conditions as in Table I part 1. The deuterium ratio of the fractions collected is less than $3 \times 10^{-6}$. Though, as in the case of basalt, $2/3$ of the last fraction of protonium is absorbed, when heating it up to $800^\circ C$, the remaining third is not more contaminated.

b) A fresh 255.6 mg aliquot of the sample is exposed for 24 hours, at room temperature to water vapor of 700 ppm deuterium content. The sample is then outgassed and analysed as above. Results are given in Table I part II.

**DISCUSSION AND CONCLUSION.**

The first small (1.1 mm$^3$) fraction is not representative because sampled during water elimination. A constant excess of deuterium (5 ppm) is then observed compared to the first experiment. This increase is much smaller than expected contamination from 700 ppm water vapor adsorbed on the fines and exchanging with inbedded hydrogen during desorption.

Further experiments to trace the origin of the deuterium found are underway on samples 14.003 - 14.259 - 10.019. Up to now our experiments show that:

a) No hydrogen can be observed to desorb above $800^\circ C$ as then it reacts rapidly with the fines.
b) The last collected fraction is richer in deuterium than all others. Our line is not the cause of this enrichment unobserved in preliminary tests.

---

**TABLE I**

<table>
<thead>
<tr>
<th>Heating temperature (± 10°C)</th>
<th>H$_2$ collected in mm$^3$ S.T.P.</th>
<th>$D_x \times 10^6$ ratio</th>
<th>Heating temperature (± 10°C)</th>
<th>H$_2$ collected in mm$^3$ S.T.P.</th>
<th>$D_x \times 10^6$ ratio</th>
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<td>480-535</td>
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<td>535-635</td>
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<td>730-825</td>
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<td>62.5</td>
<td>635-825</td>
<td>3.8</td>
<td>58.0</td>
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<tr>
<td>Total</td>
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<td>average 25.3</td>
<td>Total</td>
<td>93.8</td>
<td>average 30.6</td>
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<td></td>
<td></td>
<td>16.2 µ moles/g</td>
<td></td>
<td>first fraction</td>
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</table>
DEUTERIUM ANALYSIS OF HYDROGEN EXTRACTED FROM LUNAR MATERIAL

L. MERLIVAT

Fig. 1: Sketch of apparatus for extraction of hydrogen
A - Quartz tube containing the sample.
B-D - Liquid nitrogen traps.
C - Pirani gauge.
E - Tube containing vanadium.
F - Line to introduce reference samples.
G - Tube containing charcoal.

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Idid p. 1103 - 1109
A FIRST LOOK AT THE LUNAR ORBITAL GAMMA RAY DATA; A. E. Metzger, Jet Propulsion Laboratory, Pasadena, Calif., 91103; J. I. Trombka, Goddard Space Flight Center, Greenbelt, Maryland 20771; L. E. Peterson, R. C. Reedy, J. R. Arnold, Univ. of Calif., San Diego, La Jolla, Calif., 92037

The Apollo 15 gamma-ray spectrometer experimenters now have in hand nearly all the prime data gathered in lunar orbit, and some of that obtained during the trans-earth phase. The first look at the data has been carried out mainly by summing over broad regions of the energy spectrum on a mapping grid of 5° squares. The most instructive of these energy regions is that above the positron line, up to and including the 2.61 MeV line of thorium (taken as 0.55-2.75 MeV). The differences in count rate observed in this energy region can be accounted for mainly by changes in the Th, U, and K content of the surface layers. The highest values are found in Mare Imbrium and Oceanus Procellarum, where a good deal of local variation is observed. Levels reached in parts of this region are consistent with a high concentration of the Th- and U-rich KREEP component. The activity falls off smoothly in the adjacent regions to the east and west. By contrast little or no rise in count rate is seen in Mare Crisium and Mare Smythii on the east as compared to the highlands. The broad highland regions are low in radioactivity, but there is a measurable rise at the southernmost latitudes on the back side. To the extent that our analysis has proceeded, the data can be fitted by a three-component model for the lunar regolith, including KREEP, mare basalt, and a low-radioactivity material such as anorthosite. The results are broadly consistent with the Luna X observations (1).

Variations are also seen in other energy regions, but these are not yet analyzed. In addition to the radioactive lines, lines which can be ascribed to O, Si, Fe, Al, and other elements have been identified. Preliminary analysis indicates that the K/U ratio is lower than that seen on earth in all regions covered. A full data reduction program is underway, involving a definition of the background continuum, followed by determination of a consistent set of elementary concentrations from the line intensities, for interesting regions of the moon.

During trans-earth coast, the total cosmic gamma-ray spectrum in space was measured over the energy range 0.3-27 MeV. Previous observations from Ranger III (2) and ERS-18 (3) had indicated the presence of this astronomical component, which apparently has a nearly isotropic distribution and is therefore associated with the large structure of the universe. Data were obtained in various gain ranges, and with the detector at different distances from the spacecraft. These data verified that the spectrum measured with the boom at 25' was indeed essentially free from spacecraft contributions. We believe effects due to local instrument matter are minor. The measured energy loss spectra agree well with the previous measurements below 1 MeV. In the 1-6 MeV range, the Apollo 15 results lie up to a factor of 5 below
the singular ERS-18 measurements. The few weak gamma-ray lines observed are attributed to local and spacecraft effects. The energy loss continuum to 27 MeV will require unfolding to obtain the true cosmic gamma-ray spectrum.

The general sky background is lower by a factor from 3 to 10 than the flux observed near the moon. Thus it does not have a large effect on the determination of the lunar flux.

References:

MINERAL ASSEMBLAGES AND THE ORIGIN OF NON-MARE LUNAR ROCK TYPES.

Charles Meyer, Jr., NASA Manned Spacecraft Center, Houston, TX 77058.

The lunar soil and breccia samples contain a wide variety of small igneous rock fragments which can be related by their mineral assemblages and chemistries to three major lunar rock suites (I Mare basalt, II KREEP basalt, III Anorthosite Clans) which are inferred to correspond to three separate episodes of lunar history. A study of these lithic fragments indicates that fractional crystallization, partial melting and thermal metamorphism were important processes during the extensive chemical differentiation of the early moon.

Mare basalts (3.1-4.0 b.y.) are characterized by high FeO (≈20%), low Al$_2$O$_3$ (<15%) and low trace element contents. They have a high opaque content, high mafic mineral to plagioclase ratio (2:1) and contain large amounts of high Ca clinopyroxene. In comparison KREEP basalts (4.3-4.4 b.y., Schonfeld, 1971) have lower FeO (<10%) higher Al$_2$O$_3$ (<18%) and much higher trace element contents than mare basalts. The mafic mineral to plagioclase ratio is 1:1, there are less opaque minerals and the dominant mafic mineral is orthopyroxene. In the Apollo 12 and 14 soils and breccias the original mineral assemblages of the fragments of this component were extensively altered by various degrees of brecciation, melting and annealing. Basalt fragments with characteristic KREEP chemistry have been found in 15023,2,5 with fresh subophitic texture (Fig. 1). These fragments have the mineral assemblage 30% orthopyroxene, 15% pigeonite, 45% plagioclase An80, 5% ilmenite, and 5% whitlockite, Kspar, SiO$_2$, ZrO$_2$ and residual glass. The pyroxenes are zoned with Mg rich orthopyroxene cores and Fe rich pigeonite rims. A small amount of ferroaugite coprecipitates with the pigeonite (Fig. 2). The bulk composition of these fragments and of the silica rich residual liquid is given in Table I. The residual liquid is present as high SiO$_2$ glass containing immiscible iron rich glasses and forms a large portion of these fragments (<2%). Granitic liquid could be derived during slow crystallization and/or by filter pressing of KREEP magma. Fragments of the annealed brecciated variety of KREEP basalt have the mineral assemblage; orthopyroxene, augite, olivine, plagioclase, ilmenite, whitlockite, Kspar (Fig. 2). The texture of these fragments is generally equigranular, with relict zoned plagioclase crystals. It is apparent that the precursor rock type for these fragments was hypabyssal basalt (rather than "norite") and that they acquired their present "noritic" mineral assemblage by recrystallization. Separation of a liquid rich SiO$_2$, K$_2$O, and BaO during this high grade thermal metamorphism (≈1000°C) would provide a second mechanism for the production of granitic liquid from a KREEP parent material. Anorthositic fragments with plagioclase-rich mineral assemblages commonly exhibit brecciated and annealed textures indicating that they have had a complicated history and are non-mare in origin. Sample 14321,17E consists of about 100 equant plagioclase grains in a polygonal intergrowth. This unusual fragment consists of a
plagioclase matrix including 21 zircon, 2 whitlockite, and 2 pyroxene grains. Fragment 12033,85C is typical of the brecciated variety of anorthosite with a few large, unzoned plagioclase grains surrounded by a matrix of fine, ground-up and annealed plagioclase. The high K₂O content (Fig. 5) of the plagioclases of these anorthosite fragments indicates that they may have crystallized and accumulated from a trace element and aluminum-rich parent magma (KREEP). Plagioclases from other anorthosite fragments commonly have low K₂O contents and presumably crystallized from a low K₂O parent magma. Lack of chemical zoning in the minerals of these fragments indicates either a plutonic origin, a high annealing temperature or both. The density of calcium rich plagioclase would not be greatly different from that of a parent liquid, thus the mechanism for plagioclase accumulation may include flow differentiation in a convecting magma. Granitic fragments have been found. These include: 1) 12032,56B contains <20 Kspar grains (<100μm) interlocking with each other and with plagioclase An40 and quartz. Approximately 50 small pyroxene grains occur as rounded blebs (<5μ) within the feldspar matrix. These pyroxenes are exsolved (<2μm) and have high Fe/Mg ratios (Fig. 3). 2) 12022,83A contains potassium feldspar and quartz intergrown as long needles. Pyroxenes in this fragment as well as the light portion of 12013,7 are exsolved in a similar manner to 12032,56. This coarse pyroxene exsolution indicates that these granitic fragments cooled slowly—perhaps as veins in a thick ejecta blanket of annealed KREEP breccia.

Summary: The rock types present in the non-mare regions of the moon are aluminum-rich and include anorthosites, pyroxene-bearing granites and trace element rich high-aluminum basalts. The relative absence of mafic rock types indicates that the non-mare regions are not composed of a series of stratiform sheets (such as the Bushveld complex) but that massive bodies of aluminum-rich rocks are exposed on the surface presumably covering their more mafic equivalents to a depth of 10 km or more. Thus an Al+Ca/Fe+Mg, or plagioclase/pyroxene, gradient existed with depth in the pre-Imbrium lithosphere.

<table>
<thead>
<tr>
<th>TABLE I</th>
<th>KREEP BASALTS</th>
<th>GRANITIC SAMPLES</th>
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<tr>
<td></td>
<td>A-12 Annealed</td>
<td>Residual</td>
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<td></td>
<td>KREEP Glass</td>
<td>KREEP Brecciated 15023,2,5</td>
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<tr>
<td>SiO₂</td>
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</tr>
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MINERAL ASSEMBLAGES AND THE ORIGIN OF NON-MARE LUNAR ROCK TYPES

C. Meyer, Jr.
Knowledge of the mechanical properties of lunar soil has been accumulating since well before the first lunar landings. With continued exploration, however, it has been possible to decrease the degree of speculation and to increase the certainty with which these properties and their variability are known. Of particular importance and interest to lunar science are density ($\rho$), cohesion ($c$), and angle of internal friction ($\phi$). Knowledge of $c$ and $\phi$ are important, both because individually they reflect density, composition, and environmental influences, and because together they control the soil strength. The soil strength, in turn, dictates the maximum possible soil slope angles, the bearing capacity of the surface, and the resistance to penetration of objects into the soil.

The gradation of the lunar soil (silty fine sand with particles generally in the range of about 0.01 mm to 5 mm) is such that density and porosity are the most important characteristics controlling the physical behavior. For terrestrial soils of similar gradation it has been well established that the cohesion, the friction angle, the compressibility, and the modulus of deformation may all vary significantly for relatively small changes in density. Evidence to date indicates that the lunar soil behaves in a similar manner. Density and porosity are of importance also to those investigators concerned with seismic wave propagation, thermal conductivity, and cosmic rays. Substantial regional and local density variations and changes in density with depth have now been observed. The development of suitable explanations for these variations will be of importance in the understanding of lunar history.

A variety of methods has been used for estimating lunar soil density, cohesion and friction angle. In this paper a number of these methods are cited and the results noted in an attempt to develop a historical perspective on each of the three properties. Conclusions are drawn relative to the ranges of values and the reliability with which they are known based on data available through Apollo 15.

**Density**

Early density estimates of as low as 0.3 gm/cm$^3$ up to 3 gm/cm$^3$ were based
MECHANICAL PROPERTIES OF LUNAR SOIL: DENSITY, COHESION, AND ANGLE OF INTERNAL FRICTION

J. K. Mitchell

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largely on assumption and deduction from earth-based optical, thermal and radar measurements. First in-place density measurements were by a gamma ray device on the Soviet Luna XIII. Grain-size distribution and footpad-surface interaction as well as the results of the Soil Mechanics Surface Sampler experiment were used to obtain a density estimate of 1.5 gm/cm$^3$ in the Surveyor Program.

A rotary drill sample returned by Luna XVI yielded a density of 1.2 gm/cm$^3$. Core tube sample data have provided the best density data during the Apollo program, although additional density estimates have been possible using the results of studies on lunar soil simulants. The best available data indicate densities in the range of about 1.3 to 2.1 gm/cm$^3$. Such a range can be expected to embrace a wide spectrum of strength and compressibility behavior.

Cohesion

Cohesion estimates prior to the Apollo program were derived mainly from television observations, landing leg strain gage data, and the Soil Mechanics Surface Sampler during the Surveyor program. Best estimates were in the range of 0.3 to 0.7 kN/m$^2$ (0.05 to 0.10 psi). During Apollo cohesion evaluations have been possible on the basis of (1) analysis of blowing dust during LM descent, (2) failure of the soil mechanics trench wall, (3) penetration resistance measurements, and (4) simulation studies. Values as low as 0.03 kN/m$^2$ (0.005 psi) and as high as 1.0 kN/m$^2$ (0.15 psi) have been obtained, with the highest values associated with the finest grain sizes and highest densities.

Angle of Internal Friction

Pre-Apollo estimates of friction angle were based on analyses of slope angles (Ranger and Orbiter programs) footpad and Soil Mechanics Surface Sampler interactions with the lunar surface (Surveyor program), and boulder track records (Orbiter). Values from 10° to 55° were indicated from these results, although 35° to 40° appeared the best estimates. During Apollo the friction angle has been deduced from (1) penetration test data, (2) simulation studies, and (3) MET track (Apollo 14) data. It now appears that values as high as 50° are possible for the soil in a dense packing. The lower bound, which would be associated with the soil in its loosest state, is still undetermined.
ELASTIC WAVE VELOCITIES AND THERMAL DIFFUSIVITIES OF APOLLO 14 ROCKS; H. Mizutani, Geophysical Institute, University of Tokyo, Tokyo, Japan, now at Seismological Laboratory, California Institute of Technology, Pasadena, Calif., N. Fujii, Y. Hamano, M. Osako, Geophysical Institute, University of Tokyo, Tokyo, Japan, and H. Kanamori, Earthquake Research Institute, University of Tokyo, Tokyo, Japan.

This paper presents new results of compressional- and shear-wave velocity measurements on two Apollo 14 rocks, 14311,50 and 14313,27 under pressure up to 10 kilobars, and thermal diffusivity measurements on sample 14311,50 over the temperature range 100° to 550°K in vacuum and in one atmospheric air.

Both sample 14311 and 14313 are polymict fragmental rocks. Mean atomic weights and theoretical densities of these samples are estimated at 22.0 and 3.04 g/cm³ respectively from the chemical composition and the normative mineral composition of similar fragmental rocks. Bulk densities of these samples are measured by the Archimedes method; porosities are estimated at 6% for 14311 and 21% for 14313 using the theoretical densities and the bulk densities.

The ultrasonic wave velocity is measured by the method described by Mizutani et al. (1), using the high-pressure system described by Kanamori and Mizutani (2). At pressure above 1 kilobar, the absolute accuracy of the present measurement is better than 0.7% for P-waves and 1.5% for S-waves. At pressures below 200 bars, the onset of the signal is so blunt that the accuracy drops considerably.

The results are summarized in Table 1, Fig. 1, and Fig. 2, where velocities and densities uncorrected for changes of dimensions under pressure are used. The overall elastic and anelastic behavior of the Apollo 14 fragmental rocks are very similar to those of the Apollo 11 and the Apollo 12 rocks; the rapid increase of the velocity and Q for the initial 2 kilobars pressure increase found for the Apollo 11 and 12 rocks are also typical of the Apollo 14 rocks. The plot of velocity (at 5 kb) versus density (Fig. 3) indicates that Birch's law (V = a(m)+bp) holds for lunar rocks as well as for terrestrial rocks. The slopes of the lines (b) for the lunar rocks are surprisingly similar to those for terrestrial rocks; for P-wave b ≈ 3.3 (km/sec)/(g/cm³) and for S-wave b ≈ 1.7 (km/sec)/(g/cm³).

The thermal diffusivity of sample 14311,50 is measured in air at atmospheric pressure and at 10⁻³ ~ 10⁻⁵ mmHg pressure by the modified Angstrom method (3). Results of the present experiment are shown in Fig. 4 and summarized in Table 2. The thermal diffusivity of the lunar rock in vacuum is found to be substantially smaller than those in one atmospheric air. The difference may be explained by the thermal conduction of gas phases filled in microcracks and pores. The low thermal conductivity due to microcracks and pores may be significant in the thermal evolution of the moon because
Elastic Wave Velocities
H. Mizutani

Microcracks and pores may persist to a fairly deep region owing to the low pressure in the moon.

References

Table 1. Bulk density (in g/cm³) and velocity (in km/sec) uncorrected for dimensional change

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<tr>
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Table 2. Thermal diffusivity (in 10⁻³ cm²/sec) of sample 14311,50

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<tbody>
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<td>k (in vacuo)</td>
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<td>3.9</td>
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</table>

Figure Captions
Fig. 1. The P- and S-wave velocities of sample 14311,50 as a function of pressure.
Fig. 2. The P- and S-wave velocities of sample 14313,27 as a function of pressure.
Fig. 3. P- and S-wave velocities at 8 kilobars versus density for lunar rocks. The numbers attached to points are sample numbers and the numbers in parentheses are mean atomic weights.
Fig. 4. Thermal diffusivity of sample 14311,50 as a function of temperature. Open circles represent the data in vacuum and solid circles the data in one atmospheric air.
Elastic Wave Velocities
H. Mizutani

Fig. 1

Fig. 2

Fig. 3

Fig. 4
TOTAL CARBON, NITROGEN AND SULFUR ABUNDANCES IN APOLLO 14 LUNAR SAMPLES, C. B. Moore, C. F. Lewis, J. Cripe, W. R. Kelly and F. Delles, Center for Meteorite Studies, Arizona State University, Tempe, Ariz., 85281

The total carbon abundances in ten samples of normal dark Apollo 14 lunar fines ranged from 90 to 160 μg/g. A light gray colored fines sample from cone crater had from 42 to 80 μg/g total carbon. These samples are similar to the normal dark and light fines samples from Apollo 11 and 12. Apollo 14 basalt 14310 had a total carbon abundance of 35 μg/g. This low total carbon is similar to basalts from Apollo 11 and 12. The fine-grained fragmental rocks or breccias ranged from 21 to 150 μg/g carbon. The total carbon contents showed a direct correlation with total Neon-20 in individual samples and may be correlated with the petrologic classification outlined in the Apollo 14 Preliminary Science Report. Breccias in the first group (F-1) with friable matrices and light colored clasts had total carbon contents of about 150 μg/g. The F-2 breccias with light colored clasts and moderate coherency had variable total carbon contents (60 to 170 μg/g). The third group (F-3) of fragmental rocks with dark colored clasts and coherent matrices had low total carbon contents ranging from 21 to 80 μg/g. Four individual samples from the large F-3 breccia 14321 had consistent total carbon contents of 21, 25, 32 and 43 μg/g. It is suggested that the fragmental rocks with higher carbon contents have normal dark fines material in their matrices while the lower carbon fragmental rocks have no solar wind exposed material.

Total sulfur contents in the Apollo 14 breccias ranged from 700 to 1000 μg/g. Significant fractionation between petrologic groups was not evident. The total nitrogen content of an F-1 fragmental rock was 130 μg/g and an F-3 fragmental rock was 60 μg/g. These values correlate directly with the total carbon content for these samples.

Earlier theories regarding the distribution of carbon, nitrogen and sulfur in lunar fines are supported by the Apollo 14 data.
CARBON, NITROGEN AND SULFUR IN APOLLO 14 SAMPLES

<table>
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<th>Total N μg/g</th>
<th>Total S μg/g</th>
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<td>42</td>
<td>-</td>
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<td>-</td>
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<td>Fines</td>
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<td>-</td>
<td>930</td>
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<tr>
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<td>120</td>
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<td>89</td>
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<td>14298</td>
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<td>14301</td>
<td>F-2</td>
<td>70</td>
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<td>32</td>
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<tr>
<td>14310</td>
<td>Rock</td>
<td>35</td>
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<td>14311</td>
<td>F-3</td>
<td>44</td>
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<td>14318</td>
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<td>F-3</td>
<td>32</td>
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<td>14321 B</td>
<td>F-3</td>
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<td>14321 C</td>
<td>F-3</td>
<td>43</td>
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<td>14421</td>
<td>Fines</td>
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<td>14422</td>
<td>Fines</td>
<td>120</td>
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Samples shown in italics were analyzed at the Lunar Receiving Laboratory during preliminary examination. Fragmental rocks (F-1, F-2, F-3) are defined in text. FU is an unclassified fragmental rock.

Apollo 14. Soils and breccias from Apollo 14 are nearly 3 times richer in siderophile elements (Ir, Re, Au, Ni) than the soils from Apollo 11,12 and Luna 16. Intrinsic lunar abundances of these elements are now known for the parent rocks of Apollo 14 soils, but are probably low. It therefore seems that the meteoritic component in these soils can be estimated fairly well from gross abundances, without any correction for the lunar component.

Most of the meteoritic component seems to be ancient, pre-dating the local craters (Ganapathy et al., Science, in press, 1971). In 14 samples with prolonged surface exposure (bulk soils, magnetic soil fractions, gas-rich rocks such as 14047), the ancient component is slightly veiled by the ubiquitous micrometeorite component of primitive (~Cl chondrite) composition. Such samples are characterized by high Se, Ag, Bi, and Zn (Table 1). In 12 other samples (4 norite and 3 glass composites from 1-2 mm soils, rock 14310, and 4 microbreccias from 14321), the ancient component is present in pure form. It has a strongly fractionated composition, with volatile elements (Ag, Bi, Ge, Sb, Se, Te, Zn) depleted relative to Au by factors of up to 30, and with Ir/Au varying between 0.6 and 3.8, with similar variations in Re/Au ratios. None of the known chondrite classes has so low a volatile content and so low and variable an Ir/Au ratio. Among the remaining meteorite classes, only group IVA irons (and the presumably related eucrites) come close to matching both of these characteristics. Cone crater soil and light norite fragments from other soil samples have uniquely low Ir/Au,Re/Au ratios, perhaps representative of deeper strata at the Apollo 14 site.

It is hard to rule out the possibility that the ancient component represents mixed planetesimal debris from the Pre-Imbrian regolith. This would require, however, that all these bodies had the same peculiar composition. Following William of Ockham, we therefore assume that this component is derived from a single body: the one that produced the Imbrian basin.

Apparently the Imbrian object was similar to the parent body of IVA irons (and eucrites?). Its radius, calculated from cratering theory, was about 100 km; the same as that of the IVA parent body (P. E. Fricker et al., Geochim. Cosmochim. Acta 34,
MAJOR IMPACTS ON THE MOON...

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475, 1970). From the data at hand one cannot tell whether its metal phase was contained in a single core or scattered throughout its volume in smaller pockets (as in the IVA body). Its impact velocity can be estimated from the amount of meteoritic material in the soil. Values of 4 km/sec or 12 km/sec are obtained, depending on whether the portion of the projectile that excavated the Fra Mauro ejecta was a metal-silicate mixture or pure metal. Both velocities can be reconciled with the late date of the Imbrian impact. A low velocity is expected for an interplanetary object in an Earth-grazing orbit, or an Earth satellite, swept up by the Moon during tidal recession or capture. A high velocity is expected for a late survivor of an Earth-crossing population of planetesimals.

The Imbrian object seems to have been markedly poorer in volatiles than the Earth, about as poor as the Moon. Perhaps it was a leftover building block of the Moon. Its high abundance of siderophiles shows that it was not material spun off the Earth after core formation, a condensate or volatilization residue from a hot Earth, or a fragment of a differentiated proto-Moon disrupted during capture.

Apollo 12 Soils. Four KREEP concentrates show a fractionated meteoritic component, similar to that in Apollo 14 soils (Table 1). Ir/Au again lies between 1.0 and 1.6, below the range for chondrites. Volatiles are depleted relative to Au, but perhaps less so than in Apollo 14 breccias. This component may be derived either from the Copernican projectile (J. W. Morgan et al., submitted to Geochim. Cosmochim. Acta, 1971) or from Imbrian ejecta underlying the Copernican site.

Luna 16 Soils. Like other soils of high surface exposure age, Luna 16 soils are dominated by a meteoritic component of unfractionated (~Cl chondrite) composition, presumably representing micrometeorites (J. C. Laul et al., Earth Planet. Sci. Lett., in press, 1971). Several elements (Bi, Ge, and especially Ag, Cd) show anomalous enrichment, suggesting contributions from an Ag,Cd-rich rock type.
Table 1.

Abundances in Apollo 14,

Sample Reaber

Type

14003, 17
14163,5?
14163,57
14259,20

Contingency
Bulk
Slk
Comprehensive

11.0

0.97

4.4

13.6
11.7

0.93
1.07

5.4
5.3

18.6

1.3

6.6

Cone Crater
Surface renc.h
Middle Trench

12.e
13.7
12.7

1.26
1.34
1.11

11.0
6.9
5.3

14141.32
14148,25
114150,25
14144,41

Ir

Re

Apollo 12, and Wna 18 Semples

AU

(ppb; Zn.Rb pp.)
Ag

Br

35

11.5

330

70
25

18.8
18.4

350

50

26.5

3.1
2.8
3.6

270
340
280

.20
160
15

30
12.6
11.7

5b

Ge

se

To

2.4

310

5.7

335
350

In

ai

Z.

Cd

T

Rb

Cs

1.3

25

94

19

13.5

575

72

1.7
16

31
31

140
139

30
35

15.8
18.1

845
730

34

1.8

22

83

18

15.4

620

480
360
290

1.7
1.8
1.3

31
22
20

441
111
77

22
19
19

18.3
18.8
13.5

790
695
570

2.2

19

199

28

13.9

805

490

Bottom Trench

11.1

1.05

7.5

2.4

290

-35

11.8

430

Fract. * 1-2m
14002,3-3

Magnetic, Contingency

13.3

1.03

8.5

2.0

357

10

10,8

470

80

1.10

22

143

39

18.5

670

14268,36-3

Magnetic, Comprehensive

13.5

1.10

5.3

1.8

283

20

8.5

330

24

0.95

19

77

16

13.6

580

14146,2-3

Magnetic, Surface of Trench
Magnetic, Middle of Trench
Dark Morite, Middle of Trench
Dark Norite, Bottom of Trench

17.7
11.9
8.9
6.6

1.47
0.99
0.92
0.63

7.6
10.0
4.8
3.1

2.2
2,4
2.2
2.9

351

25
110
-40
30

12.6
10
1.4
0.94

450
310
270
220

40
26
12
20

1.3
1.5
0.63
0.63

28
23
2.9
2.4

114
78
48
38

21
20
33
30

16.0

870

273
130
130

13.2
14.5
16.4

580
735
710

4.9

0.43

5.4

1.7

91

170

1.2

150

10.8

0.47

4.1

1.2

110

-50

0.85

100

15.2
20.0

620

4.8

0.87
0.63
0.53

121
88
82

30

2.5
0.74
0.78

70
190
100

4.1
25.1
7.1

185
935
250

16

16.2

670

17
1.4

14.4
2.1

10

11.8

650
90
540

2.7
3.6

170
230

14154.2-3
141854,2-7
14151,12-7
14146.2-9
14142,1-12

Light Norite, Surface of French
Light Morite, Cone Crater

14258,36-5
Yellow Clas
Green Glans

Ropy Glass, Coprebensive
From: 14002;14258314154:14151
From: 14002;14258:14146;14154;14262

13.1
3.8
7.0

0.91
0.32
0.53

5.1
1.2
2.0

-20

14047,38

Glass Coating

11.7

1.12

5.2

2.0

315

160

14047,25
14053,26
14310,119

fragment of Rock
Bsalt
"Basalt"

11.2
0.017

1.06
0.0068

5.4
0.11

2.1
0.44

320
141

'40
15

1.02

4.31

4.5

130

120

4

14321,184-1S
14321,184-9A

Basaltic Clast
Matrix

0.044
0.71

0.0031
0.056

0.30
0.70

0.78
15.3

640
240

338
162

14321,184-14A
14321,184-15

MIcrobreccia
Dark Clait (Microbreccia)

7.8
6.9

0.70
0.64

6.06
8.08

2.2
2.1

50
500

14321,184-lOA
14321,184-19A

Mlcrobreccia
Mlcrobreccla

9.7
6.1

0.86
0.55

9.96
6.41

2.9
2.4

760
400

Apollo 12
12033,105

2 KRM

1.6

0.097

0.99

0.43

56

12033,20-4

Glassy XRKEP (+30 mesh)

1.5

0.15

1.5

0.79

84

12033,20-7
12033,20-1

Cindery KEB37 (+30 mash)
Magnetic (30-100 Mesh)

2.2
4.7

0.19
0.57

2.1
5.4

1.5
3.1

99
172

4.2

0.25

2.2

0.77

166

9.6
9.8

3.7
3.4

2.7
2.9

3.3
4.2

8.7

0.71

2.4

2.2

0o009

0.79

0.75

Fragments

12037.25

<1 me Soll

Lana 1s
L-16-A-5
L-16-G-5

<0.125m
<0.125 -

Apollo 11
.10084,49

<I

8CR-i

Columbia River Basalt

Soil
Soll

S ll

10.5

560

1300

1500

10.3

5.9
3.6
7.8

1.8

14

0.48

2.4

27

0.49
0.72
0.26

5.6
2.8
1.5

20
18
16

S0.7

9.9
10
5.2
3.0
3.5

90

300

27

1.1

23

11.0
0.60
10780

370
50
235

50
15

1.1
0.29

20
2.1

2.5

2.3

4.5

6

0.80
1.10

80
85

1.84
2.74

0.39
0.55

2.9
2.8

24
7.3

128
139

11
8

0.88
1.49

300
520

3.40
1.69

0.34
0.32

8.8
3.8

298
17

8.1
3.9

12.9 650
30.8 1150

110
92

6

2.63
0.83

270
215

1.77
1.45

0.25
0.24

2.5
3.3

106
52

1.8
1.9

10.9
9.3

640
510

3.3

70

1.25

4A

1.1

8.0

8.0

19.4

710

2.2

160

1.83

0.70

1.6

3.8

4.1

17.0

6855

2.0
2.5

140
130

1.21
1.31

0.4
0.4

3.3
3.2

7.7
13

3.3
3.3

17.0
14.9

425
535

0.70

8.1

54

2.6

4.9

210

00
10s0

1.4
1.9

1.5
1.9

8
8

340
370

40
26

270

30

92

-

25

90

1340
540

120
ISO

9.0
26

130

222
3.0
1.9

85
80

-

88

4.6
5.1
1.5
46

24
29

80
104
20

19

37

116

129

1.7
1.7

1.7
300

2.7
48

106
925

C4
-4

K

C

0 )1


VOLATILE AND SIDEROPHILE ELEMENTS IN APOLLO 14 AND 15 ROCKS

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Department of Chemistry and Enrico Fermi Institute
University of Chicago, Chicago, Illinois 60637

Five Apollo 15 soils and 4 rocks were analyzed by neutron activation analysis for 16 volatile and siderophile elements (Table 1). Four Apollo 14 samples, measured too late for inclusion in our companion paper (Morgan et al., 1972), are also listed.

Soils. In general, soils from Apollo 15 resemble those from Apollo 11, 12 and Luna 16 (Ganapathy et al., 1970; Laul et al., 1971a, b; Morgan et al., 1972). Elements largely of meteoritic origin (Ir, Re, Au, Sb, Ge, Se, Bi, Br) are present in abundances equivalent to \(1.5-2\%\) CI chondrites. Of the two Elbow Crater soils, 15071, collected 25 m from the crater rim, is systematically lower in meteoritic and other trace elements by a factor of \(0.8\) than 15081, collected 65 m from the rim. Apparently 15071 was diluted by 20-30% ejecta from Elbow Crater, probably freshly crushed (basaltic?) bedrock.

Apennine Front soil 15431 is 20-30% lower in "meteoritic" elements than the other 4 Apollo 15 soils and intercrater soils from other landing sites. Probably this implies dilution by talus from the Front. However, 15431 is highest in "largely non-meteoritic" elements (Zn, Cd, Tl, Rb, Cs), higher in fact than Apennine Front basalt 15256. Presumably the Front contains other rock types that are enriched in these elements. Such rock types apparently have contributed also to the 4 soils away from the Front, which are richer in the above elements than the Rille basalts.

No systematic difference was found between soils on and off the Aristillus-Autolycus ray (15256, 15471 vs. 15071, 15091, 15501).

Rocks. Anorthosite 15415 and Rille basalts 15555 and 15597 show some of the strongest depletions in siderophile and volatile elements seen to date, especially for Ge, Sb, In, and Zn (Ganapathy et al., 1970; Anders et al., 1971). Apennine Front basalt 15256 is generally less depleted, particularly in Rb, Cs, Tl, Zn, Cd.

Depletion in Volatiles. Apollo 14 rocks have a slightly higher Tl/Cs ratio (2\(\times\)10\(^{-2}\) vs. 0.7\(\times\)10\(^{-2}\)) than Apollo 11, 12, 15 basalts, anorthosites, and "granites" (12013). The mean ratio of all Apollo rocks, 1\(\times\)10\(^{-2}\), still is more than an order of magnitude below the terrestrial and eucritic ratios (Anders et
VOLATILE AND SIDEROPHILE ELEMENTS...

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In terms of the two-component model of planetary accretion, this suggests that the Moon contains only ~2% low-temperature material, with a nominal accretion temperature of ~450 K. If a non-volatile element such as U is used as the reference standard, the Moon is depleted in Tl by two orders of magnitude relative to the Earth.

Meteoritic Components. Elements diagnostic of meteorites are far less abundant in 15415 anorthosite than in anorthosite fractions from Apollo 11 soil (Laul et al., 1971a). This supports our earlier suggestion that the Apollo 11 anorthosite contains a fractionated meteoritic component, equivalent to ~0.5% H-chondrite or a similar meteorite type.

Apennine Front basalt 15256, though shock-melted, shows no detectable meteoritic component.

A search for KREEP-rich rocks free of meteoritic material has been unsuccessful (cf. Table 1 of Morgan et al., 1972, for data). 14053 and a basaltic clast from 14321 are low in meteoritic elements, but are much poorer in Rb,Cs than KREEP. Our sample of 14310 (a rock classified as igneous) contains meteoritic elements in exactly the same amounts and proportions as Apollo 14 soils and breccias (except for the curiously high Ag content). Perhaps our sample came from a small patch of breccia in this rock; LSPET had in fact suggested that this "rock" might actually be a large clast from a breccia. If not, we are forced to speculate that it represents remelted breccia, or an ejecta stratum initially heated above the melting point. Rock 12013 also contains a small meteoritic component (~1% Cl equivalent in the KREEP portion, ~0.2% in the granitic portion). Rock 14065 was listed in the LSPET report as having high K and low Ni, but samples from another portion of this rock (14063, Tatsumoto consortium), contained both Rb,Cs and Ir,Re,Au in abundances only slightly above basaltic levels.

<table>
<thead>
<tr>
<th>Sample Type and Description</th>
<th>1r</th>
<th>Re</th>
<th>Au</th>
<th>Co</th>
<th>Cr</th>
<th>Ge</th>
<th>Se</th>
<th>Ag</th>
<th>Br</th>
<th>In</th>
<th>Sn</th>
<th>Cd</th>
<th>Tl</th>
<th>Rb</th>
<th>Cs</th>
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<td></td>
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<tr>
<td>15265.6</td>
<td>0.022</td>
<td>0.0049</td>
<td>0.019</td>
<td>46.2</td>
<td>0.43</td>
<td>3.8</td>
<td>119</td>
<td>0.78</td>
<td>51</td>
<td>6.8</td>
<td>0.41</td>
<td>0.92</td>
<td>104</td>
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<td>Apennine Front</td>
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<td>0.139</td>
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<td>0.067</td>
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<td>156</td>
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<td>0.0081</td>
<td>0.045</td>
<td>39.6</td>
<td>1.49</td>
<td>6.5</td>
<td>117</td>
<td>0.90</td>
<td>24</td>
<td>0.59</td>
<td>0.21</td>
<td>1.7</td>
<td>0.32</td>
<td>0.72</td>
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Table 1. Trace Elements in Apollo 14 and 15 Samples

Microcrater population distributions have been determined on part or all of the surfaces of ten lunar rocks from Apollo 12 and 14. Lunar surface orientations of three rocks not documented by surface photography were determined by microcrater distribution studies.

Flux Rates: A current estimate of the influx of meteoroids is expressed by two equations (1):

\[ \phi_1 = 10^{-14.37} \text{m}^{-1.21} \left(10^{-6} \leq m \leq 10^0 \text{gms} \right) \]

\[ \phi_2 = 10^{-14.34} \text{m}^{-1.58 - 0.63 \log_{10} m} \left(10^{-12} \leq m \leq 10^{-6} \text{gms} \right) \]

where \( \phi_1 \) and \( \phi_2 \) are numbers of impacts/meter\(^2\)/sec/2\(\pi\) steradians and \(m\) is the particle mass in grams. A second estimate is provided by Whipple, Curve B (2) expressed as:

\[ \phi_3 = 10^{-13.80} \text{m}^{-1.0} \]

These estimates bracket calculated flux using the particle track age of Fleischer et al (3) and the crater counts of Horz et al (4) for the glass surface on rock 12017G which indicates a flux of approximately \(10^{-7.4}\) particles/meters\(^2\)/sec of mass \(10^{-6}\) grams and larger. Particle density is taken as 0.44 gms/cc — in keeping with Whipple's density estimates.

We set a lower limit of the influx of meteoroids on the lunar surface with crater population and exposure age data from rock 14301. This flux predicts a minimum infall of approximately \(10^{-8.8}\) particles/meters\(^2\)/sec of mass \(10^{-6}\) gms for particles with a density of 0.44 gms/cc. The rate is a minimum because the exposure age of \(6.5 \times 10^5\) years is based upon an \(Al^{26}\) measurement which does not distinguish between direct exposure of the rock and burial of a few mm. The true flux, therefore, could be higher but not lower. The data from 14301 is in fair agreement with the flux suggested by crater counts of Bloch et al (5) and particle track ages of Crozaz et al (6) for one surface of rock 12063. For the other surface of 12063, the flux obtained may be a minimum because it may be in a steady-state and therefore record fewer events than the number actually occurring. Both fluxes are less than that obtained from rock 12017G and the current estimates of flux by a factor of about 10 at a mass of \(10^{-6}\) gms. Fluxes extrapolated to 1 gram particle sizes using the equations for \(\phi_1\) and \(\phi_3\) are higher than those obtained by Latham et al (7) for that mass.

Mass Distribution: Crater size frequency and mass distributions have been determined primarily by optical methods which are unreliable for small craters on rock surfaces (3). To extend the crater size-frequency distribution below optical resolution an exterior chip of breccia 12073 was examined using a Scanning Electron...
Microcraters on Lunar Rocks

D. A. Morrison

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Microscope (SEM). A 40X mosaic of 44.22 mm² of the surface showed a total of 74 craters. The crater size distribution curve has a relatively constant slope of -1.9 down to a size of about 50 microns (pit diameter) where it levels out rapidly. The smallest crater measured has a central pit diameter of 15 microns corresponding to a mass of 10^{-10.6} gms.

The glassy liners of many of the 74 craters detected by the scan were examined for superposed microcraters. None were found that could be identified unambiguously as microcraters produced by hypervelocity impact. A further search was made of smooth crystal faces on the rock surface and a total of 37 faces with an aggregate area of 0.59 mm² were scanned at magnifications ranging from 500-2000X. No craters larger than the crater resolution limit of 5 microns were seen. Although an area this small is not ideal statistically, it sets an upper limit of 1 crater per 0.59 mm², or about 170 craters per cm², larger than 5 microns on this steady state breccia surface. This limit agrees with a projection of the distribution curve for the larger craters to a size of 5 microns. The upper limit of 170 craters/cm² larger than 5 microns is less than that obtained by Bloch et al (5).

Because the mass distribution obtained by the scanning microscope on a breccia surface is in disagreement with those obtained by other workers (5) on glass surfaces, we have made a preliminary scan of a glass chip from the surface of rock 15015. This scan showed a higher percentage of craters per unit area of less than 5 microns diameter, but it also indicated that a wide variety of impact-like features occur, some of which could be attributed to low velocity, possibly secondary particles whereas others may represent vesicles. We conclude that there is a lack of criterion for distinguishing primary microcraters from features of similar morphology, and that crater counts of glass surfaces may result in unusually high numbers of microcraters of less than 10 microns diameter per unit area if all crater-like forms are considered to be primary impact craters. Until unambiguous criteria can be established, only craters in crystalline surfaces should be used in determining flux of micron size particles. This is particularly true in the case of spherules.

The Steady State: Ideally, the cumulative frequency distribution of craters on rock surfaces becomes independent of time when surfaces reach a steady-state. Our understanding of the steady-state surface for a rock is unclear at this time. The youngest dated rock surface (12017G, 9 x 10^3 years) yields a flux rate and frequency distribution which is in reasonable agreement with current estimates of the crater production frequency for the moon. Older rock surfaces (>10^6 years) have fewer craters than expected by a factor of at least 1/5. This decrease may be the result of 1) approaching or attaining the steady state, 2) a decrease in crater production with time, i.e. the present flux rate is higher than the recent past, 3) the orientation of the rock on the lunar surface, 4) partial covering of rock by fines, or 5) any combination of these.

Because there are only two data points for flux rates, the flux of micrometeorites is not known with certainty, and the time required to attain the steady state remains somewhat speculative. We assume that all cratered surfaces which are well-rounded are in the steady state, because most or all of the original surface has been removed. By this criterion, most of the rock surfaces examined have steady-state microcrater populations. Optically determined crater-size distribution curves for such surfaces exhibit the general characteristics described by Horz et al (4). However, some rocks have steady-state populations considerably in excess of the 10% saturation value of Gault. Such large populations (e.g. 124 craters/cm² larger than 75 microns
Microcraters on Lunar Rocks

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central pit diameter on a surface of 14306) occur on surfaces which have spall diameter to central pit diameter ratios of 3.4/1.

Exposure Age: Studies of craters on several rocks indicate that an estimate of surface residence time may be obtained by combining crater populations with flux rates. We calculate that 14053 was exposed for approximately \(1.1 \times 10^6\) years assuming 1) the crater population has not been severely biased by an approach to the steady state and 2) the crater population of 14306 is the steady-state. We also calculate a surface exposure of \(1.2 \times 10^6\) years for rock 14311 with the same assumption concerning steady state. These estimates can be tested. In both cases the rocks are not rounded but do show some signs of wear. Flux rate and crater distribution data indicate that rock 14313 had a five-stage surface history with a surface exposure time \(23 \times 10^6\) years or larger. The rock assumed five distinct positions on the surface of greater than \(10 \times 10^6\), approximately \(2.5 \times 10^6\), and \(1 \times 10^6\), collectively and greater than \(10 \times 10^6\), and greater than \(1 \times 10^6\) years duration successively.

EQUILIBRIUM STUDIES WITH A BEARING ON LUNAR ROCKS A. Muan, J. Hauck and T. Löfll, The Pennsylvania State University, University Park, Pa., 16802.

Phase relations in real lunar samples as well as in simplified oxide and silicate model systems have been studied in order to determine liquidus and solidus temperatures, sequences of appearance of the various crystalline phases under equilibrium conditions, and the distribution of elements among coexisting phases at subsolidus temperatures. Particular attention in the studies of synthetic samples has been directed toward the determination of solid-liquid equilibria in relatively simple iron oxide-containing silicate systems in which occur the phases olivine, pyroxene and anorthite under strongly reducing conditions, and toward the determination of immiscibility relations in iron oxide-containing spinel phases under strongly reducing conditions.

Equilibrium studies have been carried out on rocks no. 14-310 and 14-259 in contact with metallic iron under vacuum. The former has a liquidus temperature of approximately 1285°C with plagioclase as the primary crystalline phase, and a solidus temperature of approximately 1120°C. The latter sample has a liquidus temperature of approximately 1235°C, again with plagioclase as the primary crystalline phase, and a solidus temperature of approximately 1130°C.

The dominant phases of lunar rocks are olivine, pyroxene and plagioclase. Crystallization relations among these minerals are an important source of information on consideration of the petrogenesis of lunar rocks. In order to delineate such relations in a simple model system, we have initiated studies of phase relations in diopside-fayalite-anorthite. The equilibrations were carried out in iron crucibles sealed into silica capsules under vacuum. As a starting point in this work, phase relations have been delineated for the bounding system diopside-fayalite. A temperature maximum exists on the olivine liquidus curve at approximately 1335°C. Similarly, a temperature maximum exists at approximately 1195°C on the olivine-anorthite liquidus boundary curve in the system diopside-fayalite-anorthite. The phase relations below liquidus temperatures deviate markedly from binary and ternary relations, respectively, in the two systems because of the permutation of magnesium and iron between olivine and pyroxene phases.

Among the most interesting findings from studies of lunar rocks has been the observation of coexistence of chromium-rich titanate spinels and titanium-rich chromite spinels. It is of interest to determine the conditions under which such coexistence is possible and to place some limits on the temperatures and phase compositions for such coexistence under equilibrium conditions. Toward this end, we have continued and extended our previous work on model systems of the type AO-B2O3-TiO2 (A = divalent ion, usually Mg or Fe, and B = a trivalent ion, usually Cr or Al.) by determining the essential features of phase relations in the systems FeO-Cr2O3-TiO2 and FeO-Al2O3-TiO2 at 1300°C. These FeO-containing systems show the same features as were
The directions of conjugation lines between the coexisting spinel phases within the two-phase area have been determined; they are slanted heavily toward the Mg$_2$TiO$_4$ and Fe$_2$TiO$_4$ apices of the triangular diagrams representing the two systems, respectively. We are able to rationalize the compositions of coexisting spinel phases observed in lunar rocks on the basis of the model systems described above.
observed in the analogous MgO-containing systems, viz. a miscibility gap between the titanate and aluminate spinels and a continuous solid-solution series between the titanate and chromite spinels.

In order to further elucidate miscibility relations among titanate-chromite-aluminate spinels, and to determine the distribution of cations between coexisting spinel phases, we have made additional studies within the ternary systems MgAl$_2$O$_4$-Mg$_2$TiO$_4$-MgCr$_2$O$_4$ and FeAl$_2$O$_4$-Fe$_2$TiO$_4$-FeCr$_2$O$_4$ in the temperature range of 1000-1300°C. Equilibrium in these systems is attained within a reasonable period of time (~1 week) under "dry" conditions at the highest temperatures used (1300 and 1200°C), whereas at lower temperatures a flux and/or hydrothermal conditions were used in order to promote the reactions. In each of these systems a miscibility gap originates in the aluminate-titanate join and gradually closes as the chromite end member is added as a third component. The approximate miscibility relations as a function of temperature are sketched in Figs. 1 and 2.

**Fig. 1** Diagram showing miscibility gap in the system FeAl$_2$O$_4$-Fe$_2$TiO$_4$-FeCr$_2$O$_4$ in the temperature range of 1000-1300°C and an oxygen pressure of 10$^{-10}$ atm.
GEOLOGIC SETTING OF APOLLO 16 LANDING SITE: DESCARTES
W. R. Muehlberger, E. L. Boudette, D. P. Elston,
U.S. Geological Survey, 601 East Cedar Avenue,
Flagstaff, Arizona 86001; C. A. Hodges, D. J. Milton,
U.S. Geological Survey, 345 Middlefield Road, Menlo
Park, California 94025

Apollo 16 will land at the Descartes site (8°55'S, 15°25'E) in the Central Highlands of the Moon. This site is an exceptionally favorable locality to establish the nature of the varied and complex terra volcanism and thus is the highlands counterpart of the Apollo 11 and 12 mare missions.

Three formations dominate the region: Cayley Formation; rugged and bright Kant material that extends from the landing area south to Descartes; and other materials of the Kant Plateau (east of the landing site). All are believed to be of volcanic origin and Imbrian in age. The Cayley appears to be older than the maria and younger than the rugged Kant Plateau materials. Secondary craters from Theophilus and chain craters of possible volcanic origin are identifiable in the vicinity of the landing site.

Two kilometer-sized bright-rayed craters in the traverse area expose a multi-layered sequence in their walls. Sampling of these craters and others along the traverses should permit stratigraphic reconstruction of the Cayley Formation in this area and show its relation to the rugged materials of the Kant Plateau.

NOTE: This work done under NASA contract numbers T-65253-G and T-66359-G.
MORPHOLOGY OF SPRAYS FROM THE MOON AND OTHER SOURCES. G. Mueller, Institute of Molecular Evolution, University of Miami, Coral Gables, Florida.

It was found that the statistical petrography of the rotationally elongated and flattened spray assemblages from Apollo 11 and 12 samples (1) and from Apollo 14 fines varies within relatively narrow limits. Sprays from other natural and synthetic sources can be readily distinguished from lunar sprays and from each other, as shown in Table 1, below.

Table 1. Statistical morphology and petrography of spray assemblages

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<td>97</td>
<td>99</td>
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<td>6.</td>
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<td>17</td>
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<td>32</td>
<td>1</td>
<td>112</td>
<td>0,1</td>
<td>12</td>
<td>-</td>
<td>2</td>
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<td>1,0%</td>
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<tr>
<td>7.</td>
<td>100</td>
<td>57</td>
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<td>32</td>
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<td>46</td>
<td>-</td>
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<td>8.</td>
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<td>32</td>
<td>0,2</td>
<td>28</td>
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<td>11</td>
<td>12</td>
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<tr>
<td>9.</td>
<td>100</td>
<td>51</td>
<td>3,5</td>
<td>16</td>
<td>3</td>
<td>11</td>
<td>-</td>
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<td>11</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0,5%</td>
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<tr>
<td>10.</td>
<td>100</td>
<td>8</td>
<td>2</td>
<td>1</td>
<td>0,2</td>
<td>1</td>
<td>0,5</td>
<td>4</td>
<td>-</td>
<td>2</td>
<td>-</td>
<td>0,1</td>
<td>1</td>
<td>95,0%</td>
</tr>
</tbody>
</table>


It was observed that the length/breadth ratios (L/B) of over 95% of the distinct types of regular, smooth lunar elongated particles fall within...
Morphology of sprays.
G. Mueller

non-overlapping ranges, as follows: A) prolate spheroids, \( L/B = 1-1.8 \);
B) cylindrical spheroids, \( L/B = 1.8-2.48 \); C) dumbbells, \( L/B = 2.48 \) to about 6.

In the case of dumbbells, under breadth is understood as the mean of the diameters of the two lobes. The breadth to waist (B/W) ratios of most of the smooth and symmetrical lunar dumbbells tend to increase with \( L/B \), according to the following empirical equation:
\[
L/B - B/W - 1 = D.C.,
\]
where D.C., the dumbbell constant, remains within micrometric measurement error (about \( \pm 0.2 \) units) close to the mean L/B value at which the dumbbell constriction commences, that is, to 2.48. Preliminary results indicate that the ratio of length of parallel section/length (P/L) of both cylindrical spheroids and dumbbells (in dumbbells P is the distance between the two bulges) tends to increase with \( L/B \) as follows:
\[
L/B - 3(P/L)^2 = P.C.;
\]
the parallel constant which remains close to the \( L/B \) value of 1.8, at which the parallel section appears. The following relation was found so far to hold between the dumbbell and parallel constants:
\[
D.C. = 1.4 P.C.
\]

It was observed that the dumbbell constant holds well for regular dumbbells in chemically homogeneous sprays irrespective of size, such as the basaltic sprays of Hawaii. In chemically heterogeneous sprays, such as the lunar sprays and those of Hekla, Iceland, dumbbells of anomalous constants and composition tend to appear. The dumbbell constant seems to be effected also by the physical conditions which prevail during the formation of a given spray assemblage. These factors are briefly discussed and illustrated (Fig. 1) below.

**Fig. 1.** Tracings of typical elongate particles with \( L/B = 3.0 \)

- A. Average Apollo 12,033 dumbbell of brown glass of close-to-basaltic composition, D.C. = 2.48, magnification (approximate): 1000X.
- B. Rugged (eroded) A. 12,033 dumbbell of brown, basaltic glass, 500X.
- C. Dumbbell of unusually low D.C. of 2.19 from A. 14,003 fines. The presence of droplets of nickel iron in the pale blue-green glass of this dumbbell may possibly indicate composition within the ultrabasic range, 500X.
- D. Dumbbell of unusually high D.C. of 2.9 from A. 14,003 fines. The extremely low relief in Kanada balsam of this dumbbell of colorless glass seems to be indicative of a close-to-felspatic or rhiolitic composition, 500X.
- E. Cylindrical spheroid of indochinite, the dumbbell constriction does not appear on the average indochinite with \( L/B = 3.0 \) as the mean of the D.C. is about 3.0; 1/2 natural size.
- F. Dumbbell from flame spray made of ground indochinite, D.C. = 2.56, 100X.
- G. Dumbbell from basaltic ash of Kileaua Iki, Hawaii, D.C. = 1.95, 10X.
- H. Dumbbell from rhiolitic ash, Haffelnaes, Iceland, D.C. = 2.53, 100X.

It was found that the constant increases (compare Fig. 1A and 1B) with the progressive roughening (presumably more eroded surface) of lunar dumbbells, a feature which can be geometrically derived (and demonstrated with HF...
Morphology of sprays.

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etching experiments) through equidimensional erosion of originally smooth dumbbells).

The mean of dumbbell constants of lunar sprays is 2.48 (see Fig. 1), that of microtectites estimated from photographs published by Dr. Glass (2) is approximately 2.8, and the constants of tectites measured at the British Museum average about 3.0. The foregoing figures seem to indicate that the dumbbell constant in general tends to increase with the SiO₂ percentage of the glass, from about 2.5 of the mean of lunar glasses with about 40% SiO₂ to 3.0 of the tectites with 70% SiO₂. The same general trend was observed in dumbbells (Fig. 1F-H), which are known of having rotated in the terrestrial atmosphere, such as volcanic ashes, flame sprays of lunar fines and terrestrial rocks and minerals. The constants of these dumbbells of terrestrial atmospheric rotational origin are shifted, however, about half a unit towards lower elongation ratios.

Gravitationally stretched basaltic hanging droplets with constricted waists were observed in icelandic lava caves, and similar structures were experimentally produced on melting of suspended chips of rocks in the flame of the blowpipe. The basaltic droplets produced under terrestrial gravitational stretching have breadth (B) close to 6 mm, and their dumbbell constants (when taking the distance between the tip and the waist of the droplet as L/2) are of intermediate value between the terrestrial rotated basaltic dumbbells and the lunar dumbbells of basaltic composition.

No regular macro- or micro-spray could be detected in the ashes from the Riesz crater, Germany, which is generally believed to be of meteoritic impact origin. Rather irregular dumbbells of unknown composition, described and illustrated from the meteorite impact crater of Sikhote-Alin, USSR, by Krinov (3), have constants in the region of 2.0. Dumbbells from industrial mineral shot described and illustrated by Cross (4) have constants around 3.0 which appears to be anomalously high for a glass with only 35.4% SiO₂. The composition of the glass is, however, far removed from that of normal rocks containing 34.9% CaO.

A systematic study of sprays may ultimately lead towards the distinction between sprays of incandescent cloud, of volcanic, or meteorite impact, of solar flare, and of other types of origins. Indications exist for the reconstruction of atmospheric pressure, of certain mechanical effects, and of the intensity of the gravitational field which prevailed in the course of formation of a given spray assemblage. Detailed knowledge of the entire range of morphological variability of sprays may enable us to distinguish between spray particles, organic microspheres, and micro-organisms in remote sensor photographs from Mars or other celestial bodies.

References
CHEMICALLY BOUND NITROGEN ABUNDANCES IN LUNAR SAMPLES, AND ACTIVE GASES RELEASED BY HEATING AT LOWER TEMPERATURES (250 to 500°C).

O. Müller, Max-Planck-Institut für Kernphysik, 69 Heidelberg, Germany

Chemically bound nitrogen concentrations have been determined in lunar samples from different Apollo landing sites. The major aim of this investigation was 1) to compare bound nitrogen content with total nitrogen amount and 2) to study the surface dependence of nitrogen in grain size fractions of Apollo 14 and Apollo 15 fines.

Nitrogen can occur in crystalline material and glasses in different forms either as gaseous, molecular $N_2$ or as bound nitrogen, the latter can be present as ammonium- and nitride-nitrogen.

In most nitrogen work of previous Apollo missions the total nitrogen content of lunar samples has been determined using a high temperature extraction technique (Moore et al. 1970, 1971). There exist only few measurements on chemically bound nitrogen (Hintenberger et al. 1970; Chang et al. 1971). Major differences in nitrogen content have been observed, especially for lunar basalts. Possible contamination of lunar samples by terrestrial atmospheric nitrogen has been suggested.

We have applied acid hydrolysis to convert chemically bound nitrogen to the $\text{NH}_4^+$-form. After a Kjeldahl distillation the ammonia was determined spectrophotometrically with Nessler's reagent. This technique discriminates against molecular nitrogen possibly present in lunar samples.

The results are compiled in Table 1. Bulk fines of Apollo 11, 12 and 14 have a chemically bound nitrogen content of 93, 80 and 92 ppm N, respectively. Apollo 15 fines 15101 and 15601 contain 109 and 80 ppm N, respectively. Compared to Moore's total nitrogen values of Apollo 11 fines (range 102 to 153 ppm N), and Apollo 12 fines (range 85 to 140 ppm N), our values are somewhat lower. Chemically bound nitrogen represents the major portion of total nitrogen.

To learn more about the solar wind nitrogen in lunar fines, we have separated fines 14003, 15101 and 15601 in grain size fractions and determined the chemically bound nitrogen content (see Table 1). The results illustrated in Figure 1 show a distinct increase of nitrogen by a factor of about four with decreasing grain size, i.e. nitrogen is clearly correlated with the grain surfaces and derives most likely from the solar wind. With increasing grain size the slope of the nitrogen curve turns sharply and becomes almost horizontal. This indicates a volume component of nitrogen which can be interpreted as indigenous lunar nitrogen.

The amount of bound nitrogen (131 ppm) in a dark colored chip of breccia 10046,1 is similar to total nitrogen in breccias 10002,54 and 12057,1 found by Moore et al.
Fragmental rock 14303,13 shows a relatively low nitrogen content of 31 ppm. Typical solar wind constituents such as light rare gases (Kirsten et al. 1971) and hydrogen (see below) are much less abundant in this rock than in lunar fines and breccia. Therefore, most of the nitrogen is probably not derived from the solar wind, but rather is indigenous. The vesicular basalt sample 10057 and the two crystalline rocks 12063 and 12075 differ significantly in chemically bound nitrogen content. The sample 10057 has a relatively high N value of 64 ppm, whereas in two Apollo 12 rocks practically no nitrogen was detected. We suggest that part of the gases present during the formation of the vesicles were ammonia and/or nitrogen which interacted with the magma forming chemically bound nitrogen. Recently, a chip of Apollo 15 vesicular basalt 15556 was analyzed. The N content is below 10 ppm.

It is known that the lighter solar wind noble gases are partly lost by sputtering and diffusion. Because nitrogen is chemically active, it should be less prone to such loss from surfaces and may thus be useful as a reference element for solar abundances. In Table 2, the abundance of chemically bound N in the 24 N fraction of 14003 is compared with the abundances of trapped solar wind noble gases (Kirsten et al.). Our measured 4He/N ratio (1.2) is lower by a factor of ~500 than the solar ratios of Cameron and Suess-Urey; our 20Ne/N ratio (0.02) is lower by a factor of ~50 than the solar ratio. We infer from these data that N is, as expected, more tightly bound than the noble gases and may thus be used as a reference.

The release of active gases by heating fines 14003, fragmental rock 14303 and fines 15101 at temperatures between 250 and 500 °C has been studied using gas chromatography. The predominant gases extracted by heating are H2, CO, CO2 and CH4. Minor amounts of C2H6, C2H4, C3H8 and C3H6 were only detected in the 350,400,450 and 500 °C experiments. In 14003 and 15101 solar wind hydrogen is the major component of the gases released. From 14303, however, only about one tenth of the hydrogen is evolved at the correspondent temperatures, indicating a much lower content of solar wind.

As in our previous work on Apollo 11 and 12 samples (Funkhouser et al. 1971), no molecular nitrogen was detected in 14003 and 14303 during the heating experiments. On heating fines 15101 only trace amounts of N2 were found. This is probably due to the high portion of chemically bound nitrogen the diffusion rate of which is negligible compared to that of molecular nitrogen.

To study trapped gases we have performed crushing experiments (Funkhouser et al.) on vesicular basalt 15556,25, on a small piece of a glass coated breccia and on a bubble containing glass cylinder (5x2 mm), both picked from coarse fines 14257. The eight chips of 15556 we obtained from NASA show open bubbles of mm size on the surface. Unfortunately, no interior vesicles are present as X-ray photographs have shown. On crushing a 400 mg chip a trace of methane (<10^-7 cc) was detected by gas chromatography. The two glass samples possibly also contain a trace of methane and no other active gases.

Furthermore, we have determined the concentrations of the alkali and alkaline earth elements, La and U in several Apollo 14 and Apollo 15 samples. We will report on the results elsewhere.
CHEMICALLY BOUND NITROGEN ABUNDANCES

O. Müller

Table 1. Chemically bound nitrogen (chem.b.N) concentrations in lunar bulk fines and grain size fractions, breccia and crystalline rocks determined as ammoniacal nitrogen by the Kjeldahl method. Error: ± 5 ppm.

<table>
<thead>
<tr>
<th>Sample Type and Number</th>
<th>Chem.b.N(ppm)</th>
<th>Sample Type and Number</th>
<th>Chem.b.N(ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fines 10084,31-bulk</td>
<td>93</td>
<td>Fines 15101,59-bulk</td>
<td>109</td>
</tr>
<tr>
<td>&lt; 48 µm</td>
<td>148</td>
<td>Grain &lt; 24 µm</td>
<td>272</td>
</tr>
<tr>
<td>Fines 12070,76-bulk</td>
<td>80</td>
<td>size 24-48 µm</td>
<td>113</td>
</tr>
<tr>
<td>&lt; 48 µm</td>
<td>92</td>
<td>fract. 48-60 µm</td>
<td>86</td>
</tr>
<tr>
<td>Grain &lt; 24 µm</td>
<td>194</td>
<td>60-109 µm</td>
<td>65</td>
</tr>
<tr>
<td>size 24-48 µm</td>
<td>226</td>
<td>Fines 15601,63-bulk</td>
<td>80</td>
</tr>
<tr>
<td>fract. 48-60 µm</td>
<td>149</td>
<td>Grain size &lt; 24 µm</td>
<td>157</td>
</tr>
<tr>
<td>60-109 µm</td>
<td>73</td>
<td>fract. 24-48 µm</td>
<td>83</td>
</tr>
<tr>
<td>Breccia 10046,1</td>
<td>131</td>
<td>Vesicular basalt 10057,80</td>
<td>64</td>
</tr>
<tr>
<td>Fragmental rock 14303,13</td>
<td>31</td>
<td>Micromagbro 12063,112</td>
<td>&lt; 10</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Basalt 12075,13</td>
<td>&lt; 10</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Vesicular basalt 15556,25</td>
<td>&lt; 10</td>
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</table>

Table 2: Solar wind nitrogen relation with noble gases

<table>
<thead>
<tr>
<th></th>
<th>4He/3He</th>
<th>3He/32Ar</th>
<th>4He/N</th>
<th>N/20Ne</th>
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<tr>
<td>Fines 14003,24</td>
<td>54</td>
<td>3.6</td>
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</tr>
<tr>
<td>&lt; 24µm</td>
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<td></td>
<td></td>
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<tr>
<td>Solar wind comp.</td>
<td>550</td>
<td>37</td>
<td>-</td>
<td>-</td>
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<tr>
<td>Apollo 14</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Geiss et al. 1971</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>A.G.W. Cameron</td>
<td>980</td>
<td>11</td>
<td>860</td>
<td>1.1</td>
</tr>
<tr>
<td>(1967)</td>
<td></td>
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<tr>
<td>Suess-Unev</td>
<td>394</td>
<td>67</td>
<td>466</td>
<td>0.8</td>
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<tr>
<td>(1958)</td>
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Acknowledgment. The collaboration of Mrs. S.Hasse and D. Kaether is gratefully acknowledged. I thank Dr. P. Horn and R. Schwan for making the grain size fractions.

The Rb-Sr internal isochron age of the Fra Mauro basalt 14310 is 3.93 ± 0.06 AE. The age of this sample resembles the high temperature 39Ar-40Ar ages of low K mare basalts from the Sea of Tranquillity, but is significantly greater than those of the mare basalts collected from the Apollo 12 and 15 missions. The high initial 87Sr/86Sr ratio (0.70033 ± 0.00019) suggests either derivation from source material which had a high Rb/Sr ratio since near the time of the moon's formation, or contamination with highly differentiated Rb rich materials at the time of volcanism. The abundances of the trace elements K, Rb and Ba are also distinct from those of the known mare basalts. From the combined trace element data and Rb/Sr systematics, the most reasonable interpretation leads to the conclusion that lunar basalts are in general derived from source regions of low Rb/Sr ratio, but are contaminated in varying degrees with early formed materials containing high abundances of K, Rb, Ba and other incompatible trace elements.

Three soil samples from the Apollo 14 mission also show high abundances of these trace elements relative to the soils from earlier Apollo missions. However, model ages of the Apollo 14 soils range from 4.4-4.5 AE., in agreement with model ages for soils obtained from other sites on the moon. A detailed study of various size fractions from soil sample 14149 shows that this high trace element abundance pattern is common to all size fractions. The element abundances of K and Rb show a spread of 20-50% but Sr and Ba vary by only about 10% in the various size fractions. From our leaching experiments and other studies of Apollo 11 and 12 soils, we had previously concluded that an exotic component containing highly radiogenic Sr and rich in K, Rb, Ba should be ubiquitously distributed in the lunar soils. This is confirmed in the case of the Apollo 14 soils. Further, this component is invariably present not only in different Apollo 14 soils, but in the various size fractions as well. If the exotic component had been transported over long distances and mixed with the soils, we would expect a grain size dependency, which in this case is not observed. A reasonable interpretation is that either a wholesale transport of material occurred in a collision event (Imbrium Basin excavation?) and that the soil was produced locally by lunar surficial degradation processes or that the exotic component was admixed into the Fra Mauro basalts at the time of their genesis and provided "in situ" source material for the formation of the soils.

A single sample of a mare basalt from the Hadley Rille site (15555) has
been dated at 3.30 AE by both a Rb-Sr internal isochron and K-Ar dating on a plagioclase separate from the sample. Both the trace element pattern and the Rb-Sr systematics in this sample are highly reminiscent of the Apollo 12 basalts. It is clear that extensive volcanism leading to mare flooding occurred in widely separated regions of the moon between about 3.1 and 3.9 AE ago. We hypothesize that the mare basalts may be products of the internal thermal regimen of the moon, whereas the basalts at Fra Mauro may be melts produced or triggered by the collisional event associated with the excavation of the Imbrium basin. This would explain the high abundance of elements such as K, Rb and Ba in the Fra Mauro basalts as resulting from the fusion of trace element poor surface layers and underlying material rich in these elements. In a model of early lunar differentiation, we had earlier proposed that the outer regions of the moon became mineralogically and chemically stratified at a very early time, producing an outer layer of the moon which graded from Fe and Ti rich materials in the lower parts to residual material enriched in lithophile elements near the top, overlain by an anorthositic crust. A major collisional event such as that postulated for the Imbrium basin could result in excavation and melting of material down to the trace element rich layer, and thus produce materials having the composition of the Apollo 14 basaltic rocks, as well as the exotic component present uniformly in the soils from the Fra Mauro region. If the exotic component is tapped from the pre-mare basement only in a major impact event such as the Imbrium collision, a corollary of this hypothesis would be that the exotic component in the soils from Apollo 11, 12 and 15 should reside preferentially in the smaller size fractions, as a consequence of transport from a distant source. Experiments to test this are currently underway and will be reported.

Magnetic properties of nine Apollo 14 lunar samples (2 fines, 2 crystalline rocks and 5 fragmented rocks) have been examined. The results of measurements are summarized in the following tables, where the notations of magnetic parameters are as follows: \( \chi \) initial magnetic susceptibility; \( \chi_a \) paramagnetic susceptibility; \( I_s \) saturation magnetization; \( I_R \) saturation remanent magnetization; \( H_c \) coercive force; \( H_{RC} \) remanence coercive force; \( I_n \) natural remanent magnetization (NRM); \( H_0 \) AF-demagnetization field to reduce NRM to e\(^{-1}\) of its initial value; \( H^* \) AF-demagnetization field to result in a deviation larger than 10° in angle of the remaining magnetization from its initial NRM direction; \( \theta \) Curie temperature; \( b \) coefficient of acquisition of isothermal remanent magnetization (IRM), where \( b = (IRM)/H^2 \). The types of viscous remanent magnetization (VRM) are defined in such a way that the viscous component is much smaller than the stable component in Type I while the viscous component is considerably larger than the stable one in Type II. All magnetic parameters in the tables are determined at 300°K. (See Refs. 1, 2 and 3.)

Some special points noticeable in the tables are as follows:

(a) Intense NRM: A crystalline rock, No. 14053, has an unusually intense NRM whose intensity amounts to \( 2 \times 10^{-3} \) emu/gm. The intensity of \( I_c \) of this sample also is unusually large for a lunar crystalline rock, indicating a metallic iron content of about 1%.

(b) Two types of viscous magnetic characteristics: Those samples of Type II VRM characteristics contain a considerable amount of very fine ggrains (about \( 10^2 \)A\(^0\) in mean diameter of metallic irons which behave superparamagnetically at room temperature. The superparamagnetic component is blocked at 50K. The Type II samples are mostly fines or fragmented rocks. In Type I samples, the magnetization is carried mostly by stable ferromagnetic particles. Most crystalline rocks belong to this category. (See Ref. 4.)

(c) FeNi alloy particles: The bulk of the native irons in three samples (14301, 14303, and 14311) are indentified as kamacite containing \( 3 \sim 8\% \) Ni. The ferromagnetic component in the other samples are almost pure metallic irons. Petrographic examination, electron probe measurements and Mössbauer analysis are in progress with the goal of developing a quantitative relationship between mineral content and magnetic characteristics.

(d) Thermal Demagnetization of 14053: Thermal demagnetization of the NRM of 14053 showed a remarkable decrease in the low blocking temperature range so
Table 1: Magnetic Properties of Fine Samples

<table>
<thead>
<tr>
<th>Magnetic Parameter</th>
<th>14259 (fines)</th>
<th>14161</th>
<th>14161 (selected coarse grains)</th>
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</thead>
<tbody>
<tr>
<td>$X_{o}$ x 10^{-3}</td>
<td>10.04</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>$X_{a}$ x 10^{5}</td>
<td>2.5</td>
<td>2.4</td>
<td>2.1</td>
</tr>
<tr>
<td>$I_s$</td>
<td>1.5</td>
<td>0.98</td>
<td>1.13</td>
</tr>
<tr>
<td>$I_R$ x 10^{2}</td>
<td>6.0</td>
<td>4.3</td>
<td>-</td>
</tr>
<tr>
<td>$H_C$</td>
<td>19</td>
<td>26</td>
<td>28</td>
</tr>
<tr>
<td>$H_{RC}$</td>
<td>300</td>
<td>430</td>
<td>-</td>
</tr>
<tr>
<td>$I_n$ x 10^{5}</td>
<td>-</td>
<td>-</td>
<td>4.48</td>
</tr>
<tr>
<td>$\theta^{(\circ C)}$</td>
<td>750</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>VRM</td>
<td>(II)</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Unspecified units are emu/gm for $X$ and $I$, Oe for $H$ and emu/(gm Oe^2) for $b$.

that by 350°C the NRM had decreased to less than 1/10 of the initial value. Preliminary results of thermal demagnetization of weak field TRM (1000 Y) showed somewhat similar behavior which may have important implications regarding the ancient lunar field. (See Ref. 4)

(e) NRM and its stability: NRM of 14063 is particularly stable against the AF-demagnetization. NRM of 14301 has the direction in situ of $D = 5.6^\circ$ westward from the North and $I = 39.5^\circ$ downward from the horizontal plane, which does not agree with the direction of the magnetic field measured at the Apollo 14 landing site.

(f) Magnetic properties of coarse fines, 14161: Although coarse fines, 14161, have a reasonable intense value of $I_s$ (about 1 emu/gm) as a whole, $I_s$ values of individual grains are much diverse, while $X_a$ values are practically the same in all individual grains.
Magnetism of Apollo 14 Lunar Materials
Takesi Nagata

Table 2: Magnetic properties of Rock samples

<table>
<thead>
<tr>
<th>Parameter</th>
<th>14047</th>
<th>14053</th>
<th>14301*</th>
<th>14303*</th>
<th>14310</th>
<th>14311*</th>
<th>14063</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\chi_o \times 10^3$</td>
<td>8.01</td>
<td>2.24</td>
<td>0.93</td>
<td>0.69</td>
<td>0.25</td>
<td>0.46</td>
<td>-</td>
</tr>
<tr>
<td>$\chi_a \times 10^5$</td>
<td>2.4</td>
<td>4.6</td>
<td>2.1</td>
<td>2.4</td>
<td>-</td>
<td>2.3</td>
<td>2.14</td>
</tr>
<tr>
<td>$I_s$</td>
<td>1.4</td>
<td>2.2</td>
<td>0.69</td>
<td>1.27</td>
<td>-</td>
<td>0.74</td>
<td>0.034</td>
</tr>
<tr>
<td>$I_R \times 10^2$</td>
<td>6.1</td>
<td>4.0</td>
<td>0.64</td>
<td>2.1</td>
<td>-</td>
<td>0.43</td>
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<td>(I)</td>
<td>(II)</td>
<td>(I)</td>
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Native irons in samples with (*) are kamacites.

References:

Vitrification experiments with terrestrial basalt and gabbro using various fusion techniques were conducted to simulate impact melting on the moon, and to compare spectral and chemical results arising from use of each technique. Fusion in molybdenum crucibles in inert gas yield silicate glass severely contaminated with Mo; the reflectance spectra show a very broad deep 1.1 μm and shallow 1.9 μm band and a sharp peak in reflectance at 0.6 μm indicating a strong UV absorption feature. Fusion in platinum crucibles, on the other hand, produces glasses of relatively high reflectance with broad shallow absorption bands near 1.1 μm and 1.9 μm. Rapid melting in vacuum at temperatures estimated at about 2500°C was achieved using a solar furnace with 2.89 m diameter parabolic collecting mirror focussed on a 5 cm diameter area of rock slab which also served as a crucible for the melt. This technique: (1) greatly reduces overall reflectivity, (2) increases R/B ratio, (3) reduces the depth and increases the width of the 1 μm band, and (4) produces a broad shallow structure near 1.9 μm. Chemical changes include increased Fe, Ti, and Mn concentrations, decreased Fe$^{4+}$/Fe$^{3+}$ ratio, and decreased K, Na, and Al concentrations. Spattering of the melt produces glass spherules similar in shape, size, and surface texture (no zap pits) to lunar spherules. Terrestrial gabbro with much less Ti and Fe than lunar material, when fused by this method, yields powder having a relatively flat reflectance spectrum with broad shallow bands at 1.1 μm and 1.9 μm and an albedo as low as typical dark mare fines. This experiment suggests that volatilization of less refractory elements may accompany relatively high temperatures achieved during (repeated) lunar impact melting, thus producing darker glasses, enriched in Fe and Ti, than would result from lower temperature (platinum) or closed system fusion experiments.

Spectral studies of mineral mixtures of various proportions of plagioclase, orthopyroxene and opaques (ilmenite) were made and compared with lunar sample spectra to establish dilution effects and to provide a basis for estimating mineral concentrations in lunar samples from spectra data. In general optical properties are not directly proportional to mineral abundance; e.g., 25% opaques reduces albedo of plagioclase by over 50% and 1% orthopyroxene in plagioclase produces a distinct 1 μm band. Albedo is dominated by plagioclase concentration: On ternary plots, lines of equal albedo and R/B ratio are more or less parallel to the orthopyroxene-opaque join; thus an unknown lunar material of given albedo or R/B could correspond to a wide range of compositions in the sample ternary system. Band-depth lines cross those of albedo and R/B, offering the possibility of esti-
mating mineral concentrations in this three-component mix. For example, the light grey fraction of lunar breccia 14063 compares best with a synthetic mix of about 2:1 plagioclase-pyroxene with a small amount of opaques. The reflectance of lunar fines 15531, with unusually strong 1μm and 2.1μm bands for a material of low albedo, compares best with a mix of about 1:1 pyroxene-opaques and possibly a small amount of plagioclase. The experimental result of Adams and McCord (1970) that opaques (including glass) depress albedo without removing pyroxene bands is thus confirmed.

Reexamination of the broad weak absorption band near 0.5μm in lunar rock spectra shows this structure to be composed of multiple, weak, closely spaced bands, the strongest of which are located near 0.55 and 0.51μm, with additional extremely weak depressions near 0.48 and 0.44μm. The composite nature of this feature rules out Ti⁴⁺ as the only contributing ionic species. An alternative interpretation is that the multiple structures arise from spin forbidden transitions in Fe⁺⁺ on distorted octahedral sites.
A STUDY OF THE NATURE OF THE GAS CLOUD
PRODUCED BY VOLCANIC AND
IMPACT EVENTS ON THE MOON
AND ITS RELATION TO ALKALI EROSION

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Hawaii Institute of Geophysics,
University of Hawaii,
Honolulu, Hi. 96822

The ejection of gases, volatiles and particulates from Hawaiian volcanoes has been evaluated as a model for the emission of hot vapor and ejecta clouds that might be produced during volcanic or impact events on the moon. From observations of the degassing from non-explosive lava fountains or from lava lakes, we have noted that volatiles are emitted in the initial stage as gases, which in part condense to particles ("blue haze") and ultimately to acidic steam clouds. Analyses of the main gaseous components in volcanic fountains have been accomplished using infrared techniques (1). Also recently it has been possible to collect volatiles directly from above lava fountains. Analyses of these samples gives us the composition of the volatile phase emanating from oceanic basaltic lava at the erupting temperature (circa 1200°C). Lunar basalts have been analyzed extensively so that the content of elements that would contribute to a volatile phase is known. From the elemental compositions of terrestrial lavas and volatiles, and of lunar lavas, then, by a simple proportionality one may arrive at an approximation for the elemental composition of a lunar volatile phase. This assumes that similar temperatures would obtain during lunar and terrestrial volatile emission, that the vaporizing molecules in the two systems would be similar, and that the partial vapor pressures would be proportional to the elemental fractional composition in the respective lava systems (Raoult's Law).

Volcanic volatiles have been studied successfully by treating them as an equilibrium system (2). If we assume that this same condition may obtain in the lunar situation, then from the elemental composition we can calculate the relative equilibrium concentrations of all likely volatile molecular compounds of the elements for which thermodynamic data is available—the 122 compounds in this case (3). This is done by finding the composition corresponding to the minimum free energy for this system using
Nature of Lunar Gas Cloud

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Well-known computer assisted procedures (4). The resulting calculated composition of a lunar vapor phase at 1500°K is shown in Table 1, limited here by space considerations only to major components. Similar calculations for terrestrial systems show that the same compounds do occur in the primary volcanic gaseous phase, though, of course, in very different proportions. It will be noted that a surprisingly large contribution to the gaseous phase under these conditions would be made by condensibles, which would produce about 70% of the total pressure. Wellman (5) has shown that this pressure within a vesiculating lunar lava might be as high as 100 atmospheres. CO, Ti compounds and S2 are the main contributors to the vapor, but oxides, halides, sulfides and elemental metals, and particularly the alkalis, are significant components. This last also has been shown in experimental studies of vaporization from lunar rocks (6).

The nature of the erosive attack by alkali metal vapor on polycrystalline terrestrial and lunar rocks has been studied further in an effort to elucidate the mechanism of this process. Scanning electron micrographs of samples exposed to attack by potassium give good evidence of a disruptive effect along grain boundaries, with resistance to attack by glass coated areas such as lined impact craters. This leads to pits which are elevated with respect to the surrounding areas, the so-called "pedestal pits" which have been noted on natural lunar rock surfaces (7). Other than this thus far nothing has been noted in the

Table 1. Mole fraction ($f_M$) of major equilibrium components of lunar vapor phase at 1500°K.

<table>
<thead>
<tr>
<th>Cpd.</th>
<th>Log $f_M$</th>
<th>Cpd.</th>
<th>Log $f_M$</th>
<th>Cpd.</th>
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Nature of Lunar Gas Cloud

J. J. Naughton

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morphology of surfaces attacked by potassium that would be uniquely characteristic or diagnostic in enabling one to identify a surface that has been subjected to alkali erosion.

The mechanism of release of volatiles into a vacuum from high temperature lava particles have been investigated by using Pele's tears as analogs of hot lava droplets and of the glass spheres of the lunar dust. Gaseous release occurs by decrepitation and, at higher temperatures in the softening range of the glass, by bursting of bubbles confined near the sphere surface. This latter action produces cratering in which the flow morphology usually allows the resulting craters to be differentiated from impact craters. Gases were released as pressure surges or spikes which were monitored mass spectrometrically. In terrestrial samples water vapor and CO$_2$ were the main gases identified in these surges.

References


In an earlier paper (Bloch et al. 1971) crater number densities of about 20 craters > 0.3 mm diameter and about 70 craters > 0.1 mm diameter per cm² have been reported to be on samples from the Apollo 11 and 12 missions. These crater counts have been extended on new samples for the same crater size range. The samples 12053.37, 14303.13 and 14305.36 have been scanned using optical binoculars. Similar results have been found on these samples.

One of the main emphases of this paper is the investigation of micron- and submicronsized craters by means of electron microscope technique (Stereo-scan). In these size ranges only extremely smooth surfaces are suitable for this type of investigation. Surfaces as glassy spheres or dumbbells, glass splashes and glass linings of small craters have been successfully scanned in a magnification up to 5000 x. The following results could be found for the samples 12024.8 (glass splash), 12063.41 (crater glass), from coarse fines 14257 (dumbbell), 14303.13 E2 (crater glass) and 14303.13 B2.11 (crater glass): 50-100 craters > 10 μm diameter, 1000-3000 craters > 1 μm diameter and up to 10000 craters > 0.25 μm diameter per cm² surface area.

These new number densities for micron- and submicronsized craters agree with and extend to smaller diameters the cumulative size-frequency distributions previously determined for craters > 0.1 mm diameter (Fig. 1), although counted on 3 different types of glass surfaces. This suggests that the craters have been produced predominantly by interplanetary micrometeoroids. Related simulation experiments have revealed a criterion for discriminating between high velocity (primary from micrometeoroids) and low velocity (mostly secondary from lunar debris) impact craters. Low velocity impact craters are usually shallow depressions without any rims. High velocity impact craters show a diameter to depth ratio of 2:1 with a rim, but do not exhibit any spallation patterns.

In the cumulative log crater frequency vs. log crater diameter diagram (Fig. 1) the slope in the micron- and submicronsized crater diameter distribution is -0.8 in contrary to the -2 slope of the equilibrium condition (Gault 1970). This deviation may have 2 reasons: a) besides erosion through continuous meteoroid bombardment, another erosion process by solar wind sputtering effects the number of submicronsized craters and/or b) there is a change in the meteoroid size frequency distribution and the surfaces may not have reached equilibrium condition (Hörz et al. 1971). The influence of solar wind sputtering in comparison with continuous meteoroid bombardment has been investigated in a recent paper (Neukum and Dietzel 1971). It was found that for micron- and submicronsized craters solar wind sputtering is the most dominant erosion process assuming a sputter erosion rate of 1 Å per year. Sputtering experiments on micronsized craters have demonstrated this erosion effect. Applying this re-
suit to flux calculations a mass exponent of -0.6 was found for the micron- and submicron-sized meteoroid flux in the interplanetary space at 1 AU. Samples showing a production state in certain diameter ranges of their crater statistics may be dated using these crater distributions. This dating method has been applied to sample 12024.8 (glass splash). The corresponding exposure age was found to be approximately 5000 years depending on the meteoroid flux used.

In order to compare this age with an independently determined age the particle track method was applied to the same sample. For glasses and minerals which were resting not deeper than 1 mm below the lunar surface, residence times at the very top of the surface can be calculated using the tracks produced by the heavy nuclei emitted in solar flares (Fleischer et al. 1971). These tracks are identified by their steep density gradient. The solar flare tracks were counted at different depths in the glass splash of sample 12024.8 and an orthopyroxene crystal trapped in the glass. For the glass and the orthopyroxene concordant solar flare exposure ages between 900 and 3500 years are found, depending on the fluxes of solar heavy nuclei used for the calculations (Crozaz et al. 1971, Price et al. 1971, Fleischer et al. 1971). This result is in agreement with the exposure age obtained from crater statistics.

The interdependence between crater diameter $D$ and particle mass $m$ may be described by $D = k m^{1/\beta}$ ($k, \beta$ constants). A differential particle flux $\varphi(m,t) = a m^{-\gamma} f(t)$, ($a, \gamma, f(t)$ constants, $f(t)$ function of time $t$) yields after a time $t$ a cumulative number $N$ of craters per unit area. For the crater population in any state of erosion by superposition of craters the theory (Neukum and Dietzel 1971) yields

$$N(m, t = 0, t_A) = \int \frac{\gamma^{-1} \frac{2\gamma^{-1}}{k_1 m^{1/\gamma} - \gamma + 1} \cdot m^{-\gamma} dm \left[1 - \exp \left(\frac{a k_1 m^{2/\gamma - 1}}{2k - \gamma + 1} \cdot (F(t) - F(0))\right)\right]}{d m}$$
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(kj constant, \( F(t) \) antiderivative of \( f(t) \)) at the time \( t=0 \) (today) for any age \( t_A \) of a given surface area \( A \). \( \overline{m} \) is the greatest mass which can have fallen on the given surface of the moon in the time \( t_A \) and is given by

\[
\int_0^\infty \int_{E_A} \varphi(m,t) \, dm \, dt = \frac{A}{\overline{m}}
\]

For small ages \( t_A \) this population simplifies to the production population

\[
N = N_\tau = \int_{tm}^{\omega} \alpha m^\theta (F(\omega) - F(t_A)) \, dm.
\]

The formula has been applied to crater populations in the diameter range \( > 3 \) km where the erosion by superposition is believed to be the dominant process. For a known crater population in any state of erosion the formation age \( t_A \) of the surface can be calculated if the particle flux is known. The time dependence of \( \varphi(m,t) \) has been determined earlier (Bloch et al. 1971). The production populations in the Lunar Maria are given by Baldwin (1970). Counts for the Southern Lunar Terrae have been performed by Hartmann (1966). The crater frequency of the far side of the moon has been determined by Neukum (1971). From these data mean formation ages \( t_A \) of lunar maria and highland areas are estimated to be between 3.3 and \( 4.6 \times 10^9 \) years.

References:

Rb-Sr relationships for some chemically defined lunar materials.

L. E. Nyquist, N. J. Hubbard and P. W. Gast, NASA Manned Spacecraft Center; H. Wiesmann, Lockheed Electronics Co., Houston; and S. E. Church, National Research Council Fellow, Houston, Texas 77058.

Rb and Sr systematics were determined for eight lithic fragments from the coarse fines of Apollo 11 (10085), Apollo 12 (12033), and Apollo 14 (14161), for bulk soils from Luna 16, Apollo 14, and Apollo 15 and for "whole rocks" from Apollo 14 and the Apollo 15 anorthosite sample 15415. These samples have been chemically characterized with respect to Na, Mg, K, Ca, Ba, U and REE by Hubbard et al. (1). This report deals primarily with representatives of the KREEP class of materials, including breccias 14006, 14068, 14301, 14307, crystalline rock 14310, and six fragments (27-55mg) from 14161. Analytical results are presented in Table 1.

In order to determine possible instrumental bias, five analyses of the C.I.T. seawater standard were made. Our results are biased with respect to those of Papanastassiou and Wasserburg (2), and Compston et al. (3) by about (+0.00012 ± 0.00007) and (-0.00011 ± 0.00007) respectively, and agree within the analytical uncertainty with those of Murthy et al. (4). Our results on the E. and A. standard are consistent with these biases but are less precisely defined.

Rb-Sr relationships for the Apollo 14 KREEP breccias are shown in Fig. 1. A least squares fit including all data yields an apparent age of 4.04 ± 0.11 b.y. resulting primarily from inclusion of two samples (14161,35,3 and 14006,3) with anomalously low Rb and K concentrations (3). While this apparent age is certainly somewhat in error because of including samples with differing initial Sr-87/Sr-86, it indicates the approximate time when K and Rb was lost from these samples. A self-consistent treatment of the data including only those three samples having anomalously high K/Rb ratios and low Rb concentrations (1) yields an apparent age of 3.88 ± 0.25 b.y. As this pattern of elemental concentrations has been qualitatively reproduced in a laboratory heating experiment (5) we suggest that loss of volatiles during a major brecciation event may have produced the observed Rb-Sr systematics.

An alternative hypothesis which could also explain the observations is heterogeneity of K and Rb on a mm to cm scale. A possible cause of this heterogeneity might be concentration of these elements in mineral assemblages of (K-feldspar + quartz) which occur in these breccias (6). The origin of these assemblages is uncertain but if they play a significant role in the Rb-Sr systematics the clock is most likely to be reset in breccias of high metamorphic grade. The three previously mentioned samples plus breccia 14068 are identified as high grade (7) and yield an apparent age of 3.9 ± 0.1 b.y. Although we are unable to unambiguously distinguish between the two alternatives we identify the 3.9 b.y. age with brecciation of these rocks and associated thermal mobilization/loss of volatiles occurring during the Imbrium event.
Rb-Sr RELATIONSHIPS FOR SOME CHEMICALLY DEFINED LUNAR MATERIALS
L. E. Nyquist

Model ages are given in Table 1 for the analyzed materials. The model ages of 14161,35,3 and 14006,3 are $T_{\text{BABI}} = 5.91 \pm 0.28$ and $4.93 \pm 0.10$, respectively, reflecting Rb losses of 60-70% and a resultant failure of these fragments to satisfy the closed system criteria. The mean model age for the other KREEP materials is $T_{\text{BABI}} = 4.44 \pm 0.22$ b.y. (2σ). Two stage model calculations for Sr-87/Sr-86 evolution between $T = 4.6$ b.y. and $T = 4.0$ b.y. for Apollo 14 KREEP materials show that the Rb/Sr ratio for the first stage was within a factor 2 of the present value.

As found by several investigators, Luna 16 soil is less radiogenic than other lunar soils. We find Sr-87/Sr-86 = 0.70025 ± 0.00005 and $T_{\text{BABI}} = 4.3 \pm 0.3$ b.y. For Apollo 15 soil 15101,90 we find $T_{\text{BABI}} = 4.62 \pm 0.24$. A preliminary analysis of anorthosite 15415 yields initial (Sr-87/Sr-86) = 0.6994 ± 0.00005.

Acknowledgements
We thank G. J. Wasserburg for providing the C.I.T. seawater standard and C. E. Hedge for providing the Elmer and Amend SrCO₃ standard. C. Meyer provided stimulating discussion, petrography, and microprobe analyses of some of the breccias. J. Warner classified the breccias according to metamorphic grade. E. Schönfeld provided programming assistance and helpful discussion. NASA provided financial support.

References
1. N. J. Hubbard, P. W. Gast, M. Rhodes, and H. Wiesmann, this volume.
5. E. K. Gibson and N. J. Hubbard, this volume.
7. J. Warner, personal communication.
Rb-Sr RELATIONSHIPS FOR SOME CHEMICALLY DEFINED LUNAR MATERIALS.

L. E. Nyquist

TABLE 1. Rb and Sr Analytical Results

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<tr>
<th>SAMPLE</th>
<th>wt. (mg)</th>
<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>$\frac{87Rb}{86Sr}$</th>
<th>$\frac{87Sr}{86Sr}$</th>
<th>$\frac{87Sr}{86Sr}$</th>
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<td>18.5 ± .5</td>
<td>188 ± 3</td>
<td>29.6 ± 1.0</td>
<td>0.71736 ± 56</td>
<td>4.45 ± .20</td>
<td></td>
</tr>
<tr>
<td>II. Bulk Soils</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14101,35.11</td>
<td>218.7</td>
<td>15.1 ± .2</td>
<td>185 ± 2</td>
<td>23.7 ± .3</td>
<td>0.71454 ± 22</td>
<td>4.54 ± .09</td>
<td></td>
</tr>
<tr>
<td>Luna 16</td>
<td></td>
<td></td>
<td></td>
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<td></td>
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<td></td>
</tr>
<tr>
<td>L6-43</td>
<td>15.47</td>
<td>1.82 ± .08</td>
<td>277 ± 3</td>
<td>1.90 ± .08</td>
<td>0.70763 ± 14</td>
<td>4.65 ± .58</td>
<td></td>
</tr>
<tr>
<td>L6-44</td>
<td>15.70</td>
<td>1.90 ± .08</td>
<td>275 ± 3</td>
<td>1.67 ± .08</td>
<td>0.70829 ± 12</td>
<td>4.40 ± .54</td>
<td></td>
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<tr>
<td>L6-45</td>
<td>21.49</td>
<td>1.87 ± .04</td>
<td>279 ± 3</td>
<td>1.94 ± .05</td>
<td>0.70929 ± 15</td>
<td>4.30 ± .40</td>
<td></td>
</tr>
<tr>
<td>L6-46</td>
<td>24.60</td>
<td>1.84 ± .04</td>
<td>323 ± 3</td>
<td>1.85 ± .05</td>
<td>0.70902 ± 5</td>
<td>4.23 ± .33</td>
<td></td>
</tr>
<tr>
<td>15101,90</td>
<td>53.7</td>
<td>4.32 ± .04</td>
<td>144 ± 1</td>
<td>9.1 ± 0.1</td>
<td>0.70914 ± 31</td>
<td>4.66 ± .24</td>
<td></td>
</tr>
<tr>
<td>III. Whole Rock</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14101,35.50-8</td>
<td>13.1</td>
<td>2.22 ± .07</td>
<td>98 ± 3</td>
<td>6.57 ± .29</td>
<td>0.70351 ± 35</td>
<td>4.67 ± .42</td>
<td></td>
</tr>
<tr>
<td>IV. Anorthosites</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>13081,11.47</td>
<td>3.05</td>
<td>0.27 ± .27</td>
<td>149 ± 6</td>
<td>0.32 ± .52</td>
<td>0.69097 ± 59</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14081,11.146</td>
<td>13.17</td>
<td>1.28 ± .17</td>
<td>159 ± 7</td>
<td>2.35 ± .32</td>
<td>0.70113 ± 110</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14083,97.97</td>
<td>34.17</td>
<td>2.91 ± .10</td>
<td>301 ± 8</td>
<td>2.47 ± .09</td>
<td>0.69568 ± 11</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14101,35.1</td>
<td>11.07</td>
<td>0.32 ± .08</td>
<td>162 ± 2</td>
<td>0.70 ± .17</td>
<td>0.69995 ± 23</td>
<td></td>
<td></td>
</tr>
<tr>
<td>15101,1</td>
<td>53.25</td>
<td>0.17 ± .02</td>
<td>177 ± 2</td>
<td>0.28 ± .03</td>
<td>0.69935 ± 18</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

$\frac{87Rb}{86Sr}$ and $\frac{87Sr}{86Sr}$ are normalized to 88Sr/86Sr = 8.3762.

FIGURE 1.
CONCENTRATIONS OF PRIMORDIAL RADIOELEMENTS AND COSMOGENIC RADIO-
NUCLIDES IN APOLLO 15 SAMPLES BY NONDESTRUCTIVE GAMMA-RAY SPECTROMETRY,*
G. Davis O'Kelley and James S. Eldridge, Oak Ridge National Laboratory, Oak
Ridge, Tenn., 37830; E. Schonfeld, NASA Manned Spacecraft Center, Houston,
Texas 77058; K. J. Northcutt, Oak Ridge National Laboratory, Oak Ridge,
Tenn., 37830.

Nine samples from Apollo 15 were analyzed by nondestructive gamma-ray
spectrometry. The first samples (15016 and 15101) were received from the
Lunar Receiving Laboratory (LRL) about nine days after liftoff from the moon.
Other samples were received at intervals beginning about two weeks later.

The equipment and techniques of nondestructive gamma-ray spectrometry
are essentially those we developed for use during the Apollo 11 and Apollo 12
missions at the LRL. Spectrum libraries for all samples except 15455
were obtained from replicas which accurately reproduced the electronic and
bulk densities of the lunar samples. Bulk densities also were obtained from
the known sample weights and measured volumes of the replicas. Results are
listed in Tables 1 and 2.

As we have shown previously (1,2), mass ratios K/U for the earth and
moon fall into separate groups whose average values appear characteristic of
each planet. The constancy of the ratio of K (a volatile element, depleted
on the moon) to U (a refractory element) appears to be the result of early
chemical fractionation which is not affected by later igneous processes.

Basalts 15015, 15475, and 15495 have potassium concentrations which are
similar to those of the Apollo 11 low-K basalts and the Apollo 12 basalts.
However, the Apollo 15 basalts have average K/U ratios of ~9000, somewhat
higher than the K/U ratio of ~2000 for Apollo 11 and 12 basalts of the same
potassium concentration. The range of K/U values for eucrites slightly overlaps
the data region for the Apollo 15 crystalline rocks.

As was the situation for the Apollo 12 samples (2), the soil and breccia
from Apollo 15 are much higher in K, Th, and U than the Apollo 15 basalts.
Therefore, the soils and breccia are very likely mixtures of Apollo 15 basalt
and a foreign component high in K, Th, and U. Our K/U systematics suggest
that the most likely candidate for the foreign component is the lunar material
KREEP (3). A two-component mixture requires only about 10-20% KREEP for the
Apollo 15 breccia and soil samples, lower than the range of KREEP concentra-
tions estimated for Apollo 12 samples (2). Sample 15601 and the dark portion
of 15455 are similar in their primordial radionuclide concentrations and are
lowest in KREEP (~10%) of the Apollo 15 samples we have examined.

The general concentration patterns for 22Na and 26Al resemble those
observed on previous Apollo missions. Because chemical analysis data are
lacking for most of the samples in Table 2, it is not possible to make de-
tailed interpretations of the concentrations of 22Na and 26Al. However, the
low ratio 26Al/22Na for 15475 suggests that the 26Al content did not attain
Concentrations in Apollo 15 Samples
G. Davis O'Kelley

saturation; this implies ejection onto the lunar surface within the last $10^6$ years. The cosmogenic radionuclide yields, especially of $^{56}$Co, suggest that the rocks were collected from the surface, in agreement with preliminary documentation.

The radioactivity of the soil samples can be related to depth effects. The high concentrations of $^{22}$Na, $^{26}$Al, and $^{56}$Co are consistent with a sampling depth of about 3 cm for 15101 and a somewhat shallower depth of 1.5-3.0 cm for 15601, assuming a chemical composition similar to that of Apollo 12 soils. Similarly soil 15031, taken from a 35 cm depth at the ALSEP site (Station 8), shows the low levels of cosmogenic radioactivity characteristic of deep samples. The radioactivity of 15041, taken from the top of the trench at Station 8, is consistent with sampling within the top 3 cm of the surface.

During the preliminary examination of the Apollo 11 and Apollo 12 samples in the LRL, we were able to detect in six rocks the $^{48}$V produced by solar-flare protons. From the $^{48}$V content of rock 12062, which appeared to have been buried, we inferred a yield from galactic proton bombardment of about $40 \pm 20$ dpm/kg Fe.

We determined $^{48}$V quantitatively in the first two Apollo 15 samples received; however, the results on 15016 were superior, because weak components of the gamma-ray spectra of lunar basalts suffer less interference from the Th and U decay series than spectra of lunar soils and breccia. The concentration of $^{48}$V in 15016 leads to a galactic production rate for $^{48}$V of $57 \pm 11$ dpm/kg Fe, based on an FeO concentration (4) of 22.6%. This agrees well with our earlier estimate within the experimental errors.

Although no measurements of the intensity of the solar flare of 25 January 1971 have been published, the $^{56}$Co concentrations in Table 2 suggest that this flare was slightly more intense than the well-characterized event of 3 November 1969.

Research carried out under Union Carbide's contract with the U. S. Atomic Energy Commission through interagency agreements with the National Aeronautics and Space Administration.

Concentrations in Apollo 15 Samples
G. Davis O'Kelley

Table 1. Concentrations of Primordial Radionuclides in Apollo 15 Samples.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Weight (g)</th>
<th>Density (g/cm³)</th>
<th>K (ppm)</th>
<th>Th (ppm)</th>
<th>U (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Crystalline Rocks</td>
<td></td>
<td></td>
</tr>
<tr>
<td>016</td>
<td>924</td>
<td>2.4</td>
<td>374 ± 20</td>
<td>0.52 ± 0.02</td>
<td>0.15 ± 0.01</td>
</tr>
<tr>
<td>475</td>
<td>288</td>
<td>2.9</td>
<td>354 ± 20</td>
<td>0.40 ± 0.02</td>
<td>0.12 ± 0.01</td>
</tr>
<tr>
<td>495</td>
<td>909</td>
<td>2.9</td>
<td>495 ± 25</td>
<td>0.60 ± 0.03</td>
<td>0.16 ± 0.01</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Breccias</td>
<td></td>
<td></td>
</tr>
<tr>
<td>285</td>
<td>251</td>
<td>2.4</td>
<td>1610 ± 80</td>
<td>3.4 ± 0.1</td>
<td>0.93 ± 0.05</td>
</tr>
<tr>
<td>455</td>
<td>833</td>
<td></td>
<td>900 ± 150</td>
<td>2.0 ± 0.3</td>
<td>0.53 ± 0.08</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Fines</td>
<td></td>
<td></td>
</tr>
<tr>
<td>031</td>
<td>100</td>
<td></td>
<td>1860 ± 95</td>
<td>4.3 ± 0.2</td>
<td>1.1 ± 0.05</td>
</tr>
<tr>
<td>041</td>
<td>100</td>
<td></td>
<td>1740 ± 90</td>
<td>4.0 ± 0.2</td>
<td>1.1 ± 0.05</td>
</tr>
<tr>
<td>101</td>
<td>116</td>
<td>~1.2</td>
<td>1484 ± 74</td>
<td>3.1 ± 0.3</td>
<td>0.86 ± 0.08</td>
</tr>
<tr>
<td>601</td>
<td>204</td>
<td>~1.7</td>
<td>900 ± 45</td>
<td>1.8 ± 0.2</td>
<td>0.51 ± 0.05</td>
</tr>
</tbody>
</table>

Table 2. Concentrations (dpm/kg) of Spallogenic Radionuclides in Apollo 15 Samples. Decays Corrected to 1711 Hours GMT, 2 August 1971.

<table>
<thead>
<tr>
<th>Sample</th>
<th>²²Na</th>
<th>²⁶Al</th>
<th>⁴⁶Sc</th>
<th>⁴⁸V</th>
<th>⁵⁴Mn</th>
<th>⁵⁶Co</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Crystalline Rocks</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>016</td>
<td>29 ± 2</td>
<td>82 ± 4</td>
<td>3 ± 1</td>
<td>10 ± 2</td>
<td>31 ± 4</td>
<td>16 ± 3</td>
</tr>
<tr>
<td>475</td>
<td>32 ± 3</td>
<td>40 ± 3</td>
<td>3 ± 2</td>
<td></td>
<td>23 ± 3</td>
<td>11 ± 5</td>
</tr>
<tr>
<td>495</td>
<td>29 ± 3</td>
<td>69 ± 3</td>
<td>3 ± 1</td>
<td>trace</td>
<td>25 ± 2</td>
<td>11 ± 2</td>
</tr>
<tr>
<td>Breccia and Fines</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>285</td>
<td>50 ± 4</td>
<td>85 ± 4</td>
<td>3 ± 2</td>
<td></td>
<td>30 ± 5</td>
<td></td>
</tr>
<tr>
<td>031</td>
<td>33 ± 3</td>
<td>49 ± 3</td>
<td></td>
<td>40 ± 10</td>
<td></td>
<td></td>
</tr>
<tr>
<td>041</td>
<td>57 ± 4</td>
<td>99 ± 7</td>
<td>3 ± 2</td>
<td>33 ± 10</td>
<td>7 ± 5</td>
<td></td>
</tr>
<tr>
<td>101</td>
<td>44 ± 5</td>
<td>120 ± 12</td>
<td>&lt; 4</td>
<td>9 ± 6</td>
<td>28 ± 8</td>
<td>11 ± 6</td>
</tr>
<tr>
<td>601</td>
<td>55 ± 6</td>
<td>112 ± 11</td>
<td>&lt; 4</td>
<td></td>
<td>32 ± 8</td>
<td>28 ± 9</td>
</tr>
</tbody>
</table>

Apollo 14 fines (14421, 17) subjected to metabolic tests for detecting viable organisms show chemical reactions which result in gas changes. The metabolic tests employed were $^{14}$CO$_2$ fixation, $^{14}$CO$_2$ evolution from $^{14}$C-labeled organics and gas exchange as measured by gas chromatography. The techniques were essentially that earlier reported, Oyama et al (1971), but the medium employed here was the M-4 medium for Viking Mission Application (Oyama -- in preparation).

Results of $^{14}$CO$_2$ absorption prior to determination of fixation are shown in Table 1.

<table>
<thead>
<tr>
<th>Period of Incubation (Day)</th>
<th>Solution from</th>
<th>CPM</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Autoclaved</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Unheated</td>
<td></td>
</tr>
<tr>
<td>0-14</td>
<td>57,000</td>
<td>400,000</td>
</tr>
<tr>
<td>14-28</td>
<td>98,000</td>
<td>370,000</td>
</tr>
<tr>
<td>28-42</td>
<td>5,000</td>
<td>65,000</td>
</tr>
</tbody>
</table>

* 2.5356 grams  ** 2.6697 grams

Incubation conditions: Approximately 10 mm $^{14}$CO$_2$ containing 3.5 x $10^7$ CPM generated from carbonate in 1 atmosphere of N$_2$ before each period; room temperature; 5 mls of M-4 medium in contact with lunar material through filter paper wick; medium and $^{14}$CO$_2$ freshly added before incubation period.

Less $^{14}$CO$_2$ was absorbed by the autoclaved sample solution than the unheated sample solution and absorbed $^{14}$CO$_2$ decreased during the 28-42 day period for each of the two samples. The pH of the media showed that more than 90% of the $^{14}$CO$_2$ will be absorbed and, therefore, the difference between the amount of $^{14}$CO$_2$ available (ca 3.5 x $10^7$ CPM) and the counts in the solution must be that absorbed by the wet lunar sample.

Results of the $^{14}$CO$_2$ evolution experiment show that the evolved $^{14}$CO$_2$ decreased with time and more $^{14}$CO$_2$ was evolved from the unheated lunar sample than from the autoclaved sample (Table 2).
Search for Viable Organisms in Lunar Samples
V. I. Oyama

TABLE 2. $^{14}$CO$_2$ Evolved by Lunar Samples Incubated With $^{14}$C-labeled Medium.

<table>
<thead>
<tr>
<th>Period of Incubation (Day)</th>
<th>Autoclaved Sample **</th>
<th>Unheated Sample ***</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CPM</td>
<td></td>
</tr>
<tr>
<td>0-14</td>
<td>21,580</td>
<td>31,635</td>
</tr>
<tr>
<td>14-28</td>
<td>2,600</td>
<td>3,005</td>
</tr>
<tr>
<td>28-42</td>
<td>655*</td>
<td>1,325</td>
</tr>
</tbody>
</table>

* Tube broken in handling at 28 day. Only portion of sample could be transferred to new tube for 28-42 day period.

** 2.4017 g
*** 2.6048 g

Incubation conditions: At room temperature; under N$_2$ at 1 atmosphere; 5 ml of $^{14}$C labeled nutrient medium added only at beginning contact with sample through filter paper wick; fresh KOH trap for collection before each period. $^{14}$CO$_2$-labeled supplement contains: DL-alanine-$1^{-14}$C, DL-glutamic acid-UL, glycine-UL, DL-lysine-$1^{-14}$C, DL-methionine-$^{14}$CH$_3$, Na acetate-$1^{-14}$C, and glycerol-UL (total CPM/tube was 9 x 10$^6$).

These results are not comparable to results obtained for samples of Apollo 11 & 12 (Oyama et al., 1971), where the rate of $^{14}$CO$_2$ evolution was constant for the entire period (8 weeks). If it is assumed that the rate of radio-autodegradation is constant, then the Apollo 14 fines must be capable of producing larger quantities of basic substances over a longer period of time to account for the reduction in CO$_2$ evolved.

Data for the gas exchange experiment are shown in Table 3.

TABLE 3. Gas Changes Over Apollo 14 Fines Incubated at Room Temperature

<table>
<thead>
<tr>
<th>Initial Incubation Atmosphere</th>
<th>Analyzed Mass of Fines</th>
<th>Period of Incubation (Day)</th>
<th>Incubation Condition</th>
<th>Change in Headspace Gas (Micromoles).</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>0 - 17</td>
<td>17 - 32</td>
<td>32 - 47</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Dry</td>
<td>0.5 ml</td>
<td>2.0 ml</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>medium</td>
<td>medium</td>
</tr>
<tr>
<td>Hydrogen</td>
<td></td>
<td>$^{+}$28.9</td>
<td>$^{+}$74.2</td>
<td>$^{+}$81.6</td>
</tr>
<tr>
<td>3.7996 g</td>
<td>CO$_2$</td>
<td>-2.67</td>
<td>-29.9</td>
<td>-38.0</td>
</tr>
<tr>
<td>Oxygen</td>
<td></td>
<td>$^{+}$0.079</td>
<td>$^{+}$0.121</td>
<td>$^{+}$0.155</td>
</tr>
<tr>
<td>3.7666 g</td>
<td>O$_2$</td>
<td>-6.4</td>
<td>-17.6</td>
<td>-20.6</td>
</tr>
<tr>
<td></td>
<td>CH$_4$</td>
<td>0</td>
<td>0</td>
<td>$^{+}$0.058</td>
</tr>
<tr>
<td></td>
<td>CO$_2$</td>
<td>-0.65</td>
<td>-13.4</td>
<td>-22.1</td>
</tr>
</tbody>
</table>

1. 20% H$_2$, 2% Kr, 2% CO$_2$, 76% He.
2. 1% O$_2$, 2% Kr, 2% CO$_2$, 95% He.
3. Gas exchange M-4 medium without nitrate.
Search for Viable Organisms in Lunar Samples
V. I. Oyama

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The data show that molecular oxygen reacted with dry lunar matter, but with added moisture the process was accelerated. The amount of hydrogen gas produced was dependent upon the removal of oxygen. More carbon dioxide was absorbed in the hydrogen atmosphere than in the oxygen atmosphere indicating that oxides are not easily converted to hydroxides. Oxidation was noted by reddening of solutions exposed to oxygen. Methane gas appeared sooner under hydrogen than under oxygen.

The results of these tests are consistent with the oxidation of metals such as iron by molecular oxygen or by water. The latter reaction produces molecular hydrogen and the hydroxide of the metal; the hydroxide product in turn absorbs the carbon dioxide present. The appearance of methane on the other hand, is more likely from either a release of occluded solar methane or methane resulting from the hydrolysis of carbides.

The generation of hydrogen and methane and the consumption of molecular oxygen nearly parallel biologic responses from terrestrial soils. However, every soil which produces hydrogen, also consumes this hydrogen. Further, they produce carbon dioxide beyond the capacity of even alkaline soils to retain it. The biological generation of molecular hydrogen from this medium is accompanied by the formation of ferment acids and carbon dioxide in contrast to the non-biological production of hydrogen and alkaline substances.

ACKNOWLEDGEMENT

We thank M. P. Silverman for consultation on procedures, and E. Munoz, C. Johnson and O. Whitfield for assistance.

REFERENCE

LUNAR ASH FLOWS: HOW THEY WORK; S. I. Pai and T. Hsieh, Institute for Fluid Dynamics and Applied Mechanics, University of Maryland, College Park, Maryland 20742; and J. A. O'Keefe, Laboratory for Space Physics, Goddard Space Flight Center, Greenbelt, Maryland 20771.

On the lunar surface, a dustless gas promptly expands and vanishes into the thin atmosphere; a gasless dust describes a set of ballistic trajectories and stops. But a gas-dust mixture will behave like a pseudo-fluid, consisting of a pseudo-gas on top (the dilute phase) and a pseudo-liquid underneath (the dense phase) (O'Keefe and Adams, 1965). Disentangling the dust from the gas takes hours of days for particles of the size of lunar fines, and a cloud size of a few kilometers. The present study makes these general concepts quantitative and provides models of lunar and terrestrial ash flows with consistent and physically reasonable values for the upward velocities, pressures, temperatures and scale heights for the gas and for the solid particles under a wide variety of conditions.

It is found that:

a) On the moon, unlike the earth, a dilute phase will always be present above the dense phase. (Pai and Hsieh, 1971)

b) The temperature in a lunar flow, unlike that on the earth, varies greatly with position in the flow; the upper portions of the flow are extremely cold because of the expansion of the escaping gases.

c) As a result of the low surface temperature, the dust cloud has a sharply-defined upper boundary, above which the gas density drops in a short distance from values near atmospheric to a few hundredths of a torr. (Fig. 1)

d) The solid particles are even more strongly concentrated near the base of the flow. This makes it possible to believe that a flow of this kind could carve a sinuous rille such as the Hadley Rille (Cameron, 1964; Schumm, 1970)

It is suggested that the lunar soil and breccia were chiefly produced by ash flows (possibly impact-induced) rather than by mechanical grinding and turnover (gardening). (McKay et al., 1971; Waters et al., 1971; Fredriksson et al., 1971)
Lunar Ash Flows: How They Work

Pai, S.I.

P.594

REFERENCES


Fig. 1  Pressure $\rho$ as a function of height $y$ above the lunar surface. The unit of $\rho$ is of the order of $10^{-1}$ cm$^{-2}$, the unit of $y$ is of the order of 15 meters.
MICROPROBE INVESTIGATIONS OF THE OXIDATION STATE OF Fe AND Ti IN ILMENITE IN APOLLO 11, APOLLO 12, AND APOLLO 14 CRYSTALLINE ROCKS,
M. Pavicevic, P. Ramdohr, A. El Goresy, Max-Planck-Institut für Kernphysik 69 Heidelberg, Germany

The oxidation states of iron and titanium in the Apollo samples has been a subject of concern in several lunar investigations. The presence of metallic iron in the Apollo 11 basalts and Fe Ni alloys in the Apollo 12 and 14 crystalline rocks suggests that iron is present in the divalent state in the Fe-Ti oxide minerals. However, Haggerty and Meyer (1) suggested the presence of Fe in some members of the chromite-ulvospinel series present in the Apollo 12 basaltic rocks. The principal aim of the present investigation was to develop a quantitative technique for the direct experimental determination of the oxidation state of iron and titanium in simple binary systems and to apply this technique to determine the oxidation states of these elements in opaque minerals present in the lunar samples.

In an attempt to apply the electron microprobe for semiquantitative determination of the oxidation state of Fe and Mn, Albee and Chodos (2) performed several measurements of the L spectra of these elements present in oxides and silicates at 15 kV. O’Nions and Smith (3) recently presented new results on the L and L spectra of several iron and titanium oxides over a wide range of excitation potentials and have defined several of the experimental and theoretical difficulties involved in this technique.

Our investigations of the L and L spectra were conducted using pure synthetic standards in the solid solution series ilmenite-hematite, ulvospinel-magnetite, and ferropseudobrookite-pseudobrookite. The measurements were expanded to lunar ilmenites and ulvospinels. Our probe investigations of the Fe L and L spectra indicate that any change in the oxidation state of iron among members of a solid solution series influence the spectra in the following manner: 1) As the total amount of Fe increases, the peak position shifts continuously, but nonlinearly toward shorter wave lengths. 2) The L/L peak intensity ratio decreases almost linearly with the increase in the oxidation state. 3) There is an

Fig. 1
apparent wave length shift of the spectra to higher wave lengths upon increase of the excitation potential.

The nonlinearity of the wave length shift (point 1) with change in the oxidation state of iron disqualifies this phenomenon from further quantitative evaluation. However, measurements of the $L\beta/L\alpha$ intensity ratios (point 2) of several members of a solid solution series over a wide range of excitation potentials allows the construction of calibration curves for the determination of the oxidation states of Fe in minerals belonging to the same series. Figures 1 and 2 show such calibration curves for compounds along the FeTiO$_3$-Fe$_2$O$_3$, and Fe$_2$Ti$_4$O$_9$-Fe$_3$O$_4$ joins, respectively. These calibration curves, however, can only be used for members of the binary series - i.e. the presence of high concentrations of impurities such as MgO lowers the $L\beta/L\alpha$ intensity ratio and leads to meaningless results. Fig. 1 shows the variation of the $L\beta/L\alpha$ ratio for Ilm$_{100}$-Hem$_0$, Ilm$_{96}$ - Hem$_4$, and Ilm$_{100}$ - Hem$_0$, and Fig. 2 shows these peak ratios for Usp$_{100}$ - Mt$_3$, Usp$_{80}$ - Mt$_{10}$, and Usp$_{20}$ - Mt$_{100}$. The distance between the curves in Fig. 2 is a direct function of the amount of the Fe$_2$O$_3$ molecule present. The Fe $L\beta/L\alpha$ ratios for lunar ilmenites together with the values for the standards used, as measured at 10 kV are compiled in Table 1.

Table 1

<table>
<thead>
<tr>
<th>Composition</th>
<th>$L\beta/L\alpha$ ratio</th>
<th>S</th>
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</thead>
<tbody>
<tr>
<td>1. Ilm$_{100}$</td>
<td>0.857</td>
<td>± 0.024</td>
</tr>
<tr>
<td>2. Ilm$_{96}$ - Hem$_4$</td>
<td>0.842</td>
<td>± 0.024</td>
</tr>
<tr>
<td>3. Hem$_{100}$</td>
<td>0.867</td>
<td>± 0.028</td>
</tr>
<tr>
<td>4. 10047</td>
<td>0.855</td>
<td>± 0.026</td>
</tr>
<tr>
<td>5. 12063</td>
<td>0.857</td>
<td>± 0.030</td>
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</table>

The results presented in Table 1 indicate that the iron in the ilmenites studied in Apollo 11 rock 10047, Apollo 12 rock 12063, and Apollo 14 rock 14053 is present as Fe$^{2+}$. The $L\beta/L\alpha$ intensity ratios of the lunar ilmenites agree within the experimental error with the values for pure synthetic ilmenite Fe$_2$Ti$^{4+}$O$_3$. Alée and Chodos give a $L\beta/L\alpha$ value of 0.84 obtained at 15 kV for both pure ilmenite and ulvospinel (Fig. 4 in their paper); however, our values for ilmenite and ulvospinel at 15 kV are 0.89 and 0.675, respectively. Our measurements also indicate that the
Oxidation State of Fe and Ti

M. Pavicevic

LB/La variation curve for ulvospinel is distinctly lower than the ilmenite curve. The calibration curves for members of the Fe₂TiO₄-Fe₃O₄ join cannot be applied for the majority of the lunar ulvospinels, since they contain appreciable amounts of the chromite molecule. Only one ulvospinel with low concentrations of Cr, Mg, Mn, V, and Al in Apollo 12 microgabbro 12063 was found to have a LB/La value close to that of Usp100 (El Goresy et al., 4).

We have also measured the TiLa and LB spectra in the lunar and synthetic ilmenites, and several terrestrial ilmenites using a focused clinochlore crystal (2d = 28.6 A). With one exception we have found no indication of wave length shifts or intensity variations, despite the drastic compositional differences in these ilmenites. We believe that a little to no change in the Ti valence state takes place in compounds along the join Fe₂TiO₃ - Fe₂O₃, i.e. titanium is present as Ti⁴⁺ in these compounds. Therefore measurements of Ti LB/La intensity ratios for these compounds results in an average value for the presence of 100% Ti⁴⁺.

The Ti LB/La intensity ratios in ilmenites of the same approximate composition should be identical if, and only if, Ti is solely present as Ti⁴⁺. However, as shown in Table 2, measurement of several ilmenites in Apollo 14 sample 14053, 2 resulted in LB/La ratios which are significantly higher than values obtained from the synthetic standards or Apollo 11 and 12 ilmenites. These higher values can only be interpreted as due to the presence of Ti³⁺. However, it would be difficult in our present state of development to estimate the amount of Ti³⁺ present. This first determination of the presence of Ti³⁺ in lunar minerals is, perhaps, not so surprising when other mineralogical features of rock 14053 are considered. El Goresy et al. (5) have recently described several observations made on this rock which led them to conclude that rock 14053 crystallized under more reducing conditions than any Apollo 11 or 12 rocks. That is, in an environment conducive to the formation and preservation of Ti in the trivalent state.

Acknowledgment: We are grateful to Don Lindsley and E. Woermann for supplying us with synthetic Fe-Ti oxides.

References:

The natural remanent magnetization of igneous samples returned from the moon by the Apollo 11, 12 and 14 missions is weak compared to similar terrestrial rocks, and is generally between $10^{-6}$ and $10^{-4}$ emu/gm. This is at least in part due to the relative abundance of ferromagnetic phases contained in the rocks - approximately .1% native iron in the lunar case as against several percent magnetite in typical earth basalt.

This lunar igneous NRM consists, in most cases, of two distinct components, an unstable component readily removed upon alternating field (AF) demagnetization in fields less than 100 oe, and a second component which is still recognizable after demagnetization in fields up to 400 oe. The unstable component is the stronger and its behavior to demagnetization can be duplicated by an isothermal remanent magnetization induced in the sample by a magnetic field of 10 to 20 oe. This component is thus considered to be an IRM acquired during or after return to earth.

The more stable component with an intensity of about 1 to 2 x $10^{-6}$ emu/g is probably a thermoremanent magnetization due to cooling from above 800°C in the presence of a field of a few thousand gammas. When several chips were analyzed from the same rock, they had the same direction of magnetization after AF demagnetization (that is, the stable component has a consistent direction within an individual rock).

Many breccias respond to AF demagnetization in much the same way as do the igneous rocks, although they may carry a very strong VRM associated with superparamagnetic grains or large unstable multidomain grains. As an example of this, Figure 1 shows the behavior of sample 14313,25. Curve A in the intensity plot and stereographic projection represents the decay of the NRM. Curve B refers to the decay of a VRM acquired after storage in the earth's field for several days. Curve C shows the decay of a magnetization remaining after the sample had been subjected to a field of 10 oe. and then stored in a field-free space for several days. In each case the demagnetization curve is similar to that for the NRM and all direction change curves converge to a stable position. It would seem that the NRM consists of a stable component of probable lunar origin and a stronger unstable component more likely of terrestrial origin.

Breccia 14321 does not show this easily acquired VRM and since several adjacent pieces were available for study, an AF demagnetization and a thermal demagnetization were performed on different pieces (Figure 2). An NRM component stable to 400 oe. and to 700°C has been isolated and found to be consistent in direction in the two pieces. This degree of stability suggests that the magnetization is a thermoremanent magnetization and that it is carried by single domain and small multidomain particles of metallic iron.
It would appear that some breccias carry similar paleomagnetic information as the igneous samples. Thus we may have in them a record of the moon's magnetic field beyond the age span of the igneous rocks.

Thus far laboratory tests have been discussed that demonstrate the stability of magnetization against field and temperature, but a test for stability against time would be important if possible. Such a test, the conglomerate test, is possible assuming the lunar regolith to be a conglomerate with an age of about one million years, set by the tumbling rate caused by meteorite impacts. The direction of magnetization with respect to the lunar orientation for the lunar samples should be random if the magnetization is older than one million years. Figure 3 illustrates the distribution of inclinations (only reliable inclination data is available for many of the oriented samples from Apollo 11 and 12) for 10 samples. The dashed lines represent the expected distribution for randomness. The goodness of the fit shows that the magnetization is of considerable age.

The anomaly profile provided by the Apollo 15 subsatellite magnetometer appears to be correlated with many of the craters flown over. At 110 km altitude these anomalies ranged up to 1 gamma maximum. In the particular case of the crater Van de Graaf, the anomaly can be simulated in shape and magnitude by an appropriate sized model crater with associated material carrying a remanent magnetization of $2 \times 10^{-6}$ emu/gm, a value typical of the stable intensities for the returned samples, both igneous and fragmental. This would seem to indicate that the lunar crust contains large areas of uniformly magnetized material. Further, the magnetizing field would be lunar-wide rather than local.

In summary our data imply that the moon experienced a magnetic field lasting at least from 4.0 b.y. to 3.0 b.y. ago, which represents the age range of Apollo samples and their probable last major heating. This history can be extended if other samples, particularly breccias, can be found with well determined ages outside this range.
REMANENT MAGNETIZATION OF LUNAR SAMPLES

G. W. Pearce

Figure 2

Figure 3
K-Ar DATING OF LUNAR SOILS: APOLLO 12, APOLLO 14 AND LUNA -16

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In a preliminary report on rare gases in Apollo 12 soils (1), it was noted that all soils examined - which included surface, subsurface, and breccia samples - fall closely along one or the other of two linear correlation lines on a plot of $^{40}$Ar vs. $^{36}$Ar. The two lines differ in slope by almost a factor two, revealing two clearly distinct populations of "trapped" or surface-correlated $^{40}$Ar/$^{36}$Ar at the site, but intersect at a common point on the ordinate: $^{36}$Ar = 0, $^{40}$Ar = 2000 x $10^{-8}$ ccSTP/g. Assuming that this volume correlated component is in situ produced radiogenic $^{40}$Ar, and taking the average K concentration in Apollo 12 soils as - 2700 ppm, the K-Ar age of this suite of Apollo 12 soils is $-1.3 \times 10^9$ yr., ignoring possible diffusive loss of Ar. A more detailed analysis, which however still assumes a single component system, yields an age of $1.30 \pm 0.20 \times 10^9$ yr.

Hubbard et al. (2) have derived a two-component mixing model of KREEP with average Apollo 12 basalt to account for variations in major and trace element chemical composition among the Apollo 12 soils and breccias. Application of two-component systematics to K-Ar dating of Apollo 12 soils, with endpoint K concentrations of K(KREEP) = 6640 ppm and K(basalt) = 530 ppm (2), yields the following chronology: t(KREEP) = $0.95 \pm 0.05$ x $10^9$ yr. and t(basalt) = $2.8 \pm 0.2 \times 10^9$ yr. K and radiogenic Ar in surface soil 12032, trench soil 12033 and breccia 12034, and core samples 12025-62 (~ 10 cm), 12028-101 (~ 21 cm) and 12028-140 (~ 37 cm) are consistent with a mixing curve between these endpoint ages. Surface soils 12001 and 12044 (Ar data from Hintenberger et al. (3)), 12042 and 12070 are tightly grouped along a mixing curve between KREEP and a second, significantly older basaltic component with age $4.0 \pm 0.2 \times 10^9$ yr. in this model. There is therefore evidence that the Apollo 12 soils are mixtures of materials with at least three distinct K-Ar ages. The values of these ages, approximately 0.95, 2.8 and $4.0 \times 10^9$ yr. as calculated here by the bulk K-Ar method, are based on the assumptions of complete outgassing of radiogenic Ar accumulated prior to the outgassing events, and of quantitative retention of radiogenic Ar following the events.

The K-Ar age of $0.95 \pm 0.05 \times 10^9$ yr. for the KREEP component in these soils is in quite good agreement with the "third event" ages of $0.8-1.0$ x $10^9$ yr. derived from U-Pb measurements on Apollo 12 materials (4,5,6,7); the age may be taken as the formation age of the crater Copernicus on the assumption that KREEP is material from the Copernican ray which crosses the Apollo 12 site. The agreement between the K-Ar and U-Pb ages suggests that the age is not an artifact of diffusive loss of either Ar or Pb. However, loss of Ar from the basaltic components of the soils is quite possible; for

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example, the apparent age of $2.8 \pm 0.2 \times 10^9$ yr. may represent either an Ar-gassing event or about 30% Ar loss from Apollo 12 basaltic rock debris with crystallization age near $3.2 \times 10^9$ yr. (8,9).

$^{40}\text{Ar} - 36\text{Ar}$ correlation of Ar from $\sim 1\mu$ to $150\mu$ grain size separates of Apollo 14 soil 14259 ($< 1$mm) defines an intercept age of $\sim 2.8 \times 10^9$ yr., which could define an episodic outgassing event. However, K-Ar ages of the three grain size separates for which K contents are known (V. R. Murthy, personal communication) correlate inversely with K, suggesting a two-component mixture with different end-point ages. Data presently on hand are insufficient to define either K contents or ages of the components. However, there is evidence that the Apollo 12 KREEP soil component is present in this Apollo 14 soil. The coarse, 150-1000$\mu$ grain size fraction from 14259 lies far off the correlation defined by the $<150\mu$ fractions; assuming that its trapped (40/36) ratio is that characteristic of KREEP-rich Apollo 12 soils, its apparent K-Ar age is $0.98 \times 10^9$ yr., and it falls precisely on the mixing curve between 0.95 and $2.8 \times 10^9$ yr. defined by Apollo 12 soils.

Grain size separations and gas analyses have been carried out on small samples from both levels of the Luna-16 core. Ar correlation diagrams yield intercept ages of $2.8 \pm 0.4 \times 10^9$ yr. and $4.1 \pm 0.4 \times 10^9$ yrs. for the $\sim 6$ cm and $\sim 30$ cm levels respectively.

K-Ar intercept ages calculated from data for Apollo 11 soils 10084 (3, 10) and 10087 (3) are $3.8 \pm 0.2 \times 10^9$ yr. and $3.1 \pm 0.4 \times 10^9$ yr. respectively. Taking the results of Ar analyses of soils from all four sites together, it appears that there may be evidence for rather large-scale outgassing events on the lunar surface at times very roughly $2.8$ and $4 \times 10^9$ yrs. ago, and a rather more precisely dated and more local event, apparently the formation of Copernicus, just under $1 \times 10^9$ yrs. ago.

REFERENCES

K-Ar DATING OF LUNAR SOILS: APOLLO 12, APOLLO 14 and LUNA-16
R. O. Pepin

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Far infrared spectroscopic investigations and Raman light scattering measurements have been made on a large variety of lunar samples and terrestrial minerals. These studies were mainly conducted at room temperature and covered the frequency range 2000-20 cm\(^{-1}\) (5-500 microns in wavelength). This region encompasses all the optically active natural vibrational frequencies of the internal and lattice modes found in lunar and most terrestrial materials.

The infrared specular reflectance spectra of all the polished bulk samples and pressed discs of the fines were compared with the reflectance of a freshly aluminized mirror. The reflectance data were subjected to a Kramers-Kronig analysis from which the real (\(\varepsilon'\)) and imaginary (\(\varepsilon''\)) parts of the dielectric dispersion were obtained. The analysis also provided values of the refractive index (\(n\)) and the extinction coefficient (\(K\)) from 5-500 microns.

The numerical frequencies of the infrared active modes of vibration that are characteristic of the chemical constituents of the materials are the frequencies corresponding to peaks in the imaginary part of the dielectric function. These are listed in Table 1 for the lunar materials currently investigated and a few examples of terrestials are shown for comparison purposes.

Transmission techniques were also employed on the fines. Approximately 2-3 mg/cm\(^2\) of sample was suspended in a KBr matrix and two rather broad spectral bands were observed at \(\sim 400\) cm\(^{-1}\) and 900 cm\(^{-1}\) corresponding to 'silicate' bending and stretching regions respectively. A thin layer of the powder material on a conducting substrate was used to obtain the frequency where the transmission passes through a maximum. This frequency is called the Christianson frequency (\(\omega_C\)) and corresponds to the state where \(n \approx 1\) and \(K \approx 0\). It has been found in terrestrial materials that \(\omega_C\) varies with the degree of basicity and the values obtained in the lunar samples are listed in Table 1. \(\omega_C\) was also obtained from the Kramers-Kronig analysis and the frequencies were generally in reasonable agreement (\(\pm 5\) cm\(^{-1}\)) with the transmission data. From Table 1 it can be seen that \(\omega_C\) varies from \(\sim 1270-1200\) cm\(^{-1}\) in the lunar samples investigated. This region characteristically lies between the terrestrial intermediate rocks like Andesite and Syenite and the ultra-basic rocks like Chondrites, Peridotites and Olivene. As well as providing the optical constants, the Kramers-Kronig analysis also gave values of the dielectric constant (\(\varepsilon_\infty\)) at \(\sim 10^{11}\) Hz. These are also listed in Table 1. It would appear that generally the values of the dielectric constant measured by other researchers (e.g., D. Chung et al.) at frequencies less than \(10^6\) Hz are approximately a factor of two larger than our results. As this discrepancy is

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outside the experimental errors it would appear that there may be other relaxation mechanisms or free carrier absorption effects occurring between \(10^6\) and \(10^{11}\) Hz.

The Raman studies performed in this work principally utilized three laser exciting lines namely 6328Å, 5145Å and 4880Å with an oblique-incident back scattering geometry. The overall success of these studies have been restricted mostly to the bulk materials due to the inherent weakness of the Raman effect and the high level of elastically scattered light (especially from the fines). However, identifiable spectra have been obtained from individual crystalline grains or inclusions in about ten lunar samples by direct comparison with the spectra of terrestrial minerals. Spectra corresponding to olivene, ilmenite, pyroxene and quartz grains could be distinguished using the 50μ diameter probe of the laser beam. Spectral bands characteristic of the olivene bearing basalts have been observed in 12009,48 and 14301,20. In samples 10058,56; 12065,115; 12073,42; 14301,20; 14310,76; and 14310,184 ilmenite and pyroxene grains give well defined bands and quartz bands are observed in 12009,48; 14320,76; 14310,184 and 14321,98.

The small diameter (∼50μ) laser spot allows quite small grains to chemically and mineralogically identified. This analytical method can be contrasted with complementary infrared measurements. In the latter case the spectra are more representative of the distribution of the major constituents throughout the bulk material. As such these results should provide significant information for comparison with possible future infrared remote sensing measurements of the lunar surface.
Table 1
Comparison of the infrared vibrational frequencies, Christianson frequency ($w_c$) and $\varepsilon_0$ at $\sim 10^{11}$ Hz for the lunar samples and some terrestrial materials

<table>
<thead>
<tr>
<th>Lunar Sample</th>
<th>$w_c$ cm$^{-1}$</th>
<th>Vibrational frequencies (cm$^{-1}$) ($\pm$ 10 cm$^{-1}$)</th>
<th>$\varepsilon_0$ at $10^{11}$ Hz</th>
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</thead>
<tbody>
<tr>
<td>12008,23</td>
<td>1260</td>
<td>1210(w) 1040(w) 890(m) 770(w) 620(w) 550(w) 470(m) 390(m)</td>
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<td>12065,115</td>
<td>1240</td>
<td>1065(?) 990(m) 920(s) 725(w) 580(w) 530(m) 380(m) 240(w)</td>
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<tr>
<td>12073,42</td>
<td>1240</td>
<td>1100(w) 970(w) 830(w) 730(w) 570(w) 530(w) 480(w) 400(w)</td>
<td>2.7</td>
</tr>
<tr>
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<td>1240</td>
<td>1070(w) 920(m) 800(w) 730(w) 620(w) 490(m) 400(w) 350(w)</td>
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<tr>
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<td>14259,38</td>
<td>1225</td>
<td>1040(w) 825(w) 730(w) 530(w) 470(w) 395(w) 245(w)</td>
<td>2.4</td>
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<tr>
<td>12002,186</td>
<td>1220</td>
<td>1080(w) 1010(w) 980(w) 930(m) 830(w) 760(w) 520(m) 400(m) 340(m) 250(w)</td>
<td>3.9</td>
</tr>
<tr>
<td>10058,56</td>
<td>1220</td>
<td>1160(w) 1070(w) 995(w) 930(m) 710(w) 630(w) 605(s) 505(s) 340(m) 330(m) 275(w)</td>
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<td>14313,51</td>
<td>1215</td>
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<td>14321,98</td>
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<td>14161,36</td>
<td>1200</td>
<td>1020(w) 720(w) 530(w) 470(w) 240(w)</td>
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<tr>
<td>Orthoclase (sanidene)</td>
<td>1250</td>
<td>1100(w) 985(m) 945(m) 630(w) 580(w) 465(w) 410(m)</td>
<td>3.4</td>
</tr>
<tr>
<td>Pyroxene (augite) ±20</td>
<td>1200</td>
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<tr>
<td>Olivene</td>
<td>1170</td>
<td>1030(w) 900(m) 610(w) 510(s) 410(s) 370(m) 300(m)</td>
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<tr>
<td>Ilmenite</td>
<td>1140</td>
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<td>9.6</td>
</tr>
</tbody>
</table>
RADIATION DAMAGE IN SOILS FROM FIVE LUNAR MISSIONS: P.P. Phakey, I. D. Hutcheon, R.S. Rajan and P.B. Price, Dept. of Physics, Univ. of Calif., Berkeley, CA 94720

1. Optical microscopic comparison of radiation damage in young or active areas and in maria:

<table>
<thead>
<tr>
<th>Sample</th>
<th>Site</th>
<th>Quat. track density at ctr. of grains (cm^-2)</th>
<th>Frac. of grains with &gt;10^8/cm^2 at edge of grain gradients</th>
<th>Fraction of grains with optical gradients</th>
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<tbody>
<tr>
<td>14141</td>
<td>Conelet</td>
<td>5 x 10^6</td>
<td>.32</td>
<td>.41</td>
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<tr>
<td>15301</td>
<td>Spur</td>
<td>-10^8</td>
<td>.7</td>
<td>?</td>
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<tr>
<td>15271</td>
<td>Apennine</td>
<td>3 x 10^7</td>
<td>.78</td>
<td>.58</td>
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<td>15501</td>
<td>Scarplet</td>
<td>5 x 10^7</td>
<td>.9</td>
<td>.86</td>
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<table>
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<th>Sample</th>
<th>Site</th>
<th>Quat. track density at ctr. of grains (cm^-2)</th>
<th>Frac. of grains with &gt;10^8/cm^2 at edge of grain gradients</th>
<th>Fraction of grains with optical gradients</th>
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<tr>
<td>10084</td>
<td>Tranquil.</td>
<td>&gt;10^8</td>
<td>.96</td>
<td>?</td>
</tr>
<tr>
<td>12025</td>
<td>Procell.</td>
<td>&gt;10^8</td>
<td>.88</td>
<td>?</td>
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<td>14259</td>
<td>Fra Mauro</td>
<td>&gt;10^8</td>
<td>high</td>
<td>?</td>
</tr>
<tr>
<td>15006</td>
<td>Hadley</td>
<td>&gt;10^8</td>
<td>.95</td>
<td>?</td>
</tr>
<tr>
<td>Luna 16</td>
<td>Fecund.</td>
<td>&gt;10^8</td>
<td>1.0</td>
<td>?</td>
</tr>
</tbody>
</table>

Track densities in soils from rims of fresh craters (e.g. Bench; a 10 m crater near the rim of Cone; a 10 m crater near Scarp) and on sloping ground (Apennine front) are considerably lower than in soils on level mare ground. Using the analysis of Arrhenius et al. (1971), we infer an age of ~5My for "Conelet," ~50My for Scarplet and ~100My for Spur. Because of gradual soil mixing the latter two ages are quite uncertain, but at least the order is the same as that assigned by the field geology team on the basis of crater morphology.

A large fraction of grains from Conelet, the Apennine front and Scarplet, including those several 100, in diameter, have strong track density gradients on all surfaces showing that they were irradiated isotropically (last column of table). The track densities in the other soils are too high to be studied optically. We draw two inferences from these observations: a) Since the soils from Conelet, the Apennine front and Scarplet were collected from sloping surfaces (>10°), the isotropic irradiation may have resulted from a surface transportation down the slope, driven by gravity and perhaps assisted by
Electrostatic charging. b) The case for a parent-body origin of the gas-rich meteorites should now be taken more seriously. The weight of current opinion is that this class of meteorite formed by accretion of grains in space. The strongest evidence is the existence of isotropically irradiated grains inside the gas-rich meteorites. Our observations show that isotropic irradiation commonly occurs on a parent body as well as in space.

2. High voltage electron microscope observations of radiation damage in "micron-size" soil grains.

<table>
<thead>
<tr>
<th>D.P.</th>
<th>Strong spots</th>
<th>Weak spots; diffuse</th>
<th>No spots</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Tracks or blobs</td>
<td>Low contrast</td>
<td>Amorphous</td>
</tr>
<tr>
<td>10084</td>
<td>55%</td>
<td>29%</td>
<td>16%</td>
</tr>
<tr>
<td>12028</td>
<td>43%</td>
<td>39%</td>
<td>18%</td>
</tr>
<tr>
<td>14259</td>
<td>29%</td>
<td>45%</td>
<td>26%</td>
</tr>
<tr>
<td>15501</td>
<td>65%</td>
<td>24%</td>
<td>11%</td>
</tr>
<tr>
<td>15000</td>
<td>60%</td>
<td>27%</td>
<td>12%</td>
</tr>
<tr>
<td>41 cm</td>
<td>65%</td>
<td>15%</td>
<td>20%</td>
</tr>
<tr>
<td>83</td>
<td>64%</td>
<td>12%</td>
<td>24%</td>
</tr>
<tr>
<td>124</td>
<td>67%</td>
<td>21%</td>
<td>12%</td>
</tr>
<tr>
<td>207</td>
<td>75%</td>
<td>12%</td>
<td>12%</td>
</tr>
<tr>
<td>246</td>
<td>75%</td>
<td>16%</td>
<td>9%</td>
</tr>
<tr>
<td>L-16</td>
<td>7%</td>
<td>33%</td>
<td>60%</td>
</tr>
<tr>
<td>7 cm</td>
<td>9%</td>
<td>23%</td>
<td>68%</td>
</tr>
<tr>
<td>30 cm</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The table divides micron-size grains into 3 categories on the basis of relative radiation damage: a) Those with good single-crystal electron diffraction patterns. Track densities are invariably high, ranging from \(-10^9/cm^2\) to un-resolvably high (>3 x 10^{11} /cm^2). The latter show a mottled pattern of blobs and the diffraction spots are fragmented. b) Those with faint diffraction spots on a diffuse background and low contrast in the micrographs. c) Those with no diffraction spots and an amorphous structure that we attribute to intense radiation damage. The Luna 16 soil had a strikingly higher radiation exposure than did the others (Phakey and Price, 1971).

Very few grains from either the 7 cm or 30 cm layer of Luna 16 showed good, crystalline diffraction patterns and individually resolvable tracks. Most had been converted into nearly or completely amorphous grains. A possible explanation of the long surface radiation exposure of the Luna 16 grains is that, being at high longitude near the east limb, they were not directly exposed periodically to the energetic electrons in the earth's magnetospheric tail which, laboratory experiments (T. Gold, private communication) have shown, may stir and transport the soil by electrostatic charging. On this model the radiation damage would be concentrated near the surface of Mare Fecunditatis but would extend to a greater depth at the other landing sites.

3. Depth dependence of radiation damage in the Apollo 15 deep core.

In the graph below, the micron-size grains monitor the solar wind and low energy (<1 MeV/nucleon) solar flare particles; the 50-micron grains monitor the multi-MeV/nucleon flare particles. Our observations indicate that any gradient in radiation damage is slight in the upper 250 cm of the mare at the
Apollo 15 site. At the greatest depth we find that the micron-size grains typically contain $\geq 10^{10}$ tracks/cm$^2$. A statistical mixing model (Comstock et al, 1971) faces difficulties in reproducing the shallowness of the observed depth gradient, whereas a "successive burials" model would be consistent with our data.

It would be extremely interesting to obtain a deep core from a site near the back side of the moon (e.g. Marius Hills) and see if the radiation damage falls off more steeply than in the Apollo 15 core.

References


THE SEARCH FOR AMINO ACIDS IN THE APOLLO 12 AND APOLLO 14 SAMPLES.
Cyril Ponnamperuma, Laboratory of Chemical Evolution, University of Maryland, College Park, Md. 20742, Charles Gehrke, Experimental Station Chemical Laboratories, University of Missouri, Columbia, Missouri 65201, Keith Kvenvolden, NASA, Ames Research Center, Moffett Field, Ca. 94035.

A cardinal tenet of the hypothesis of chemical evolution holds that molecules of importance to life could be synthesized in the abiotic milieu. The availability of samples from the lunar surface has given us an opportunity of testing this statement which is the cornerstone of our studies in exobiology.

In this report we present the results of our investigations relevant to one class of such compounds, the amino acids, which play a fundamental role in the terrestrial biosphere. Three sets of experiments were performed. The first involved the Apollo 12 sample 12023. The second used the Apollo 14 sample 14298. In the third investigation the Apollo 14 SESC sample 14240 was studied.

In the first two investigations the extraction procedures were identical. Approximately 1 gram of material was treated with water for 24 hours at 100°C. Any amino acids present in the extracts were converted to the N-trifluoro acetyl n-butyl esters prior to gas chromatographic analysis. Compounds having retention times of glycine, alanine, serine, aspartic, and glutamic acid appeared to be present in concentrations of 3 - 4 ng/gram in the case of glycine and less than 1 ng/gram of each of the others.

In the case of the Apollo 14 SESC sample, 6 grams were extracted with water. The water extract was examined by both gas chromatography and ion exchange chromatography. Both techniques showed that compounds having the retention times of glycine, alanine, and aspartic acid, glutamic acid were present. As in the previous experiments the concentrations were 3 - 4 ng in the case of glycine and less than 1 ng of each of the others.

It must be concluded that in the absence of mass spectrometric data, these identifications appear to be inconclusive. Furthermore, the indigenous nature of these materials cannot be established until the amino acids identified can be separated into their enantiomers.
CARBON IN THE APOLLO 14 SAMPLES

Cyril Ponnamperuma

The nature, quantity, and isotopic composition of carbon in an Apollo 14 sample (14259) were determined by the techniques of pyrolysis, hydrolysis, and crushing. Aliquots of approximately 1 gm, 2 gm, and 3 gm, were used in our experiments.

The total carbon in the sample was determined by complete combustion and had a value of 128 micrograms per gram with a carbon isotopic abundance $\delta^{13}C$ (relative to the Pee Dee Belemnite standard) of $+11$ per mil. In comparison the Apollo 12 sample 12023 had a value of 110 micrograms per gram with a $\delta^{13}C$ figure of $+12$.

In three sets of experiments, the pyrolysis temperature was varied in a stepwise manner up to 1100°C. Variations of the $\delta^{13}C$ value indicated the presence of carbon with more than one range of isotopic values from $-5.7$ to $+11.9$. In one determination the condensable material on pyrolysis at 400°-600° gave a $\delta^{13}C$ value of $+31.8$. The solar wind "hydrogen stripping" mechanism discussed previously may be one of the factors causing this wide variation in the $\delta^{13}C$ value of the assemblage of carbon atoms in the mineral matrix.

The results of the crushing and hydrolysis experiments were, in general, similar to those we reported for the Apollo 12 fines.
PREDOMINANT TRAPPING OF SOLAR-FLARE GASES IN LUNAR SOILS
G. Poupeau, J.L. Berdot, G.C. Chetrit, and P. Pellas
C.N.R.S., and Lab. Mineralogie du Museum, Paris, France

Assuming crystals and lithic fragments (grain-size: 100-600 μ) having high cosmic-ray track densities in their centers (>10⁸ cm⁻²) are solar-flare irradiated fragments (Track-Rich Fragments = TRF), then there is a correlation between the percentages of TRF and Ar³⁶ content for bulk soils. For each soil sample reported here, 80 to 140 fragments have been counted. Our data (underlined), and data from the literature, are shown in Fig. 1. A distinct correlation is also reported for gas-rich achondrites (straight line on the left side of Fig. 1). If one considers that the light rare gases in Pesyanoe aubrite are almost purely of solar-wind origin, it is of some interest, after appropriate corrections (median grain size, % TRF), to normalize the solar gas content of lunar soils to the maximum meteoritic values. The following ratios are obtained for a grain size of 100 μ using the data of Eberhardt et al. (1):

<table>
<thead>
<tr>
<th></th>
<th>Ne²⁰</th>
<th>Ar³⁶</th>
<th>Ar³⁸</th>
</tr>
</thead>
<tbody>
<tr>
<td>10084 ilmenite/Pesyanoe</td>
<td>~21</td>
<td>~7</td>
<td>~8</td>
</tr>
<tr>
<td>10084 bulk soil/Pesyanoe</td>
<td>~23</td>
<td>~58</td>
<td>~56</td>
</tr>
</tbody>
</table>

We believe the largest fraction of these "excess" gases was solar-flare implanted. However, this does not explain the large Ar³⁶ and Ar³⁸ "excesses" observed in the bulk soil as compared to the ilmenite. Another process therefore must be considered in order to explain these "excess" amounts of Ar isotopes. We suggest spallation reactions with calcium as the main target by solar-flare particles and high-energy neutrons emitted by the Sun 4.0-4.6 AE ago. According to this hypothesis, the Ca-rich material of the lunar highlands was strongly irradiated. Thus a soil containing a larger fraction of highlands material shows larger "excesses" of Ar isotopes. To check this working hypothesis, we have computed the Ar⁴⁰ "excess" for a number of soils, taking into account the weighted average of their different components (2,3,4,5,6). Our own data are reported in Fig. 2, along with other data from the literature (7,8).

In the 114 fragments studied, we have been unable to find even one fragment of Luna 16 (C-119, 20-22 cm depth in the core) which was not TRF. Correspondingly, Luna 16 soil shows the largest Ar⁴⁰ "excess", although it has the lowest K content of all the soils studied to date. Conversely, 12032 and 12033 show...
the smallest fraction of TRF (3 and 17%, respectively) and the smallest amount of trapped solar gases. If it is assumed that soil 12032 has been essentially irradiated by galactic cosmic rays, then an "exposure age" of 700-800 Myr is found. For the same soil, assuming that all the Ar\(^{40}\) is radiogenic in K-REEP, a bulk K-Ar age of 1.5 AE is found, whereas Pepin and Nyquist (9) have computed an age of 0.95±0.05 AE for K-REEP in 12033. Thus the lower limit (exposure age) and the highest limit (all Ar\(^{40}\) is radiogenic) bracket an age of 1.2±0.4 for the impact event which formed the K-REEP fragments. This latter value (within the error limits stated) is the same as that obtained by Hartmann (10) using lunar cratering chronology for Copernicus. For Apollo 14 and 15 soils we have no rare gas data yet, except 14259 (11).

In Fig. 2, we have thus assumed that the Imbrian impact (3.9 AE) governs the age of the noritic component. In such a case, our data (in Fig. 2) have to be considered as "predicted values" for the Ar\(^{40}\) "excess".

It is of interest to note that Funkhouser et al. (12) in their study of Apollo 11 and 12 soils and breccias have shown that the Ar\(^{40}\) "excess" follows the trend: Apollo 11 breccia > (Luna 16 soil) > Apollo 11 soils > Apollo 12 soils > Apollo 12 breccia and 12033 and 12032 soils. Summarizing this sequence reflects: 1) the early intense cratering rate which produced the Apollo 11 breccia; 2) the early intense flux of solar particles (protons, neutrons) which irradiated the highlands material in which strong turnover and stirring processes must have acted; 3) the large fractions of highly irradiated highlands material contained in Luna 16 and Apollo 11 breccia and soils due to their proximity to nearby highlands.

If our hypothesis is correct, then differences in isotopic K ratios produced by bombarding solar particles should be found in Apollo 11 breccias or in the finest grain size of Luna 16 soil. If these differences are found, then the strange mechanisms proposed to explain the so-called "excess" Ar\(^{40}\) in lunar soils (1,13,14) would have to be rejected.

3) Apollo Soil Survey (1971), ibid. 12, 49.
5) Jakes et al. (1972), Earth Planet. Sci. Lett. 12, 257.
6) Wood et al. (1971), Meteoritics 6, 181.
12) Funkhouser et al. (1971), preprint.
14) Manka and Mitchel (1970), Science 169, 278.
PREDOMINANT TRAPPING ...

G. Poupeau

FIG. 1 \( \text{Ar}^{36} \times 10^{-4} \text{ cc. STP/g} \)

% of TRF \((10^9/\text{cm}^2)\)

FIG. 2 LUNAR SOILS (grain-size < 1 mm)

LUNA 16

APOLLO 11

(APOLLO 11 et al. 1971)

AP. 15

Imbrium: 3.9 By.

AP. 14

AP. 12

Copernic: 1.3 ± 0.5 By.

Crystal fractions with track-density, \(10^9 \text{ cm}^2\)

C data from others

\(\bullet\) this work

\(\oplus\) this work — no noble gas data

(Maximum radiogenic content subtracted)
Soil samples from two collection sites of the Apollo 14 mission (14161, -162, -163, -258), both from smooth terrain material of the Fra Mauro formation, were studied by optical microscopy and electron microprobe techniques. Lithic particles (0.2-4 mm) were classified on the basis of mineralogy, chemistry, and lithology, and a population study was made on the basis of a systematic petrographic survey of polished thin sections of several hundred particles. Specific rock types were then studied in detail to determine their petrogenesis and place in lunar geology.

Most lithic particles are fragmental in character (though often recrystallized) or show other evidence of impact metamorphism, including partial fusion. Only ~4% have primary igneous textures. Six broad groups were established: A. basalts with igneous textures; B. dark soil breccias and glass-rich particles with basaltic affinities (many contain clasts of group C type); C. feldspar-rich crystalline rocks with recrystallized fragmental textures, including anorthosites, norites, troctolites; D. complex (multigenerational) microbreccias with strong affinity to Group C; E. complex microbreccias with roughly equal parts of A-B and C components; F. "ultramafic" olivine- and/or pyroxene-rich particles. The distribution of lithic particles >500μ among these 6 groups is shown in Table 1.

Because of their petrologic significance groups A, C and F were studied in greatest detail. Group A rocks are homogeneous fine- to medium-grained hypocrystalline basalts with textures ranging from intergranular to ophitic. Major phases include calcic plagioclase (An75-90) and clinopyroxene (augite, subcalcic augite and pigeonite). Other phases present include olivine,

<table>
<thead>
<tr>
<th>Group</th>
<th>A</th>
<th>B</th>
<th>C</th>
<th>D</th>
<th>E</th>
<th>F</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>161-163</td>
<td>14</td>
<td>107</td>
<td>71</td>
<td>46</td>
<td>24</td>
<td>4</td>
<td>266</td>
</tr>
<tr>
<td>14258</td>
<td>-</td>
<td>15</td>
<td>20</td>
<td>28</td>
<td>14</td>
<td>-</td>
<td>77</td>
</tr>
<tr>
<td>TOTAL</td>
<td>14</td>
<td>122</td>
<td>91</td>
<td>74</td>
<td>38</td>
<td>4</td>
<td>343</td>
</tr>
<tr>
<td>(% of Total)</td>
<td>4</td>
<td>36</td>
<td>26</td>
<td>22</td>
<td>11</td>
<td>1</td>
<td>100</td>
</tr>
</tbody>
</table>

Table 1. Distribution of rock types in size range 0.5-4 mm.
orthopyroxene, ilmenite, troilite, Fe-metal, Cr-spinel, apatite, K-feldspar, and two immiscible glasses (high Si, K and high Fe, low Si). The glasses have compositions similar to those in rock 14310. Some basalts closely resemble rock 14310. Others more closely resemble Apollo 12 coarse basalts. One microporphyritic olivine basalt has unusually magnesian olivine (Fo88) and clinopyroxene (Wo45En50Fs5), the latter previously unreported in lunar basalts. Like the Apollo 12 basaltic rocks, the Apollo 14 examples are poor in ilmenite and rich in olivine by comparison to Apollo 11 basalts. Apollo 14 basalts generally have a higher plagioclase content than either Apollo 11 or 12 basalts. The presence of orthopyroxene in Apollo 14 basalts is a further distinction.

Group C includes rocks with anorthosite (plagioclase), norite (plagioclase + orthopyroxene), and troctolite (plagioclase + olivine) mineralogies. Minor and accessory phases present in most particles are clinopyroxene (usually Ca-poor), ilmenite, Cr-spinel, Fe metal, apatite, K-feldspar. Some particles contain accessory glass. These three subgroups have a common affinity: all are feldspar-rich and types grade into one another. All have fragmental textures, variably obscured by recrystallization, which in many cases has produced a granulitic texture. The relatively large size of many mineral grains in these particles indicate they were derived from rocks very coarse-grained by comparison to mare basalts. By further contrast to the basalts the pyroxenes in group C rocks are Mg-rich, Ca- and Ti-poor, much less strongly zoned, and paler in color. Some are relatively rich in Cr2O3 (~2%). Pyroxenes in the least recrystallized varieties often show relatively coarse exsolution lamellae. Clasts of group C rocks are common in breccias of group B and predominant in group D. One anorthosite particle has primitive ferromagnesian phase compositions. The particle consists of ~97% plagioclase (An86), with ~2% diopside (Wo46En48Fs6) and ~1% magnesian ilmenite (8.2% MgO). Although the plagioclase is not unusually calcic the pyroxene and ilmenite compositions imply this material is more primitive than most feldspathic materials previously reported, in which the ferromagnesians are generally more Fe-rich.

Group F includes rocks of ultramafic character. Particles and clasts occur which consist of essentially 100% olivine. Textures reveal intergranular comminution not unlike that typical of some terrestrial dunites. Shock deformation lamellae also occasionally occur. One particle was found to consist of ~70% clinopyroxene and 25% olivine with ~5% Cr-spinel (one large euhedral grain). It cannot be said with certainty that these particles are pieces of true ultramafic rocks and not merely non-representative samples. If the former is not true then at least they were derived from coarse-grained rocks which crystallized in a slow cooling environment as implied by pyroxene exsolution.
One special particle was discovered worthy of particular attention. The assemblage includes 30% subhedral clinopyroxene grains (Wo\textsubscript{24}En\textsubscript{15}Fs\textsubscript{51}), 55% euhedral fayalitic olivine (Fa\textsubscript{58}), 5% plagioclase (An\textsubscript{68} with 1.3% BaO), 5% ilmenite (0.7% MgO) and 5% mesostasis (mixture of SiO\textsubscript{2} and K-feldspar with >1% BaO). All phases show evidence of shock deformation, but primary igneous textures are preserved. The euhedral olivines poikilitically enclose highly rounded blebs of high Si + K material compositionally similar to the mesostasis. The pyroxene contains moderately coarse exsolution lamellae (15µm), indicative of slow cooling implying a deep-seated origin. All phase compositions reveal the late-stage character of this assemblage. The abundance of such Fe-rich olivine and the size of such Fe-rich pyroxene are unusual for lunar rocks, as is the relatively sodic plagioclase composition. The association of K, Si-rich mesostasis with Fe-rich ferromagnesians plus the poikilitic olivine texture suggest liquid immiscibility was important in the petrogenesis of this rock. (Such immiscibility is described by Roedder and Weiblen in lunar basalts.) This particle implies that lunar differentiation processes exist which are capable of producing appreciable volumes of late-stage liquid, which in this case subsequently split into two liquids, one of which (high Fe) crystallized as euhedral olivine enclosing blebs of the other (high Si + K), which ultimately crystallized fine K-feldspar and SiO\textsubscript{2}. The textures and specialized bulk composition suggest a cumulate process involving settling of Fe-rich ferromagnesians which trapped minor amounts of high Si + K liquid. Such processes could produce a range of compositions (including KREEP), depending on the ratio of cumulus to intercumulus phases. In any case this particle is significant for its late-stage character and the information it provides regarding lunar differentiation trends and liquid lines of decent. We have drawn the following conclusions about the origin and evolution of rocks in the Fra Mauro formation: (1) The regolith is very mature by comparison to other sampled sites, as shown by the very low abundance of primary, undegraded lithic particles. (2) Many materials in the formation have been subjected to slow annealing causing recrystallization of fragmental textures. (3) The dominant rock types are of specialized (differentiated) character and show evidence suggestive of origins as cumulates crystallized at depth in the lunar crust. (4) The above conclusions are consistent with the interpretation of the Fra Mauro formation as a thick ejecta blanket from an early major impact event (Imbrium). (5) The mare-type basalts have a probable origin as flows from Oceanus Procellarum, deposited in the Fra Mauro formation by post-Imbrian impact events.
LUNAR CRYSTALS AS DETECTORS OF VERY RARE NUCLEAR PARTICLES, P.B. Price and I.D. Hutcheon, Dept. of Physics, Univ. of Calif., Berkeley, CA 94720: D. Lal, Tata Institute of Fundamental Research, Bombay, India; V.P. Perelygin, Joint Institute for Nuclear Research, Dubna, USSR.

The potential of lunar crystals for detecting rare, heavily ionizing particles such as trans-uranic cosmic rays or magnetic monopoles (if they exist) is until now largely untapped. We have initiated studies of the response of lunar minerals to heavy ions from the Berkeley Hilac and the Dubna Heavy Ion Cyclotron, so that long tracks of heavily ionizing particles can be properly identified.

We have obtained new etching rate data for Zn and Kr ion tracks in several different minerals with and without subsequent annealing treatments. Data obtained for $^{84}$Kr tracks in olivine and augite are shown below.

1. Olivine particle detectors.
   The curves show how the length $L$ of that portion of a track revealed during a given etching time varies with the residual range $R$ of the particle. Near the end of its range ($R \to 0$) the ionization rate of a slowing ion reaches a maximum and one might expect the chemical reactivity to be greatest at small $R$. For olivine the etching rate is indeed a strong function of ionization rate
Detectors of Rare Nuclear Particles
P.B. Price

and continues to increase even at the highest ionization rate of a Kr ion. The etch rate is undetectably small for Kr ions of energy greater than \( -10 \text{ MeV}\)/nucleon and presumably for other ions of comparable ionization rate. This dependence of \( L \) on \( R \) is qualitatively similar to that observed in glasses and plastics and suggests that olivine can be developed into a particle identifier. Notice, however, that the etching rate is significantly reduced after relatively mild annealing treatments. The slope of the response curve is made more negative and its intercept with the \( L \) axis decreases with annealing. This means that the etching rate is reduced much more for lightly ionizing particles such as Fe ions in the cosmic rays than for heavily ionizing particles such as fission fragments. The technique of preferentially annealing away undesired tracks left by lightly ionizing particles in order to study tracks of more heavily ionizing particles was first developed by Kapuscik et al. (1966) and by Maurette (1970).

From our preliminary results, it appears likely that Fe tracks will have a limited lifetime in an olivine grain on the lunar surface (maximum temperature \( -135^\circ C \)). With further studies it may be possible to use olivine as a thermometer, so that comparisons of cosmic ray track densities in olivine and augite or feldspar at various depths in a core may provide novel information on thermal profiles or stirring models. In this connection it may be relevant that Crozaz et al. (1970) found considerably lower mean track densities in olivine than in feldspars and pyroxenes in Apollo 11 fines.

2. Augite particle detectors.
Augite is more sensitive to Kr ions than olivine and is far more resistant to track-annealing. Note that for Kr ions of any energy up to \( 10 \text{ MeV}/\text{nuc} \) the etching rate is the same in the absence of annealing. This saturation effect appears to us to account for the observations by Price et al. (1971) of extremely long tracks in pigeonite and augite crystals of particles whose etching rates were constant along their lengths. There is no longer a need to consider as possible candidates particles such as magnetic monopoles, which have the property of ionizing at a constant rate. The tracks were most likely left by ultra-heavy cosmic rays. However, if monopoles do traverse lunar rocks at a finite rate, one could attempt to detect them in large pigeonite or augite crystals that have been annealed prior to etching. Then any tracks that show no change in etching rate along their intersections with several cleavage planes would be considered as serious candidates for monopoles.

3. Applications of selective annealing of tracks in lunar minerals.
a) Fission track dating of common minerals. By a careful choice of annealing temperatures and times we have been able to erase the vast majority of solar and galactic cosmic ray tracks — those due to Fe-group nuclei — without affecting fission tracks or ultraheavy cosmic ray tracks. This technique greatly extends the applicability of fission track dating to common minerals such as feldspar that sometimes contain greater than \( 100 \text{ ppb} \) of uranium heterogeneously distributed. Typical annealing temperatures for erasure of Fe tracks without altering fission tracks are: anorthite, \( 650^\circ C \) for 1 hr and pigeonite, \( 525^\circ C \) for 1 hr. The optimum temperature depends strongly on chemi-
Detectors of Rare Nuclear Particles
P.B. Price

P.621

cal composition, and one has to be careful not to interpret hastily a heterogeneous distribution of tracks surviving the annealing treatment as only fission tracks. Often the remaining tracks are located in zones where the composition is different from that of the remainder of the crystal. To be safe one must anneal at several temperatures so that even the thermally resistant zones lose their cosmic ray tracks.

b) Ultraheavy cosmic rays. By further annealing studies on lunar minerals irradiated with various heavy ions of known energy, we expect to be able to establish simple criteria for the identification of extremely heavy nuclei. For example, in augite crystals annealed 30 min at 675°C, Kr ion tracks will probably look similar to Fe ion tracks in unannealed augite. The threshold for recording ultraheavy cosmic ray tracks can be set wherever one wants by adjusting the annealing temperature.

c) The mystery of the Fe tracks. We have found that Zn ions with ranges of ~45μ are fully recorded in enstatite, oligoclase, bytownite, hypersthene, and diopside. We calculate that Fe ions should then be recorded over at least 25μ of range, yet there is an experimentally observed peak at 11.5μ in the length distribution of cosmic ray tracks in lunar minerals. It is very unlikely that this peak is other than at Fe. Can there be a relaxation in damage density occurring over long times even at low temperature? The peak is reproducible both in lunar and meteoritic minerals which have had different thermal histories. Further laboratory annealing studies may clear up this mystery.

References


The cooling rate of lunar basalts is of interest to understand in a quantitative sense the large amount of mineralogical and petrological data for these rocks. For this purpose we have calculated the variation of temperature with time of lunar lava flows of different thicknesses.

The temperature variation with time inside the basalt flow is given by the equation

\[ \rho C_p (T) \frac{dT}{dt} = V \cdot K(T) \nabla T \]

Because of the non linear boundary conditions and the temperature dependences of the specific heat \( (C_p) \) and the thermal conductivity \( (K) \), the above equation was solved numerically by means of an explicit finite difference method. At its top the flow loses energy by radiation into space according Stefan's law and at the bottom of the flow energy is lost by conduction into the lunar basement. The flow is emplaced, being totally liquid (1200°C), subsequently it crystallizes over the temperature interval 1200°C-1100°C, while liberating latent heat of fusion, and finally it cools down to the lunar ambient temperature. For the temperature dependence of the thermal conductivity we have used the measurements of Murase and Mc Birney (1970); the thermal conductivity of the liquid is taken to be infinite; i.e. the liquid portion of the flow is supposed to be isothermal. The temperature dependence of the specific heat has been taken from Horai et al. (1970). Further it has been assumed that the temperature of the liquid portion of the flow is given by the relation

\[ T = T_s + (T_l - T_s) F \]

where \( T_s \) - solidus temperature (1100°C)
\( T_l \) - liquidus temperature (1200°C)
\( F \) - the fraction which has solidified
The proportionality expressed by this equation has been observed during the cooling of Hawaiian lava lakes. (Wright and Weiblen, 1967; Peck et al., 1966).

Using the methods of Bottinga and Weill (1970; 1971) to calculate the density and the viscosity of lavas, we conclude that in flows of less than or equal to two meters thickness one should not expect to find the development of cumulate textures. The reason being that the upper crust of the flow grows at a rate greater than the rate at which formed crystals can sink to the bottom. This holds even for the formation of the early ilmenite crystals in the Apollo 11 basalts. The absence of cumulate textures in the Apollo 11 basalts has been noticed by James and Jackson (1970) and Weill et al. (1970).

A one meter thick flow will be totally solid after about 5 days. This time does not depend very much on the melting interval of the basalt but depends of course critically on its thermal conductivity and heat capacity. The radiation at the top surface of the flow is only important, as far as the rate of cooling is concerned, directly after eruption of the basalt; later on conduction in the solid silicate phase is rate determining for the cooling process.


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Fe AND Mg IN BYTOWNITE-ANORTHITE PLAGIOCLASE IN LUNAR BASALTIC ROCKS
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Fe and Mg are present in relatively large amounts in bytownite-anorthite plagioclase in lunar basaltic rocks. Fe is primarily in the ferrous state and is located in Ca\(^{++}\) sites and as an impurity in effective tetrahedral sites (not regular Al\(^{+++}\) sites) coupled with oxygen vacancies and lattice defects (Hafner et al., 1971). It is assumed here that Mg is similarly situated. The contents of Fe and Mg in these plagioclases are not only high but are systematically related to the anorthite content, or, as illustrated in Fig. 1, to the orthoclase content. These are the feldspars of the high iron trend of Smith (1971).

Fig. 1. Weight percent Fe and Mg in plagioclase as a function of mole percent orthoclase in rock 14053. Tie lines connect core and rim compositions. Plagioclase enclosed in pigeonite zoned to ferroaugite are labeled P; those enclosed in cristobolite and ilmenite, C and I, respectively; those bordered or partly enclosed by mesostasis, M.
Textural and compositional evidence indicates that Fe enrichment and Mg depletion was progressive during crystallization of rock 14053. The same is true of Fe and Mg in plagioclase in the rock. Brown and Carmichael (1971) noted a similar Fe enrichment in plagioclase of this compositional range in lavas of the Lake Rudolf region and, for this case where Fe was presumed to occupy tetrahedral sites, concluded that early formed plagioclase may have cooled more slowly permitting preferred Al\(^{+++}\) ions to occupy most tetrahedral sites, and as cooling became more rapid with less time for diffusion, more Fe\(^{+++}\) entered. In the present case, however, changes in the extent of diffusion do not explain the observed simultaneous Fe enrichment and Mg depletion. It is suggested that given conditions where Fe and Mg can enter the plagioclase structure as accidental constituents (rapid crystallization at high temperature) their abundances are a function of their concentrations in adjacent liquids. If this is true, rocks with strong Fe enrichment will have strong progressive Fe enrichment trends in plagioclase formed at successive times whereas rocks with arrested Fe enrichment should have arrested Fe enrichment in the plagioclase. This is the relationship observed for lunar basaltic rocks illustrated in Fig. 2.

Fig. 2. Fe content of plagioclase as a function of percent orthoclase for rocks 12038 (very strong Fe enrichment), 14053 (Fe enrichment) and 12036 (arrested Fe enrichment with abundant cumulus magnesian olivine with which residual liquids reacted). Data for rocks 12038 and 12036 from Busche et al., (1971).
Thus, whereas rapid crystallization at high temperature favors entrapment of Fe and Mg in basaltic plagioclase, changes in concentrations of these constituents in successively formed plagioclases reflect changes in their concentrations in residual liquids. If these conclusions are correct, the fractionation trend of Fe with respect to Mg can be determined for lunar basalts by measuring the Fe and Mg contents of selected plagioclase crystals. Such a fractionation trend for rock 14053 is illustrated in Fig. 3 where molecular percent orthoclase is assumed to provide a measure of the extent of fractionation.

Fig. 3. Fe/Fe + Mg as a function of molecular percent orthoclase in sequentially formed plagioclase in rock 14053.

References


Regolith fines from smooth terrain (14163, 14260) have median diameters comparable to those of mare surface samples. Although they are more poorly sorted than comparable Apollo 11 and 12 samples, their size distributions have approached the steady state distributions of mare deposits. Size distributions of most components are similar to those of mare fines, but glassy spherules are more abundant in all size categories suggesting a greater degree of reworking and hence greater age for these deposits. The spherules are compositionally distinct from spherule populations of other Apollo samples with normative compositions spread about clustered norms of breccia samples. They are thus provincial and were derived primarily by impact against local rocks. Analyses of 64 glassy spheroids (63-125 μm) are divisible into three categories: 15% are high Fe glasses (>15% FeO), most are Fra Mauro glasses (7%-15% FeO, Al₂O₃ < 21%-23%) and 6%-15% are high Al glasses (>21%-23% Al₂O₃). Fra Mauro and high Al glasses together are separable into six groups on the basis of FeO, MgO and K₂O + Fe₂O₃ contents, Table 1. It is not certain, however, whether or not these groups are classification artifacts. Furthermore, the origin of glassy spheroids may be so complexly varied as to preclude simple interpretation of their analyses. All that can be said with certainty is that most Fra Mauro glasses could have been derived by impact against coherent breccias and their contained igneous clasts. Scatter among their compositions may be due in part to the volume of melt contributing to each liquid jet and the degree of mixing and the extent of partial fusion and fractionation that took place during melt formation and prior to jetting. Certainly some spheroids are exotic and the Fe rich group is a prime candidate with possible source areas in mare basalt terrains. Any of the others may also have distant origins, but the variety of crystalline lithic fragments in the breccias, ranging in composition from olivinite to anorthosite, suggests that any of the types observed could have local sources.

The bulk composition of Fra Mauro fines is also distinct from Apollo 11 and 12 materials (LSPET, 1971). The fines are notably enriched in Al₂O₃, MgO and K₂O, relative to Apollo 12 sample 12070, and depleted in FeO, MnO, TiO₂, Cr₂O₃ and Na₂O. The differences are reflected in the mineralogy. Plagioclase is consistently more abundant than in other Apollo samples and its abundance is reflected in the high alumina content of the Fra Mauro samples. The enhanced MgO content and the Fe depletion is in part a reflection of the abundance of bronzite among pyroxene grains. Pigeonite, subcalcic augite, subcalcic ferroaugite, ferroaugite and augite are also present. The relative depletion of TiO₂ and Cr₂O₃ is consistent with the paucity of opaque phases, fewer by a factor of four than in Apollo 12 samples.

Fines are compositionally similar to breccias and must have been derived from certain of them. Some breccias contain abundant glassy aggregates and
Fra-Mauro Fines and Breccias

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spherules and must have originated by lithification of local regolith deposits (14313, 14055). They are termed regolith breccias here and are characterized in section by high proportions of clastic glass, lack of a true matrix, continuum of grain sizes, abundance of grain-to-grain contacts and a light color. Analyses of regolith breccias (LSPET, 1971) reveal further, probably inherited, characteristics. They have high Ni, total carbon and solar wind implanted noble gas contents. Jackson and Wilshire (1972) indicate that the two rocks here called regolith breccias have different degrees of cohesion, rock 14313 is a coherent type (F2) whereas 14055 is noncohesive (F1). Wilshire and Jackson (1972) suggest that the light colored F1 types are weakly lithified regolith derived either from disaggregation of F2 types or from originally unconsolidated material at the top of the Fra Mauro deposits. They further speculate that F2 types may be samples of a near surface, rapidly chilled facies of the Fra Mauro Formation. The abundance of glassy spheres in the two samples is comparable, however, (14313 = 1.55% and 14505 = 1.7% of grains >20µm) and both must have originated from regolith materials with comparable old exposure ages. Rock 14055 is simply a porous unshocked microbreccia and 14313 a shock compressed microbreccia, both having been formed from the local regolith.

Other breccias, less friable, contain little or no glass apart from coatings and fracture fillings and consist of mineral grains, crystalline rock fragments, dark and light breccia clasts and recrystallized angular glass fragments floating in a fine, even-grained dark matrix which exhibits various degrees of recrystallization in different samples. This group is termed annealed breccias (14006, 14066, 14304, 14306 and 14320). Differences in texture between dark and light breccia clasts within these samples may parallel those between large samples observed at the site. Light clasts consist of a continuum of particle sizes with few but large opaque grains whereas dark clasts contain large grains in a recrystallized matrix with abundant small crystallites of opaque minerals. The abundance of the finely crystalline opaque minerals appears to reflect the volume of recrystallized matrix and may be the prime factor controlling the color. The abundance of large fragments of annealed breccia on the ejecta apron of Cone Crater suggests that this is the material of the deep substrate. The impactoclastic texture, abundant evidence of high shock pressures and annealed characters of these breccias suggests that they were deposited as heated impact ejecta. The debris must have been heated during passage of impact-generated shock waves, ejected and laid down in avalanche deposits which retained sufficient heat to produce annealing. Different degrees of annealing may have been attained at different depths in the deposit or at different locations where the beds were of different thickness. If the color of the breccia does reflect the degree of annealing as suggested above, upper layers of the Fra Mauro Formation may consist of light colored less thoroughly annealed breccias. This could account for the prevalence of light clasts in the regolith and regolith breccias.

The event which produced the annealed breccia deposits had to be extremely energetic, and the nature of the deposit is consistent with the hypothesis that the Fra Mauro Formation is composed of ejecta from the Imbrium Basin. If this hypothesis is correct, the annealed breccias represent a sample of part of the pre-Imbrium crust and the character of the included crystalline clasts...
Fra Mauro Fines and Breccias
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has a special significance. Types recognized include subophitic plagioclase-orthopyroxene rocks; sheared and partly recrystallized plagioclase rocks; granulitic plagioclase with minor amounts of either 1) orthopyroxene, ilmenite and native Fe, 2) clinopyroxene or 3) olivine; polycrystalline plagioclase mosaics; orthopyroxene rocks; highly shocked augite-ilmenite rocks, olivinite, and ophitic and subophitic clinopyroxene-plagioclase rocks. Thus, rocks represented probably include noritic, anorthositic and troctolitic types, orthopyroxenite, pyroxenite, dunite and basalt. Relative abundances cannot be determined because of small clast and sample sizes, but the preponderance of low Fe, high Mg, highly feldspathic rocks suggests that the principal rocks of the pre-Imbrium crustal areas were of noritic composition. The presence of basalt clasts (14053, 14310 and smaller fragments) is nonetheless significant, indicating that the pre-Imbrium basin terrain was overlain by basaltic rocks, some not greatly different in composition from mare basalts erupted in later times. Further complexity of the source area is indicated by the presence of both dark and light breccia clasts in the annealed breccias. They are probably primary constituents, suggesting that the pre-Imbrium terrain was highly cratered and contained thick deposits of impactoclastic debris, some of which was itself annealed. This interpretation is in keeping with a high, early cratering rate and the possible mantling of the Imbrium region by welded Serenitatis ejecta.

Table 1. Average composition of groups of Fra Mauro and high Al spheroids classified on the basis of FeO, MgO and K₂O+P₂O₅.

<table>
<thead>
<tr>
<th></th>
<th>A</th>
<th>B</th>
<th>C</th>
<th>D</th>
<th>E</th>
<th>F</th>
<th>Av.</th>
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<td>18.5</td>
<td>15.2</td>
<td>22.8</td>
<td>16.5</td>
<td>21.1</td>
<td>17.5</td>
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<td>11.3</td>
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<td>6.39</td>
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<td>.54</td>
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<td>7</td>
<td>4</td>
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</table>

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Cosmogenic radionuclides produced in the lunar surface by solar and galactic protons are employed to study both the recent and ancient cosmic ray irradiation history and to characterize lunar surface processes such as erosion of rocks and mixing of soil. Radionuclide measurements in Apollo 11 and 12 samples indicated the relative contributions of solar and galactic protons to the cosmic ray flux at the lunar surface. Measurements of \( ^{26}\text{Al} \) and \( ^{22}\text{Na} \) depth gradients in whole rock, rock slices and core tubes from Apollo 12 have provided the necessary information to define the average intensity and energy spectrum of solar protons incident on the moon. The best fit for incident protons above 10 MeV, using the kinetic energy power law to describe the solar proton spectrum is:

\[
\frac{dJ}{dE} = kE^{-3.0}
\]

Where \( J \) is the proton flux, \( E \) is proton energy and \( K \) a constant related to the particle intensity. From radionuclide and cosmic ray energy spectrum considerations, the average proton flux with energy greater than 10 MeV incident on the moon for the last few million years has been 60 p/cm\(^2\)-sec. The fact that the observed \( ^{26}\text{Al} \) and \( ^{22}\text{Na} \) depth concentration gradients can be described by the same proton energy and intensity spectra indicates the solar activity has remained relatively unchanged for at least the last million years.

Although the Apollo 11 and 12 lunar sample studies have greatly increased our knowledge of lunar processes and the cosmic radiation history, it is far from complete. A study of the suite of Apollo 15 samples near the St. George crater at the Apennine Hadley site has been particularly illuminating in revealing the history of some of the processes which occurred at this site as well as confirming the rates of lunar soil mixing. Rock chips from the surface of a large boulder and samples of surrounding and underlying soil were studied. Two chips (15205 & 15206) from the top surface of the boulder together with soil samples from beneath the boulder (15231), adjacent to the boulder (15211), and about 0.5 (15231) and 10 meters (15091) from the boulder were studied. The results of the radionuclide analyses are summarized in Table 1. Conclusions based on these analyses must be qualified somewhat because chemical analyses have not yet been reported; however, the following history of the boulder and soil appears evident. The boulder has been on the lunar surface in its present position for about one million years. This is evident from the fact that while the \( ^{22}\text{Na} \) concentrations in the boulder chips are at saturation, the \( ^{26}\text{Al} \) concentrations are at approximately one-half to two-thirds of their saturation value. This rather short lunar surface life is also attested to by the fact...
that the soil (15231,1) directly beneath the boulder has a higher \( ^{26}\text{Al} \) concentration than its saturation value for its rather highly shielded location (minimum of 20 g/cm\(^2\)). This high concentration requires that it was exposed to saturation before the boulder covered it about one million years ago. The "soil fillet" (15211,2) adjacent to the boulder has a primordial radionuclide composition which is essentially identical to that of nearby soil samples, yet is far different from that of the boulder. This requires that the soil fillet originated almost entirely from nearby soil rather than from erosion of the boulder. This well developed soil fillet around the boulder helps to confirm our previous conclusions\(^{1}\) that soil mixing through the first two or three centimeters of depth occurs rather well on a time scale of 100,000 years through micrometeorite bombardment. Soil projectiles which were ejected by micrometeorite bombardment and which struck the boulder would form the observed fillet.

Two very interesting lunar specimens were collected at Station 9 of the Hadley Apennine site (see Table 2). Sample 15501,2 was a soil clod from the ejecta blanket of an apparently young crater.\(^{2}\) Our measurements show saturation with respect to \(^{22}\text{Na} \), however the \(^{26}\text{Al} \) concentrations indicate a surface age of only 0.75 to 1 million years. A possible alternate explanation for the low \(^{26}\text{Al} \) concentration is a very high erosion rate; however, such a low concentration would require an erosion rate on the order of centimeters per million years.

Rock 15556, a highly vesicular basaltic rock, showed an unusually high \(^{26}\text{Al} \) concentration which was adequately justified by its low density and the fact that it was only very slightly recessed in the lunar soil. Soil samples which were taken from trenches at stations 6 and 8 have been analyzed (see Table 2). Samples 15041,1,4 and 15031,1,4 represent top and bottom soil samples from the trench at station 8. Our previous observations of both rock and soil cores have shown that where the chemical composition of the major target elements for \(^{26}\text{Al} \) and \(^{22}\text{Na} \) are reasonably constant, the \(^{26}\text{Al} \) to \(^{22}\text{Na} \) ratio decreases by as much as twofold in the first 20 g/cm\(^2\) of depth. The fact that the \(^{26}\text{Al} \) to \(^{22}\text{Na} \) ratio is the same at the surface as at the trench bottom suggests a very different chemical composition for soil from the two levels. Such a difference, however, is certainly not suggested by the primordial radionuclide concentrations. The approximate twofold increase in \(^{54}\text{Mn} \) concentrations with depth, and the approximately fourfold decrease in the \(^{56}\text{Co} \) with depth are in accord with the expected galactic cosmic ray flux buildup with depth and the solar cosmic ray attenuation with depth, respectively. The rather low concentration of \(^{46}\text{Sc} \), \(^{48}\text{V} \) and \(^{56}\text{Co} \) were to be expected because of the lack of major solar flares for several months prior to the Apollo 15 mission.

Table 3. The comparison of the observed \(^{26}\text{Al} \) content of Rock Slice 14310,1,87 with the predicted value based on the average incident proton spectrum, the rock's chemical composition and the \(^{26}\text{Al} \) excitation functions indicates this rock was at the surface for at least the last two or three million years. This is of particular interest, since preliminary track measurements\(^{3}\) suggested that the rock might have had a relatively short lunar surface
age. Rock sample 14321,40 is a 700 gram surface specimen of the 9 kilogram football sized breccia. The $^{22}$Na and $^{26}$Al concentrations in this section indicate saturation, however, the $^{26}$Al is somewhat low and may be the result of losses of surface material which are known to have occurred during the cutting of this specimen. Soil sample 14163,0 has substantially lower $^{22}$Na and $^{26}$Al concentrations than are present in the top two or three centimeters of lunar soil. This indicates that the sampling depth for this specimen was approximately 8 centimeters. Some of the Apollo 15 samples were counted one to two weeks after return to Earth. Measurements of the $^{4}$He (16.1d) in these samples indicated that it could be entirely accounted for by galactic cosmic ray production, thus demonstrating a negligible solar cosmic ray contribution over the past several months. In the case of $^{56}$Co its concentration is about 10-fold higher than could result from galactic cosmic ray bombardment, indicating a substantial residue from the January 24, 1971, solar flare.

The primordial radionuclide concentrations of Apollo 15 soil samples fall in a range between those at the Apollo 11 and Apollo 12 sites. (Table 4) The K, U and Th content of soil samples near Hadley Rille (stations 2 and 9) are lower than those at the landing site (station 8) and station 6, indicating that the wide differences which were observed between the Apollo 11 and 12 sites can also occur locally on a scale of a few kilometers. The K, U and Th content of the Apollo 14 soil sample is two to four-fold higher than those at the Apollo 12, and 15 sites while the K/U ratios (1200-1500) are comparable suggesting that the magmatic differentiation indicated by these ratios is characteristic of a large portion of the lunar surface. The Th to U ratios of Apollo 14 and 15 soil are about 4, in agreement with observation on Apollo 11 and 12 soils.

REFERENCES:


STUDY OF THE OPTICAL PROPERTIES OF LUNAR GLASS SPHERULES by K. J. Rao, S.K. Sarkar, L. Klein, and A. R. Cooper, Division of Metallurgy and Materials Science, Case Western Reserve University, Cleveland, Ohio 44106

Optical absorption studies have been performed on glassy spherules obtained from lunar fines returned from Apollo XI, Apollo XII, and Apollo XIV by employing microspectrophotometry. Absorption maxima at 430 and 650 nm have been noted. The glasses seem to fall into roughly three categories--possessing either or both of the absorption peaks--though the overall spectral features are essentially similar. Electron probe micro-analytical data has been obtained on these spherules and the probable origin of the absorptions has been discussed. Predominantly the absorption is due to iron. The absorption data, in the light of the known spectral effects of Fe and Cr in silicate glasses, seem to indicate the Fe and Cr are present largely as Fe$^{3+}$ and Cr$^{3+}$. It is suggested that the high optical densities result from Fe$^{3+}$ entering substitutionally into Si$^{4+}$ sites in the silica poor alkali rich lunar glasses providing increased transition probabilities. Usually heating for more than 1 hour above 600°C in air or vacuum causes noticeable absorption increases. Accurate refractive indices of glass spherules measured by interference microscopy permitted the variations of refractive indices with successive heat treatments and differing thermal histories to have been determined. An arrangement by which a glassy spherule can be rapidly cooled (~200°C/sec) from temperatures (~900°C) well above the transformation range is discussed. It permits heat treatments which do not noticeably change the optical absorption of the spherules. Spherules given this rapid heat treatment invariably show birefringence which is dependent on the rate of quenching. It is difficult to explain the existence of birefringence in such small samples based on residual stress. Yet the possibility that spherulic crystal growth is responsible must be ruled out based on the observation that birefringence disappears upon annealing at ~600°C.

The rapid quench necessary to cause the birefringence nearly reproduces the as received density of the glassy spheres in contrast to thermal treatments with slower cooling rates which typically cause an increase in density.

Based on these observations the following tentative approach to thermal history of glasses has been made:

a) Glass spherules--particularly those which are homogeneous have been formed by quenching of low viscosity lunar melts from high temperatures.

b) The quenched spherules originally had birefringence and hence internal stress.
Optical Properties of Lunar Glass Spherules
K.J. Rao

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c) Under lunar surface conditions the stresses have relaxed gradually.
d) Lunar surface conditions have not been conducive for significant volume relaxation.

If this surmise can be verified two important consequences occur:

a) Birefringence and density may be used to give a rough estimate of the date of particular lunar glass spherule was melted.
b) Lunar glasses provide opportunity for studying stress relaxation at lower temperatures and longer times than has been possible.
DIRECT MEASUREMENT OF THE LUNAR PHOTOELECTRON LAYER
D. L. Reasoner and W. J. Burke, Dept. of Space Science, Rice University, Houston, Texas 77001

The Charged-Particle Lunar Environment Experiment (CPLEE), a part of the Apollo 14 ALSEP, is an ion-electron spectrometer capable of measuring ions and electrons with energies between 40 eV and 50 keV. Accordingly, the instrument, with apertures 26 cm above the surface, has detected a layer of photoelectrons, or photoelectron gas above the sunlit lunar surface with energies ranging up to 200 eV. The experimental data for periods when the moon was in the earth's magnetotail for electron energies between 40 and 200 eV follows a power-law spectrum of the form \( j(E) = j_0 (E/E_0)^{-\mu} \), with \( j_0 = 2.5 \times 10^5 \) electrons/cm\(^2\) - sec., \( E_0 = 40 \) eV, and \( \mu = 3.5 \). The implications of this measurement are two-fold, in that the lunar surface potential can be immediately determined to be at least 200 volts, and a value of the photoelectron yield of the lunar surface material for photon energies above 40 eV may be computed.

A numerical solution for the variation of electron density and potential above the lunar surface was obtained. The method of solution was based on solving Poisson's equation, and computing the electron density at a height \( x \) by using the Liouville Theorem along with the conservation of energy equation for the single-particle trajectories. Two parameters of the solution are the solar photon spectrum \( I(h\nu) \) and the photoelectron yield function \( Y(h\nu) \) of the surface materials. The solar photon spectrum was obtained from various experimental sources, and the solution for the energy spectrum at the height of the measurements (26 cm) was computed for various values of \( Y(h\nu) \) until a fit to the experimental data was obtained. We used a functional form of \( Y(h\nu) \) to be \( Y(h\nu) = a(h\nu - W) \) for \( 6 \) eV \( \leq h\nu < 9 \) eV, and \( Y(h\nu) = Y_0 \) for \( h\nu \geq 9 \) eV. \( W \), the lunar surface work function, was chosen to be 6 eV. This procedure resulted in a value of \( Y_0 \) of 0.1 electrons/photon. The solution also showed that for the condition of low ambient plasma density in the magnetotail, the photoelectron density falls by 5 orders of magnitude within
10 meters from the surface, but that the layer actually terminates several hundred meters above the surface.

A hydrostatic model of the lunar photoelectron gas was developed assuming a polytrope law in conjunction with Poisson's equation and the conservation of momentum equation for the gas. With a constant polytrope index, the analytic solutions were inadequate. Modified to include a polytrope index that varied with height, we found that the variations of the index were related to the shape of the solar photon spectrum. This suggested that a hydrostatic model made up of a multicomponent gas does describe the physical situation.

The detailed temporal history of the photoelectron intensity during the total lunar eclipse on February 10, 1971 was studied in order to determine the source distribution of solar photons in the 40 ev to 200 ev range over the solar disc. It was found that the temporal behavior of the photoelectron intensity exhibited a classical penumbral-umbral behavior. Hence it is concluded that at the time of the eclipse the emission of these higher energy photons was uniform over the solar disc.
CONCENTRATIONS AND LABILITY OF THE HALOGENS, PLATINUM METALS AND MERCURY IN APOLLO 14 AND 15 SAMPLES. G. W. Reed, Jr., S. Jovanovic and L. H. Fuchs, Argonne National Laboratory, Argonne, Ill. 60439.

Four sets of elements were measured, the halogens, the electropositive elements, Li, Ba and U, the platinum metals, Ru and Os and the element Hg. Neutron activation is the technique used. Besides studying the concentration, distribution and hot water leachability of these elements our attention has turned to the volatilization of some of them in the heating experiments in which Hg is extracted.

Because of the complexity of the fragmental rocks some preliminary mineralogical characterization was necessary as a guide for sampling. The mineralogy of the dark fractions of 14305, 64 and 14321, 185 is generally similar to that of the light fractions, but there are marked contrasts between the texture of the dark and that of the light. The dark of both samples consists of relatively large sub-angular fragments of plagioclase, pyroxene and olivine, up to 0.1 mm in diameter, embedded in a fine-grained (0.01 mm) matrix of silicates and abundant needles or plates of smaller (1-3 µ) opaque grains. The hard and porous dark clasts appear to be partially recrystallized breccias. The light fraction, interstitial to the well-defined dark clasts, contains an assortment of less consolidated angular mineral and glass grains and both recrystallized and unaltered rock fragments. Thus, at least two thermal events were involved. Rock 14310 contains orthopyroxene; this is the first occurrence of the mineral in any of the lunar crystalline rocks we have examined and may in some respects be related to its unique chemistry.

Cl contents cluster around 45 ppm. F contents are similar to those previously measured: of the order 100-300 ppm in Apollo 14 soils [1,2,3] and <100 in igneous rock 14310, 124 fragmental rocks and in Apollo 15 soils. I concentrations span the range (2-200 ppb) previously reported [2]. Br concentrations as high as a ppm were observed in a number of samples, including soil, fragmental and igneous rocks. Previous results placed Br in Apollo 12 soil and breccia and Apollo 11 soil, breccia and igneous rocks at a few tenths ppm and that in Apollo 12 igneous rocks at a few tens ppb. The Cl/Br ratios for most of the Apollo 14 samples cluster near 50 as was the case for Apollo 11 [1]. The ratios of insoluble/soluble Cl and Br at all four sites show a distribution similar to terrestrial basalts. Another result of interest is the larger concentrations of Cl and Br in the dark relative to the light separates from 14305 and 14321.

Ru and Os are expected to be depleted in lunar surface material as are most siderophile elements. The Ru concentrations, Table 1, are similar to those reported for Apollo 12 rocks [2]. No significant enrichment is apparent in the soils. On the other hand the amounts of Os decrease from soil to fragmental to igneous rock in the order >67+24+11 ppb. The Os concentration in rock 14310 is similar to that in Apollo 11 and 12 soils. This rock has been a closed system possibly since the Imbrium event. Its enhanced (x10) Os content
relative to Apollo 11 and 12 basalts suggests that it may be remelted breccia. The still larger Os contents in Apollo 14 fragmental rock and soil require further additions of an exogenous component. The fragmental rocks which contain basaltic clasts could have acquired the additional Os between the time of formation of the basalts and consolidation. The soils either continued to acquire Os or originated from a more Os rich strata or region.

Hg concentrations in general are less than 2-3 ppb Hg, with the notable exception of the trench samples which contain 7-9 ppb Hg. The trench samples also contain the higher concentrations of labile Hg (volatilized at <130°C). These trench samples do not show a variation with depth probably because of mixing during sample collection. The 15231 soil sample from under a boulder should have been cold and has a large fraction of labile Hg compared to that in 15091, a nearby soil sample exposed to lunar daytime temperatures.

The most striking aspect of the Apollo 14 Hg measurements is the large release above 450°C during stepwise heating. In samples from other sites only a small fraction of the Hg remained above this temperature; in Apollo 14 most samples still retained 30% or more Hg. We interpret the presence of Hg in highly retentive sites as being due to high temperature equilibration [4]. Since these samples are known to be associated with the impact that excavated the Imbrium basin both high temperature quenching and shock implantation may account for the Hg in retentive sites.

The cold traps in which Hg was collected during the stepwise heating were observed to contain other radioactivities, notably Br. Losses due to volatilization had not been considered serious before since Br results obtained on both heated and unheated samples agreed quite well. Most of these samples, however, were Apollo 12 igneous rocks. Experiments are now conducted so as to account for all possible losses. It has been established that not only Br but other elements, F, Ru, and Os for instance, may be volatilized. In Apollo 14 soils and fragmental rock 14305 up to 10% of the Br was volatilized below 450°C and most of the remainder between 450° and 1200°C. In igneous rock 14310 20 to 60% of the Br volatilized. It would be interesting to measure 14053 which is more like mare basalts in which little loss appears to have occurred. Ru and Os volatilization was low, few % or less. 15231 and 15091 are exceptions; up to 75% of the Os and 29% of the Ru were mobilized. F results may also be affected when F was determined on samples from which Hg had been extracted.

Apollo 15 soils from the front and the mare region have significantly lower Cl and Os, but not Br, I and Ru, than the Apollo 14 soils. Such differences between coherent elements probably requires a lunar fractionation process.

We have not completely assessed our data in the light of meteoritic contributions. The examination of the samples did reveal several possibilities. Metal particles in the dark parts of 14321,185 and 14305,64 have Ni and Co contents within the range of meteoritic metal. The metal of 14321,185 contains inclusions of schreibersite and a carbide, probably cohenite. The light part of 14305,64 contains polycrystalline aggregates of olivine and three chondrule-Like objects from 0.8-1.5 mm in diameter. Two of these "chondrules" were sectioned; they resemble meteoritic chondrules in shape only.

Our trace element results make resolution of an external contribution difficult. For instance the Ru/Os in 14310 and the Apollo 15 soils fall near
the meteoritic ratio of 1.7 whereas the Apollo 14 soil and fragmental rocks have much lower ratios and Apollo 12 soil and rocks have higher ratios (2). Data for other elements may elucidate this problem; for instance, we observed a linear correlation between the Ni and Os contents of 14259, 14305 and 14310, and also between 12033, 12070 and 10084.

Table 1. Halogens, ruthenium, osmium, mercury and other trace elements in Apollo 14 and 15 Samples

<table>
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<th>Cl</th>
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References
Work performed under the auspices of the U.S.A.E.C. and NASA
CHEMISTRY OF HIGHLAND AND MARE BASALTS AS INFERRED FROM GLASSES IN THE LUNAR SOILS. Arch M. Reid, W.I. Ridley, Jeff Warner, Russell S. Harmon, Robin Brett, NASA Manned Spacecraft Center; Petr Jakes, Lunar Science Institute and Roy W. Brown, Lockheed Electronics, Houston, TX 77058

If a sufficient number of glasses in lunar soils are analyzed, preferred compositions can be distinguished and used as a guide to the composition and abundance of lunar rock types. Over 2000 electron microprobe analyses of phases in the lunar soils supplemented by data from the literature have been used to determine such preferred compositions. Average compositions and abundances of these glass types, given in the table, are based on random surveys. The classification is based on natural breaks or minima in overlapping populations.

Three major glass populations can be recognized in the Apollo 14 soils. (1) Mare-derived glasses, 11 percent of the total, have compositions like the Oceanus Procellarum basalts. (2) The most abundant glass type has the major element composition of a high alumina tholeiite. These Fra Mauro basalts (KREEP, norite) are the major component at the Apollo 14 site, but with their high content of K U and REE they are unlikely to be an abundant rock type on the moon. The equivalent rocks have plagioclase more abundant than pyroxene, and are characterized by orthopyroxene, minor K feldspar, phosphate, and zirconium phases. Two major subgroups occur in nearly equal abundance, one quartz normative and the other olivine normative. (3) A high Al, high Ca group, low in Fe, Cr, K with a Ca/Al ratio like that of the Fra Mauro basalts but higher Mg/Fe. These glasses are equivalent in composition to feldspathic basalt or anorthositic gabbro with the norm of the average having 70 percent feldspar, 20 percent pyroxene, 9 percent olivine and 1 percent ilmenite. Minor glass types are equivalent in composition to gabbroic anorthosite and potash granite.

Glasses in the Luna 16 soil are of three major types. (1) The most abundant glass type, Fecunditatis basalt A, is equivalent in composition to a mare basalt but with higher Al and lower Fe than the Apollo 11 or 12 basalts. The norm has approximately equal amounts of plagioclase and pyroxene and is olivine normative. (2) A minor glass group, Fecunditatis basalt B, has compositions high in Fe, Ti, low in Al and resembles Mare Tranquillitatis basalt. (3) The third group is high in Ca, Al with the composition of feldspathic basalt. The average composition and the norm with 70 percent feldspar is almost identical with the Fra Mauro group 3 glasses described above. Few gabbroic anorthosite, anorthosite or Fra Mauro type glasses are present and granitic glasses are rare.

To obtain comparable data on the Apollo 11 and 12 soils we have worked with glass analyses from the literature. These are not based on random surveys and reliable estimates of relative abundances of glass types cannot be made. Four types of glass appear to predominate in the Apollo 11 soil. (1) The major glass type, Tranquillitatis basalt A, is rich in Fe, Ti with low Al and K. The norm of the average has 12 percent ilmenite, approximately equal amounts of pyroxene and plagioclase, and 10 percent olivine. (2) A second mare type glass, Tranquillitatis basalt B, has extremely high Fe, Ti and much lower Al. The norm has
only 17 percent feldspar and is mostly pyroxene, ilmenite and olivine. (3) Feldspathic basalt is abundant, and identical in composition with the Apollo 14 and Luna 16 feldspathic basalts. (4) Gabbroic anorthosite and anorthosite glasses are present but less abundant than the feldspathic basalt glasses. Rare granitic glasses have been reported. The glasses reported from the Apollo 12 soils are of three main types (1) Glasses with the composition of mare basalts (Procellarum basalts) resemble the Tranquillitatis basalts but contain higher Si, lower Ti. Pyroxene is more abundant than feldspar and olivine and ilmenite are less abundant. (2) Higher K higher Al basaltic glasses, Fra Mauro type glasses, are a major constituent of the soil. (3) Feldspathic basalt glasses are also present at the Apollo 12 site, but are less abundant than at the other sites. Minor gabbroic anorthosite, anorthosite and granite have been reported.

Discussion. The nature of the non-mare lunar rocks can be determined from a study of the one non-mare site (Apollo 14) and from a study of the exotic fragments in the mare soils. Non-mare samples are characterized by high Al (Al2O3 greater than 14 percent), low FeO (less than 14 percent), low Cr (Cr2O3 less than .2), and low CaO/Al2O3 (less than 0.7). Two major types of non-mare material are recognized in mare soils and at the Apollo 14 site. Fra Mauro basalts predominate at the Apollo 14 site, are prominent at the Apollo 12 site, but are rare in the Apollo 11 and Luna 16 soils. The other non-mare component is present in significant amounts in soils at all four sites and has essentially the same composition at each site. These data are consistent with this component being derived from the lunar highlands and since it has the composition of a feldspathic basalt we have called it Highland basalt. In comparison, glasses with gabbroic anorthosite and anorthosite compositions are much less abundant. If, as seems probable, there is a highland-derived component common to all 4 sites the best candidate for this component is the Highland basalt.

Mare basalts are characterized by high Fe, Cr, Ca/Al, low Al, Ni, in comparison with the non-mare samples. These 'mare' characteristics systematically vary in the sequence Luna 16, Apollo 11, 12, and 15. Luna 16 basalts have some chemical characteristics, such as Al content, that are intermediate between non-mare and the other mare samples. Mare basalts appear to have formed at depth by partial melting of a dominantly pyroxenic mantle (Ringwood, 1970). The two major non-mare rock types must derive from a more Al-rich, more feldspathic source at shallower depth. Fra Mauro basalts are formed as partial melts or as a residuum from fractional crystallization. The origin of the Highland basalts, a rock whose primary texture we have not seen, is highly speculative. The highly feldspathic composition and the high Mg/Fe ratio is suggestive of a cumulate origin. However, the constancy of composition at four widely separated sites does not appear consistent with formation as a coarse-grained heterogeneous cumulate and we cannot discount the possible existence of highly aluminous basaltic liquids liquids early in lunar history.

Lunar surface rocks are derived from a feldspathic outer layer (non-mare basalts) and a deeper pyroxenitic source (mare basalts). Our model calls for a heterogeneous layered moon, high in Ca, Al near the surface, more mafic at depth; with both layers depleted in volatile and siderophile elements, and enriched in certain trace elements. This layering may be due to differentiation early in lunar history, or may reflect an accretionary stratigraphy.
### LUNAR ROCK TYPES AS ESTIMATED FROM GLASS COMPOSITIONS IN THE LUNAR SOIL

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<td><strong>Al₂O₃</strong></td>
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*Abundance data not based on random surveys; rare granitic glasses have also been reported.
THE PETROLOGY OF BASALTIC PARTICLES IN THE LUNA 16 SOIL by J. B. Reid, Jr., Smithsonian Astrophys. Obs., Cambridge, Massachusetts 02138

Of some 2380 Luna 16 soil particles surveyed, roughly 20% are basaltic. Of those, four in five are primary igneous rocks with nearly holocrystalline ophitic to subophitic texture and with a limited range in grain size (50 to 175 μ). Recrystallized basaltic breccias, very fine-grained devitrified basaltic glasses, and rapidly quenched vitrophyres constitute the remaining 20% of the basaltic particles.

The primary igneous fragments vary from mafic (less than 30% modal plagioclase) to rather feldspar-rich (50% modal feldspar). Textural evidence suggests that plagioclase is the liquidus phase in the felsic group, while olivine, chromite, and pyroxene begin crystallizing before feldspar in the more mafic fragments. The majority of the basaltic particles falls between these extremes, and in them the crystallization of pyroxene and plagioclase seems to have been simultaneous. Late-stage interstitial phases include troilite, native Fe, phosphates, ulvöspinel, zircon, and immiscible K- and Fe-rich glasses.

Luna 16 basalts tend to fall into two groups when their pyroxene compositions are plotted on diagrams showing TiO₂ contents as a function of Fe/(Fe + Mn + Mg) (Figs. 1 and 2). TiO₂ contents in type 1 pyroxenes (Fig. 1) are high in grain cores but fall sharply toward grain boundaries, probably reflecting the removal of TiO₂ from the melt by the simultaneous crystallization of ilmenite. Type 2 pyroxenes (Fig. 2) have lower and less variable TiO₂ contents than type 1 pyroxenes but are more strongly zoned in Fe/Mg and reach more Fe-rich compositions. The more mafic basalts tend to have type 1 pyroxenes, while the more feldspar-rich fragments contain type 2 pyroxenes.

The ilmenites of a given basaltic particle are quite homogeneous with respect to Mg content, although there is variety in Mg/(Mg + Mn + Fe) in ilmenites of different particles. When Mg/(Mg + Mn + Fe) of the ilmenite of a given particle is plotted as a function of the Fe/(Fe + Mn + Mg) ratio of the coexisting pyroxenes (normalized to a constant TiO₂ content) (Fig. 3), some of the basalts ((306-215), (306-77), (306-70)) fall along a well-defined linear trend, suggesting that they may represent successive flows from a differentiating magma chamber. Three type A Apollo 11 basalts plotted on this diagram also fall on a linear trend and may be internally related in a similar fashion. Two Luna 16 particles, (306-241) and (306-236), do not fall on this trend and may be parts of a second flow sequence.
Basaltic fragments with type 2 pyroxenes may be parts of later flows genetically related to the type 1 fragments or may represent an unrelated, more feldspar-rich magma generation.

Fig. 1. TiO$_2$ vs. Fe/(Fe+Mn+Mg) in type 1 Luna 16 basalt pyroxenes.
Fig. 2. TiO$_2$ vs. Fe/(Fe+Mn+Mg) in type 2 Luna 16 basalt pyroxenes.

Fig. 3. Average Mg/(Fe+Mn+Mg) in ilmenites as a function of Fe/(Fe+Mn+Mg) in coexisting type 1 pyroxenes having the same concentration of TiO$_2$ (taken arbitrarily to be 2.60 wt. %).
ANALYSIS OF FRAU MAURO SURFACE FINES FOR Porphyrins

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Abstract

Lunar surface fines samples 14163, 179-181 totalling 10 grams were pooled and Soxhlet extracted with benzene: methanol 3/2 v/v for 24 hours. The extract was concentrated to a volume of 0.3 ml and its fluorescence spectrum examined for free base porphyrin with negative findings. The extract was then carried through a demetallation procedure as described previously by Rho and co-workers (1971). The resulting solution was again examined for fluorescence. The amount of solution carried through to final analysis corresponded to 6 grams of lunar fines. A similar solution resulting from the processing of 200-mesh ground optical quartz was examined under the same conditions as were the reagents blank and a solution resulting from demetallation of 0.2 ng of Ni-mesoporphyrin IX. Fluorescence spectra were obtained using an Aminco Bowman Spectrofluorometer with excitation set at 400 nm. Averages of 128 scans of fluorescence spectra from 500 to 750 nm were recorded on a Hewlett-Packard 5480 signal analyzer at 10 seconds/scan. One hundred eighty data points between 555 and 735 nm were processed by the least squares method described by Cohen and Rho (1971). For comparison a second data reduction scheme was used to confirm the absence of computational artifacts. The latter method replaces each data point by the difference between itself and the average of two points equidistant from it in wavelength. When the interval between the two outer points is chosen to be a linewidth the background due to scattering is suppressed to a much greater degree than the spectral features. For the present case a 20 nm interval gave good suppression of background without much loss of the spectrum.

Irrespective of the computational procedure followed, the Apollo XIV extract, the quartz extract and the reagent blank gave features of approximately one fifth the intensity of those of the solution made from 0.2 ng Ni-mesoporphyrin IX. After extensive examination of other solutions and independently constructed blanks and after studying the purity of light from the excitation monochrometer, it was found the weak features resulted from undispersed xenon light being scattered by the sample. The intensity of these
weak features was, in fact, proportional to the amount of light scattered by the sample.

Identification of the weak features as an instrumental artifact justified the subtraction of known blanks from the standard porphyrin spectrum and the Apollo XIV sample. The resulting traces showed that a signal to noise ratio of 2/1 would have been obtained from $10^{-13}$ moles extractable metallo porphyrin in the 6 grams of soil treated. No fluorescence above that of a blank was distinguishable from the Apollo XIV sample extract. A conservative limit of detection on the amount of extractable metallo porphyrin per gram fines is $2 \times 10^{-14}$ moles.

Conclusion: It is apparent from this work and previous results that porphyrins of any interpretable significance are absent from lunar surface fines. If organic molecules of any complexity are to be found, they will probably come from core samples and rock interiors where the effects of solar radiation are much less than on the surface. These potential sources of organic matter should be exploited prior to concluding an absence of porphyrins on the Moon.

References


PETROLOGY OF LUNAR BASALT 14310, W. I. Ridley, Richard J. Williams, Robin Brett*, Hiroshi Takeda, NASA Manned Spacecraft Center, Houston, TX 77058 and Roy W. Brown, Lockheed Electronics Corp., Houston, TX 77058.

Rock 14310 is one of two crystalline rocks greater than 50 gm returned from the Apollo 14 mission, and is particularly important as it represents the only large crystalline rock at the Fra Mauro site whose chemistry resembles that of the Fra Mauro breccias, and the KREEP component in Apollo 11, 12 soils.

Rock 14310 is quite variable in texture, ranging from intergranular to subophitic to ophitic. The rock also is variable in grain size, especially with respect to plagioclase, containing some 10 percent large equant anhedral plagioclase grains; the remainder of the plagioclase is lath-like and covers a range of sizes down to felted plagioclase laths in the groundmass. Some small plagioclase laths are partly included in the equant plagioclase crystals. A mode of the rock is: plagioclase 66%, pyroxene 31%, opaques 2% (ilmenite 72%, chromian ulvöspinel 8%, troilite 16%, Ni-Fe 4%, mesostasis 0.5%). Trace amounts of tranquillite and baddeleyite also occur.

Electron microprobe analysis of large anhedral plagioclase phenocrysts indicates a large compositional variation (Fig. 1); some are unzoned (about An94) others are normally zoned from An91 to An75, whereas others are irregularly zoned with a sodic rim (An75) on one side, a calcic rim on the other (An5), and an intermediate core. Plagioclase laths of intermediate size are also of variable composition (Fig. 2), ranging from unzoned to relatively zoned sodic plagioclase with one rim as sodic as An58. The composition of the groundmass plagioclase laths is fairly regularly zoned with a calcic (An5) to sodic (An72-80) rim (Fig. 3).

It is difficult to explain the compositional variation of the plagioclase grains in terms of crystallization in situ. The variation can readily be explained, however, if one assumes that the 14310 melt contained xenocrysts of plagioclase when it was intruded or extruded. This indicates that the composition of rock 14310 is not that of a pristine liquid. Glasses of 14310 composition are rare in the Apollo 14 soil (Apollo Soil Survey, 1971) which is supporting evidence for the above conclusion.

The pyroxenes of rock 14310 and their compositional trends differ from those of mare basalts. The cores of many pyroxene grains are bronzitic, zoning out through Mg-pigeonite to intermediate pigeonite. The intermediate pigeonite rims (Wo15En40Fs55) commonly contain exsolution lamellae of Ca-poor ferroaugite (Wo32En28Fs40). Some pyroxene grains are composed totally of augite (Wo39En41-Fs20)(Fig. 4). Unlike mare basalts, whose pyroxenes contain metastable compositions and assemblages, it appears that crystallization of pyroxenes in 14310 has been stable, defining the limbs of the pyroxene solvus.

Using x-ray and microprobe data, and the approximate crystallization tempera-
Petrology of lunar basalt 14310

W. I. Ridley

Characteristics of orthopyroxene and clinopyroxene from melting experiments under reduced conditions, the crystallization history of 14310 pyroxenes can be constructed and compared with the Ca-free binary system En-Fs after Kuno (Fig. 5). Our data are remarkably consistent with phase relations in this system, confirming the overall stable crystallization paths of these pyroxenes.

Ratios of Ti/Al in pyroxenes from 14310 (Fig. 6) differ from those of pyroxenes in mare basalts. Orthopyroxenes contain a wide range of Al2O3 content, ranging up to 5%. Magnesian pigeonite, ferropigeonite and augite follow a trend between the 1:2 and 1:3 line explained by a coupled substitution similar to Apollo 11 pyroxenes if some Al is accounted for by Cr. The high alumina content of some orthopyroxenes cannot be totally explained by Ti-Al and Cr-Al coupling, some Al must be octahedrally coordinated.

The high Al2O3 contents of some orthopyroxenes may be explained by crystallization from an aluminous melt with high silica and alumina activities. Bronzites in soil sample 14359 (Apollo Soil Survey, 1971) are not commonly aluminous although clearly derived from an aluminous environment. Crystallization of aluminous orthopyroxene in 14310 prior to extrusion cannot be discounted, but the overall fine-grained texture argues against this explanation. The titanium-aluminium distributions in pigeonite and augite probably indicates increased titanium activity during crystallization. This would be consistent with the lack of precipitation of a high titanium phase prior to and during most of the clinopyroxene crystallization interval.

The order of crystallization of phases is a spinel phase, plagioclase, orthopyroxene, and clinopyroxene + olivine. The presence of plagioclase close to the liquidus and the high temperature crystallization of pyroxene are consistent with our interpretation of the texture of 14310, the x-ray diffraction work, and Fig. 5 (after Kuno). The narrow solidus tends to preclude large scale gravitational differentiation from a 14310 melt, although the density of the synthetic glass (2.85gr/cc) does not preclude plagioclase floatation.

Metal grains are rarely included in pyroxene and never in plagioclase; indicating that metal precipitated late in the crystallization sequence. Therefore the fO2 at least during the bulk of crystallization was lower than for Apollo 11 basalts but higher than for Apollo 12 basalts. The assemblage ulvospinel-ilmenite-Fe was observed indicating subsolidus reaction at an fO2 of approximately 10^-14.5 at 1100°C.

We conclude (1) that the composition of 14310 does not represent that of a pristine silicate melt, (2) the rock type is not abundant in the Apollo 14 soil and is not a major rock type on the lunar surface or interior, (3) the fO2 during crystallization of 14310 was intermediate between that of Apollo 11 and 12 basalts, (4) the bulk of crystallization occurred at higher temperatures than in mare basalts and was largely an equilibrium process, (5) the 14310 melt probably crystallized before the Imbrian event.
ZONAL STRUCTURE AND ORIGIN OF THE MOON

A.E. RINGWOOD, Australian National University

Experimental petrological investigations of mare and highland rocks combined with constraints imposed by the moon's density and moment of inertia suggest that the moon is characterized by radial chemical and mineralogical zoning (1). An outer 'crust', 50-100km thick (ρ≈2.95 g/cm$^3$) is composed of plagioclase + low-Ca pyroxenes ± olivine and is rich in Al$_2$O$_3$ and perhaps also in other involatile components, e.g. rare earths, U,Th(2). This layer, which is the source of highland basalts has lower Fe/Mg and K/U ratios than the underlying pyroxenite "mantle" (ρ≈3.4 g/cm$^3$) which is the source of the mare basalts. The pyroxenite mantle does not contain plagioclase and has lower abundances of Al$_2$O$_3$, CaO and perhaps other involatile components, compared to the crust. Although complicated by subsequent episodes of magmatic differentiation, this zoning appears to represent a primary accretional feature of the moon (1,2).

If confirmed, this inferred zoning, characterized by relative enrichment of involatile elements towards the outer regions of the moon, would constitute a key boundary for theories of lunar origin. It is difficult to reconcile with current versions of the binary planet or capture hypotheses (3,4) which maintain that the moon formed as a separate planet by inhomogeneous accretion in a cooling solar nebula and that relatively volatile elements should be strongly concentrated in the outer layers of the moon. On the other hand, the precipitation hypothesis (5) which maintains that the moon formed from a sediment-ring of earth-orbiting planetesimals precipitated from a massive, hot, primordial terrestrial atmosphere (~0.2 to 0.5 M$_E$) offers a simple explanation of the zoning. The primitive atmosphere was spun out to form a disc in which the least volatile components (e.g. CaO,Al$_2$O$_3$) were precipitated at the highest temperatures closest to the earth whilst components of intermediate volatility (Fe,Mg silicates) were precipitated as planetesimals at lower temperatures and further out. The most volatile components - (e.g. K,Rb,Pb,Zn,Bi, Tl, etc.) were precipitated at the lowest temperatures (<1000°C) and furthest from the earth as sub-micron sized smoke particles. These were too fine to be decoupled from the gases of the primitive atmosphere (CO and H$_2$) and were lost when the massive atmospheric disc was dissipated when the sun passed through the T-Tauri stages. As the atmosphere was lost, the radii of the orbits...
of planetesimals increased \((R_{\text{z}}/M_E)\) whilst Roche's limit decreased somewhat \((R_{\text{z}}M_E^3)\). This caused the zoned ring of planetesimals to expand successively beyond Roche's limit, where sequential coagulation to form the moon occurred. The planetesimals of intermediate volatility (FeMg silicates) coagulated first, producing the lunar mantle whilst the least volatile planetesimals were the last to accrete and thereby formed the lunar crust. The precipitation hypothesis has also been shown to provide a possible explanation of important chemical differences between the earth and moon - particularly the fractionation of metallic iron, the overall depletion of volatile elements in the moon, the differences in \(\text{Fe/Mg}\) and \(\text{Ca+Al/Mg+Si}\) ratios between earth and moon, and the differences in redox states \((5,6)\). Some of these differences are not readily accounted for by alternative hypotheses of lunar origin \((6)\).

Geochemical and geochronological studies indicate extensive near-surface magmatic activity on the moon prior to 4 b.y. ago \((2)\). A possible source of heat may have been partial conservation of gravitational energy during accretion \((6)\). This would require that the moon accreted in 100-1000 years which is consistent with the time-scale for coagulation of a terrestrial sediment ring. If, however, the moon accreted as an independent planet from sun-orbiting material, the accretion period would be orders of magnitude longer and gravitational heating negligible.

It is often claimed that the large discrepancy in relative masses of moon to earth \((1.2\%)\) compared with the relative masses of the satellites of the outer planets to their primaries \((0.01\%\) to \(0.03\%)\) implies a fundamentally different origin for the moon, than for the satellites of the outer planets. This argument is fallacious. When appropriate corrections are applied for the losses of \(\text{H}_2\) and \(\text{He}\) from the satellites of the outer planets, the primordial relative masses of the satellites of Jupiter, Saturn and Neptune to their respective primaries are also found to be on the order of \(1\%). Likewise, when corrections for loss of \(\text{H}_2\) and \(\text{He}\) from the planets of the solar system are applied, the minimum mass of the primordial solar nebula is found to be about \(1\%\) of the mass of the sun. These considerations suggest that the origin of the earth-moon system may not have been fundamentally different from those of the outer planet-satellite systems, nor indeed, of the solar system itself.

If the total angular momentum of the earth-moon system is combined into one body, and likewise for the solar system, it is found that these bodies, the larger asteroids, the outer (giant) planets, and young, massive, O and B stars have rotational periods within factors of 2 to 4 (average 3) of the rotational instability limit, despite the fact that these bodies range in mass over 16 orders of magnitude \((7,8)\). If corrections are made for the
hydrogen and helium which was lost when Neptune and Uranus were formed, the period of a single body representing the combined solar system would exceed the instability limit. The solar system is believed to have evolved via processes which transferred angular momentum from the sun to a nebula from which planets condensed and from which a large mass of gas (carrying angular momentum) escaped. The O and B stars are likewise presumed to have formed at the rotational stability limit and to have lost some angular momentum (presumably by magnetohydrodynamic coupling) to surrounding nebulae. The formation of the regular satellite systems of the giant planets is believed to have been closely analogous to the formation of the solar system. If corrections are made for the light gases lost from the "rocky" regular satellites of Jupiter and Saturn, and from the primordial cloud from which Uranus and Neptune formed, it can be shown that the rotational instability criterion played a key role in the formation of these systems, and that the existing planets must have lost angular momentum to escaping gases (7,8). The larger asteroids must also be primary condensations (7). It is possible that they also formed at the rotational instability limit and lost angular momentum when icy mantles of \( \text{NH}_3, \text{H}_2\text{O} \) and \( \text{CH}_4 \) were evaporated from the sub-solar points when the sun passed through the T-Tauri phase (8).

The fact that the combined earth-moon system possessing a period within a factor of about 2.5 of the rotational instability limit belongs to the hierarchy discussed above strongly suggests similarities between the mode of origin of the earth-moon system and the other systems (as is also indicated by the earlier comparison of mass ratios of primary and satellite systems). It appears likely that the moon and earth were born in a common region of space (9) and are genetically related. Moreover, the overall relationships suggest that the earth-moon system originally formed close to the rotational instability limit and angular momentum was subsequently transferred to a massive primitive atmosphere which subsequently escaped, and that the origin of the moon is linked with the loss of this primitive atmosphere. This is precisely the theory of lunar origin which has been advocated primarily on geochemical grounds by the author since 1960 (10,11,5).

EXPERIMENTAL PETROLOGY AND PETROGENESIS OF APOLLO 14 BASALTS
A.E.RINGWOOD, D.H.GREEN and N.G.WARE, Australian National Univ.

SUB-SOLIDUS MINERALOGY. At 1100°C, basalt 14310 crystallizes to a plagioclase+pyroxene+ilmenite+minor olivine (gabbro, \( \rho \approx 2.95 \text{g/cm}^3 \)) assemblage up to 9kb, to a garnet+pyroxene+plagioclase (granulite) assemblage from 9 to 21kb and to Ca-rich garnet+pyroxene+kyanite+quartz (eclogite, \( \rho \approx 3.5 \text{g/cm}^3 \)) assemblages at pressures >21kb. The KREEP composition behaves similarly except that the transition to a plagioclase-free assemblage occurs at 18kb and the garnets are poorer in Ca. The experimentally determined P,T,\( \rho \) parameters for the gabbro-eclogite transition in KREEP and 14310 in relation to the range of P,T conditions existing within the moon demonstrate that rocks resembling these compositions cannot be representative of the lunar interior. A moon made of such rocks would possess a substantially smaller coefficient of moment of inertia than is observed. This suggests that ultimately, these rock types are differentiates from more primitive source rocks in the lunar interior.

MELTING RELATIONSHIPS: Melting studies of basalt 14310 (in Fe-capsules within evacuated SiO\(_2\) tubes) at 1 atm. show that plagioclase is the liquidus phase (at \( \approx 1300°C \)) joined by olivine (\( \text{Fo}_{88} \)) at \( \approx 1230°C \) and low calcium clinopyroxene at \( \approx 1180°C \). Plagioclase (\( \geq \text{An}_{91} \)) remains the liquidus phase to at least 10kb (1350°C) but at 5kb is joined at 1260°C by olivine (\( \text{Fo}_{84} \)), orthopyroxene (\( \text{Wo}_5 \text{En}_{86} \text{Fs}_{15};8.5\% \text{Al}_2\text{O}_3 \)) and possibly clinopyroxene. Orthopyroxene is absent at lower temperature, its place being taken by low-calcium clinopyroxene. At pressures >7kb clinopyroxene (\( \text{Wo}_1 \text{En}_{72} \text{Fs}_{17};8.0\% \text{Al}_2\text{O}_3 \)) joins plagioclase as the second phase to appear and olivine is absent.

Experimental study of the KREEP composition shows that olivine (\( \text{Fo}_{80} \)) and plagioclase occur on the liquidus (1225°C) at 1 atm. but olivine, plagioclase, sub-calcic augite and pigeonite (or orthopyroxene) occur on or very near the liquidus at 5kb and pyroxenes occur on the liquidus (1250°C) at 7kb joined by plagioclase at 1240°C. Analyses of subcalcic augite-pigeonite pairs at 5kb, 1220°C gave compositions \( \text{Wo}_{16} \text{En}_{55} \text{Fs}_{17};5.7\% \text{Al}_2\text{O}_3 \) and \( \text{Wo}_9 \text{En}_{72} \text{Fs}_{19};3.4\% \text{Al}_2\text{O}_3 \). Runs at 7kb, 1250°C to which orthopyroxene seeds were added showed growth of the seed crystals whereas runs in which only clinopyroxene seeds were present produced only low-calcium clinopyroxenes. The experiments indicate the existence of complex relationships between pigeonite, sub-calcic augite and orthopyroxene which are very sensitive to composition and to temperature.
PETROGENESIS OF APOLLO 14 BASALTS
A.E. RINGWOOD

PETROGENESIS: Optical and electron-probe studies of the natural basalt 14310 demonstrate that the cores of plagioclase laths range from Or0.3Ab4.8An94.9 to Or0.8Ab8.4An90.8 consistent with nucleation over a temperature range during cooling from above the liquidus. Magnesian olivine was not observed in the natural rock and the pyroxenes are complex. The cores of a number of pyroxene crystals (which are in ophitic relationship to plagioclase laths and crystallized with or after the plagioclase), consist of orthopyroxene, containing sparse clinopyroxene exsolution lamellae, especially near their edges. The orthopyroxene cores are rimmed by pigeonite and then by augitic clinopyroxene. Electron microprobe studies show that the orthopyroxene cores range from Wo5 En7 Fs17 2.6% Al2O3 to Wo5 En7 Fs22 2.2% Al2O3 and overlap in composition with pigeonite (Wo5 En7 Fs19 2.1% Al2O3 + Wo5 En7 Fs24 1.9% Al2O3 + Wo5 En7 Fs34 0.8% Al2O3). The natural orthopyroxenes differ significantly from that crystallized at 5kb, 1260°C particularly in Al2O3 content and do not provide evidence for transport of phases formed at high pressure to the lunar surface. Basalt 14310 has MgO/MgO+FeO ratio of 0.62 and partition relationships between liquid and orthopyroxene show that if orthopyroxene crystallizes as the first ferromagnesian phase to appear from 14310, then it should have MgO/MgO+FeO ~0.86. The most magnesian natural orthopyroxene has MgO/MgO+FeO = 0.82 and the experimental study yielded olivine (MgO/MgO+FeO = 0.84) at 1200°C just prior to the appearance of pyroxene. We suggest therefore that the natural rock no longer preserves evidence of olivine crystallization, which in a basalt with ~0.3% normative quartz is of a transitory nature, the olivine reacting with liquid at T<1180°C to precipitate more Fe-rich pigeonitic pyroxene. It is inferred that the orthopyroxene occurring in 14310 results from inversion (at T slightly less than 1180°C) of the most magnesian pigeonite cores of the zoned crystals. More Fe-rich, higher-calcium pigeonites did not invert to orthopyroxene - similar selective inversion of zoned pigeonite crystals is well documented from terrestrial hypabyssal intrusives. We consider that 14310 was originally liquid at T>1300°C and cooled at a rate permitting development of ophitic rather than quench texture, elimination of olivine by reaction with residual liquid, and partial inversion of magnesian pigeonite to orthopyroxene.

RELATIONSHIP TO MARE BASALTS: Electron probe studies and major element chemistry of samples 14053, 14072 and two fragments from breccia 14321, demonstrate that these are high iron, high Cr, olivine basalts of mare type, resembling the Apollo 12 suite. A third fragment from breccia 14321 is a sparkling, white troctolite (35% olivine, Fo85.7; 63% plagioclase An95.6 with accessory chrome-spinel, ilmenite, armalcolite, whitlockite, bronzite and Ba-rich K-feldspar). The rock is interpreted as an equilibrated (?)hornfels), recrystallized accumulate related to basalts resembling 14310 rather than mare basalts.
The high MgO/MgO+FeO ratio, Cr concentration and abundances of K, U, Th and other trace elements suggest that liquids of composition like 14310 are products of small degrees of partial melting of the lunar interior. However, the large field of plagioclase crystallization at pressures up to 10kb shows that 14310 composition could not itself be a partial melt of a polymineralic source rock but must be enriched in plagioclase. Either 14310 represents a partial melt enriched in plagioclase by flotation during ascent of a magma or it may represent an impact melt (?)Imbrium impact) of a complex lunar regolith composed dominantly of basaltic material but enriched in plagioclase. We prefer the second hypothesis. Insofar as the 14310 and KREEP compositions or their progenitors characterize the lunar highlands, we recognize important differences between 'highland basalts' on the one hand and mare basalts (2) on the other. The role of plagioclase on the liquidus at pressures up to and exceeding 5kb, in conjunction with MgO/MgO+FeO ratios, pyroxene compositions, Cr and other trace element abundances suggest that highland basalts may have formed by partial melting of a source region containing plagioclase, orthopyroxene (MgO/MgO+FeO>0.8), subcalcic clinopyroxene ± olivine. In contrast, earlier investigations (2) indicated that mare basalts were formed by small degrees of partial melting of a pyroxene -rich source region (MgO/MgO+FeO<0.8) in which plagioclase was absent. Source regions of highland basalts also appear to be richer in Al₂O₃ than the source regions of mare basalts. These considerations indicate that the source regions of mare basalts were generally deeper (>200km) than the source regions of highland basalts (<200km). Since mare basalts are believed to be younger on the average than the highlands, this implies that the depths of magma formation in the moon, on the average, increased with time. Moreover, it implies that the moon is chemically zoned, the outer layer having a higher MgO/FeO ratio and a higher Al₂O₃ content than the deeper region beneath 200km. The enrichment in Al₂O₃ in the outer layer may also be accompanied by corresponding enrichment of other elements possessing involatile oxides, e.g. Ca, U, Th, Zr, rare earths (3). Occurrence of chemical zoning controlled by volatility is also suggested by K/U ratios which are systematically higher in mare basalts than in highland basalts and KREEP(4).

(4) LSPET reports on Apollo 11, 12, 14 and 15 rock samples.

Compositions Studied Experimentally

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<th>Component</th>
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Crystalline igneous rock samples 14053, 14310, 15058, 15076, 15415, 15475, 15495, 15555; breccia samples 14303, 14305, 14319, 14321, 15405; and soil samples 14162 and 14163 from the Apollo missions, and soil samples 301, 303, 304, 311, 315, 318, and 320 (Smithsonian Astrophysical Observatory numbers) from the Luna 16 mission were examined by microscope and electron microprobe. They exhibit many features seen in Apollo 11 and 12 samples, and some novel ones as well. Our main goal was not to make standard petrographic descriptions and analyses of the major minerals, but rather to obtain crystallization and differentiation trend data from silicate melt inclusions. Coincidental with this examination for minute inclusions various other small-scale features were observed, including some that are not easily interpreted but are reported because they may be significant to other workers.

Silicate melt inclusions in olivine are abundant and occasionally large (300μm) in some clasts in 14321. Olivine in 15555 has many small melt inclusions, mostly <10μm. After trapping, many inclusions in olivine formed blebs of sulfide and daughter crystals of ilmenite and plagioclase (both epitaxial), and pyroxene. Nucleation and growth (epitaxial or random) of these phases varies with inclusion size, gross composition, and probably with cooling history. Solid inclusions in olivine consist of relatively large chrome spinel euhedra, and in part of the olivine crystals rows of very minute chromium-rich crystals have decorated what are apparently dislocations, otherwise invisible. The decorations could have formed during metamorphism of these breccias. The inclusions in olivine of the Luna sample are very similar in appearance, but these olivines also have external epitaxially-arrayed ilmenite plates growing in the groundmass, a feature not seen in Apollo 11 or 12. The gross composition of melt inclusions in olivine varies from about 48 to 62 percent silica.

Melt inclusions in coarse anorthosite rock 15415 are scarce and small (∼1μm). Most occur along healed fractures within individual plagioclase grains, indicating the presence of at least a small amount of melt at the time of fracturing.

Actual melt inclusions in ilmenite (as opposed to the apparent inclusions of matrix phases so common in these skeletal crystals) range in composition from the anomalous low-silica, high-potassium type reported in Apollo 11 to those that are essentially identical with the late-stage, high-silica residual melts. In several samples, melt inclusions in ilmenites suggest trapping at successive stages in a differentiation sequence; these show a systematic distribution of MgO between the inclusions and the host ilmenite.

Melt inclusions related to the onset of immiscibility are common in all the samples. Although somewhat variable, these two immiscible melts are essentially like those in Apollo 11 and 12 -- potassic rhyolite and ferropy-
roksenite in composition. These are best seen as interstitial inclusions of late-stage residual melt (mesostasis) in 14310, where there are many excellent examples of menisci between spherical blebs of colorless high-silica melt (now containing a few minute acicular crystals, presumably feldspar), protruding into almost opaque, high-iron melt (now a very fine grained crystalline aggregate) which fills the interstices between late plagioclase crystals. Some of the later plagioclase and many of the later pyroxene crystals contain melt inclusions in their outer zones, generally of high-silica melt. A few of the melt inclusions within late plagioclase exhibit both immiscible melts and a vapor (shrinkage) bubble, although none of these are as large, clear and abundant as those recently found in 12035,24.

Most of the high-iron melt has crystallized, as in the Apollo 11 and 12 samples. In slowly cooled rocks this now consists of single crystals of fayalite (or pyroxferroite?) containing intergrown, wormy inclusions of high-silica melt. In addition, some of this late-stage material contains blebs of iron metal and (or) troilite, euhedral crystals of apatite (low in rare earths and yttrium), potassium feldspar, tridymite, and cristobalite, plus minor amounts of tranquillityite and other, unidentified phases. The mesostasis in 14053 has some unusual textures, which may indicate the simultaneous occurrence of four immiscible melts: the two silicate melts, plus iron-rich metal and iron sulfide melts.

The two immiscible silicate melts at the several sites have similar compositions for the major components, except for some very significant differences in the alkali ratios. There is an inverse relationship between the $K_2O/Na_2O$ ratio in the bulk rock and its residual high-silica melt.

In any given cogenetic rock series the early melt inclusions in olivine, the later melt inclusions in plagioclase, ilmenite and ulvospinel, and the late-stage immiscible melt inclusions should form a continuous series, delineating the liquid line of descent (with its split into two lines at the onset of immiscibility). The limited number of samples studied and the small volumes available for analysis have frustrated efforts to obtain a complete series of successive melts representing the liquid line of descent for any given magma. However, when the compositions of all silicate melt inclusions that we have analyzed, from all hosts and all five sites, are plotted on a silica variation diagram the results are surprisingly consistent, showing a generally similar liquid line of descent for most of these samples.

Many of the above features are present in the breccias and soil samples, but less suitable for study. These latter also exhibit many petrologically important features seen poorly or not at all in the crystalline igneous rocks. Individual pyroxene crystal fragments show a variety of exsolution textures, including oriented platelets of ilmenite, as in some terrestrial rocks. In some of the metamorphosed breccias, olivine grains are surrounded with a reaction rim of pyroxene. Shock damage to plagioclase or anorthosite, followed by later recrystallization, has yielded some novel textures.

Many grains of a reddish-purple spinel were found in several samples of soil and breccia from the Apollo 14 site. Two major composition groups have been found by electron microprobe, one approximately $(Mg_{0.55}Fe_{0.47})(Al_{0.91}Cr_{0.09})_2O_4$, and the other $(Mg_{0.58}Fe_{0.42})(Al_{0.98}Cr_{0.02})_2O_4$, with only minor
amounts of other ions. These differ greatly from all previously recorded lunar spinels, and no spinels of intermediate composition were found. A darker rim on some of these spinels, and a narrow zone of almost pure plagioclase, free of pyroxene and opaques, in the breccia matrix at the contact indicate reaction between the spinel and the breccia matrix. One crystalline rock fragment was found in which relatively coarse grained spinel occurs in a possibly cumulus texture with olivine and plagioclase, which suggests that such a rock may be the source of the spinel. Several others have spinel embedded in single plagioclase crystals. Another clast, itself a breccia, in the Apollo 14 breccias, contains many grains of this spinel, along with olivine and plagioclase but no pyroxene, implying the existence of a spinel-rich regolith somewhere. As only a few grains of this spinel have been recognized in all the work on the Apollo 11, 12, 15 and Luna 16 samples, despite its striking appearance, and as it is abundant in the Apollo 14 samples, it seems reasonable to assume that exposure(s?) of this spinel-rich rock must be very restricted on the Moon, and must be near to the source of the Apollo 14 materials. These data are also in agreement with the concept of essentially local origin for most soil fragments, i.e., relatively little lateral mixing.

Glass fragments and spherules in the lunar fines from all three sites show a range of crystallization textures which vary with the individual cooling history and composition. Using the microscope crushing stage (Roedder, 1970, Schöiz, Min. Pet. Mitt., v. 50, p. 41-58), the bubbles in some of these glasses are now found to contain $10^5$-$10^{10}$ molecules of (unidentified) noncondensable gas (pressure <0.01 atm); presumably these are mainly solar wind gases from the soil that entered (or formed?) the bubbles during the fusion to form the glass.

A very small fraction of the glass spheres, and nearly 10% of the glass dumbbells in lunar soil 14163 show two immiscible glasses, one light and the other dark red brown in color. The latter was apparently higher in density as a melt, as those dumbbells that have dark glass in one end only are asymmetrical, with that end smaller, as would be expected during rotation while in free fall. More significant, however, is the occurrence of several areas and fragments of glass of potassic rhyolite composition in the Apollo 14 breccias. These glasses have slightly lower CaO and FeO, and appreciably higher $K_2O$ than the high silica immiscible melt. We believe, however, that these represent fragments of a rhyolitic phase that separated as a result of liquid immiscibility and was concentrated somewhere at the surface of the Moon relatively near to the source of the materials for the Apollo 14 site.
COMPOSITIONAL DATA FOR FIFTEEN FRA MAURO LUNAR SAMPLES,
H.J. Rose, Jr., F. Cuttitta, C.S. Annell, M.K. Carron,
R.P. Christian, E.J. Dwornik, and D.T. Ligon, Jr., U.S. Geological

The major, minor, and trace element compositions of fifteen
lunar samples (4 soils, 4 basalt rocks, and 7 breccias) collected
at the Fra Mauro Formation by the Apollo 14 Mission were
determined by combined semi-micro chemical, X-ray fluorescence,
and optical emission techniques.

SOILS

The Fra Mauro lunar soils show a greater homogeneity than the
soils from the Tranquility Base (Apollo 11 Mission) and the Ocean
of Storms (Apollo 12 Mission) sites, with little variation being
found in either the major, minor, or trace elements. Si, Na, K,
P, Rb, Li, Ba, Be, Nb, La, and Zr display notable concentration
trends characterized by Apollo 11 soils < Apollo 12 soils < Apollo
14 soils. Ti, Fe, and Sc show a reverse trend. There is also a
notable increase in the Pb content of the Fra Mauro highland
materials (average about 10 ppm) as compared to that found
previously for the mare lunar materials (about 1 ppm). The Fe/Ni
ratio of the Apollo 14 soils (218) is about one-third of that
found for the Apollo 11 (661) and 12 soils (617). The Ba/Sr
ratios found for the lunar soils show a trend increasing from
Apollo 11 through Apollo 14 (Apollo 11, 1.6; Apollo 12, 4.3; and
Apollo 14, 5.5).

BASALT ROCKS

Comparison of the Fra Mauro basalt composition with those
from Tranquility Base and Ocean of Storms shows that Si, Al, Ca,
Na, P, and Ni content increases from Apollo 11 through Apollo 14
with Fe, Ti, Cr, and Sc having the reverse trend. Fra Mauro
basalts are highest in K, Pb, Ba, Sr, Be, Nb, La, and Y, but are
lowest in Ca, Co, and V. The Fe/Ni ratios drop sharply in samples
from each successive site (Fe/Ni: 21,000, 2520, 504) while Ni/Co
increases (Ni/Co: 0.23, 0.98, 8.9). Fra Mauro basalts show the
highest Rb/Sr, Cr/V, Ba/V and Ba/Sr ratios and the lowest Cr/Ni
ratio.
The breccia samples from Fra Mauro also show the highest Si, Al, Mg, Na, K, Zn, Cu, Li, Rb, La, Ba and Zr concentrations when compared to the Tranquility Base and Ocean of Storms breccias (Apollo 14 breccias > Apollo 12 > Apollo 11). Fe, Ti, Sc and Mn show a reverse trend. Fra Mauro breccias show the lowest Cr and V content. Compared to materials returned by earlier Missions, the trace elements Ni and Y are highest in Apollo 14 breccias. The Fra Mauro breccias also have the highest Ba/V, Ba/Sr, and Rb/Sr ratios (Apollo 14 > Apollo 12 > 11). Cr/Ni and Fe/Ni ratios are lowest in Fra Mauro breccias while Ni/Co is highest.

Preliminary evaluation of these and published elemental, photographic, and mineralogic data suggests:

1. Most of the Apollo 14 samples are breccias. This strongly suggests that the breccias that characterize the Fra Mauro Formation are a gigantic ejecta blanket generated by the impact which formed the largest of the lunar seas - the Mare Imbrium.

2. Optical microscopic and photographic examination shows that the Fra Mauro highland material is more complex mineralogically than the Apollo 11 and 12 materials implying a different manner of formation or a history different from rocks returned from lunar mare regions.

3. The Ni content of the Fra Mauro fragmental rock is significantly lower than that of the regionally associated soils in contrast with Apollo 11 samples where fragmental rocks are similar in Ni content to that of the soils.

4. The Fra Mauro highland rocks differ from the mare basalts in having much lower contents of Fe, Ti, Mn, and Cr and higher concentrations of Si, Al, Na, K, Pb, Li, Rb, Be, Nb, La, and Zr.

The average major, minor and trace element composition for the basalt, breccia and soil samples we have analyzed is given in the following table:
DATA FOR FIFTEEN FRA MAURO
H.J. ROSE, JR.

Average composition for Apollo 14 Samples analyzed (in %)

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Basalt1</th>
<th>Breccia 2</th>
<th>Fines 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO$_2$</td>
<td>47.70</td>
<td>47.83</td>
<td>47.93</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>21.44</td>
<td>16.79</td>
<td>17.60</td>
</tr>
<tr>
<td>Fe$_2$O$_3$</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>FeO</td>
<td>7.78</td>
<td>9.96</td>
<td>10.37</td>
</tr>
<tr>
<td>MgO</td>
<td>7.29</td>
<td>11.03</td>
<td>9.24</td>
</tr>
<tr>
<td>CaO</td>
<td>13.05</td>
<td>10.45</td>
<td>11.19</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.70</td>
<td>0.87</td>
<td>0.68</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>0.48</td>
<td>0.56</td>
<td>0.55</td>
</tr>
<tr>
<td>H$_2$O$_2$</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>1.16</td>
<td>1.53</td>
<td>1.74</td>
</tr>
<tr>
<td>P$_2$O$_5$</td>
<td>0.42</td>
<td>0.55</td>
<td>0.53</td>
</tr>
<tr>
<td>MnO</td>
<td>0.11</td>
<td>0.13</td>
<td>0.14</td>
</tr>
<tr>
<td>Cr$_2$O$_3$</td>
<td>0.25</td>
<td>0.21</td>
<td>0.25</td>
</tr>
<tr>
<td>Total</td>
<td>100.38</td>
<td>99.91</td>
<td>100.19</td>
</tr>
<tr>
<td>ARC</td>
<td>0.32</td>
<td>1.36</td>
<td>2.80</td>
</tr>
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</table>

Average trace element composition (in ppm)

<table>
<thead>
<tr>
<th>Element</th>
<th>Basalt</th>
<th>Breccia</th>
<th>Fines</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pb</td>
<td>11</td>
<td>16</td>
<td>10</td>
</tr>
<tr>
<td>Zn</td>
<td>4</td>
<td>15</td>
<td>25</td>
</tr>
<tr>
<td>Cu</td>
<td>9.0</td>
<td>22</td>
<td>18</td>
</tr>
<tr>
<td>Ga</td>
<td>3.7</td>
<td>4.7</td>
<td>5.5</td>
</tr>
<tr>
<td>Li</td>
<td>22</td>
<td>25</td>
<td>23</td>
</tr>
<tr>
<td>Rb</td>
<td>14</td>
<td>14</td>
<td>13</td>
</tr>
<tr>
<td>Co</td>
<td>13</td>
<td>35</td>
<td>35</td>
</tr>
<tr>
<td>Ni</td>
<td>116</td>
<td>345</td>
<td>370</td>
</tr>
<tr>
<td>Ba</td>
<td>740</td>
<td>800</td>
<td>1030</td>
</tr>
<tr>
<td>Sr</td>
<td>170</td>
<td>147</td>
<td>189</td>
</tr>
<tr>
<td>V</td>
<td>37.5</td>
<td>50.0</td>
<td>55.6</td>
</tr>
<tr>
<td>Be</td>
<td>4.1</td>
<td>7.0</td>
<td>6.6</td>
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<tr>
<td>Nb</td>
<td>31.0</td>
<td>48.0</td>
<td>55.3</td>
</tr>
<tr>
<td>Sc</td>
<td>22.5</td>
<td>22.0</td>
<td>26.7</td>
</tr>
<tr>
<td>La</td>
<td>59</td>
<td>76</td>
<td>74</td>
</tr>
<tr>
<td>Y</td>
<td>192</td>
<td>253</td>
<td>276</td>
</tr>
<tr>
<td>Yb</td>
<td>16.0</td>
<td>22.0</td>
<td>25.0</td>
</tr>
<tr>
<td>Zr</td>
<td>615</td>
<td>790</td>
<td>813</td>
</tr>
</tbody>
</table>

1) Basalts: samples 14276, 8 and 14310, 114
2) Breccias: samples 14066, 21(203), 14318, MF and 14318, 27A
3) Fines: samples 14003, 30; 14049, 37; 14163, 54; 14240, 9; 14259, 12; and 14421, 23
The isotopic compositions of uranium and thorium (Table 1) have been determined on three samples from the Apollo 14 mission: 14116 (bulk fines), 14259 (comprehensive fines), and rock 14310. The Pa$^{231}$/U$^{235}$ ratio was determined in 14163 by the neutron activation technique described by Rosholt and Szabo (1969). The daughter product isotopes of uranium, thorium, and protactinium analyzed (Table 1) are in radioactive equilibrium with parent nuclides in these samples. Dissolved lunar samples from which uranium and thorium were purified were obtained from M. Tatsumoto. Neutron irradiation of a separate portion of fines for Pa$^{231}$ determination was done in the GSTR TRIGA facilities as was neutron irradiation of selected portions of highly purified thorium previously extracted from Apollo 11, 12 and 14 samples.

Clues to Existence of an Isomer of Th$^{232}$. Of more interest are observations resulting from neutron irradiation of purified thorium from both lunar and terrestrial rocks. Previous to irradiation with a total flux of approximately $5 \times 10^{18}$ neutrons, purified thorium had been electrodeposited on pure titanium or molybdenum counting discs. The purpose of this investigation is to determine if a suspected isomer of Th$^{232}$ (Th$^{232_m}$) can be distinguished from Th$^{232}$ by a difference in the reaction products produced from neutron irradiation of thorium; the existence of such an isomer was suggested from interpretation of anomalous Th$^{232}$/Th$^{230}$ activity ratios in crystalline rocks from Apollo 11 and Apollo 12 missions (Rosholt and Tatsumoto, 1971).

The Th$^{232}$ (n,γ) Th$^{233}$ β$^-$ Pa$^{233}$ β$^-$ U$^{233}$ reaction will produce a measurable quantity of alpha-particle emitting U$^{233}$ on the counting disc. In addition to the expected U$^{233}$ produced, a relatively large and unexpected activity of a 5.3 MeV alpha emitter was also produced. Preliminary measurements indicate that a portion of the 5.3 MeV alpha activity decays with a half-life of approximately 40 days. A significant fraction of the activity in the alpha peak is from Po$^{210}$ (5.30 MeV, 138.4-day half-life) and a very small fraction may be from U$^{232}$ (5.31 MeV, 72-year half-life) produced by the reaction: Th$^{232}$ (n,2n) Th$^{231}$ β$^-$ Pa$^{231}$ (n,γ) Pa$^{232}$ β$^-$ U$^{232}$ (Hyde et al., 1964). Neutron capture by bismuth impurities in the thorium or in the metal discs could produce Po$^{210}$ by the reaction: Bi$^{209}$ (n,γ) Bi$^{210}$ β$^-$ Po$^{210}$.

Activity at the 5.3 MeV alpha energy has been produced in all irradiated thorium that has been measured but in variable quantities from one type of thorium to the other. If indeed this unidentified alpha activity was produced
Isotopic Composition of Th, Po and U in Lunar Samples

John N. Rosholt

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as a reaction product from neutron irradiation of Th$^{232m}$, it will provide a very sensitive technique to estimate the abundance of Th$^{232m}$, even in very small proportion, in lunar and terrestrial thorium. The U$^{233}$ alpha activity produced can be used as an internal flux monitor in the thorium itself and the 5.3 MeV/U$^{233}$ activity ratio can be used to estimate the abundance of Th$^{232m}$ in the thorium irradiated. Preliminary measurements by alpha spectrometer indicate that this ratio varied from about unity in lunar thorium from sample 12033 to about 2 in thorium from 14310 and #3633 terrestrial granite reference, to about 2.5 in thorium from terrestrial obsidian reference. Based on Th$^{230}$/Th$^{232}$ activity ratios in Apollo 12 samples (Rosholt and Tatsumoto, 1971), 12033 should have only a very small amount of Th$^{232m}$ relative to lunar crystalline rocks and data in Table 1 indicate that 14310 may also have only a small amount of isomer. Apparently our terrestrial reference samples have a slightly greater Th$^{232m}$ abundance ratio than that in most lunar fines and breccia samples that have been analyzed.

REFERENCES


Isotopic Composition of Th, Po and U in Lunar Samples
John N. Rosholt

Table 1. Isotopic ratios of uranium, protactinium, and thorium.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>$\text{U}^{238}/\text{U}^{235}$ (atom ratio)</th>
<th>$\text{U}^{234}/\text{U}^{238}$ (activity ratio)</th>
<th>$\text{Pa}^{231}/\text{U}^{235}$ (activity ratio)</th>
</tr>
</thead>
<tbody>
<tr>
<td>14163</td>
<td>Fines</td>
<td>137.6+3.3</td>
<td>1.00+0.02</td>
<td>0.97+0.04</td>
</tr>
<tr>
<td>14259</td>
<td>Fines</td>
<td>137.9+3.3</td>
<td>0.99+0.02</td>
<td>n.d.</td>
</tr>
<tr>
<td>14310</td>
<td>Rock</td>
<td>137.6+3.3</td>
<td>0.98+0.02</td>
<td>n.d.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Sample</th>
<th>$\text{Th}^{232}/\text{U}^{238}$ (atom ratio)</th>
<th>$\text{Th}^{232}/\text{Th}^{230}$ (activity ratio)</th>
<th>Expected $\text{Th}^{232}/\text{Th}^{230}$</th>
<th>Measured $\text{Th}^{232}/\text{Th}^{230}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>14163</td>
<td>3.66</td>
<td>1.15+0.02</td>
<td>1.02+0.03</td>
<td></td>
</tr>
<tr>
<td>14259</td>
<td>3.66</td>
<td>1.19+0.02</td>
<td>.98+0.03</td>
<td></td>
</tr>
<tr>
<td>14310</td>
<td>3.65</td>
<td>1.14+0.02</td>
<td>1.03+0.03</td>
<td></td>
</tr>
</tbody>
</table>

1/ Values for $\text{Th}^{232}/\text{U}^{238}$ atom ratios were furnished by M. Tatsumoto and his coworkers.

2/ Expected $\text{Th}^{232}/\text{Th}^{230}$ activity ratio was calculated from $(\text{Th}^{232}/\text{U}^{238})_\text{atom} \times (\lambda_{232}/\lambda_{238})$, where $\lambda_{232} = 4.88 \times 10^{-11} \text{yr}^{-1}$ and $\lambda_{238} = 1.537 \times 10^{-10} \text{yr}^{-1}$. 
IMPLICATIONS OF THE MAGNETISM AND FIGURE OF THE MOON

Professor S.K. Runcorn, Department of Geophysics and Planetary Physics, School of Physics, University of Newcastle upon Tyne, Newcastle upon Tyne NE1 7RU, England.

The study of the magnetic properties of the Apollo samples shows that there is a primary component of the natural remanent magnetization which has considerable stability. In some cases there is also a secondary component of low stability and obscure origin. The stable component seems to have been acquired in the early history of the rock formation of which the Apollo samples were once part. The conclusion that permanent magnetism is rather widespread in the lunar crust can reasonably be concluded from the Apollo sample studies. The fields of the order of $10^{-4}$ over 10 -100 kms scale near the surface of the Moon inferred from the disturbance in the geomagnetic tail and measured by magnetometers at and near the surface led also to the same conclusion. Fields of this order from a uniformly magnetized layer only occur at its ends. The observation that such fields are common on the highlands and are particularly associated with the large craters fits therefore the view that the highlands are generally magnetized and produce 'fringing' fields where the highlands have been 'punctured' by impacts. The 36 $\gamma$ field observed by the Apollo 12 magnetometer seems related to the fact that Oceanus Procellarum is filled with the Fra Mauro formation and the Apollo 12 landing site is on a basalt flow filling an ancient valley in this formation. Similarly the absence of such a field reported by the Apollo 15 magnetometer is a consequence of the uniform layering of lava at the Hadley site.

Whether the stable component of magnetization is of thermoremanent origin has yet to be determined but the presence of a steady field of about 1,000 gamma at the time of magnetization seems necessary. Unless the magnetization is obtained instantaneously, and this seems unreasonable in view of the evidence referred to of uniformity of magnetization on a large scale, only that component of an external field parallel to the axis of rotation of the Moon would be effective. This consideration seems to rule out an appeal to a supposedly enhanced solar wind magnetic field as a magnetizing agent, for in the long run its mean vector must lie in the plane of the ecliptic. This leaves the dipole fields of the Sun and Earth as the only possible external causes of the magnetization. The former possibility seems impossible for the solar polar field would have had to be about 200,000 G – an order of magnitude greater than that observed in any star. While a close approach of the Moon to the Earth is possible dynamically, it appears that the Moon could only have been at the required distance, i.e. 2 - 3 earth radii, for a very short time whereas the ages of the magnetized rocks on the Moon now spread over approach 10$^7$ years. The conclusion is reached therefore that the Moon possessed a field of internal origin from its early history until some time later than 3,200 million years ago.
Thermal convection in the Moon has been postulated to explain its marked non-hydrostatic figure. Deviations of the gravity field from that corresponding to hydrostatic equilibrium are due to two causes, finite strength and density differences associated with convection. It is apparent that the relatively small scale gravity anomalies on the Moon, particularly the mascons, are maintained by the finite strength of the lunar crust. The low harmonics may, however, result from convection in the deep interior. The discrepancy between the actual external shape of the Moon and its gravity field gives important information about the departure of its interior from uniformity in density and this is held to be clear evidence that the shape of the Moon was not due to distortion of an essentially uniform body in its early history. The shape of the global surface of the uplands is very similar to that of the mare and quite different from the equipotential gravitational surface. It is concluded therefore that when the very fluid lavas filled the maria basins, the present low harmonics in the gravitational field were absent. This reasoning suggests that at that time the external shape of the Moon was very close to the hydrostatic value and has since been distorted. Thermal convection explains this. Analysis of the heights of the lunar surface from the geometrical librations shows the surface of the mascons to be lower than the other mare: in particular Mare Crisium being lower than the others. This suggests that the added load has caused the lithospheric shell to be depressed in the region of the mascons. The origin of the mascons requires the lithospheric shell to have had finite strength at the time of its formation as well as subsequently, and it requires material to be transported from the site. This is possible if it is assumed that about half a billion years after the Moon formed the temperatures at a few hundred kilometres depth rose to values of around 900°C at which partial melting allowed the separation of the anorthositic highlands. This more or less complete shell was then punctured by the great impacts which form the circular mare and about one billion years after the Moon's origin these basins filled with lava from cracks in the lithospheric shell extending down to a region between 200 and 300 kilometres which by then had reached temperatures of about 1200°C at which basaltic lava can separate. The filling of the mare basins through the 'hydrostatic head' from this region would lead to isostatic equilibrium between the mare and the highlands had the lava in the basins remained molten. It can be shown that the 10 percent contraction of lava on solidification allows the magma to flow out in the mare basins resulting in an additional 1.5 kilometres of lava required to explain the mascons, provided that the depth of the lava in the mare basins is about 20 kilometres. The absence of gravity anomalies associated with the other mare must mean that before this process began they were already filled with debris. The second degree harmonic in the convection pattern required to give the observed difference in moments of inertia requires the existence of a small iron core, probably about 0.2 of the lunar radius. In such a small core dynamo action could once have produced a magnetic field, but convection cools the Moon and the field could have been removed either by solidification of the iron core or because the magnetic Reynolds number fell below the critical value.
IMPLICATIONS OF THE MAGNETISM AND FIGURE OF THE MOON

S.K. RUNCORN

The various implications of these ideas for the thermal history thus almost require thermal convection to be occurring as this would be a much more effective means of cooling the Moon than conduction.
MAGNETIC PROPERTIES OF LUNAR ROCKS AND FINES.
S.K. Runcorn, D.N. Collinson, W.O. Reilly and A. Stephenson,
Department of Geophysics and Planetary Physics, School of
Physics, The University, Newcastle upon Tyne, England.

1. Remanent Magnetism. This has been investigated in one
sample of crystalline rock and several samples of the
fragmental material (Apollo 14) and a further sample of Apollo
11 crystalline rock. Intensity of NRM is in the range 2.0
- 220 x 10^-6 emu/g, the highest value being observed in the
crystalline sample 14053,35.

Alternating field demagnetization of the NRM reveals
an apparently complicated history of magnetization, with two
components present, and possibly more. All samples show a rapid
initial decay of NRM in fields of up to 75 Oe (with the
exception of 10020,4F, which was highly stable) and changes of
direction of several tens of degrees. Higher fields indicate
the presence of a harder component both through a change of
slope in the intensity curve and a more persistent direction
of NRM. In 14053,35 and a fragmental sample (14305,56) a rise
of intensity occurs around 150 Oe, followed by a region of low
and variable intensity but essentially constant direction up
to peak fields of 500 Oe. The intensity of the harder component
in the four samples tested is 1 - 5 x 10^-6 emu/g. The NRM of
14053,49 during thermal demagnetization shows a steady decrease
up to 800°C, which suggests that the hard component is carried
by iron; the heating was done in a vacuum of 10^-5 torr, and no
oxidation was detected.

Growth of viscous magnetization has been measured in
several samples and is significant in all of them; 14053,35
shows the most rapid growth (s = 3.0 x 10^-6 emu/g/Oe), but in
no sample can the observed NRM be attributed to a VRM acquired
in any reasonable time and field on the moon. Rather low fields
are required to give the samples an isothermal remanence equal
to their NRM, varying from 7 to 22 Oe; however, the stability
of this IRM to both thermal and A.F. demagnetization is
significantly less than that shown by the NRM of the rocks.
Some samples easily acquire a 'shock' magnetization in a weak
field (~ 1 Oe); very moderate, mechanical shocking of 14053.35
produces a magnetization of about 10^-6 emu/g along the ambient
field direction.

All samples tested show that magnetic particles are
present which are capable of carrying a hard magnetization,
by demagnetization of the saturated IRM.
The properties of single magnetic particles extracted from both fragmental and crystalline rock have been examined, using a new small-scale magnetometer. Two groups of particles are distinguishable in 14053,35, both of which have high susceptibility averaging about $4 \times 10^{-3}$ and $3 \times 10^{-2}$ emu/g/0e; only the latter group appear to possess any NRM (typically 0.1 emu/g) which, in the particles measured, is very soft and can be removed in less than 100 Oe. Saturation IRM of this group averages 1.5 emu/g and is acquired in 600 Oe or less. The particles are not generally homogeneous and are not easy to identify. They do, however, rust in a moist atmosphere, and this feature, combined with their strong magnetic properties is evidence that they consist at least partly of native iron. This work is proceeding, but it now seems probable that the hard and soft component of NRM is carried by iron in this rock.

2. Rock Magnetism. The Apollo 14 fines have properties very similar to those returned from previous missions. A single Curie point of 775°C is observed, an initial susceptibility of $3.0 \times 10^{-3}$ emu/g/0e, and the magnitude of induced and isothermal remanent magnetization and saturating field again indicate the presence of a range of fine-grained iron particles which contribute the magnetic properties.

The Apollo 14 rock samples have initial susceptibilities in the range 120 - 650 $\times 10^{-2}$ emu/g/0e, the highest values being in the crystalline sample; all the values are considerable higher than in the Apollo 11 and 12 rocks. 14053,35 is remarkable in having rock magnetic properties comparable with the fines, which is presumably due to its possessing a similar range of fine-grained iron. The Curie temperatures of all samples so far measured indicate native iron to be the most important magnetic mineral present.

Electron microprobe analysis is being carried out on a selection of the 'metallic' grains present in the rocks to establish any correlation between composition and magnetic properties. The analysis of 9 such grains from 14306,63 shows that they are homogeneous and predominantly of iron, with nickel content of 1 - 18%.

Low Temperature measurements. Some rock chips have been cooled to -196°C in zero field to observe any changes in NRM. Sample 10049,1 (crystalline) showed a reversible increase by a factor of 2.7, whereas 14053,35 showed an irreversible decrease to about 50% of the original NRM. A similar result occurred when the saturated IRM acquired at room temperature was cooled in zero field to -196°C. The decrease produced by this treatment was 6%, but on warming to room temperature a further decrease of 6% was observed. This behaviour was also seen in 14318,38 and is tentatively ascribed to
irreversible domain wall movement, presumably in multi-domain iron grains.

The TRM acquired on cooling to -196°C in a field of 1.2 Oe has been investigated in 14053,35 and 14318,38. The TRM in the former showed a sharp decrease at about -150°C during re-warming to room temperature, almost certainly indicating the presence of ulvöspinel; the fragmental sample (14318,38) showed a much weaker TRM, with no discontinuity on warming, suggesting a much smaller quantity, or absence of ulvöspinel. The behaviour of saturated IRM between -196°C and 20°C shows the absence of detectable amounts of magnetite in these rocks.

3. Discussion. Native iron, with or without small amounts of nickel, is confirmed as the dominant magnetic mineral in lunar rocks, and it appears to carry the NRM.

A great variability of stability of the NRM of the rocks has become increasingly apparent, but several samples do possess a component of remanent magnetization which is stable in the range 300 - 500 Oe and of 1 - 5 x 10^-6 emu/g in intensity. Thermal demagnetization of the NRM and A.F. demagnetization of laboratory-produced TRM is consistent with the origin of the hard component being thermoremanence acquired in a weak field on the moon.

The additional soft component which is usually present may be attributed to one or more of several processes. In the rocks examined, its direction is within the same octant as that of the hard component, which may have some significance; a puzzling feature is the decay of the NRM on moderate heating (~150°C), since this component would be expected to be eliminated by thermal cycling on the moon. Although it seems unlikely, it is perhaps necessary to consider a soft component of NRM being acquired by the rocks during or after their return to Earth, in addition to a lunar origin.

In the absence of knowledge about the effects of such lunar features as prolonged thermal cycling, radiation, and meteoritic shocks, the evidence from the Apollo 14 rocks is that they possess a weak, stable NRM with properties suggesting a thermoremanent origin in a weak field of probably less than 0.1 Oe. They subsequently acquired a substantial component of soft magnetization, for which the most likely origin seems to be a shock or isothermal remanent process associated with meteoritic impact(s) which distributed the rocks over the lunar surface.
DISTRIBUTION OF CARBON, SULFUR AND NITROGEN IN APOLLO 14 AND 15 MATERIAL

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University of California
Institute of Geophysics and Planetary Physics
Los Angeles, California 90024

Three samples have been analysed at the time of writing this report, one Apollo 14 fine (No. 14298), one Apollo 15 fine (No. 15271) and an inside fragment of Apollo 14 (coarse breccia, (No. 14321). In the latter sample, the dark mineral fraction was dominant in the fragment analysed. Results are shown in Tables 1 and 2.

The carbon content of the two fines is in the same range as that analysed on Apollo 11 and 12 samples (between 100-150 ppm). The Apollo 14 sample was found to be isotopically lighter (+11.0%) than the Apollo 15 sample (+15.8%). Both are significantly enriched in C\textsuperscript{13} with respect to PDB carbonate standard and fall in the range of previous measurements on Apollo 11 and Apollo 12 material. The breccia contained only 27 ppm carbon and yielded an isotopic value of -17.3\% This value is somewhat heavier than previous measurements of S\textsuperscript{13} on igneous lunar rocks and may represent a true isotopic value for indigenous lunar carbon.

Acid hydrolysis of the lunar fines 14298, with hydrochloric or sulfuric acid yielded non volatile carbon gases in the range of 20 to 30 ppm. It is believed these gases are probably methane, derived from hydrolysis of metal carbides. A sieved aliquot of the fines (<50\(\mu\)) was treated with 6N H\textsubscript{2}SO\textsubscript{4} at three different temperatures (-12\(^\circ\),+13\(^\circ\),+95\(^\circ\)C), and the gases collected for analysis. Table 2 shows that \(\delta\text{C}^{13}\) became progressively heavier during hydrolysis, the weighted average being a little lighter than \(\delta\text{C}^{13}\) for the total unsieved sample.

The fraction of methane released was least at the first temperature and greatest at the intermediate temperature. In contrast, the largest amount of hydrogen sulfide was released during the first low temperature hydrolysis (Table 3). Several experiments indicated that from 25 to 50\% of the total sulfide can be rapidly released at temperatures of -20 to -10\(^\circ\)C by acid hydrolysis. The sulfide cannot be removed by water hydrolysis and is therefore not in the form of an alkaline earth sulfide (i.e. CaS, oldhamite). As in the case of carbon gases, the last sulfide fraction released is the heaviest. It is apparent from
Tables 1 and 3, that considerable enrichment in $S^{34}$ has occurred in the sulfide of the finest size fraction. It is further apparent, that sulfide in the fines may have arisen from two sources. One highly enriched in $S^{34}$ and the other with an isotopic value closer to the sulfur of lunar crystalline rocks.

Analysis for hydrolyzable nitrogen by heating the sample at 400°C with 6N H$_2$SO$_4$ (Table 4), indicates that a large amount of the nitrogen previously reported by Moore and coworkers may be accounted for either as nitride or ammonia.

During hydrolysis, hydrogen gas and helium were captured from the 50-μm fraction of 14298. The volume of He (0.116 cc/g) is low, probably indicating it has not been efficiently released during hydrolysis. However, 5.6 cc/g H$_2$ was released. On the assumption that it almost completely arises from solution of native iron, we calculate that it represents 1.36% Fe. Thus in this fraction of the fines the Fe/FeS ratio is 5.8.

**Table 1** Concentration and Isotopic Ratio of Total Carbon and Sulfur

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>C (ppm)</th>
<th>$\delta^{13}C$</th>
<th>S (ppm)</th>
<th>$\delta^{34}S$</th>
</tr>
</thead>
<tbody>
<tr>
<td>14298.4 (fines)</td>
<td>138</td>
<td>+11.0</td>
<td>920</td>
<td>+9.7</td>
</tr>
<tr>
<td>14321.49 (breccia)</td>
<td>27</td>
<td>-17.3</td>
<td>337</td>
<td>+9.4</td>
</tr>
</tbody>
</table>

**Table 2** Non-condensable Carbon Gases by 6N H$_2$SO$_4$ Hydrolysis of 14298.4 (<50μm) at Three Temperatures

<table>
<thead>
<tr>
<th>Exp. No.</th>
<th>Temp. °C</th>
<th>Period hr</th>
<th>Carbon Released</th>
<th>$\delta^{13}C_{PDB}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-12</td>
<td>1</td>
<td>20.6%</td>
<td>-2*</td>
</tr>
<tr>
<td>2</td>
<td>+13</td>
<td>1</td>
<td>47.1</td>
<td>+7.2</td>
</tr>
<tr>
<td>3</td>
<td>+95</td>
<td>60</td>
<td>32.3</td>
<td>+13.9</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Weighted average</td>
<td>+8.3</td>
</tr>
</tbody>
</table>

* Questionable value. Measured by isotope dilution
Distribution of Carbon, Sulfur and Nitrogen in Apollo 14 and 15 Material

H. Sakai

Table 3 Hydrogen Sulfide by 6N H$_2$SO$_4$ Hydrolysis of 14298,4 (<50μ) at Three Temperatures (the same run as Table 2)

<table>
<thead>
<tr>
<th>Exp. No.</th>
<th>Temp. °C</th>
<th>Period</th>
<th>H$_2$S Released</th>
<th>S$_3$4</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-12</td>
<td>1 hr</td>
<td>38.2% + 6.7</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>+13</td>
<td>1</td>
<td>27.2 +15.1</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>+95</td>
<td>60</td>
<td>34.6 +17.8</td>
<td></td>
</tr>
</tbody>
</table>

Weighted average +12.8

Table 4 Hydrolyzable Nitrogen (as ammonia) by Concentrated H$_2$SO$_4$

<table>
<thead>
<tr>
<th>Sample</th>
<th>Nitrogen Released</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hawaii Basalt</td>
<td>0,1,3 and 2* ppm</td>
</tr>
<tr>
<td>Allende Meteorite</td>
<td>87</td>
</tr>
<tr>
<td>14298,4 Sieved Fines</td>
<td></td>
</tr>
<tr>
<td>250 - 110 μm</td>
<td>164</td>
</tr>
<tr>
<td>110 - 50 &quot;</td>
<td>110</td>
</tr>
<tr>
<td>50 &quot;</td>
<td>217</td>
</tr>
<tr>
<td>14321,49 Breccia</td>
<td>181</td>
</tr>
<tr>
<td>15271,24 Fines</td>
<td>111</td>
</tr>
</tbody>
</table>

* by 6N HCL Hydrolysis

The $^{40}$Ar-$^{39}$Ar ages of 16 basaltic and fragmental samples from Fra Mauro have been measured by standard methods using, as a standard, a hornblende separate from the Northern Light Gneiss of Northern Minnesota. This hornblende has been precisely dated at $2.61 \pm 0.06 \times 10^9$ yrs. (1).

Samples from Fra Mauro show a characteristic temperature release pattern different from rocks from Mare Tranquillitatis and Oceanus Procellarum in that the $^{40}$Ar/$^{39}$Ar ratio decreases in the highest temperature fractions. This raises the question of just how to interpret the release pattern in order to derive a valid age. For example, an age determined from the plateau will be greater than one calculated for the total high-temperature release fractions if there is an appreciable decrease at the highest temperatures. (See Figure 1). We have reported in Table 1 both plateau and high-temperature ages in order to facilitate a direct comparison with ages reported in the same manner from other laboratories.

The Fra Mauro high-temperature ages in Table 1 are depicted as a histogram in Figure 2. The majority of ages fall between $3.7 - 4.0 \times 10^9$ yr. The experimental uncertainties are sufficiently small such that this clustering of ages represents several different events, one of which, at $3.8 - 3.9 \times 10^9$ yrs., most likely represents the time of the Imbrium Impact Event. Two of the three samples that make up the distinctly lower age spike at $3.5 - 3.6 \times 10^9$ yr. can be classified as non-mare type basalts according to their mafic-plagioclase ratio (2). The third sample is a fragmental rock with a high proportion of clasts (3). Cosmic-ray exposure ages appear to group at about 30 and $300 \times 10^6$ yr. for all but one of the Fra Mauro samples analyzed.

Six samples from the Hadley Rille-Apennine Front area have now been dated. The gas release patterns of all these samples exhibit a well-defined plateau. Excluding the $4.09 \times 10^9$ yr. age of the anorthosite, 15415, these basalts define a crystallization age closely around $3.25 \times 10^9$ yr., similar to the age of rocks returned from Oceanus Procellarum. Cosmic-ray exposure ages vary and range from $50 - 500 \times 10^6$ yr.

Thus we note further volcanic and impact activity on the lunar surface from $3.2$ to $4.1 \times 10^9$ yr., with nothing, so far, indicative of any earlier or later activity.

* Present address: Ohio State University, Columbus, Ohio
AGES OF LUNAR MATERIAL

O. A. Schaeffer

P. 676


(2) LSPET, A preliminary description of the Apollo 15 lunar samples, preprint.


Figure 1

Figure 2
AGES OF LUNAR MATERIAL
O. A. Schaeffer

Table 1
40Ar-39Ar Ages from Fra Mauro

<table>
<thead>
<tr>
<th>Sample</th>
<th>Plateau</th>
<th>High-temperature</th>
</tr>
</thead>
<tbody>
<tr>
<td>14310,101</td>
<td>3.78±.03</td>
<td>3.74±.04</td>
</tr>
<tr>
<td>14053,34</td>
<td>3.92±.08</td>
<td>3.83±.09</td>
</tr>
<tr>
<td>14152,1-1</td>
<td>3.82±.10</td>
<td>3.77±.15</td>
</tr>
<tr>
<td>14152,1-2</td>
<td>3.75±.05</td>
<td>3.75±.05</td>
</tr>
<tr>
<td>14193,2-1</td>
<td>3.82±.10</td>
<td>3.75±.10</td>
</tr>
<tr>
<td>14167,6-1</td>
<td>3.79±.03</td>
<td>3.76±.04</td>
</tr>
<tr>
<td>14167,6-4</td>
<td>3.72±.03</td>
<td>3.70±.03</td>
</tr>
<tr>
<td>14257,12-1</td>
<td>3.87±.06</td>
<td>3.78±.10</td>
</tr>
<tr>
<td>14257,12-3</td>
<td>3.50±.12</td>
<td>3.50±.12</td>
</tr>
<tr>
<td>14161,34-2</td>
<td>4.03±.04</td>
<td>3.85±.06</td>
</tr>
<tr>
<td>14161,34-4</td>
<td>3.99±.03</td>
<td>3.96±.03</td>
</tr>
<tr>
<td>14161,34-5</td>
<td>4.01±.04</td>
<td>3.93±.05</td>
</tr>
<tr>
<td>14161,34-6</td>
<td>3.90±.04</td>
<td>3.86±.04</td>
</tr>
<tr>
<td>14161,26-11</td>
<td>3.59±.03</td>
<td>3.59±.03</td>
</tr>
<tr>
<td>14161,26-12</td>
<td>3.54±.03</td>
<td>3.52±.03</td>
</tr>
<tr>
<td>14303,13,R5-1</td>
<td>3.91±.04</td>
<td>3.89±.04</td>
</tr>
</tbody>
</table>

Table 2
40Ar-39Ar Ages from Hadley Rille-Apennine Front Area

<table>
<thead>
<tr>
<th>Sample</th>
<th>K-Ar</th>
<th>40Ar-39Ar Age, x 10^9 yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>15415,9</td>
<td>3.95</td>
<td>4.09±.19</td>
</tr>
<tr>
<td>15555,33</td>
<td>2.87</td>
<td>3.28±.06</td>
</tr>
<tr>
<td>15385,3</td>
<td>3.28</td>
<td>3.32±.06</td>
</tr>
<tr>
<td>15668,5</td>
<td>2.50</td>
<td>3.15±.08</td>
</tr>
<tr>
<td>15678,4</td>
<td>2.51</td>
<td>3.28±.06</td>
</tr>
<tr>
<td>15683,3</td>
<td>2.86</td>
<td>3.27±.06</td>
</tr>
</tbody>
</table>
A kinetic treatment of glass formation will be presented. This treatment is based on the construction of time-temperature-transformation curves for the crystallization of particular liquids under consideration. From such curves corresponding to some barely-detectable degree of crystallinity, the minimum cooling rates required to form glasses of various materials can be estimated. Also estimated are the maximum sizes of the materials which can be obtained in the glassy state. Adopting this analysis, the most important factors determining the glass-forming characteristics of different liquids are suggested to be the magnitude of the viscosity at the melting point or liquidus temperature and the rate of increase in viscosity with falling temperature below the melting point. It is suggested that relations between the glass forming behavior of materials and their melting points or various bonding characteristics should be regarded either as specific cases of limited generality or as relations between these parameters and the flow behavior.

The analysis has been applied to lunar compositions 14259 and 14310, using the viscosity vs. temperature relations to be reported in a companion paper together with present determinations of crystallization kinetics for these compositions. The kinetic data were obtained in each case over a wide range of temperature from 835 to 1219°C. All samples were prepared synthetically, being melted at elevated temperatures in molybdenum crucibles in a reducing atmosphere. The crystallization runs were carried out, also in a reducing atmosphere, for various periods of time at each temperature; the extent of growth after each time was measured by subsequent microscopic examination; and the growth rate was determined from the slopes of such extent of growth vs. time plots. Several crystallization runs, carried out in a vacuum rather than a reducing atmosphere, yielded similar values for the growth rates.

The results of the crystallization studies will be discussed in their own right as well as being related to the problem of forming glasses in the lunar environment. In the former discussion, attention will be directed to the nature of the interface process in these materials and to the redistribution of solute species during crystallization.
ESTABLISHMENT OF A SELENODETIC SYSTEM, Lawrence A. Schimerman,
USAF Aeronautical Chart and Information Center, 2nd & Arsenal Streets,
St. Louis, Mo. 63118

The unique characteristics of the Apollo 15 Metric Camera System provide
an opportunity for establishment of an improved selenodetic system. Existent
lunar coordinate systems related to the center of lunar mass were developed
in support of circumlunar navigation and mapping of sites of scientific or
mission interest. Their area of coverage is restricted and the Lunar Orbiter
and Apollo 8-14 photographic mission data on which they are based has
inherent accuracy limitations.

As a first step in development of an improved selenodetic system, the
USAF Aeronautical Chart and Information Center (ACIC) has undertaken the
analytical photogrammetric reduction of Apollo 15 Metric Camera photography
for NASA. A prime objective of this project is the establishment of a
consistent system of lunar positions and elevations throughout the area of
coverage. This new positional data will enable improved geodetic-
cartographic support of the lunar scientific community. Also, intermediate
results reflecting the photogrammetric solutions' measure of the effects of
uncertainties in lunar geodetic and geophysical parameters will provide new
data for investigators involved in these specialties.

The analytical photogrammetric solution under development employs NASA's
generation of a spacecraft ephemeris for initial orbital position, measures
of some 900 vertical terrain photos to project and extend lunar surface
positions, measurement of companion stellar photographs for determination of
terrain camera orientation and the first direct measurement of lunar orbit to
surface distances through laser altimetry. Reduction of this mass of data is
scheduled for completion in 1973 and this paper's primary concern is
presentation of test results achieved t: date.

A photogrammetric test was designed to provide a miniaturization of
the total photogrammetric triangulation, enabling evaluation of the Apollo 15
Metric Camera System results, identification of problem areas caused by data
variations and definition of attainable accuracies. Twenty-one terrain and
related stellar exposures accomplished during Apollo 15's 22nd, 27th and 60th
orbital revolutions were selected for generation of test data. (Figure)
They cover an area of approximately 80,000 square kilometers surrounding the
Apollo 15 landing site and provide a sample of the unique pattern of
photography provided by a lunar photographic mission constrained to land
astronauts at 26°N lunar latitude.
ESTABLISHMENT OF A SELENODETIC SYSTEM
SCHIMERM AN, ETC.

Analytical photogrammetric computer programs LOSAT and LOBAT providing computation of orbitally constrained photo strip and block assemblies, are the basic testing tools. Weighted input data for these programs includes terrain photo measurements, camera orientation components and spacecraft state vectors (position and velocity), as well as time of exposure. The resulting computation provides the projected positions of measured lunar surface points as well as an updating of the initial orbital and orientation data based on a best fit of these data. While information produced to date is based on a small sample, and testing of basic camera calibration and compatibility of spacecraft positions with other data is continuing, some conclusions and comparisons may presently be drawn.

a. Measurements of terrain photo images show standard errors of 5-9 micrometers, with range of error being dependent upon the sun angle of photography and its effect on precise identification and transfer of image points.

b. The process of deriving terrain camera orientation through measurement and reduction of stellar photography reflects a standard error of 20 seconds of arc in derived orientation.

c. An uncertainty of approximately 3 micrometers exists in relating terrain photo measurements to the camera fiducial system due to the combined effect of the camera's image motion compensation system and residual film distortion. It is anticipated that this displacement will become inseparable from definition of camera position in the photogrammetric solution, resulting in an error of approximately 5 meters in defining distance between successive exposure stations. The 1 millisecond accuracy of time information would otherwise have limited error in this dimension to 2 meters.

d. Comparison of Spacecraft Inertial Measurement Unit (IMU) orientation values with those obtained from reduction of stellar photography shows a bias of approximately 8 minutes of arc. The lesser accuracy inherent in the onboard sextant star sightings upon which IMU values are based, necessitates acceptance of base orientation values derived from stellar photography. However, comparison of rate of orientation change between the two methods showed good correlation and a standard deviation of 46 seconds of arc.

e. Present photogrammetric solutions have not been able to exploit the laser altimeter's one meter accuracy in defining camera station to lunar surface distance. As the point of reflection is not imaged on the photography, an approximate point identification can only be defined by establishing its photographic location with reference to camera calibration information. Measurement and transfer of this ill defined image has incorporated increased error. Not surprisingly, 7 altimeter observed distances from orbital revolution 22, show a mean difference of 35 meters.
(maximum 63 meters) with respect to photogrammetric distance determinations. Three comparisons of laser to photogrammetrically determined distance on revolution 27 show systematic 82-105 meter differences. The systematic nature and magnitude of these latter differences are inconsistent with evaluated error of the preliminary photogrammetric solution and are indicative of an editing role for laser altimetry data in this project.

Extrapolation of observed residual photographic errors to orbital and lunar surface positional determinations, defines these components at a 20 meter and lower level. Based on this limited sample, it appears that the acquired materials are adequate for fulfillment of the Apollo Orbital Photographic Science Team's concept of developing an improved selenodetic system and providing a basis for evaluation of presently accepted lunar geodetic and gravimetric values.
CHEMICAL COMPOSITIONS OF APOLLO 14, APOLLO 15, AND LUNA 16 MATERIAL:


Work in this laboratory, since the Second Lunar Science Conference, has been on Apollo 14, Luna 16 and Apollo 15 samples. Analyses include Si, Ti, Al, Fe, Mg, Ca, Na, K, Mn, P and Cr by atomic absorption spectrophotometry and colorimetry; Li, K, Rb, Sr, Ba, nine rare-earth elements, Zr, and Hf by stable isotope dilution; and Fe, Co, Na, Mn and Sc by instrumental neutron activation analysis.

Soils from the Apollo 14 Fra Mauro site have quite similar chemical compositions, both laterally and with depth; the regolith at this site has apparently been well mixed. The composition of igneous rock 14310 resembles the composition of the 14 soils; it may represent remelted soil. Apollo 14 materials most closely resemble in composition the exotic fragments, such as KREEP or norite, found in Apollo 11 and 12 soils, or the dark portion of rock 12013. Similarities between chemical characteristics of Apollo 12 soils and breccias, rock 12013, and Apollo 14 soils and igneous rock 14310 suggest that extensive quantities of Fra Mauro-like material have been transferred to the Apollo 12 mare site.

Isotope dilution analyses were obtained on two <125μ Luna 16 regolith fines samples from different depths in the core, and four rock-chips, including both igneous rocks and breccias. These materials have similar trace-element concentrations. This suggests a largely local derivation of the Luna 16 regolith, unless the Luna 16 igneous rocks are remelted soil. The Luna 16 materials differ from Apollo 11, 12 and 14 lunar samples in a manner suggesting the presence of excess feldspar; i.e., the Luna 16 samples have higher K/Rb, lower Rb/Sr and lower Sm/Eu ratios.

To date, only two Apollo 15 samples have been analyzed: igneous rock 15555 and soil 15531. The igneous rock has the lowest concentration of rare-earth elements and smallest negative Eu anomaly of any Apollo 11, 12, 14 or Luna 16 sample. Plagioclase/whole-rock trace-element distributions from this rock are similar to those obtained from Apollo 11, 12 and 14 plagioclase-whole-rock pairs. Soil 15531 Ba and rare-earth concentrations are two to four times higher than the igneous rock, requiring another component in the soil. Not more than 7 to 8% KREEP in the soil would account for the differences in composition.

Six portions of the Apollo 15 nine-meter core tube are being analyzed for major and trace elements; the results will be reported at the Conference.
The abundance of each component in soils and breccia was determined by using a linear mixing model and the weighted least squares method (Gauss). The weights are inversely proportional to the variance of each chemical element. The elements used in this mixing model are in most cases: Na, K, Rb, Sr, Ba, La, Ce, Sm, Eu, Yb, Y, Nb, Si, Al, Ca, Fe, Ti, Cr, Mn, Zr, U, Th, V, Co, Ni, and Sc. Several components have been identified from rocks and by direct observation in the coarse fraction of the soils (Fig. 1). The components used in this analysis include mare basalts from all sites (data from many authors: 1st and 2nd Lunar Sci. Conf. Proceedings), anorthosites (data from Wood, Keil, Schmitt and Hubbard), KREEP (data from Meyer, Gast and Hubbard), silicic or "granitic" material (light portion of 12013; data from Hubbard, Keil, Schmitt, Schnetzler, and Wood), ultramafic material (data from Chao and Keil), and meteoritic component (Anders, data from CC-1). The chemical data for the soils are from many authors (1st and 2nd Lunar Sci. Conf. Proceedings).

The Apollo 11 soils and breccia can be described as a mixture of local mare basalts, KREEP (0 to 6%), and anorthosites (20 to 25%), with only minor granitic, ultramafic, and meteoritic components. The Apollo 12 samples can be described as a mixture of Apollo 12 mare basalts and KREEP (30 to 68%), with only minor granitic, anorthositic, and meteoritic components (Meyer et al., 1971). By using the preliminary data on Luna 16 (Vinogradov, 1970) the soil from that mission can be described as a mixture of local mare basalts, anorthosites (20%), and minor contributions from KREEP, meteoritic, ultramafic, granitic, and meteoritic components. The Apollo 14 soils and breccia are very rich in KREEP (60 to 95%) with relatively minor contributions of anorthosites, mare basalts, ultramafic, granitic, and meteoritic components. There is a correlation between KREEP and the granitic component. The ratio of KREEP to granite is about 35 and suggests a petrologic relationship between those two components.

The mixing model was also applied to the preliminary chemical data on soils and breccia from the Apollo 15 mission (M. Rhodes et al.; LSPET, 1971). Elements included in the analysis are: Si, Ti, Al, Fe, Mn, Ca, Na, K, P, Cr, Sr, Zr, Nb, Rb, Th, Ni, and Y. Some of the results are shown in figure 2. They show that the samples closer to the Apennine Front contain an increased amount of "anorthositic gabbro." The abundance of mare basalts follows an opposite trend. Again this site is quite rich in KREEP. Samples that represent deeper locations such as sample 15601 (from the edge of Hadley Rille) and 15471 (from the rim of Dune Crater) have the smallest amount of KREEP, suggesting that the mare area was covered with a blanket rich in KREEP. The high concentration of anorthosite close to the Apennine Front is consistent with a model of the lunar highlands (Wood, 1970). In all 10 samples the granitic
COMPONENT ABUNDANCE AND AGES IN SOILS AND BRECCIA

E. Schonfeld

APOLLO 15 COMPONENT ABUNDANCE IN SOILS AND BRECCIA

FIGURE 1

APOLLO 15 SAMPLE NUMBER

Source similar to source of mare basalts.

FIGURE 2

FIGURE 3

FIGURE 4
component was equal or less than about 1%. The meteoritic component was in all cases about 2%. All samples indicated a few percent of a component with chemical composition similar to an ultramafic component. The highest value was about 20% in samples 15923 (green clod) and 15301 (close to Spur Crater).

Absolute ages determined for lunar samples from the Apollo missions range from 3.2 to 4.0 billion years. Single stage, or model ages of the lunar soil and breccia range from 4.3 to 4.8 billion years, but since soil samples and breccias are mixtures of different components, these model ages are difficult to interpret. The mixing model was used to help explain the model ages of soils and breccias. By using the results of the mixing model on samples 10084, 10052 and 12070 the material balance of the elements Rb, Sr, and the radiogenic $^{87}\text{Sr}$ was calculated. The results show that there is a very good material balance (within the experimental error measured by Wasserburg et al., 1971). These results also indicate that in those sites there is little or no volatility of Rb. KREEP is the most important radiogenic component in the soils. It contributes radiogenically 40% in the Apollo 11 soil, about 90% in the Apollo 12 soils, about 95% in the Apollo 14 soils, and about 80% in the Apollo 15 soils. In contrast, the Apollo 11 samples show an excess of radiogenic Pb and in the Apollo 12 samples a depletion of radiogenic Pb indicating that the closed system hypothesis is a first approximation (Silver, 1971). The mixing model was also used to determine the Pb-U-Th systematics for KREEP from the Apollo 12 site using data from Cliff et al., Silver, and Tatsumoto et al. (shown in Fig. 3 with the calculated $4.4$ and $4.6$ continuous Pb diffusion curves; Tilton, 1960). The correction is not large and the plot indicates a third event (Cliff et al., Silver, Tatsumoto et al.) at $0.9\pm0.1$ b.y. with an initial apparent age of about $4.43$ b.y. The gas retention ages of 12034 and 12033 (data from LSPET, O'Kelley et al., and Pepin et al.) are $1.0\pm0.2$ b.y. These ages plus the fact that KREEP in these samples is mainly in the form of impact glasses (Meyer et al.) excludes the Pb continuous diffusion explanation. The average model ages of the KREEP fragments from Apollo 11, 12 and 14 is close to $4.41\pm0.04$ b.y. (data from Gast, Hubbard, Nyquist, and Wasserburg et al.). A two-stage model was used to calculate the formation age of KREEP (Fig. 4). These calculations show that if KREEP was formed from a source material that had a Rb/Sr ratio similar to that of the source material of the Apollo mare basalts ($0.003$ to $0.010$; Wasserburg et al.) or Eucrites ($0.0045$), then the formation age of KREEP is very close to its model age of about $4.41$ b.y. and compares very well with the initial apparent age of KREEP by the Pb-U-Th method. KREEP is the oldest type of moon rock found so far. The widespread occurrence of this material suggests, that a large surface portion of an initially cold moon underwent melting in its very early history.
Measurements of electrical conductivity of samples from 10024, 10048 and 12053 at temperatures from 300°C to 800°C have been reported (1). These data indicate that complex changes occur during heating; in particular, the conductivity at 300°C, $\sigma(0)$, increases by several orders of magnitude after heating above 500°C in vacuum or in Ar, He, or He-2% H₂ atmospheres—the effects of oxidizing atmospheres had not been studied. Recent data for terrestrial augite show that heating at 250°C in reducing environments increases $\sigma(0)$; whereas, subsequent heating in oxidizing atmospheres decreases $\sigma(0)$ to its original value. A lunar conductivity sample (12053), previously heated to 800°C in He-2%H₂, was heated to 500°C in air and $\sigma(0)$ decreased by four orders of magnitude, returning almost to its original value. Subsequent heating at 800°C in air increased $\sigma(0)$, probably indicating secondary oxidation effects. Consequently, it appears that part of the reported hysteresis in $\sigma(0)$ is due to high-conductivity surface layers formed under reducing conditions. Detailed studies of atmosphere effects and separation of bulk and surface contributions to the conductivity are in progress.

Mössbauer spectra were obtained at 300K, 78K and 4.2K for 12053 and 10048 in the virgin state and after high temperature conductivity runs. For both samples, heat treatment caused broadening of the pyroxene peaks and transfer of ~2-3% of the Fe from M2 to M1 sites. For 12053, ilmenite peaks were unaffected by heating; however, in a heated sample of 10048 ilmenite peaks were increased in intensity by ~30% relative to an unheated sample. Since spectra were taken for different specimens from 10048, these differences may be due to sampling errors; however, in view of conductivity and magnetic data (1), this increase is probably a real effect.

Mössbauer studies have been made of Apollo 14 samples including samples to be used in conductivity studies (see table). Salient features are (i) 14053 has unusually high metallic Fe and FeS content and contains a previously unreported magnetically-ordered phase, possibly a spinel, (ii) 14063 has high olivine content, (iii) 14047 and 14259 (fines) exhibit a broad paramagnetic peak near 1.35 mm/sec, possibly from a glassy phase, (iv) Apollo 14 samples in general are more heterogeneous than Apollo 11 and 12 samples.

Spectra were obtained at 2 K for samples in magnetic fields of 0 and 900 Oe (2). Large hyperfine fields exist due to long spin relaxation times in pyroxenes at this temperature. Although the spectra are complex, they are
potentially useful in determining electronic energy levels of Fe atoms in pyroxene and, consequently, valuable in understanding physical properties including conductivity. In contrast to higher temperature spectra, these low temperature spectra show considerable variation among samples. For example, in 0 magnetic field 12053 has an extremely complex spectrum but at fields of 900 Oe the spectrum sharpens into one which can be described as three or four sets of six-line magnetic peaks. By contrast 14303 and 14047 have simpler magnetic hyperfine spectra at 0 field and show relatively small changes when field are applied. Since most of the pyroxene Fe is in M2 sites in these two samples, while nearly half of it is in M1 for 12053, this leads to the tentative conclusion that only one component of the g tensor is large for M2 Fe, which in turn might imply a large splitting between d electron states of |x y⟩ and |x z⟩, |yz⟩ symmetry at that site (3). The small temperature dependence of the M2 quadrupole splitting supports this suggestion. Further studies of this type are in progress on lunar pyroxene separates, and on terrestrial and synthetic pyroxenes.


Relative Distribution of Fe Atoms in Apollo Lunar Samples

<table>
<thead>
<tr>
<th>Phase</th>
<th>14053-48</th>
<th>14047-47</th>
<th>14301-65</th>
<th>14303-35</th>
<th>14063-47</th>
<th>14259,69</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>(Fines)</td>
<td>(Fines)</td>
<td>(Fines)</td>
<td>(Fines)</td>
<td>(Fines)</td>
<td>(Fines)</td>
</tr>
<tr>
<td>Silicate</td>
<td>.774</td>
<td>.849</td>
<td>.885</td>
<td>.851</td>
<td>.831</td>
<td>.827</td>
</tr>
<tr>
<td>(P)</td>
<td>(P,0,gl.)</td>
<td>(P, -)</td>
<td>P</td>
<td>P</td>
<td>P, .25</td>
<td>0</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(P,0,gl.)</td>
</tr>
<tr>
<td>Ilmenite</td>
<td>.062</td>
<td>.031</td>
<td>.075</td>
<td>.114</td>
<td>.169</td>
<td>.027</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>.088</td>
<td>.060</td>
<td>.040</td>
<td>.020</td>
<td>~0</td>
<td>.086</td>
</tr>
<tr>
<td></td>
<td>.05</td>
<td>≤.01</td>
<td>~0</td>
<td>.015</td>
<td>~0</td>
<td>-</td>
</tr>
<tr>
<td>FeS</td>
<td>.026</td>
<td>.06</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>.06</td>
</tr>
<tr>
<td>Other</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(broad,para.)</td>
</tr>
<tr>
<td>(Magnetic)</td>
<td>(broad,para.)</td>
<td>(broad,para.)</td>
<td>(broad,para.)</td>
<td>(broad,para.)</td>
<td>(broad,para.)</td>
<td>(broad,para.)</td>
</tr>
</tbody>
</table>

P - Pyroxene
O - Olivine
gl - glassy phase
ELECTRON MICROSCOPY OF EXPERIMENTALLY SHOCKED ROCK-FORMING MINERALS WITH APPLICATION TO LUNAR SAMPLES. Charles B. Sclar and Stephen P. Morzenti, Department of Geological Sciences, Lehigh University, Bethlehem, Pa. 18015.

Thinned samples for transmission electron microscopy have been prepared from experimentally shocked terrestrial minerals and their unshocked counterparts by the ion milling technique using an argon plasma. These samples have been examined in a conventional 100 kV electron microscope (RCA EMU-3G) in bright and dark field and by selected-area electron diffraction for shock-induced features such as dislocation arrays, transformation textures, and high-pressure phases.

Particulate samples (-48+150 mesh) of olivine (Fa11) from the Admire pallasite and olivine (Fa17) from Hawaii (USNM #44770) were subjected to explosive shock loading at approximate peak pressures of 200 kb and 400 kb, respectively. The shocked meteoritic olivine contains what appears to be a very fine dispersed phase (particle size <800 Å) of low electron transparency relative to the olivine matrix. This may be a high-pressure phase with either the spinel structure (ringwoodite) or the "β-phase" structure. The shocked Hawaiian olivine has been largely recrystallized to fine polycrystalline aggregates (average grain size of about 1000 Å), probably as the result of the intense thermal pulse accompanying the shock event. Similar effects in the olivine of dunite shocked in the 1 Mb range were reported by Carter et al. (1968).

A particulate sample (-48+150 mesh) of enstatite from the Mt. Egerton enstatite achondrite was shock loaded at a peak pressure of about 250 kb. Transmission electron microscopy of the shocked sample reveals the presence of a dispersed equiaxed phase with a maximum particle size of about 1000 Å. Many of the particles are euhedral with a hexagonal outline. The dispersed phase is not transparent to the electron beam at 100 kV despite the fact that the pyroxene matrix is transparent and gives electron diffraction patterns. This dispersed phase is not transparent to the electron beam either because the particles are thicker than the matrix due to differential sputtering and/or because the particles have a significantly higher density than the matrix. It is suggested that the lack of electron transparency of the dispersed phase is due to its higher density and that this phase is a shock-induced high-pressure phase which results from either the pyroxene-garnet transformation.
\( \Delta d/d \approx 10-15\% \) or the pyroxene-ilmenite transformation
\( \Delta d/d \approx 15-20\% \). The pyroxene-garnet transformation is in accord with the Hugoniot data of Ahrens and Gaffney (1971) obtained on Bamle enstatite whereas the pyroxene-ilmenite transformation is in accord with the shock-wave data reported for Stillwater bronzitite by Ahrens, Anderson, and Ringwood (1969) and Davies and Anderson (1971). The extreme difference in electron transparency between the dispersed phase and the matrix suggests that the dispersed phase may be the hypothetical polymorph of MgSiO$_3$ with the ilmenite structure.

A particulate sample (-48+150 mesh) of Durango apatite shock loaded at a peak pressure of about 250 kb shows dislocations not evident in the unshocked equivalent. The optically resolvable planar features previously observed in this shocked sample by Sclar and Usselman (1970) are revealed by transmission electron microscopy; the interplanar distances are as small as 1000 Å.

Attempts are being made to apply the shock-induced features observed in the experimentally shocked samples to the lunar minerals.
CHEMICAL ANALYSES OF LUNAR SAMPLES 14003, 14311 and 14321

J. H. Scoon, Department of Mineralogy and Petrology,

Three samples 14003.33, 14311.67 and 14321.65 of the Apollo 14 collection were analysed by classical chemical methods for major elements. A small, but significant, amount of water is reported on 14003.33 and 14321.65. The comparatively large amount of water reported on sample 14311.67 (sawdust) might have been acquired during the sawing operation. This sample also contained numerous small shiny white metallic fragments. Some fibrous material was also noted on microscopic examination. Similar contamination was noted on sample 14321.65 (homogenized fines) but to a much less extent. Iron again is present only in the ferrous state. $P_{2}O_{5}$ is appreciably higher than in the Apollo 12 samples.

The analytical results are shown in the accompanying table. The unacceptably high total for 14311.67 is probably due to the metallic fragments being reported as oxide.
Chemical Analyses of Lunar Samples

J. H. Scoon

Chemical analyses of lunar samples (wt%)

<table>
<thead>
<tr>
<th></th>
<th>14003.33</th>
<th>14311.67</th>
<th>14321.65</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>47.99</td>
<td>47.24</td>
<td>47.78</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>17.54</td>
<td>18.05</td>
<td>15.20</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>nil</td>
<td>nil</td>
<td>nil</td>
</tr>
<tr>
<td>FeO</td>
<td>10.68</td>
<td>11.13</td>
<td>12.25</td>
</tr>
<tr>
<td>MnO</td>
<td>0.13</td>
<td>0.14</td>
<td>0.17</td>
</tr>
<tr>
<td>MgO</td>
<td>9.24</td>
<td>9.59</td>
<td>10.73</td>
</tr>
<tr>
<td>CaO</td>
<td>10.99</td>
<td>10.16</td>
<td>9.94</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.75</td>
<td>0.85</td>
<td>0.78</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.52</td>
<td>0.81</td>
<td>0.62</td>
</tr>
<tr>
<td>H₂O⁺</td>
<td>0.04</td>
<td>0.29</td>
<td>0.05</td>
</tr>
<tr>
<td>H₂O⁻</td>
<td>0.01</td>
<td>0.03</td>
<td>nil</td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.94</td>
<td>1.81</td>
<td>2.06</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.41</td>
<td>0.76</td>
<td>0.41</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.19</td>
<td>0.15</td>
<td>0.26</td>
</tr>
<tr>
<td>S</td>
<td>0.08</td>
<td>0.06</td>
<td>0.07</td>
</tr>
<tr>
<td>F</td>
<td>nil</td>
<td>nil</td>
<td>nil</td>
</tr>
<tr>
<td></td>
<td>100.51</td>
<td>101.07</td>
<td>100.32</td>
</tr>
<tr>
<td>Less S = 0</td>
<td>0.02</td>
<td>0.02</td>
<td>0.02</td>
</tr>
<tr>
<td></td>
<td>100.49</td>
<td>101.05</td>
<td>100.30</td>
</tr>
<tr>
<td>Total iron as Fe₂O₃</td>
<td>11.76</td>
<td>11.91</td>
<td>13.59</td>
</tr>
<tr>
<td>ARC</td>
<td>1.15</td>
<td>1.27</td>
<td>0.34</td>
</tr>
</tbody>
</table>

ARC = Total reducing capacity less the reducing capacity attributable to the FeO content. (assuming all iron is present as FeO)
Metallic Al extracted from 14311.67 by 5% NaOH solution 1.3%.
SOIL EROSION DURING LUNAR MODULE LANDING, by R. F. Scott, Division of Engineering and Applied Science, California Institute of Technology, Pasadena, California 91109

As the lunar module approaches to within a few hundred feet of the lunar surface, the effects on the lunar surface of the exhaust stream from the descent engine become apparent. Initially, fine-grained material is removed from crater rims and ridges and is blown away in the dispersing gas cloud. Erosion becomes more intense as the spacecraft approaches more closely; channeling or streaking of the gas and dust flow sheet occurs as it flows around protuberances on the lunar surface. Eventually, when the spacecraft is of the order of a few meters or less from the lunar surface, the concentration of lunar surface particles entrained in the gas flow become great enough to result in almost complete obscuration of the surface. At this stage the film from the descent camera shows rock fragments up to 10 cm across being carried along in the gas flow at velocities of the order of 30 meters/sec.

Although this sequence of events during descent has been similar for all four of the manned landings to date, there have been large differences in the details of the processes involved. Obscuration of the surface did not develop during the Apollo 11 and 14 landings to the degree observed in the 12 and 15 descents, for example. It is desirable to determine the causes of these differences, from both scientific and mission safety points of view.

To analyse the astronauts' observations and the features of the erosion process visible in the descent films, it is necessary to consider a number of factors. These include the mechanism of exhaust gas impingement on, and deflection by the surface, gas flow into, through, and out of the permeable lunar soil, the effect of the variability of lunar soil properties as determined to date on the soil removal processes, and the effect on the visibility of the dust cloud of the solids concentration, and the lighting and viewing conditions from the Lunar Module. Preliminary examinations of the effects of these factors on the observed dust conditions in the four landed missions have been performed and are reported in the paper.

The results indicate that, although the variation of soil properties and visibility considerations may play a part in the amount and appearance of the erosion which occurs, a large contribution to the quantity of dust removed appears to develop as a consequence of gas flow into and through the lunar soil. Descent observations and analysis show that the longer the exhaust gas from a particular landing impinges on the same area of soil, the greater is the erosion developed. Thus, the greatest obscuration observed to the present occurred in the Apollo 15 mission, in which the final descent occurred almost vertically from a height of about a hundred meters. The descent was also slower than in previous missions. The
Soil Erosion During Lunar Module Landing

R. F. Scott

Paper includes a brief discussion of the amount and distribution of the soil removed from the landing site, since this may play a part in sample selection and analysis.

The sub-125 μm fraction of two samples (L-16-A-10) and L-16-G-9) of the Luna 16 regolith core are being studied petrographically and mineralogically in our laboratory. The samples came from the 6-8 cm and 29-31 cm depth intervals of a core collected from the northeastern part of Mare Fecunditatis.

The samples were split and sorted under the binocular microscope according to their size and their lithologic or mineralogic character. The size classes used were 63-125μm, 32-63μm, and <32μm. Particle identification's were confirmed by x-ray diffraction photographs or semi-quantitative analysis in the electron microprobe.

The particles in the sub-32μm fraction were not hand-sorted because of their great numbers and their small size. Polished mounts were made and population studies of the particles were carried out under the petrographic microscope and in the electron microprobe.

The bulk of the work to date has been done on the lowermost of the two core samples, L-16-G-9. Table 1 shows the distribution of lithologic and mineralogic types of particles found in two of the size fractions from L-16-G-9. The data are preliminary.
Mineralogic and petrographic description of the sub-125 micrometer fraction of two Luna 16 core samples
G. A. Sellers

Table 1. Distribution of particles in two size fractions of Luna 16 sample G-9 (numerical percent)

<table>
<thead>
<tr>
<th>Particles</th>
<th>63-125μm</th>
<th>32-63μm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Microgabbro (ave. grain size &gt;0.05mm.)</td>
<td>10.3</td>
<td>1.0</td>
</tr>
<tr>
<td>Basalt (ave. grain size &lt;0.05 mm.)</td>
<td>2.8</td>
<td></td>
</tr>
<tr>
<td>Pyroxene</td>
<td>15.5</td>
<td>33.8</td>
</tr>
<tr>
<td>Olivine</td>
<td>0.5</td>
<td>1.3</td>
</tr>
<tr>
<td>Feldspar</td>
<td>11.3</td>
<td>21.5</td>
</tr>
<tr>
<td>Opaques</td>
<td>1.4</td>
<td>1.1</td>
</tr>
<tr>
<td>Breccia and glazed aggregates</td>
<td>54.0</td>
<td>39.8</td>
</tr>
<tr>
<td>(includes slaggy glass)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Colored glasses</td>
<td>4.2</td>
<td>1.5 (all spherules)</td>
</tr>
<tr>
<td>Total</td>
<td>100.0</td>
<td>100.0</td>
</tr>
<tr>
<td>Total number of particles</td>
<td>213</td>
<td>1566</td>
</tr>
</tbody>
</table>

The distribution of particle types in the sub-125μm size fractions differs somewhat from that reported by Vinogradov (1971) for the coarser size fraction. The olivine content of this sample is lower than those reported by Vinogradov for material examined by Mössbauer spectroscopy. Nickel metal flakes found in the sample are judged, by their low iron content, to be contamination.

Among the phases confirmed in the samples are olivine, pigeonitic and augitic pyroxene, bytownite, ilmenite, aluminian chromite, and β-cristobalite. The presence of orthoclase is indicated but it will have to be confirmed by further work.

Reference


* Publication authorized by the Director, U. S. Geological Survey
CRATERING HISTORY AND EARLY EVOLUTION OF THE MOON, Eugene M. Shoemaker, Caltech, Pasadena, Calif., 91109

On the basis of Rb-Sr ages of crystalline rocks summarized by Papanastassiou and Wasserburg (1971) fairly narrowly defined ages may be derived for two mare surfaces; a well defined upper limit also may be set on the age of the Fra Mauro formation, which underlies a surface that is widely exposed in the highlands around the Imbrium basin. In addition, an age for the crater Copernicus and a lower limit on the age of the "Surveyor" crater may be established by inference from U-Pb isotopic studies of samples of fine-grained material (ray material) by Leon T. Silver and from detailed considerations of the geologic relations of these and other samples at the Apollo 12 landing site. When the isotopic ages are compared with the crater abundances on the mare surfaces, the surface of the Fra Mauro formation, and the surface of the rim of Copernicus (fig. 1), rather stringent bounds may be set on certain specific models of the cratering history of the Moon.

The available data suggest that, beginning at about 3.2 AE, the rate of cratering on the Moon averaged over periods of time of the order of $10^8$ to $10^9$ years has remained relatively steady at about 1500 to 3000 craters $Z.1$ km diameter/10$^6$km$^2$/10$^9$ yr. This rate of cratering is consistent with the observed slopes of the "Surveyor" crater (Shoemaker, Jackson, and Hait, 1971) and with cosmic ray exposure ages of crystalline lunar rock fragments (Shoemaker, in press). When traced backward in time beyond 3.2 AE, however, the rate of cratering increases very rapidly; the average rate between 3.65 and 3.85 AE was at least 150,000 craters $Z.1$ km diameter/10$^6$km$^2$/10$^9$ yr, and may have been considerably higher.

In the model adopted in this paper, the craters are assumed to have been formed by impact of two discrete families of extralunar objects and by secondary lunar fragments. The average rate of impact for one family of objects has remained nearly constant during geologic time; the primary and secondary craters produced by these objects account for nearly all of the craters formed since 3.2 AE. This family of objects can be shown, with considerable confidence, to be composed primarily of comets and fragments of comets. The second family of objects is assumed to have impacted the Moon at an exponentially decaying rate. Between 3.2 and 3.85 AE, the half-life of this family could not have been greater than about 110 my. If an exponentially increasing rate of impact is extrapolated back from 3.2 AE to 4.65 AE, and the total mass of objects that impact the Earth in this time period at rates corresponding to those on the Moon is assumed not to exceed the mass of the Earth, then the half-life of this family of objects cannot be less than about 45 my. This range of half-lives is consistent with objects in relatively eccentric and inclined orbits around the Sun that are being swept up primarily by collision with the Earth; I will call such objects "planetesimals".

If the second family of objects is assumed to be in orbit about the Earth and an exponentially increasing rate of impact is extrapolated back to 4.65
AE, the total mass of objects impacting the Moon would exceed the mass of the Moon if the half-life is less than about 54 my. This relatively high lower limit for the half-life suggests the objects were not satellites of the Earth.

With the shortest permitted half-life of the "planetesimal" family of objects (45 my), the surface of the Moon will have been saturated with craters larger than 1 km that have been formed since 4.0 AE. This means that no part of the lunar surface significantly older than 4.0 AE would have survived or would be recognizable. Most of the area of the lunar surface would be occupied by fairly large younger craters. With the 110 my half-life, the oldest surviving parts of the lunar surface will have ages between 4.2 and 4.3 AE.

Rock fragments may survive as closed isotopic systems somewhat longer than surfaces. If the true half-life of the "planetesimal" family of objects is close to the 45 my limit, however, it will be surprising to find more than a few rocks that have remained isotopically undisturbed for more than about 4.2 AE.

A simple explanation is suggested for the distribution of ages of the recognizable maria. Basaltic lava flows may have poured out extensively on the lunar surface beginning at some point in time very early in the history of the Moon, perhaps as early as 4.5 AE or earlier. Because of the high rates of cratering early in lunar history, however, the only extensive lava surfaces, such as the maria, that have survived were formed after about 3.85 AE. The observed distribution of "cratering ages" for the maria (Soderblom, 1970) suggests that basaltic eruptions were episodic and that the average frequency of eruption gradually decreased with time. The ages of the oldest recognizable mare surfaces, found in Mare Fecunditatis, are probably close to 3.85 AE. There is no need to postulate a sudden onset of volcanism at about 3.85 AE to explain the maria, although there may have been a change in magma types. The Moon probably was hot very early in its history, at least to a depth of several hundred kilometers if not all the way to the center, and the outermost part of the Moon has slowly cooled with time; the rate of volcanism decreased as the thickness of the lunar lithosphere increased. This same model, when applied to the Earth, can explain the apparent absence of rocks older than about 3.5 AE.

The unexpectedly long half-life of the residual planetesimal swarm indicated by this proposed model of cratering history may provide an explanation for several dynamical peculiarities of the Earth-Moon system. The present low total angular momentum of the Earth-Moon system can no longer be taken as a serious objection to hypothesis of the origin of the Moon by fission from the Earth. Fission of the Moon could have occurred at a time long before the Earth reached its present mass. If about 1/3 of the Earth accreted after fission, then the Earth could have been despun by an amount required by the fission hypothesis. Furthermore, the Moon could have lost substantial mass and angular momentum by bombardment, at the same time that the Earth was gaining mass. At the average velocities of impact associated with the eccentric and inclined orbits required for the long half-life of the planetesimal swarm, more mass is ejected at escape velocity from the Moon than falls on it during each impact. The amount of mass lost from the Moon could be equal to or greater than the present mass of the Moon.
Objections raised by Goldreich (1966) to the origin of the Moon by either fission or by accretion in a close Earth orbit are applicable only if the Earth-Moon system has remained essentially closed. These objections are based on tracing the history of the lunar orbit and obliquity of the Earth backward in time. With the high rates of bombardment and possible mass gains and losses suggested here, the early history of the orbit of the Moon and the obliquity of the Earth become indeterminate.

It is clear that accretion of neither the Earth or the Moon could have begun from objects following orbits like those of the residual planetesimal swarm. In the initial stages of accretion, only objects with closely similar orbits can collide and stick together. Probably the planetesimals with low orbital eccentricity and inclination were swept up very rapidly to form the protoearth and possibly the protomoon. It is only the remaining planetesimals with initially high eccentricities or inclinations or which were perturbed into eccentric and inclined orbits by close encounters with the growing protoearth, for which we have a record on the lunar surface.
BERYLLIUM AND CHROMIUM ABUNDANCES IN FRA MAURO LUNAR MATERIAL
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Lunar samples returned by Apollo 14 from the Fra Mauro Formation have been analyzed for beryllium and chromium using electron capture gas chromatography. The very sensitive and selective method employed, which involves the conversion of the respective metal to its β-diketonate chelate, has been previously described by Sievers, et al. Beryllium and chromium abundances have been determined for lunar fines 14163, fragmental rocks 14311, 14301, 14047, 14321 (interior and exterior chips), and the crystalline rock 14310. The beryllium concentration of the lunar fines (14163) has been found to be 6.64 ± 0.21 ppm while the fragmental rocks vary in beryllium content from 5.97 ± 0.29 ppm to 11.72 ± 0.57 ppm. The one crystalline rock of basaltic nature available to us for analysis contains 5.78 ± 0.24 ppm of beryllium. The beryllium abundances are significantly higher than those found in the Apollo 11 and 12 lunar material, having a range of 0.65 ± 0.06 ppm to 0.90 ± 0.06 ppm for crystalline rocks and 1.60 ± 0.16 ppm to 3.21 ± 0.12 ppm for lunar fines and fragmental rocks. The beryllium content of all the Apollo 14 samples is higher than that of the Apollo 11 and 12 samples. The fines and fragmental rocks of Apollo 14 generally show a higher beryllium abundance than in the crystalline rock 14310, consistent with the trend observed in previous Apollo mission samples. It should be noted that the LSPET has found the crystalline rock 14310 to be very similar to the Apollo 14 fragmental rocks and also that it has the highest nickel content of any lunar basaltic rock so far analyzed. This rock has generated special interest since it does differ in composition substantially from previously described basaltic rocks. The beryllium concentration found in 14310 is approximately seven-fold higher than that found in the Apollo 12 crystalline rocks.

In contrast with the beryllium results, the chromium content of the Apollo 14 samples has been found to be significantly lower than in previous Apollo mission samples. The one sample of Apollo 14 fines (14163) contains 0.148 ± 0.003% chromium while the fragmental rocks vary from 0.103 ± 0.006% to 0.138 ± 0.002% chromium. The chromium content of Apollo 11 and 12 fines varied from 0.209 ± 0.009% to 0.270 ± 0.016% and the
crystalline rocks varied from 0.208±0.019% to 0.621±0.03%. Crystalline rock 14310 shows a concentration of 0.115±0.006% which is significantly lower than the 0.30% chromium concentration in 14053 (the only other crystalline rock returned by Apollo 14), as determined by the LSPET. This finding supports the beryllium data and the observation of the LSPET that lunar rock 14310 is indeed unusual and that it closely resembles the fragmental rocks in chemical composition.

Analyses of exterior and interior chips of fragmental rock 14321 have shown definite heterogeneity in the beryllium and chromium abundances. The beryllium concentration varies from 1.77±0.13 ppm to 6.99±0.14 ppm while chromium has a range of 0.177±0.005% to 0.410±0.006%. Due to the heterogeneity of this rock, the beryllium and chromium values are not included in the ranges given for the fragmental rocks. The range in values for this one rock is not surprising since most of the Apollo 14 rocks have been found by the LSPET to be fragmental and of complex structure. The heterogeneity may be a result of this complexity and varying degrees of crystallinity within the rock. The analyses of the seven different chips of 14321 reflect the general trend of the beryllium concentrations varying inversely with chromium concentrations, just as has been observed on a macroscopic scale for various rocks and fines from Apollo 11 and 12 sites. Thus the individual chips from 14321 showing an enrichment in one element show a depletion of the other.

Examination of the results of this study concerning beryllium and chromium abundances in the Fra Mauro region of the moon reveals several interesting phenomena. This highland represents a region enriched in trace elements, e.g., beryllium, which could have been transported to nearby regions of the lunar surface. If redistribution of material to preferentially contaminate mare fines occurs, this could account for the beryllium and other trace element enrichment in the fines relative to the crystalline rocks at the Apollo 12 site.

References
LEAD VOLATILIZATION AND VOLATILE TRANSFER PROCESSES ON THE MOON

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The striking differences in Pb-U-Th systematics and apparent ages between individual Apollo 11 basalt fragments and lunar regolithic debris presumably derived from them, led Silver (1970a) to infer the presence in the soil of lead components with unusually high \( \frac{\text{Pb}^{207}}{\text{Pb}^{206}} \) radiogenic ratios and suggested the possibility of unexpected fundamental lunar processes. A combination of leaching and volatilization experiments established the extraordinary heterogeneity and lability of lead isotope systems in 10084 (Silver, 1970b) and identified the presence of parentless leads which Silver concluded were produced by volatile transfer reflecting the numerous thermal events recorded on the lunar surface. Tatsumoto (1970 and 1971) showed that \( \text{Pb}^{207} \)-rich leads tended to be concentrated in the finer-size fractions of the Apollo 11 and 12 soils and concluded that volatilization indeed had enriched the dust in an older component of surface-related lead. Silver (1971) and Huey et al. (1971), reporting initial systematic studies of volatilization of lead from Apollo 11 and 12 soils, identified the presence of a significant volatile lead component mobile at quite low temperatures (below 700° C). The lowest temperature yields were, indeed, most enriched in \( \text{Pb}^{207} \).

This work abstracts comprehensive studies on the vacuum volatilization of eight soils and two soil breccias from all four Apollo sites. A technique of extended step-heating in vacuum has been used to define lead sites in the soil with characteristic release activation energies. In every sample (see Silver, 1970b, for 10084 data), 70 - 80% of all the total lead can be volatilized at temperatures less than 1000° C. The most interesting lead yield occurs in all samples at temperatures between 475° and 600° C, peaking in yield at about 550° C.

In Figure 1, the release of data for 14163 soil is plotted against exposure time in a vacuum furnace. The lead extracted at 555° C shows an abrupt release and decline and represents 9% of the total Pb in the soil. Two successive increments (1 and 2 hours), differing by a factor of 2 in quantity, gave \( \frac{\text{Pb}^{207}}{\text{Pb}^{206}} \) values 1.272 and 1.267 indicating a lead component in a unique low activation energy site characterized by a uniform value of about 1.27. The very small lead release at 595° C (~0.1%) for 20 hours probably is a relict of the same site and has a \( \frac{\text{Pb}^{207}}{\text{Pb}^{206}} \) ratio of 1.255.

This accessible lead has been shown to be surface-area correlated by several independent experiments we have performed. Most important is the fact that it is precisely related to the slope of the line fitted through the entire family of Apollo 14 soils which is shown in Figure 1 of the accompanying paper by Silver. Parentless lead of this general composition is unambiguously the major variable responsible for the linear array. In comparing yield against isotopic composition, it is clear that while there may be one or two small intermediate release sites, the next major release takes place above 800° C, below and close to 970° C at which temperature more than 18% of the lead was released in one hour and 45% more in the next 20 hours.

The leads released in these increments have \( \frac{\text{Pb}^{207}}{\text{Pb}^{206}} \) ratios of 0.76 and 0.69, close to and below the values typical of the total sample. This second major site clearly
LEAD VOLATILIZATION AND VOLATILE TRANSFER
Leon T. Silver

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has a very different type of lead as well as activation energy. We believe this site to be within the silicate phases and we have performed several experiments on ground lunar basalt to demonstrate this. At the end of 85 cumulative hours of heating up to the 970° C temperature, the 14163 sample is still an incoherent powder showing almost no sintering or any other effects as drastic as the lunar events which have converted at least half of the 14163 soil to various glasses in the lunar environment.

The existence of these two major lead sites in soils have been confirmed in all of the samples we have studied. The higher temperature site (hts - probably peaking at 900±°C) yields 60-75% of all the lead in each soil sample. Still higher temperature site(s) release essentially all remaining lead as fusion temperatures are approached at 1100-1200° C. Quantities of lead in the lower temperature site (lts <600° C) vary with each sample and appear to be related to the local lead concentrations and regolith history. In all samples, Pb204 is preferentially concentrated in this site over higher temperature sites.

The isotopic compositions of the lts leads extracted from each sample are extremely variable from mission site to mission site. They are variable in a much more restricted range among different samples from the same mission site and do not tend to overlap values from other sites. The following data shows the best maximum value for the radiogenic Pb207/Pb206 value (corrected for contamination of Pb204) of the low temperature lead from a soil from each mission site: 10084 (207/206 = 0.97 ± .02); 12070 (0.78 ± 0.03); 14163 (1.27 ± 0.02); 15021 (0.81 ± 0.02). It is clear that there is a positive correlation between the ages generally assigned to the surfaces on which these soils were developed and the isotopic composition of the accumulated low temperature lead. A more subdued variation and correlation also occurs in the 900+° C site.

The low temperature volatile lead observations reported here have major implications and remarkable potential for understanding and explaining lunar surface history. (1) They clearly support volatile transfer as a major geochemical process on the lunar surface. They imply that many other volatile elements (we have some direct observations on Bi and Rb migration, for example) must be redistributed in this manner and that all geochemical distinctions reported between lunar soils and rocks must explicitly consider the possible effects of volatilization. (2) The provinciality observed for the lts composition of samples studied so far establishes unequivocally that there has been no moonwide comprehensive mixing or saturation by "magic component" for the Pb-U-Th systems since these regolithic sites formed. However, one can reasonably expect some moonwide volatile transfer effects from very large surface thermal events on the moon. The earliest lunar history with its enormous flux of cratering impacts and profound magmatic events may well have produced moonwide atmospheres which were recorded in the quantities and isotopic properties of volatile lead. (3) The lts lead, because of its peculiar isotopic properties and origins can be a major geographic and temporal tracer of all types of surface lunar thermal events and the transient atmospheres produced by them. (4) The earliest records of lunar history possibly may be found by exploring the volatile leads in deep drill cores on the oldest lunar surfaces we can sample. This is because some recycling of the enormous quantities of lead released to the surface in the early stages of lunar history probably continued long after access to their sources have been buried or modified by subsequent lunar events. For example, the value of 1.27 observed in the 14163 lts lead could not have formed instantaneously more recently than 4.05 b.y. ago and undoubtedly includes some much older contributions.

Step-heating vacuum extraction of volatile lead in Apollo 14 bulk soil, plotting release temperature; yield per unit time; and radiogenic Pb$^{207}$/Pb$^{206}$ ratio corrected with contamination lead. Note broken scale for yield curve. Pb$^{207}$/Pb$^{206}$ data is shown as observed increments, and as a smoothed curve for projection to zero yield.
U-Th-Pb ABUNDANCES AND ISOTOPIC CHARACTERISTICS IN SOME APOLLO 14 ROCKS AND SOILS AND AN APOLLO 15 SOIL
Leon T. Silver, Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California 91109

Two breccias, a basalt, and four soils returned from Fra Mauro by the Apollo 14 mission have been analyzed by mass spectrometry and isotope dilution, in replicate, for their lead isotopic composition and the concentrations of Pb, U and Th (Tables 1 and 2). All samples display extremely radiogenic leads at significantly higher concentrations than previous mission collections. The uranium (3.0 to 4.9 ppm) and thorium (11.2 to 18.7 ppm) represent the highest average concentration levels found at any lunar site sampled thus far, and are clearly responsible for most of the radiogenic lead observed. Several lines of evidence indicate that these enrichments (and similar enrichments observed in light-colored breccias and soils from Apollo 12 site) are unusually high for the moon. They apparently reflect a significant regional radioactivity anomaly in the crust of the southern Imbrian basin region where geologic evidence suggests both the Apollo 14 and 12 breccia materials were derived. The geochemical process(es) responsible for the enrichments of U and Th have not fractionated them; the Th/U ratio is constant at about 3.74 ± 0.10 in both rocks and soils, a value typical of Apollo 11 and 12 samples, also. In contrast, the U238/Pb204 ratio, an index of original U-Pb abundances is higher by a factor of two or more in the rocks compared to the soils, suggesting preferential enrichments of Pb204 in the soil (barring contamination effects). However, U238/Pb204 values for these rocks are not significantly different from those observed in the Apollo 11 and 12 basalts which have from 1/4 to 1/20 of the uranium levels found here (Tatsumoto 1970, 1971; Silver, 1970, 1971). Pb-U isotope systematics and apparent ages are represented on the concordia diagram of Figure 1. They fall into two distinct classes, separating soils and soil breccias from rocks.

The basalt, 14310, the most radioactive lunar igneous rock we have analyzed, yields concordant apparent ages in the range 4.27 ± 0.03 b.y. This once again is distinctly older than ages suggested by Ar39-Ar40, and Sr87-Rb87 isotope relations (3.9 ± 0.1 m.y., Wasserburg et al., 1971). This rock is probably a clast from the Fra Mauro breccias, although other interpretations are possible. When compared with the coherent dark-clast breccia, 14303, a rock containing clasts of several generations of breccias as well as basalt fragments similar to 14310, the basalt is only slightly older than the breccia which also yields essentially concordant ages at 4.20 ± 0.03 m.y. A single clast of dark older breccia from within 14303, yields apparent ages of 4.12 ± 0.04 b.y., slightly younger than the composite breccia. This data strongly suggests either that the U-Th-Pb systems in the breccia have been extensively but not completely re-equilibrated by events more recent than 4.1 b.y., or that the breccia is comprised of a mixture of materials with a variety of ages and was aggregated in present form more recently than 4.1 b.y. Whether this significant event (which cannot be much less than 4.0 b.y.) is the Imbrian impact requires extended consideration of the available geological and geochemical information on the sampling at the Fra Mauro site.

The second breccia, 14047, is a light gray friable breccia containing significant amounts of glass spheres, agglutinates and fragments indicating it is a soil breccia with a
U-Th-Pb ABUNDANCES AND ISOTOPIC CHARACTERISTICS
Leon T. Silver

prior interval of regolithic reworking. The most abundant coarser fragments are light-gray to white plagioclase-rich lithic fragments, up to 2 mm in diameter, comprising about 3-5 percent of the rock. The light-colored clasts are similar to those found in many other breccias (H. Wilshire and E.D. Jackson, Apollo 14 PET, pers. comm.). Analysis of the composite breccia sample shows it, indeed, to be similar in Pb-U systematics to the soils (Table 1, Figure 1), whereas an analysis of about 10 mg of the light clasts handpicked from 14047 is indistinguishable in Pb-U ratios and apparent ages from 14303, and very similar to 14310.

Soil analyses (Table 2) reveal the same systematic enrichments in old leads relative to uranium and thorium compared to local rocks that were observed at Apollo 11 and 12 sites, and they occupy a distinctly older region of the concordia diagram. Bulk soil sample (<1 mm), 14163, was analyzed several times after being sampled from the original allocation in different ways. The analyses vary sufficiently to indicate that size and aliquoting procedure can affect bulk values significantly by varying the ratio of fines to coarse material. Fines contain surface area-related enhancements of a Pb207-rich lead (discussed in an accompanying paper). 14163 analyses produce a linear array (Figure 1), which lies on and near concordia in the general apparent age region of 4.87 b.y. These values are markedly different from those observed in Apollo 11 and 12 soils in their radiogenic Pb207/Pb206 ratios, patterns of discordance and model "ages". They demonstrate a significant regional provinciality in U-Pb isotope properties which has most important implications. They further support our interpretation (Silver, 1970) that the composite U-Th-Pb analysis of an individual soil cannot be argued to represent the age of the moon. Three other soils from trench and comprehensive sample areas, 14140, 14149, 14421, and breccia 14047, fall on or near the linear array on "Concordia", reflecting different contributions of the surface-correlated Pb207-rich leads. The projected lower intersection, about 2.90 ± 0.05 b.y. may be the artifact of several different mechanisms, and cannot be given a unique geologic time significance at this point. It is clear, however, that components other than the rock materials analyzed in this work must be present in the family of soils to explain the pattern observed.

Analysis of the Apollo 15 contingency sample shows it to have also a distinctly different set of U-Th-Pb systematics, apparently related to its site. A more systematic study of soils and rocks from the Hadley-Apennine region is underway.

References:
Table 1
U-Th-Pb SYSTEMATICS IN SOME APOLLO 14 ROCKS

<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>Pb Composition*</th>
<th>Concentrations (ppm)</th>
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</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>206 207 208 Pb</td>
<td>U  Th Th/U</td>
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<tr>
<td>14310,70</td>
<td>Basalt</td>
<td>865.5 432.71 839.0</td>
<td>6.113 3.03 11.20 3.70</td>
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<td>14302,12A</td>
<td>Single dark clay</td>
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<td>14302,12B</td>
<td>Dark clay breccia composite</td>
<td>925.0 453.49 996.0</td>
<td>8.43 4.35 16.16 3.72</td>
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<tr>
<td>14047,14</td>
<td>Light clay soil breccia composite</td>
<td>383.18 260.15 370.3</td>
<td>7.77 3.34 12.39 3.71</td>
</tr>
</tbody>
</table>

*Corrected for chemistry blanks of 10-20 ng

Table 2
SOME APOLLO 14 AND 15 SOIL ANALYSES

<table>
<thead>
<tr>
<th>Sample</th>
<th>Pb Composition</th>
<th>Concentrations (ppm)</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>206 207 208 Pb</td>
<td>U  Th Th/U</td>
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<tr>
<td>14163</td>
<td>Expt #1</td>
<td>368.4 274.6 330.4</td>
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<td>Expt #1</td>
<td>510.51 429.60 480.00</td>
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<td>14148</td>
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<td>Expt #3</td>
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<td>10084</td>
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</tr>
<tr>
<td>12070</td>
<td>Expt #1</td>
<td>209.1 123.4 214.8</td>
</tr>
</tbody>
</table>

Figure 1
U-Th-Pb Isotope Relations in Some Apollo 14 Breccias and Soils

- Rock fragments, coherent breccias and breccia clasts
- Soils and soil breccias

Error figures show dimensions of uncertainty related to isotopic errors and choice of initial lead.
- Modern contamination lead
- Primordial meteoritic lead
Radio tracking data obtained from the command and service module while in low orbit (≈ 15 km periapsis altitude) has revealed a detailed gravity profile across the lunar front face near the Apollo 15 trajectory path providing new information about the mascons in Serenitatis and Crisium.

The data are essentially spacecraft speed measurements derived from a Doppler system having an accuracy of 0.65 mm/sec. The raw data represent or contain many components of motion, and these are removed before gravity analysis proceeds. Factors that are accounted for include the tracking-station rotation about the Earth spin axis; the spacecraft motion perturbed by accelerations from the Sun, Earth, Moon, and planets, and atmospherics and signal-transit times. All these quantities are known apriori and are removed to accuracies well beyond those required to evaluate local gravitational effects.

The reduction was performed using the Jet Propulsion Laboratory orbit-determination computer program, which contains the theoretical model with all the dynamical constraints and parameters previously mentioned. Each orbit of data (≈ 65 min) was evaluated independently, with the doppler observations least-squares-fitted using a spherical Moon and solving for (adjusting) only the six state parameters of initial position and velocity. The resulting systematic residuals (i.e., real observations minus theoretically calculated observations) are then attributed to lunar gravitational effects. Analytic patched cubic splines with continuous second derivative are least-squares fitted to the residuals. These functions are then differentiated and the line-of-sight accelerations ("gravity") are analyzed and correlated with the subspacecraft lunar track and existing topography.

The most striking features in Figure 1 are the large positive gravity anomalies over Mare Serenitatis and Mare Crisium. To analyze these features for gross shape parameters simulations were made using two different models. The first was to generate tracking data with a theoretical moon having surface disk features at the Serenitatis and Crisium locations. Then the same fitting process used on the real data was applied to the simulated data and acceleration profiles were obtained. The second simulation was the same as the first except that the theoretical moon had spherical bodies 100 km deep rather than the surface disks. A comparison of these simulations for Serenitatis with the real data is shown in Figure 2. The surface disk does remarkably well in matching the real data profile while the deeply buried spherical body does a very poor job. Similar results were obtained for Crisium. This was precisely the same result noted for Nectaris from Apollo 14 data reduction. Moreover, the mass per unit area in these simulations was held at the Nectaris value of 800 kg/cm², which strongly suggests that the mass distribution per unit area is about the same for all three mascons. The broadness of the profile depends on the radius of the surface disk which was 245 km for Serenitatis centered at
GRAVITY MEASUREMENTS FROM APOLLO 15
W. L. Sjogren

19°E longitude and 26°N latitude and 210 km for Crisium centered at 59°E longitude and 17.5°N latitude. The spherical body had a mass of 9 x 10⁻⁶ the lunar mass for both mascons.

Another interesting feature is the definite shoulder in the Serenitatis gravity anomaly at 12°E longitude. (There is also a hint of the same shouldering at 20°E longitude). Compensating for geometry again and possible surface feature correlation, these locations are shifted closer to 13.5° and 21.5°E longitude respectively. This is an indication of an abrupt excess mass change, implying a change in surface or subsurface structure. Possibly it is due to subsurface ring structure usually associated with circular maria but ill defined for Serenitatis. This arc structure is visible in Whitaker's work (The surface of the Moon, 1965, John Hopkins Press, pp 79-90) about halfway between Apennines and the Tycho ray.

The Apennines show up quite clearly as a local high with approximately an 85 milligal positive anomaly (obtained by smoothing Serenitatis curve into the low at 0° longitude). Assuming a 4 km height and a 100 km width and 200 km length for a non-isostatic feature having a density of 3.0 gm/cc, the anomaly should be 300 milligals, implying that there has been partial isostatic compensation.

The Marius Hills at 50°W longitude exhibit a 60 milligal positive gravity. This is confirmation of the Lunar Orbiter data which also reveals a positive anomaly in this region. It seems odd that the highlands terrain south of Archimedes at 5°W longitude doesn't reveal any distinct signature while being apparently the same extent as the Marius Hills. These two areas are, however, different geologic blocks as shown on U.S. Geological Survey Maps.

The large negative gravity anomalies east of Crisium and Serenitatis are possibly part of a negative ring structure similar to that determined for Orientale from the Lunar Orbiter data. This will be confirmed or denied once the solid coverage from the subsatellite is obtained.

Other areas of interest such as the low at 65°E longitude, the low amplitude but slightly positive gravity for Procellarum and the low at 0° longitude are generally consistent with Lunar Orbiter data results. Imbrium doesn't appear as a significant high, for the trajectory path is outside the inner ring. However, at 18°W longitude there is a positive gravity tongue which extends out of the Imbrium basin and this shows up mildly as 50 milligals.

The subsatellite from Apollo 15 has provided dense coverage at relatively low altitudes of 60 km to 40 km and will definitely reveal new anomalies especially around the 30° latitude band where previous data from the Lunar Orbiters were at 200 km or more. This reduction is still in progress and will require several more months before a new gravimetric map can be produced. Work is also proceeding on estimating a variable mass distribution in the form of circular surface bands for Nectaris, Serenitatis and Crisium mascons.
Figure 1. Gravity Profile along Apollo 15 Orbit No. 4.

Figure 2. Mare Serenitatis Gravity Profiles.
VAPOUR PHASE GROWTH OF FELDSPAR CRYSTALS AND FRACTIONATION OF ALKALIS IN FELDSPAR CRYSTALS FROM 12038.22.

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Rock 12038, an ophitic basalt, provides convincing evidence of the movement of small amounts of material in the vapour phase during cooling and crystallization. The rock itself is similar to a number of basalts returned on the Apollo 12 mission, though it is the most aluminous of those analysed by Kushiro and Haramura (Science, 171, p. 1235-1237). The rock is an assemblage of anorthite-rich plagioclases, pyroxenes, ilmenite and minor quantities of olivine and chromite-rich spinel. The texture of the rock is also characteristic of many lunar basalt samples from both the Apollo 11 and 12 missions, being relatively coarse grained and vuggy. The vugs are very irregular in shape and are bounded by crystals of the host rock which are terminated by well-formed crystal faces where they project into the opening. The grain dimensions of the well-formed crystals projecting into the vugs is identical to that of the matrix.

In addition to the vugs, 12038 contains a number of perfectly preserved vesicles. These are so smooth-sided and spherical in shape that at first glance they seem to be lined with glass. Close examination shows that the walls are in fact lined with innumerable grain surfaces of the host minerals.

Within both the vugs and the vesicles, perfectly formed whisker-shaped crystals of plagioclase are found. They are extremely elongated with dimensions ranging from 2 to 30 microns across, and as much as 250 microns in length. The feldspar whiskers are not crystallographically continuous with matrix phases, but nucleate on each of the three major matrix phases without preference. They do not contain solid inclusions nor overgrow other phases as observed in the matrix feldspars and prove to be compositionally different from the matrix feldspars.

While it may be argued that whisker crystals growing in vugs were simply deposited by late-stage liquids which later drained away, those in the vesicles cannot be so explained, and must have been formed by vapor phase deposition after the vesicle was frozen in. The whiskers are all elongated parallel to the a-axis, are untwinned and despite exposure times of up to two weeks, do not show evidence of c-type reflections on x-ray
precession photos. The whiskers therefore appear to have the space group $\overline{1}$, while the matrix feldspars clearly show c-type reflections and have space group $\text{C}1$. The whisker crystals display an unusual property of flexibility and even after bending through as much as 90° will return to their origin position. After a few flexings, however, the usual brittle properties of feldspars assert themselves and the crystal fragments.

Microprobe analyses of matrix feldspars shows them to average An$_{83.5}$ Ab$_{16.0}$ Or$_{0.5}$, with a range of An$_{81.4}$ to An$_{87.1}$. The whisker crystals, on the other hand, are consistently richer in alkalis and average An$_{74.0}$ Ab$_{24.0}$ Or$_{2.0}$, with a range of An$_{70.1}$ to An$_{81.0}$ and an Or range of Or$_{3.4}$ to Or$_{0.8}$.

The mineral textures and assemblage of the matrix are apparently primary, and do not allow the interpretation that the rock has been metamorphosed. Nor is there any evidence of alterations and depositions as might be expected if a post-magmatic vapor phase were introduced from the outside. We therefore suggest that the whisker feldspars are deposited from vapors released by the cooling igneous rock late in its crystallization history, perhaps at the stage when the alkali-rich residual glass reported from most lunar basalts had been reached. We have no direct evidence on the bulk composition of the transporting vapors, but they must have been alkali-rich relative to the parent rock.
ELECTRON MICROSCOPE STUDY OF APOLLO SAMPLE 14310 Deane K. Smith, Peter A. Thrower and Wesley P. Hoffman, Pennsylvania State University, University Park, Pa. 16802.

Two sections of sample 14310 have been thinned by ion bombardment to thicknesses allowing transmission electron optical studies. One of the observed phases has been positively identified as a transitional anorthite. Other phases have been observed which show a variety of substructure, but whose diffraction patterns cannot be identified with any phase reported from the sample by other investigators.

The transitional anorthite was identified by obtaining several sections of reciprocal space through diffraction patterns rotated about a common axis. Using the d-spacings and nature of the observed reflections as strong-sharp, weak-sharp, diffuse, and absent according to the classification of anorthite reflections by Gay (Mineral. Mag. 30, 169 (1953), the reciprocal lattice was indexed. The resulting unit cell yielded the following d-value comparisons with transitional anorthite (Ribbe and Megaw, Norsk Geol. Tidssk. 42, 158-167 (1963)).

<table>
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Interplanar angular relations between observed diffraction maxima were all within 0.8% of the values calculated from the Ribbe and Megaw cell.

The corresponding electron micrographs of the transitional anorthites were essentially featureless. No evidence could be found for twinning, domain structure, or second phase precipitation on the scale investigated using bright field or dark field techniques.

Several other areas have been studied which show well defined lamellar and cross-lamellar structures. These lamellae are variable in size from about 200 to 5000 Å in width, and the interfaces are planar and parallel within the observational limit. Dark-field micrographs show an alternation of bright-dark contrast indicative of polysynthetic twinning. Within the twin lamellae, several structures are observed. Some show a well defined second phase precipitation as small spherical homogeneously distributed particles around 200 Å in diameter. Others show a structure which appears to be a segregation of phases or twinning with ill defined interfaces but with a general planar character that is at an angle to the twin interfaces. These features are more clearly observed using dark-field illumination and are 100 to 200 Å in thickness. Using different diffracted beams for the dark-field images, the character of these segregations may be seen to change from the lamellar appearance described above to more complex structures.
Diffraction patterns of these areas have been complex and have not yet been satisfactorily identified. Although pyroxene and anorthite are the dominant phases reported from this sample and are the most likely phases to be encountered, the observed patterns do not appear to correspond to patterns expected from either of these phases. Nor are they comparable with any other phase reported from sample 14310. The twinning complicates the diffraction patterns leading to difficult interpretation, but with these complications in mind diffraction patterns obtained from different areas showing similar twinned structures do not appear to be interrelated. Thus more than one phase may be indicated. Continued work using diffraction patterns of several sections of reciprocal space as for anorthite should help identify these phases.
The solar wind spectrometer on the Apollo-12 ALSEP has operated continuously since its deployment, measuring the properties of the plasma in the solar wind and in the magnetopause. For lunar science the primary goal of the experiment is to determine the nature and magnitude of the modification of the plasma by the presence of the moon. Information pertinent to this goal can be obtained from simultaneous measurements by a probe on the lunar surface and another away from the moon but in the same region relative to the Earth's magnetosphere. Such simultaneous measurements have been rare, but they indicate that the velocity of solar wind protons at the moon is usually significantly less than in space. The difference is attributed to the interaction between the plasma and the 38-gamma magnetic field at the Apollo-12 site. The deceleration is less than that predicted for the compression of a uniform field, probably because of the relatively small spatial extent of the field.

With the Apollo-15 solar wind spectrometer now also in operation, simultaneous measurements at two points on the moon are possible. At this time magnetic tapes for only 20 hours of simultaneous data have been received—too small a quantity to provide any firm conclusions. A much larger body of data should be available for discussion at the conference.

Study of the data on the electrons in the plasma has recently begun. Considerable directional anisotropy is observed, and rapid changes in flux occur at times when the proton fluxes are constant. Possible interpretations of these observations will be discussed.

Using Apollo 12 data our earlier determination (Sonett et al., 1971a and 1971b) of the lunar electromagnetic transfer function from Fourier transforms of both the Lunar Surface Magnetometer (LSM) data and the Explorer 35 interplanetary magnetic field data has been extended downwards in frequency by approximately one decade. The frequency range of the transfer function now extends from half the Nyquist limit of Explorer 35 (f=4x10^{-2}Hz) down to periods of about 2 hours (f=5.0x10^{-4}Hz). The transfer function still displays the asymmetric response between the north-south and the east-west directions even at the lowest frequencies. The data has been inverted using the Newton-Raphson technique and least squares fitting as before. The very low response in the neighborhood of two hour periods suggests that a limit has been reached such that no significant data can be obtained from extensions of the analysis to lower frequencies.

Detailed consideration of the response of the Moon to higher order modes of excitation by the solar wind indicates that the high frequency plateau region behavior of the transfer function, including its slope and variability, may well be explained by excitation of higher order modes (Schubert and Schwartz, 1972). In order to assess this effect it is important to consider the polarization and direction of propagation of the incident waves on the Moon and the position of the LSM with respect to the incident wave vectors. This problem is still under investigation but it seems likely that the low frequency asymmetry in the lunar transfer function cannot be explained in this way. Another potential source for the asymmetry, i.e., transverse magnetic (TM) excitation, also does not seem likely since the very low frequency response is symmetric. Electric multipole radiation due to TM excitation appears to be below the level of detectibility.

Another candidate for the source of the asymmetry lies in excitation of the permanent magnetic field at the Apollo 12 site by variations in the dynamic pressure of the solar wind (Clay et al., 1971). Tests for this effect are presently under way and begin with rotation of the basic coordinate system in the plane tangent to the surface of the Moon. Directions of the maximum in the lunar response appear to be related to the direction of the permanent field at the site in a complicated manner. Above 1.7 mHz a monotonic change in the direction of maximum tangential response takes place as the frequency is increased. This indicates that if a pressure effect is present it must be frequency dependent or mixed with wave polarization effects which vary
Progress Report: LSM
C.P. Sonett

with frequency. Minimum variance determinations of the wave vector directions in the solar wind have been made to assess the effects of polarization and direction of propagation. There is a definite tendency for the solar wind to contain fluctuations with k vectors aligned with the mean magnetic field, while the central angle between the wave vector and the LSM tends to cluster broadly about 90 degrees showing that polarization effects and higher order modes of excitation in the Moon may all be important in a final assessment of the induction. Study of plasma pressure effects and separation of this from higher order mode excitation effects is still under way.

Time series taken from the lunar darkside period of the first lunation comprising 18 swaths of 1 and 2 hour lengths have been analyzed. Theoretical calculations have been completed of the response for a vacuum Moon so that comparison can be made between data and theory. The tangential component of the transfer function near the antisolar point deviates strongly at low frequency from the theoretical vacuum response. The source of the departure is likely currents lying in the diamagnetic cavity boundary or possibly flowing in the lunar cavity above the surface of the Moon. The radial transfer function seems to track the theoretical value closely with minor departures.

The determination of the heat flux from the Moon based upon thermal gradient calculations show that the dominant effects are due to field line compression, and that such effects as change in the multipolarity spectrum (relative dipole and higher order excitation) are masked. The angular resolution on the Moon is very poor due to smearing of the relative contributions as well as the dominance of the magnetic dipole interaction. Therefore the results are a good quasi-global average, but still subject to contamination from effects connected with field line containment or additional noise sources. Such sources have not been identified but could arise from the current sheath which conceptually may add or subtract noise from the scattered wave field depending upon coherence and phase. Tests for this effect are associated with the angular dependence of the complex transfer function and are still under way. Although a noise contaminant of this type could alter our values for the heat flux, the present values are consistent with the determination of the deep interior temperature and the latter rests upon very low frequency data which is less likely to suffer this type of contamination.

The permanent magnetic field value from Apollo 15 site is 5±5 gamma; a final value will depend upon normalization to Explorer 35 and perhaps the Apollo 15 subsatellite. The approximate value reported here is substantially lower than any previous report from other sites. The geological complexity of the Hadley site makes any guess as to the reason for the low value speculative especially in view of the very local nature of the measurement. On the other hand Explorer 35 shows that large regions of the Moon are magnetized at a presumably low level with a preference shown statistically for localities in the highlands, therefore favoring the backside of the Moon. A map has been
prepared showing the incidence of these features (Mihalov, et al., 1971; Sonett and Mihalov, 1972). The existence of large scale fossil magnetism ubiquitously over the Moon together with the evidence for a background field present over the whole span of Rb/Sr ages indicates the possibility that the Moon was endowed with a dynamo field till at least 3.2 Aeons ago. For a conventional planetary dynamo then the spin history becomes an important factor. Finally this leads to speculation that spin damping of the Moon took place less than some 3.2 Aeons ago. In that event a capture hypothesis based upon a time of capture some 3.2 Aeons ago becomes attractive. Using the Singer "soft" prograde capture a potential strong dynamical coupling with the earth at that time becomes possible in accord with the marginal geological record available from that time, though serious difficulties remain with a dynamo source (Sonett and Runcorn, 1971c).

REFERENCES


THE ROLE OF ISOTOPIC MASS FRACTIONATION IN THE PRODUCTION OF NOBLE GAS ANOMALIES IN LUNAR FINES FROM THE APOLLO 15 MISSION, B. Srinivasan, E. W. Hennecke, D. E. Sinclair, and O. K. MANUEL, Department of Chemistry, University of Missouri, Rolla, Mo. 65401

The abundance and isotopic composition of helium, neon, argon, krypton and xenon have been measured in the gases released by the stepwise heating of lunar fines, # 15601.64. The extraction temperatures were varied from 100°C to 1500°C in 100°C steps. The release pattern shows a preferential release of the lighter weight isotopes at low extraction temperatures.

Due to the role of isotopic fractionation, it is not possible to resolve any of the five noble gases into a mixture of spallation and fission components with a unique solar wind-implanted component.

In order to compare noble gases in # 15601.64 with noble gases in meteorites and in the atmosphere, the following equation (1, 2) is used to calculate the enrichment, \( r \), of the heavy isotope of mass \( m_2 \) relative to a lighter isotope of mass \( m_1 \):

\[
\log r = \left( \frac{m_2 - m_1}{m_2 + m_1} \right) \log \left( \frac{\text{Initial Volume}}{\text{Final Volume}} \right) \ldots (1)
\]

Due to the strong depletion of noble gases in meteorites and in the earth's atmosphere, the isotopic composition calculated from the above equation shows a marked deviation from linearity across the isotope ratios of xenon and krypton.

When corrections are applied for the spallation component, our results indicate that the variations observed in the isotopic composition of neon in the different temperature fractions of lunar fines define a pattern which fits eq. (1) and extrapolates to atmospheric neon. This same fractionation pattern seems to fit the variations observed in trapped neon from gas-rich meteorites (3) and carbonaceous chondrites (4).

Variations observed in the isotopic composition of krypton from different temperature fractions of lunar fines, after subtraction of the spallation component, also fit closely to the pattern described by eq. (1), and this same
isotopic fractionation pattern extrapolates to atmospheric (5) and AVCC (6) krypton.

The isotopic composition of xenon observed in the gases released from lunar fines at various extraction temperatures also appears to result from spallation reactions and from isotopic mass fractionation. Although the fractionation effects across the isotopes of xenon are minimal, the fractionation pattern appears to extrapolate better to AVCC xenon (6) than to atmospheric xenon (7), particularly for $^{134}$Xe and $^{136}$Xe. Our results seem to confirm the suggestion that solar and AVCC xenon may be related by isotopic fractionation (8), but that atmospheric xenon is distinct from xenon observed in other parts of the solar system. This observation is in conflict with the suggestion that solar and atmospheric xenon are related by simple isotopic fractionation (9). No evidence for decay products from $^{129}$I or $^{244}$Pu was observed.

The helium and argon released from # 15601.64 also appear to be a mixture components from spallation, radioactive decay, and isotopic fractionation of solar wind implanted gases. The role of isotopic fractionation is difficult to resolve due to the small number of nonradiogenic isotopes of these two gases. The $^{40}$Ar/$^{36}$Ar ratio varied from 0.57 to 33.1 and the $^{38}$Ar/$^{36}$Ar varied from 0.172 to 0.208. The $^{4}$He/$^{3}$He ratio varied from 1356 to 2169 (10).

References and Notes


ROLE OF ISOTOPIC MASS FRACTIONATION

B. Srinivasan


10. This research was supported by the National Aeronautics and Space Administration, NASA-NGR-26-003-057.
Lithic fragments from 1-2mm fines 14002,7 and 14258,33 were selected for igneous character (no clasts, only residual glass) and sufficient grains for satisfactory statistics. Oxide variation diagrams (Fig. 1 and 2) plus textures and mineralogy (Fig. 3) indicate 3 major groups: I high-Al basalt (14310 type); unlabeled x in Figs. 1 and 2 and cols. 1, 2 in Table 1; low FeO and MgO and high CaO and Al2O3 resulting from 50-65% plagioclase (An 80-95, Fe .1-.4) and 45-30% pyroxene (augite and pigeonite); mean Fe/Mg (atomic) from bulk analysis 0.6: II high-Fe basalt (14053 type); actually measured in rocks 14072,10 and 14053 since coarse pyroxene and olivine laths (>1mm) preclude accurate measurement in fines; two labeled x in Figs. 1 and 2 and col. 3, Table 1; matrix plagioclase, ilmenite and finer pyroxene; pyroxenes mostly pigeonite but augite rim on phenocrysts; mean Fe/Mg (atomic) from bulk analysis 0.7: III high-Mg basalt (norite); dots in Figs. 1 and 2 and cols. 4,5 in Table 1; fine grain (<.05mm) and poikilitic to sub-ophitic pyroxenes (single population En 75 Wo 5 Fs 20, see Fig. 3); mean Fe/Mg (atomic) from bulk analyses - 0.3. Bulk composition of fines (LSPET, 1971) (col. 6) fits best with 39% I + 26% II + 35% III (calc. anal. col. 7, Table 1). Breccia analyses (LSPET, 1971) probably contain more of III because of higher MgO values than fines. Rare lithic fragments are dunite (labeled P, Fig. 1,2; Fo 90, chromite), high-silica (Q, Fig. 1,2; SiO2 = 60%, K2O = 2.3, BaO = .4); and anorthosite (R, Fig. 1,2; Al2O3 = 30%, Na2O = 1.7%).

Apollo-14 glasses (Reid et al., 1971) designated type B, C and E on Fig. 1 (inner lines enclose 50% of analyses) show little correspondence with the lithic fragments. Of the three dominant ranges of glass composition, only Type E corresponds even approximately to group I. The glass compositions are too low in Mg to account by themselves for bulk soil and breccia analyses (LSPET, 1971). The high-Mg lithic fragments of group III are not represented by glass compositions. It appears that some factor causes noritic fragments not to appear as glass in Apollo 14 soil either because of differing source or transport mechanism or both. If distinction between glass and lithic fragments can be ascribed to the source, perhaps one can argue that some compositions result from the pre-Imbrium crust and some from a low-velocity impacting body.

Groups I and III lie almost between olivine and plagioclase (see Figs. 1, 2) thereby suggesting models of crystal cumulation of either or both minerals. Dunite and anorthosite fragments may represent cumulates if original texture was destroyed by impact and thermal metamorphism. Group II contains ilmenite and augite requiring a different model.
Red spinels in breccia 14063,14 are dominantly MgAI₂O₄ with minor chromite and hercynite: Al₂O₃ 45-63, MgO 12-21, Cr₂O₃ 3-20, FeO 11-20 wt.%. These spinels are ascribed to primary crystallization from a high-Al, low-Fe magma similar to Group I composition since they plot in or near the primary spinel field in the system Di-Fo-An. High positive correlation between Al₂O₃ and MgO content of the spinel is ascribed to the parent magma composition. Resorption of spinel is prevented by rapid cooling, or crystal separation as suggested by a spinel-olivine (Fo 85) - anorthite fragment in 14321,76.

Thin section 15415,14 and fines 15415,28 show 97% plagioclase (An 96-99; K .02, Fe .09, Mg .03 wt.%) and two pyroxene compositions. (En 40 Wo 35 Fs 15; En 55 Wo 2 Fs 43) with little variation from these compositions. Precession photographs show sharp (b) and (c) diffractions consistent with primitive anorthite. The plagioclase composition lies at the extrapolated intersection of the Fe vs. Ab content of plagioclase plot for basaltic and exotic fragments from Apollo 11 and 12 (Fig. 4) (Steele and Smith, 1971). The high Ca content is consistent with the formation from an early differentiate. Pyroxene compositions indicate equilibrium to a low temperature.

Three Luna 16 fragments (about 1mm across) were studied. One(G-37) is an ilmenite olivine basalt (SiO₂ 42.8, Al₂O₃ 15.4, TiO₂ 4.3, P₂O₅ .4, Cr₂O₃ .3, MgO 6.4, FeO 18.1, MnO 0.3, CaO 11.2, Na₂O 0.7, K₂O 0.1: 47% plagioclase, 30% pyroxene, 14% olivine, 6% ilmenite, and minor ulvöspinel and zircon). The higher Al₂O₃ and lower MgO than most Apollo 11 and 12 basalts result from a higher plagioclase content and lower Mg content of pyroxene and olivine. Olivine is strongly zoned from Fo 60 to as low as Fo 10. Ilmenite is low in MgO relative to Apollo 11 and 12 basalts due to the low bulk MgO. The plagioclase was high Fe (0.6 wt.%) characteristic of mare basalt. A second fragment (G-46) is a cinder with approximately the same bulk composition as G-37 and probably results from partial fusion of a mare basalt. The third fragment (G-54) is a light-colored fragment of plagioclase whose composition is consistent with exotic material from Apollo 12 (Fig. 4).


We thank A. T. Anderson and NASA grant 14-001-171.
MINERALOGY, PETROLOGY, BULK ELECTRON-MICROPROBE
Ian M. Steele

Table 1

<table>
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*Analysis of average fines.
**Calculated average fines. See text.
We are measuring K/Ar ages with the Ar\textsuperscript{39}/Ar\textsuperscript{40} method (Merrihue and Turner 1966) on several Apollo 11, 12, and 14 rocks and also on the Apollo 15 rock 15415. The Ar\textsuperscript{39}/Ar\textsuperscript{40} temperature release curves obtained on rocks 12051 and 15415 are shown in Figure 1.

Both measurements on rock 12051 show a very good high temperature plateau, corresponding to ages of \(3.15 \times 10^6\) y and \(3.19 \times 10^6\) y. This age is in satisfactory agreement with the high temperature age of \((3.27 \pm 0.5) \times 10^6\) y given by Turner (1971 a).

Rock 15415 is a white anorthosite with very low K abundance and high Ca content. From the absolute amounts of Ar\textsuperscript{37} and Ar\textsuperscript{39} in our irradiated samples we deduced Ca and K contents of 14.1% and 110 ppm respectively. The Ca-correction (cf. Turner 1971 b) of the measured Ar\textsuperscript{39} for this rock is substantial (up to 40%) and the Ar\textsuperscript{39}/Ar\textsuperscript{37} ratio resulting from the irradiation of Ca must be exactly known. Therefore, we have irradiated two high purity Ca\textsubscript{CO\textsubscript{3}} samples simultaneously with the 15415 aliquots. In these two samples Ar\textsuperscript{39}/Ar\textsuperscript{37} ratios of 0.68 \times 10\textsuperscript{-3} and 0.69 \times 10\textsuperscript{-3} were measured and this ratio was used for the correction of the measured Ar\textsuperscript{39} in rock 15415. The same ratio was obtained from an irradiated Ca-metal sample.

Our sample of rock 15415 consisted of a large piece (0.3 g) and some finer material. The large piece was mechanically and ultrasonically cleaned from any possible original surface material or lunar dust and a 100 mg and a 160 mg sample were irradiated. This procedure was chosen to avoid any contamination with trapped gas containing material, which could interfere with the evaluation of the Ar\textsuperscript{40} results. From the finer material two 30 mg samples were also irradiated and used for establishing the optimal degassing temperature steps for the larger samples. In a third sample of the finer material all the noble gases were determined. The finer material had indeed a much higher trapped gas content than the specially cleaned large piece.

A very good high temperature Ar\textsuperscript{39}/Ar\textsuperscript{40} plateau was obtained for the 160 mg sample (cf. Figure 1). The smaller sample gave a somewhat less satisfactory plateau. From the 160 mg sample we calculate an Ar\textsuperscript{39}/Ar\textsuperscript{40}
high temperature age of \((3.92 \pm 0.08) \times 10^9\) y for rock 15415. The last four points of the 100 mg sample would correspond, on the average, to an 
0.08 \times 10^9\) y higher age.

The K/Ar age of less than 4 billion years of the anorthosite 15415 does not support the hypothesis that this rock is the remnant of an anorthositic original crust of the moon.

This research was supported by the Swiss National Science Foundation.

References

Figure 1. \(^{39}Ar/^{40}Ar\) release pattern for lunar rocks 12051 and 15415. Errors are smaller than size of symbols used.
MINERALOGY AND PETROLOGY OF LUNAR ANORTHOSITE 15415; D.B. Stewart
M. Ross, B.A. Morgan, D.E. Appleman, J.S. Huebner and R.F. Comnoue,

Lunar anorthosite 15415 consists of more than 99 volume percent plagioclase, less than one percent pyroxene, and traces of ilmenite, silica, and spinel. All minerals show evidence of solid-state recrystallization and deformations which have extensively obscured most of the original magmatic textures.

The plagioclase that makes up most of chip 15415,22 has a nearly homogeneous composition, judging by microprobe and optical measurements. Its average composition (Table 1) is An96.5 Ab2.9 Or0.3. Optical properties measured by immersion methods and sodium light are \( \alpha = 1.574(1) \), \( \beta = 1.581(1) \), \( \gamma = 1.585(1) \), optically(-), 2V large, \( r>v \) weak. Least squares refinement of the lattice parameters from powder diffraction data taken with CuK\( \alpha \) radiation (\( \lambda = 1.54056 \) Å) and BaF\( \text{2} \) internal standard (a=6.1971 Å) gave \( a=8.179(1) \), \( b=12.879(1) \), \( c=14.179(2) \), \( \alpha=93.18(1) \), \( \beta=115.91(1) \), \( \gamma=91.19(1) \), \( V=1339.7(2) \) Å\(^3\) for a pattern containing 44 indexed a-reflections. Single crystal X-ray and electron diffraction photographs indicate that the space group is \( \text{P1} \). Sharp reflections of all types were observed. No irregularities in peak shapes could be detected by traverses across more than 1000 peaks of all types with an automated diffractometer. Long exposures with an oscillating precession X-ray camera and electron diffraction photographs taken at 200 kV for several minutes revealed no elongation of reflections or streaking between reflections. All these properties indicate that the anorthite has a high degree of short and long range order.

Pyroxenes of many types were analyzed by X-ray single crystal and electron microprobe methods (Tables 1 and 2). A microprobe mode indicated that 87% of 57 pyroxene grains (84% by volume) had high Wo contents (Wo>20%). Calcic augite is the predominant pyroxene and contains 2-5% pigeonite lamellae on (001) and 5% hypersthene on (100), as well as reddish brown plates of exsolved ilmenite (a=5.09 Å, \( c=14.17 \)) and Cr and Al-bearing spinel (a=5.39 Å). Most low calcium pyroxene occurs as discrete grains of clear, colorless hypersthene (Table 2) as well as lamellae in the other pyroxenes.

Single grains of subcalcic augite with only unmixed pigeonite, pigeonite with unmixed calcic augite, and composite grains of subcalcic augite and pigeonite were observed (Table 2). Rare grains of ferropigeonite (45<Fe<50) were found only in chip 15415,22, and have not been observed in polished thin sections (15415,17 and 15). We interpret the first crystallized pyroxene to have been augite or subcalcic augite, followed shortly after by pigeonite. Both subcalcic augite and pigeonite become more Fe-rich by fractional crystallization of trapped magma. Stepwise homogenization heating experiments of Apollo 12 augite identical in composition to subcalcic augite A5 suggests a temperature of crystallization of 1150°C for this pyroxene. A ferropigeonite crystal homogenized at 1030°C in agreement with the fractionation trend proposed.

Subsolidus recrystallization at temperatures between 1150 and 1000°C is interpreted by the same criteria to have caused (1) extensive intra-granular unmixing of magmatic subcalcic augite and pigeonite to form calcic augite and low calcium pigeonite within the same grain and (2) intergranular recrystallization...
zation of the magmatic clinopyroxenes to form discrete grains of calcic augite and hypersthene as well as oxide phases (and silica?). Possibly this stage of recrystallization resulted from thermal buffering by physical proximity with the crystallizing magma but chemical isolation from it. The distribution coefficient \(D\) of Fe and Mg between calcic augite and hypersthene (~1.7) is consistent with that for pyroxene from terrestrial granulite facies rocks, and \(D\) between the subcalcic augite and pigeonite is consistent with \(D\) (~1.3) for pyroxenes from mare basalts. The final recorded thermal history is represented by unmixing of small amounts of additional pyroxenes, notably low calcium pigeonite and hypersthene from calcic augite and augite from hypersthene.

Phase diagrams for anhydrous silicate systems so far as they are pertinent to the proportions of anorthite and pyroxene observed in rock 15415 indicate the liquidus temperature at one bar will likely be in excess of 1500°C. This precludes an origin by crystallization of a magma of its own composition. This rock formed by fractional crystallization and flotation in a gabbroic magma which had already precipitated magnesian phases.

Table 1. Electron microprobe analyses of 15415 anorthite and pyroxenes.

<table>
<thead>
<tr>
<th></th>
<th>Average</th>
<th>Average</th>
<th>Average</th>
<th>Average</th>
<th>Average</th>
<th>Subcalcic</th>
<th>Ferro-</th>
<th>Pigeonite</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>anorthite</td>
<td>calcic</td>
<td>calcic</td>
<td>hyper-</td>
<td>augite</td>
<td>pigeonite</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>15415.22</td>
<td>15415.22</td>
<td>15415.17</td>
<td>15415.17</td>
<td>15415.22</td>
<td>15415.22</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
<td>43.36</td>
<td>51.54</td>
<td>52.55</td>
<td>52.63</td>
<td>49.75</td>
<td>49.31</td>
<td>50.37</td>
<td></td>
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<tr>
<td>TiO₂</td>
<td>0.04</td>
<td>1.02</td>
<td>0.49</td>
<td>0.31</td>
<td>0.94</td>
<td>1.08</td>
<td>0.76</td>
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<tr>
<td>Al₂O₃</td>
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<td>0.91</td>
<td>1.05</td>
<td>0.73</td>
<td>3.12</td>
<td>1.30</td>
<td>1.90</td>
<td></td>
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<tr>
<td>Cr₂O₃</td>
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<td>0.17</td>
<td></td>
<td></td>
<td>0.98</td>
<td>0.23</td>
<td>0.59</td>
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<tr>
<td>FeO</td>
<td>0.08</td>
<td>10.73</td>
<td>9.86</td>
<td>25.54</td>
<td>13.09</td>
<td>27.34</td>
<td>20.02</td>
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<td>MnO</td>
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<td></td>
<td>0.24</td>
<td>0.41</td>
<td>0.33</td>
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<tr>
<td>MgO</td>
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<td>14.11</td>
<td>13.73</td>
<td>20.07</td>
<td>15.11</td>
<td>13.06</td>
<td>17.05</td>
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<tr>
<td>CaO</td>
<td>19.54</td>
<td>21.67</td>
<td>21.76</td>
<td>1.73</td>
<td>15.75</td>
<td>7.26</td>
<td>7.63</td>
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<td>Na₂O</td>
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<td></td>
<td>0.05</td>
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<td>K₂O</td>
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<td></td>
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<td>100.55</td>
<td>99.44</td>
<td>101.01</td>
<td>99.03</td>
<td>100.01</td>
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<td>1.887</td>
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<td>0.014</td>
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<td>0.032</td>
<td>0.022</td>
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<tr>
<td>Al</td>
<td>1.979</td>
<td>0.040</td>
<td>0.047</td>
<td>0.032</td>
<td>0.139</td>
<td>0.060</td>
<td>0.086</td>
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<tr>
<td>Cr</td>
<td>nil</td>
<td>0.005</td>
<td></td>
<td></td>
<td>0.029</td>
<td>0.007</td>
<td>0.018</td>
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<tr>
<td>(\sum_{Si+Al}Cr)</td>
<td>4.000</td>
<td>2.007</td>
<td>2.035</td>
<td>2.013</td>
<td>2.032</td>
<td>2.029</td>
<td>2.056</td>
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<tr>
<td>Fe</td>
<td>0.003</td>
<td>0.337</td>
<td>0.310</td>
<td>0.800</td>
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<td>0.895</td>
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<tr>
<td>Mn</td>
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<td>0.011</td>
<td></td>
<td></td>
<td>0.003</td>
<td>0.001</td>
<td>0.011</td>
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<tr>
<td>Mg</td>
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<td>0.791</td>
<td>0.769</td>
<td>1.121</td>
<td>0.854</td>
<td>0.762</td>
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<tr>
<td>Ca</td>
<td>0.965</td>
<td>0.870</td>
<td>0.876</td>
<td>0.069</td>
<td>0.640</td>
<td>0.304</td>
<td>0.313</td>
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<tr>
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<td>0.001</td>
<td></td>
<td></td>
<td>0.003</td>
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<td></td>
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<td></td>
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<td></td>
<td></td>
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<tr>
<td>(\sum_{Mg+Ca+K})</td>
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<td>2.010</td>
<td>1.955</td>
<td>1.990</td>
<td>1.920</td>
<td>1.976</td>
<td>1.940</td>
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<tr>
<td>Wo</td>
<td>43.6</td>
<td>44.8</td>
<td>3.5</td>
<td>33.6</td>
<td>15.5</td>
<td>16.2</td>
<td></td>
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<tr>
<td>En</td>
<td>39.6</td>
<td>39.4</td>
<td>56.3</td>
<td>44.7</td>
<td>38.9</td>
<td>50.5</td>
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<td>16.8</td>
<td>15.8</td>
<td>40.2</td>
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Table 2. Representative unit-cell parameters and proportions of host and exsolved pyroxenes in 15415,22

Calcic augite, grain A4, probable bulk composition in Table 1

<table>
<thead>
<tr>
<th></th>
<th>Host augite (92%)</th>
<th>Pigeonite (3%)</th>
<th>Hypersthene (5%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>9.764</td>
<td>9.702</td>
<td>18.29</td>
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<tr>
<td>b</td>
<td>8.949</td>
<td>(8.949)</td>
<td>(8.949)</td>
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<tr>
<td>c</td>
<td>5.256</td>
<td>5.208</td>
<td>5.249</td>
</tr>
<tr>
<td>α</td>
<td>106°32'</td>
<td>108°15'</td>
<td>90°</td>
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</tbody>
</table>

Hypersthene, grain H1, probable bulk composition like that in Table 1

Augite (1%)

<table>
<thead>
<tr>
<th></th>
<th>Host hypersthene (99%)</th>
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<tbody>
<tr>
<td>a</td>
<td>18.290Å</td>
</tr>
<tr>
<td>b</td>
<td>8.906</td>
</tr>
<tr>
<td>c</td>
<td>5.217</td>
</tr>
<tr>
<td>α</td>
<td>90°</td>
</tr>
</tbody>
</table>

Subcalcic augite, grain A5, bulk composition in Table 1

<table>
<thead>
<tr>
<th></th>
<th>Host augite (90%)</th>
<th>Pigeonite (10%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>9.732</td>
<td>9.732</td>
</tr>
<tr>
<td>b</td>
<td>8.925</td>
<td>8.925</td>
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<tr>
<td>c</td>
<td>5.260</td>
<td>5.227</td>
</tr>
<tr>
<td>α</td>
<td>106°32'</td>
<td>108°39'</td>
</tr>
</tbody>
</table>

Pigeonite, grain P3, bulk composition in Table 1

<table>
<thead>
<tr>
<th></th>
<th>(001) augite (25%)</th>
<th>Host pigeonite (75%)</th>
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</thead>
<tbody>
<tr>
<td>a</td>
<td>9.696</td>
<td>9.714</td>
</tr>
<tr>
<td>b</td>
<td>8.914</td>
<td>8.914</td>
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<tr>
<td>c</td>
<td>5.247</td>
<td>5.233</td>
</tr>
<tr>
<td>α</td>
<td>105°58'</td>
<td>108°50'</td>
</tr>
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</table>

Ferropigeonite, grain P1, bulk composition in Table 1.

<table>
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<th>(001) augite (40%)</th>
<th>Host pigeonite (60%)</th>
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<tbody>
<tr>
<td>a</td>
<td>9.733</td>
<td>9.745</td>
</tr>
<tr>
<td>b</td>
<td>(8.961)</td>
<td>8.961</td>
</tr>
<tr>
<td>c</td>
<td>5.264</td>
<td>5.253</td>
</tr>
<tr>
<td>α</td>
<td>106°34'</td>
<td>108°49'</td>
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Composite grain AP6 of subcalcic augite 80% and pigeonite 20%

<table>
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<th>Host subcalcic augite (75%)</th>
<th>Pigeonite (25%)</th>
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</thead>
<tbody>
<tr>
<td>a</td>
<td>9.752</td>
<td>9.752</td>
</tr>
<tr>
<td>b</td>
<td>(8.961)</td>
<td>(8.961)</td>
</tr>
<tr>
<td>c</td>
<td>5.274</td>
<td>5.274</td>
</tr>
<tr>
<td>α</td>
<td>106°12'</td>
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</tr>
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</table>

<table>
<thead>
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<th></th>
<th>(001) augite (25%)</th>
<th>Host pigeonite (75%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>9.733</td>
<td>9.745</td>
</tr>
<tr>
<td>b</td>
<td>(8.961)</td>
<td>8.961</td>
</tr>
<tr>
<td>c</td>
<td>5.273</td>
<td>5.253</td>
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<tr>
<td>α</td>
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The lunar fines on the surface of the moon have unique adsorptive and emanation properties that are important to an understanding of the composition of the lunar atmosphere. These surface properties are essentially preserved inside the Sample Return Container (SRC) up to the time of opening at the Lunar Receiving Laboratory. Therefore it is of interest to measure the gases present in the box prior to opening to derive the emanation properties of the lunar fine material.

The gases present in the sample return container (SRC) were collected by adsorbing them on charcoal at LN$_2$ and LHe temperatures. Samples were collected from a single SRC from each of the Apollo 12, 14 and 15 missions. A separation of the rare gases and hydrogen was performed and the separated fractions were counted in low-level proportional counters to obtain the amounts of 35-day 37Ar, 269-year 39Ar, 3.8-day 222Ra and 12.6-year tritium radioactivities. The results are given in the following table.

<table>
<thead>
<tr>
<th>Mission</th>
<th>Pressure in SRC (torr)</th>
<th>Wt. Fines in SRC (kg)</th>
<th>37Ar (dpm)</th>
<th>39Ar (dpm) $\times 10^3$</th>
<th>Tritium (dpm)</th>
<th>222Rn (saturation dpm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>12</td>
<td>0.030</td>
<td>2.7</td>
<td>0.040 ± 0.002</td>
<td>2.3 ± 1.0</td>
<td>---</td>
<td>6.4 ± 0.4</td>
</tr>
<tr>
<td>14</td>
<td>0.070</td>
<td>0.60</td>
<td>0.040 ± 0.002</td>
<td>0.8 ± 0.5</td>
<td>0.28 ± 0.02</td>
<td>2.42 ± 0.03</td>
</tr>
<tr>
<td>15</td>
<td>0.032</td>
<td>2.30</td>
<td>0.106 ± 0.003</td>
<td>3.0 ± 1.2</td>
<td>0.30 ± 0.01</td>
<td>0.86 ± 0.05</td>
</tr>
</tbody>
</table>

The total quantity of gas recovered in the Apollo 15 mission was 2 cm$^3$. A mass-spectrometer analysis was performed showing there was very little air. The mass pattern was dominated by H$_2$O, CO$_2$, SO$_2$ and hexane.

The radioactive 37Ar, 39Ar, 222Rn and tritium present in the gas can be attributed to the loss of these products by the lunar fine material present in the box during transit to the earth. The emanation via recoil processes of 222Rn from lunar fines derived from the above data is in the range (0.5 to 4)$\times 10^{-5}$ 222Rn atoms g$^{-1}$ sec$^{-1}$. This rate corresponds to a radon loss of 0.05 to 0.2 percent based upon the U composition of the fine material in each box.

The cosmic ray produced 37Ar, 39Ar, and tritium activities observed in the gas in the SRC were released from the fine material in the box by diffusion processes. From measurements of the amounts of these radioactivities in lunar fines we find that the fraction diffusing out is in the range 10$^{-5}$ to
Radioactive Gases in the SRC
R. W. Stoenner

10\(^{-4}\) per day. These fractions will be compared to those observed in bombardments of stimulated lunar fines. Of particular significance is the low amount of tritium observed in these measurements. The amount of tritium observed is approximately that expected from diffusion losses, and indicates little or no solar derived tritium was present during the Apollo 14 and 15 missions.

\(^{37}\)Ar, \(^{39}\)Ar and tritium radioactivities were measured in two Apollo 14 surface fine samples by a vacuum melting and extraction procedure. The results are as follows: 14259,84 \(\pm\) 3.4 dpm \(^{37}\)Ar/kg and 9.1 \(\pm\) 0.5 dpm \(^{39}\)Ar/kg; 14163,116 \(\pm\) 46.4 \(\pm\) 3.4 dpm \(^{37}\)Ar/kg, 10.2 \(\pm\) 0.6 dpm \(^{39}\)Ar/kg, and 203 \(\pm\) 4 dpm tritium/kg. The \(^{37}\)Ar observed in these samples is higher than observed in previous missions, and will be interpreted in terms of bombardment cross sections and K, Ca, Ti and Fe contents of the samples.

Two pairs of Apollo 15 soils were measured to obtain the production rate of \(^{37}\)Ar, \(^{39}\)Ar, and tritium as a function of soil depth. One pair of samples was the surface (15041,16) and bottom (15031,16) of a trench approximately 40 cm deep. A second pair of samples were a surface sample (15221,17) and a sample taken below a 1 meter diameter boulder (15231,18). The pair of samples from the trench were heated in stages 300°C, 600°C, 900°C and melted to observe the diffusion of these radioactivities from the soil samples. The samples from the boulder location were held at 150°C for sufficient time to observe the emanation of \(^{222}\)Rn. Following this 150°C heating, these samples were melted to release the argon, radon, and tritium radioactivities. The results will be interpreted in terms of a thick target of simulated lunar material bombarded with 600 MeV protons.

The 35-day $^{37}$Ar, 269-year $^{39}$Ar and 12.6-year tritium radioactivities were measured in two Apollo 14 surface fine samples by a vacuum melting and extraction procedure. The results are as follows: $14259.84 - 78.2 \pm 3.4$ dpm $^{37}$Ar/kg and $9.11 \pm 0.54$ dpm $^{39}$Ar/kg; $14163.116 - 46.4 \pm 3.4$ dpm $^{37}$Ar/kg, $10.15 \pm 0.58$ dpm $^{39}$Ar/kg, and $203 \pm 4$ dpm tritium/kg. The $^{37}$Ar observed in these samples is higher than observed in previous missions, and will be interpreted in terms of bombardment cross sections and K, Ca, Ti and Fe contents of the samples.

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Geology of the Apollo 14 Landing Site


Apollo 14 astronauts Alan B. Shepard and Edgar D. Mitchell landed on the moon on February 2, 1971, in an area referred to as the Fra Mauro region approximately 550 km south of the mountainous southern rim of the Imbrium Basin. The area is underlain by the Fra Mauro Formation represented by ridges and valleys roughly radial to the Imbrium Basin. The Fra Mauro is interpreted as material excavated by the large impact event that produced the Imbrium Basin (refs. 1, 2).

The LM landed in a broad valley 1,100 m west of Cone crater, which is located on a Fra Mauro ridge. Cone crater, 340 m in diameter, is relatively young and sharp rimmed. Blocks of ejected material as large as 15 m across occur near the rim. A ray of blocky ejecta extends westward beyond the landing site (fig. 1; ref. 3). Cone crater ejecta are presumed to have been excavated from depths as great as 60 to 80 m, well below the regolith, which is estimated to be 10 to 20 m thick in the Fra Mauro region. In the valley, the surface at the landing site is relatively smooth and fine-grained except where larger craters such as Doublet and Triplet have also penetrated through the regolith into what is considered to be Fra Mauro material.

Figure 1 shows traverse routes from the LM for the two EVAs: the first, westward to deploy the ALSEP and to collect samples; and the second eastward to within 20 m of the rim crest of Cone crater. The distribution of rock fragments on the surface is shown by symbols and patterned areas on the map and by circular insets that represent 10 m radius areas around photographic panoramic stations in which all rock fragments larger than 10 cm were mapped from the panoramic photographs.

A primary geologic objective of Apollo 14 was to sample and photograph the ejecta from Cone crater, because it most probably represents the "bedrock" Fra Mauro Formation beneath the local regolith. A secondary sampling objective was to collect soil and rock fragments from the regolith in the valley west of Cone ridge. Apollo 14 returned approximately 43 kg of lunar samples, fulfilling both sampling objectives. The samples included 35 rocks that weigh more than 50 g each, and about 30 smaller rocks between 10 and 50 g, as well as rock chips and fines collected with a scoop and with drive tubes. As expected, samples from the valley regolith included a finely fragmented, well mixed assemblage along with "soil breccias" that were formed by shock induration of regolith material by relatively small meteorite impact (refs. 4, 5).
Rock fragments collected in the blocky ray ejected westward from Cone crater between the so-called North Boulder Field (station H) and Cone crater rim (station Cl) were expected to provide a radial sample that would indicate a rough correlation between rock types and depth of excavation within Cone crater. The size of boulders increases toward the rim of Cone crater. Light colored rocks unique to the crater rim were recognized at station Cl. Like all other large boulders photographed on the EVA 2 traverse they appeared to be layered and/or fractured clastic breccias that include fragments of older rocks that are both lighter and darker in color than their surrounding groundmass (ref. 6). The white rocks include primarily dark clasts. Because the white rocks are rare in occurrence, and were seen only on the crater rim, it is suggested that they are among the deepest of those excavated from Cone crater. In the discontinuous blocky ray west of Cone crater, it is reasonable to expect some mixing of ejecta from different depths.

References and notes


4. Wilshire, H. G. and Jackson, E. D., Petrology of the Fra Mauro Formation at the Apollo 14 landing site, this Conference.

5. Jackson, E. D. and Wilshire, H. G., Classification of the samples returned from the Apollo 14 landing site, this Conference.


7. Work done under NASA contract number T-65253-G.
Figure 1. Traverse map of the Apollo 14 site. Circles are 10-m radius areas at panorama stations showing distributions of rocks larger than 10 cm (ref. 3).
GEOLOGY OF THE APOLLO 15 LANDING SITE (Ref. 1)

Three major geologic features were visited during the Apollo 15 mission. These were defined on pre-mission maps compiled from Lunar Orbiter photographs and include: 1) the mare surface of Palus Putredinis in an area that is crossed by a faint ray from either of the craters Aristillus or Autolycus, 2) the Apennine front, and 3) Hadley Rille (fig. 1, ref. 2). It was expected that the mare material would be primarily basalt with some contribution of material from the ray that crosses the mare surface, and that the front might be blanketed by ejecta, probably breccias, from the Imbrium basin. It was also postulated that some material of the front may have been derived from bedrock beneath the blanketing material which was excavated by cratering processes.

The mare surface in the landing site is situated such that a large part of the surface probably has been contaminated from other sources. In addition to the contribution of foreign materials from the ray, the Apennine mountain massifs, being topographically higher, should also contribute material to the mare surface by crater ejecta transported off steep slopes. Therefore, it is logical to assume that the mare surface in the vicinity of the landing site has a significant percentage of exotic material from diverse sources; it is also reasonable to expect these materials to be breccias derived from Autolycus or Aristillus, and from the Apennine front.

A simple four-fold categorization of the returned rock samples that weigh more than 20 g shows a logical relationship of the sample types to the map units. The categories are: 1) basalts that contain brown pyroxene (Bn Px), 2) basalts that contain greenish gray pyroxene (Gy Px), 3) breccias containing Bn Px basalt clasts, and 4) breccias that do not contain Bn Px basalt clasts. Table 1 shows the distribution of these rock types.

Table 1. Distribution of rock samples weighing more than 20 g

<table>
<thead>
<tr>
<th></th>
<th>Front</th>
<th>Mare</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bn Px-type basalts</td>
<td>0</td>
<td>21</td>
</tr>
<tr>
<td>Gy Px-type basalts</td>
<td>1(?)</td>
<td>0</td>
</tr>
<tr>
<td>Breccias containing Bn Px-type basalt clasts</td>
<td>7</td>
<td>7</td>
</tr>
<tr>
<td>Breccias that do not contain Bn Px-type basalt clasts</td>
<td>17</td>
<td>4 (near LM)</td>
</tr>
</tbody>
</table>
Breccias with Bn Px-type basalt clasts that were collected from the front were all on the down-range side of the South Cluster of secondary craters, suggesting that the secondaries are a possible source of these breccias. The breccias without Bn Px-type basalt clasts that were collected around the LM are likely to be part of the ray material; alternatively, some may be material ejected from craters on the Apennine front.

The Bn Px basalts were collected primarily from the rille rim where the regolith is thin, and from the rims of Elbow and Dune craters which are large enough to have penetrated through the regolith into the underlying mare basalts. Four of the basalt samples from the rille were collected from two large rocks that almost certainly represent local bedrock. The basalt samples from Elbow and Dune crater have a high likelihood of being derived from local bedrock. The distribution of basalts that were collected form the rille edge and from Elbow and Dune craters suggests that pyroxene-rich basalts are the dominant mare rock types to a depth of 80 or 90 m. At station 9a, however, it appears that an olivine-phyric basalt unit may overlie, at least locally, the pyroxene-rich basalts.

The zone of mixing between the mare material and the front material is shown by an albedo (15-19%) that is intermediate between that of the typical mare surface (9-15%) and that of the front (19-21%). Part of the area in the mare has a slightly higher albedo (13-15 %) than that of the surrounding mare (9-13%), which probably reflects lighter colored ray material mixed with regolith that was derived from the underlying mare basalts.

Differences between at least some of the breccias collected from the Hadley Apennine site from those collected from the Fra Mauro site (ref. 3) suggest differences in the histories of the breccias of the two sites. This may indicate that the breccias from the Hadley Apennine site were derived from a different depth within the Imbrium basin than those of the Fra Mauro site, or that some of the breccias of the Apennine front may be a part of the massif material that underlies the Imbrium ejecta.

References and notes
1. Much of the material in this report is derived from the report submitted by the Apollo Lunar Geology Investigation Team to NASA for the Apollo 15 Preliminary Science Report, and from information about the samples that was generously supplied by the Preliminary Examination Team. The authors assume full responsibility, however, for the descriptions and interpretations presented here.
3. Wilshire, H. G. and Jackson, E. D., 1971, Petrology of the Fra Mauro Formation at the Apollo 14 landing site, this Conference.
4. Work performed under NASA contract T-65253-G.
Figure 1. Traverse map of the Apollo 15 landing site.
Pyroxenes in Apollo 11 and 12 mare basalts are dominantly augite and pigeonite. Rare orthopyroxene, an indicator of non-mare rocks found in Apollo 12 coarse fines has been studied crystallographically (Meyer et al., 1971). In the Apollo 14 soil, orthopyroxene represents as much as 41% of all analyzed pyroxenes (Reid et al., 1971). The composition range in Mg/(Fe+Mg) is from .85 to .61. It is also found to be a common pyroxene in crystalline rock 14310 (Ridley et al., 1971).

The exsolution and overgrowth patterns of those orthopyroxenes, together with those of the coexisting pigeonites and augites were studied by single crystal x-ray diffraction methods to provide information on crystallization and cooling conditions. 12 pyroxene grains were separated from rock chip 14310,90. The crystals were mounted along c*, and precession photographs of h01 and Ok1 nets were taken using MoKα radiation with Zr filter. When the patterns were found to be complicated due to twinning or stacking faults, over-exposed h01 photographs with CuKα radiation using Ni filter were taken on
the same crystals. Subsequently, each single crystal was studied by electron microprobe analysis.

Based on their chemical compositions, exsolution patterns, twinning and colors, these pyroxene grains can be classified into three major groups:

(1) Orthopyroxene (4 crystals), pale yellow or almost colorless but not transparent. The compositions are bronzitic (Wo4En80Fs16). Neither exsolution of augite nor diffuse streaks along α*, indicative of the disordered orthopyroxene commonly found in meteorites (Pollack, 1968), have been detected within usual exposure time. A longer exposed photograph did show exsolution.

(2) Twinned magnesian pigeonite (5 crystals), has almost the same appearance as the orthopyroxenes, and is invariably twinned on (100) or [001]180°. The intensities of twinned pair reflections vary from grain to grain indicating that twinning is not as fine as is found in twinned clinoenstatite transformed from protoenstatite. The class b reflections (h+k odd) of pigeonites are sharp. Exsolution of augite both on (100) and (001) is observed on longer exposed photographs. The intensities of the exsolved augites are proportional to the intensities of the host twin individuals (pigeonite). The β angle of pigeonite (Table 1) is the same as for a fully exsolved one, and the β angle of augites (106°023' for one exsolved on (100) and 106°12' for (001)) are not as large as rapidly cooled pyroxenes such as those of rock 12052 (Takeda, 1971). In one of the twinned pigeonites (Wo8En71Fs20) an orthopyroxene core (Wo4En80-Fs16) is present with common (100). Another pigeonite without orthopyroxene is more iron-rich (Wo8En58Fs34).
(3) Colored intermediate pigeonite (3 crystals), the grains are pink to brownish, and more Fe-rich than the twinned pigeonites. All the grains we studied are untwinned. The diffraction patterns show predominant exsolution of augite on (001). The intensity ratio, pig./aug. is roughly 3/1. The lamellae or patches of augite (Wo31En32Fs17) are barely detectable by electron microprobe in the matrices of pigeonites (Wo7En35Fs58). The separation of the a* axes of pigeonite and augite are close to the maximum values common for volcanic pigeonites. Both facts indicate relatively slow cooling. The class b reflections of pigeonite are sharp. Very weak diffuse streaks along a* in some part of h02 rows were observed on longer exposed photographs. One augite grain (Wo2qEn44Fs27) has exsolved about equal amounts of pigeonite.

The possibility that the twinned Mg-rich pigeonite is twinned clinoenstatite inverted from protoenstatite can be ruled out by the fact that the composition of pigeonite is more Fe-rich and Ca-rich than the bronzite and the coarser twin individuals are sometimes observed optically. The diffuse streak along a* found in clinoenstatite was not observed in the Mg-rich pigeonite. The compositional relation of core bronzite to twinned pigeonite and lack of exsolution of augite also excludes the possibility of the bronzite being an inverted pigeonite.

According to composition-temperature relations of pyroxenes (for example, Kuno, 1966), orthopyroxene separates from the magma in the earlier stages, and when it reaches the ortho- to high clino-pyroxene inversion boundaries (ca 1200°C) its place is taken by pigeonite which crystallizes continuously. The pattern for a twinned pigeonite with orthopyroxene having common (100) is similar to that obtained on an orthopyroxene heated up to the stability field of pigeonite by Ross et al. (priv. comm.). The monoclinic pigeonite growing on or in the orthopyroxene may have one of twin orientations, thus producing the twinning.

The diffuse streaks observed in the intermediate pigeonites, have been reported neither in terrestrial nor in pigeonites from mare basalts. Similar diffuse streaks have been found in pigeonites coexisting with secondary orthopyroxene from some eucrites or eucritic portions of some mesosiderites. The intensities of the streaks are stronger where the strong reflections of orthopyroxene are present in reciprocal space, e.g. between 102 and 202; and 202 and 302. Those streaks thus indicate the presence of stacking faults.

Cation distribution in M1 and M2 sites and the crystal structures of orthopyroxenes, one separated from a glass matrix breccia fragment (12033,97) and the other from an Apollo 14 crystalline rock (14310,90) and of magnesian pigeonite have been refined using intensity data measured by a Picker automated single crystal counter diffractometer. The crystal data are given in Table 1. Due to small size of the orthopyroxene crystals only 339(A-12) and 866(A-14) reflections were observable, for which preliminary least-squares refinement resulted in R-values of 0.060 and 0.056 respectively.

The site populations and mean M-O bond distances for the A-12 orthopyroxene from KREEP material are: M1(0.91Mg+0.09Fe)2.097Å, M2(0.06Ca+0.50Mg+0.44Fe)2.204Å; and for A-14 orthopyroxenes: M1(0.88Mg+0.07Fe+0.05 others)2.083Å, M2(0.09Ca+
Crystallography & Mineralogy of Pyroxenes...
Hiroshi Takeda

The temperature estimated from the distribution isotherms (Virgo and Hafner, 1969; Saxena and Ghose, 1971) is about 600°-700°. The temperatures were similar to those found in orthopyroxenes from volcanic rocks. The aluminum content of the orthopyroxene of rock 14310 used for structural work is not as high as those of 14310 orthopyroxenes. The structural features found in an aluminan bronzite formed at high pressure (Takeda, 1971) were not noticeable in the 14310 orthopyroxene.

The crystallographic features of the pyroxenes from Fra Mauro basalts are distinct from those of Apollo 11 and 12 mare pyroxenes and thus should be useful in characterizing such rocks and their crystallization trends. The non-mare pyroxenes reveal cooling rates fairly slower than many of the mare pyroxenes. Some similarities between these pyroxenes and some achondritic pyroxenes as discussed above will provide useful information in deducing the thermal history of parent bodies of these meteorite groups.

References
TABLE 1. Cell dimensions and chemical compositions of orthopyroxene (Opx), pigeonite (Pig) and augite (Aug).

<table>
<thead>
<tr>
<th>Sample</th>
<th>a, Å</th>
<th>b, Å</th>
<th>c, Å</th>
<th>β, deg.</th>
<th>Wo-En-Fs</th>
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<tbody>
<tr>
<td>12033,97 Opx</td>
<td>18.302(6)</td>
<td>8.884(2)</td>
<td>5.212(7)</td>
<td>---</td>
<td>3-68-29 *</td>
</tr>
<tr>
<td>14310,90 Opx</td>
<td>18.301(3)</td>
<td>8.869(2)</td>
<td>5.215(1)</td>
<td>---</td>
<td>5-72-23 *</td>
</tr>
<tr>
<td>14310,90 Pig</td>
<td>9.673(3)</td>
<td>8.896(2)</td>
<td>5.228(3)</td>
<td>108.65(4)</td>
<td>8-62-30 *</td>
</tr>
<tr>
<td>14310,90 Pig</td>
<td>9.715(1)</td>
<td>8.963(1)</td>
<td>5.239(1)</td>
<td>108.64(2)</td>
<td>7-41-52</td>
</tr>
<tr>
<td>14310,90 Aug</td>
<td>9.713(2)</td>
<td>8.964(3)</td>
<td>5.266(2)</td>
<td>105.93(2)</td>
<td>43-27-30</td>
</tr>
</tbody>
</table>

* microprobe analysis by A. M. Reid.
U-Th-Pb AND Rb-Sr MEASUREMENTS ON SOME APOLLO 14 LUNAR SAMPLES


U-Th-Pb. The U-Th-Pb and Rb-Sr systems (Table 1) have been determined for selected samples collected by the Apollo 14 mission to the Fra Mauro locality of the moon: basalts 14053 and 14310; soils 14003 (contingency sample), 14163 (bulk fines), and 14259 (comprehensive fines). U-Th-Pb data are discussed in terms of their $\Delta^{(207}\text{Pb})/\Delta^{(206}\text{Pb})$ model ages, of concordia diagrams, and of some of the trace element characteristics.

Our values of $^{206}\text{Pb}/^{204}\text{Pb}$ are about equal to or greater than those reported by Wasserburg et al. (9), but the values of the all-important ratio for age purposes, $^{207}\text{Pb}/^{206}\text{Pb}$, are in excellent agreement. All the samples analyzed have $\Delta^{(207}\text{Pb})/\Delta^{(206}\text{Pb})$ model ages greater than 4750 m.y. except for the basalt sample 14310 whose model age is 4270 m.y. The slopes were determined using "primordial lead" as the initial lead isotopic composition. These old model ages greater than 4750 m.y. are older than any previously found in lunar whole-rock or whole-soil samples. The configuration of the data on the U-Th-Pb system on samples from previous Apollo missions has been such that depletion of lead relative to uranium must have occurred 3 to 4 billion years ago. It is therefore not too surprising that localities would eventually be found on the moon where lead would be enriched relative to uranium in order to maintain a material balance. The Apollo 14 locality at Fra Mauro predominantly appears to be just such a locality.

The figure shows the lead data plotted on a concordia diagram along with a discordia line (solid) connecting 3900 m.y. [the $^{40}\text{Ar}/^{39}\text{Ar}$ whole-rock and Rb-Sr mineral isochron age of Apollo 14 rocks (3, 4, 5, 8)] and 4660 m.y., the working model of the age of the moon (4, 6, 7). Our best analyses of 14163, 14003, and 14310 are within analytical uncertainties of that discordia line and therefore are considered to be in agreement with the other dating methods. Our best analyses of 14053 and 14259 lie beneath the discordia line and are probably reflecting the effects of "third events" such as have been previously noted for soils from the Apollo 12 locality (2, 7). The "third event" involved in 14053 may conceivably have been sample handling. Our analyzed sample may not be representative of the whole rock, because the basalt is medium grained, the sample is small (0.1 grams was used in the concentration work), and because, as Burnett et al. (1) have shown on Apollo 12 rocks, the uranium is primarily in trace phases. For 14259, there is also the suggestion of an alignment of the data of 14259, 14003, and 14163 on a discordia line (dashed) connecting an "age" of about 2000 m.y. with 4800 m.y. Our interpretation of this alignment would be: (1) enrichment of lead relative to uranium at about 3900 m.y. for materials from a source about 4660 m.y. old, (2) fractionation of the U/Pb in that source about 2000 m.y. ago wherein 14259 preferentially lost lead relative to uranium (14163 may have gained some lead relative to uranium).
The uranium and thorium contents of the basalts from the Apollo 14 mission are intermediate between those of the low-potassium, Group II, basalts of Apollo 11 and the breccia sample 12034 of Apollo 12. Sample 14053 is rather reminiscent of the high-potassium, Group I, basalts of Apollo 11, and 14310 is rather reminiscent of Apollo 12 and 14 breccias and fines as if 14310 may be fused soil or breccia. There appears to be a tendency arising for the contingency sample having the lowest value of $^{206}\text{Pb}/^ {204}\text{Pb}$ for any soil sample from a single mission. The contingency sample 12070, for example, has a raw value of $^{206}\text{Pb}/^ {204}\text{Pb}$ of 401, whereas the same ratio for 12033 is 986. Correspondingly, the contingency sample 14003 has a raw value of $^{206}\text{Pb}/^ {204}\text{Pb}$ of 155, whereas the same ratio for the other two soil samples we have analyzed is greater than 300. This characteristic suggests lead contamination of the contingency sample on the moon (by rocket exhaust?) and values of $^{204}\text{Pb}$ contents in the contingency sample are suspect for use in studies of "volatile elements" - i.e., elements that are volatile above 1000°C.

Rb-Sr-K. From the data in Table 1, Rb-Sr model ages for whole-soils 14259, 14163, and 14003 are calculated to be 4,680, 4,650, and 4,600 m.y., respectively (assuming an initial $^{87}\text{Sr}/^ {86}\text{Sr}$ of 0.69898). Similar analyses of 14259 and 14163 have been made (4) and the agreement on 14259 is good (4,680 versus 4,720 m.y.). Our model age for 14163 is somewhat older (4,650 versus 4,540 m.y.). We believe that these model ages probably reflect the age of the moon and that subsequent soil-forming events have little affected the values of Rb/Sr.

REFERENCES

U-Th-Pb AND Rb-Sr MEASUREMENTS ON SOME APOLLO 14 LUNAR SAMPLES

M. Tatsumoto

P. 743

Table 1: Data on the U-Th-Pb and Rb-Sr systems for selected Apollo 14 lunar samples

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Concentrations (ppm)</th>
<th>Raw ratios (atomic)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Pb</td>
<td>U</td>
</tr>
<tr>
<td>14053</td>
<td>1.64</td>
<td>0.59</td>
</tr>
<tr>
<td>14310</td>
<td>6.18</td>
<td>3.10</td>
</tr>
<tr>
<td>14003</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14163</td>
<td>9.74</td>
<td>3.51</td>
</tr>
<tr>
<td>14259a</td>
<td>7.97</td>
<td>3.38</td>
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<td>0</td>
<td>3.67</td>
<td>12.35</td>
</tr>
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</table>

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Concentrations (ppm)</th>
<th>Raw ratios (atomic)</th>
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</thead>
<tbody>
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<td></td>
<td>K</td>
<td>Rb</td>
</tr>
<tr>
<td>14003</td>
<td>2800</td>
<td>11.56</td>
</tr>
<tr>
<td>14163*</td>
<td>6250</td>
<td>14.61</td>
</tr>
<tr>
<td></td>
<td>84.05</td>
<td>281.4</td>
</tr>
<tr>
<td>14259a</td>
<td>3990</td>
<td>14.06</td>
</tr>
<tr>
<td>0</td>
<td>14.38</td>
<td>182.7</td>
</tr>
</tbody>
</table>

*Numbers in italics from Papastassou and Wasserburg (1971) are given for comparison.

![Graph](image-url)
Data are presented for 40 elements in 12 Apollo 14 samples (whole rock and clasts) and 4 mineral separates. Basaltic and anorthositic clasts from 14321, two basaltic clasts from 14072, the fine-grained matrix from 14063, and matrix, light and dark portions of 14306 were analysed. Whole rock samples 14310, 14047, and soils 14163 and 14003, all from the western part of the traverse area form a coherent compositional group. The average of these analyses is given in Table 1, which represents a first attempt to provide an average composition for the Fra Mauro Formation. These data resemble the Apollo 14 soil compositions and the detailed discussion of the rare earth patterns, comparisons with chondritic abundances, and with Apollo 11 and 12 samples have already been discussed by Taylor et al. (Geochim. Cosmochim. Acta, 35, 975, 1971). The most noteworthy feature is the striking enrichment in large cations (Cs, Rb, Ba), rare earths, and large high valency cations (Th, U, Zr, Hf), in comparison with mare basalts. How widespread or typical of the lunar surface is this material? The diversity in composition of the fragments and clasts, combined with the uniformity of the whole rocks suggests a thorough mixing process. The presence of similar material from the Apollo 11 (Luny rock 1) and Apollo 12 (dark 12013, light grey fines 12033, Kreep or norite components) together with the large occurrence at the Apollo 14 site shows that such strongly fractionated material, may constitute a significant primitive lunar surface layer. In Table 1, the composition is normalised to the abundances in Type 1 carbonaceous chondrites. This comparison is shown in Fig. 1 for the NON-VOLATILE elements. Those elements close in ionic radius and valency to Fe\(^{2+}\) and Mg\(^{2+}\) which will be major constituents of the lunar interior, are depleted in the Apollo 14 samples. A progressive enrichment occurs for those elements which differ in ionic radius or valency or both, from Fe\(^{2+}\) and Mg\(^{2+}\). Thus, provided that there has been no internal fractionation within the involatile elements (eg. REE) between the moon and the chondritic abundances, a strong outward concentration of elements, based on properties susceptible to crystal-liquid fractionation has occurred. This evidence suggests that the internal fractionation of the involatile elements occurred after accretion, in contrast to the loss of volatile elements at or before accretion.
Table I. Average composition at Apollo 14 site, Fra Mauro formation.

1. Abundances in parts per million (wt) or wt.%
2. Values in column I divided by abundances in Type I carbonaceous chondrites giving relative enrichment or depletion.
Figure 1. Relative enrichment and depletion in Apollo 14 samples relative to abundances in Type 1 Carbonaceous Chondrites. Data from Table 1.
Uranium Distribution and Mössbauer Studies of Apollo 11 to 14 Samples

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The U-distribution in chips of 10084, 21; 12021, 57; 12053, 39; 14305, 36 and 14321, 174 was studied by fission track and microprobe element mapping. Uranium has a relatively uniform distribution in the crystalline fragments selected from soil 10084, in the sections of rock 12053 and in the breccia 14305. In these samples an enrichment of U is indeed found in P-, Zr- and Ti-rich mineral phases. The bulk material of feldspars and pyroxenes however contribute a considerable amount of the total U-content, obviously concentrated in very small inclusions (≪1μm) of uncertain composition.

In contrast to this relatively uniform distribution, U in the rock 12021 and the breccia 14321 turned out to be highly concentrated in clusters of small apatites located in the corners of larger feldspar laths as well as along the rims of certain pyroxene grains.

The grain size of the apatites investigated in 14321 ranges from ≪10⁻⁶ cm² to about 7x10⁻⁶ cm² and contains about 46 to 81 ppm U with an average of 52 ppm. There was no correlation found between grain size and U-content.

The apatites in rock 12021 show a much greater variation in the U-contents up to several hundreds of ppm U. In this rock also appears an incompletely defined inhomogeneous silica glass, forming bubble-like inclusions and containing comparatively high amounts of U. U-contents in the bulk material of 12021 and 14321 are lying between 18 and 110 ppb U in the plagioclase feldspars and at about 20 ppb U in the clinopyroxenes.

U-rich phases containing U up to several percent, as reported for Apollo 12 sample 12013 (1) were not observed here.

Additionally, nuclear γ-resonance spectroscopy was applied to 9 different samples in order to characterize the Fe-containing minerals. The main absorption lines (77 to 92 % of the resonance area) belong to Fe²⁺ in silicates: olivine, pyroxine M1 and M2 positions and Fe-bearing glasses. With respect to ilmenite, 3 groups are distinguished ranging from 5 to 20 %. Troilite, probably of meteoritic origin, is a minor constituent of soils and breccias only (3 to 4 %), metallic iron however can be detected in all samples (Tab. 1).
Uranium Distribution and Moessbauer Studies of Apollo 11 to 14 Samples.
K. Thiel

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<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>Fe-silicates</th>
<th>Ilmenite</th>
<th>met. Fe</th>
<th>Troilite</th>
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<td>77.0</td>
<td>19.8</td>
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<td>12021, 57</td>
<td>rock</td>
<td>89.8</td>
<td>7.7</td>
<td>2.5</td>
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<td>rock</td>
<td>87.2</td>
<td>10.1</td>
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<tr>
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<td>rock</td>
<td>92.1</td>
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<td>2.9</td>
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<tr>
<td>12070, 87</td>
<td>soil</td>
<td>91.2</td>
<td>5.1</td>
<td>3.7</td>
<td>*</td>
</tr>
<tr>
<td>14163, 118</td>
<td>soil</td>
<td>81.5</td>
<td>14.2</td>
<td>4.3</td>
<td>*</td>
</tr>
<tr>
<td>14259, 86</td>
<td>soil</td>
<td>87.5</td>
<td>5.0</td>
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<td>3.5</td>
</tr>
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<td>14305, 36</td>
<td>breccia</td>
<td>91.4</td>
<td>6.2</td>
<td>2.4</td>
<td>*</td>
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<tr>
<td>14321, 174</td>
<td>breccia</td>
<td>86.9</td>
<td>6.2</td>
<td>3.9</td>
<td>3.0</td>
</tr>
</tbody>
</table>

* below detection limit of 1%

Tab. 1. Iron-containing components of Apollo 11, 12 and 14 samples. Given figures in atom-% of total Fe-content.

SULPHUR CONTENTS AND ISOTOPE RATIOS IN LUNAR SAMPLES,
H. G. Thode and C. E. Rees, Dept. of Chemistry, McMaster Univer-
sity, Hamilton, Canada.

Sulphur contents and $\delta^{34}\text{S}$ (1) values for lunar materials
have been reported for Apollo 11 and 12 samples (2-6). Thode
and Rees (7) reported similar data for Apollo 12 and, in addi-
tion, gave values for $\delta^{33}\text{S}$ and $\delta^{36}\text{S}$ indicating that sulphur
isotope abundance variations were essentially mass dependent.
Further determinations are reported here, as is a chemical
fractionation experiment on Apollo 14 fines.

Experimental procedures have been described elsewhere (7).
Sulphur contents have an error of $\pm 3\%$ while standard deviations
for individual determinations of $\delta^{34}\text{S}$, $\delta^{33}\text{S}$ and $\delta^{36}\text{S}$ are
$\pm 0.07\%$, $\pm 0.07\%$ and $\pm 0.7\%$ respectively. Experimental results
are shown in the table. $\delta^{33}\text{S}$ and $\delta^{36}\text{S}$ values have been convert-
ed to $\Delta^{33}\text{S}$ and $\Delta^{36}\text{S}$ values, which are residues after allowance
has been made for mass dependent isotope fractionation processes,
according to the relations $\Delta^{33}\text{S} = \delta^{33}\text{S} - 0.515 \delta^{34}\text{S}$ and $\Delta^{36}\text{S} =
\delta^{36}\text{S} - 1.90 \delta^{34}\text{S}$ (8).

The pattern of sulphur contents and $\delta^{34}\text{S}$ values is in accord
with previous determinations. Rocks have $\delta^{34}\text{S}$ values close to
zero while fines are 8-10% enriched in $\delta^{34}\text{S}$. For Apollo 12 and
14 samples this latter enrichment does not seem to be accompanied
by any marked difference in sulphur content as was the case for
Apollo 11. There is inhomogeneity in both sulphur contents and
$\delta^{34}\text{S}$ values. For rocks and breccia (12021-60, 12053-71; 14321-
220) there is some indication of a negative correlation of total
sulphur concentration and $\delta^{34}\text{S}$ value while for the fines 14163-
132 the correlation is positive.

The mean value of $\Delta^{36}\text{S}$ is $-0.2 \pm 0.2\%$, and indicates clearly
the absence of mass independent effects within the present limits
of precision. The mean value for $\Delta^{33}\text{S}$ is $-0.06 \pm 0.02\%$. The
deviceation from zero is almost certainly not significant. The
mean $\delta^{34}\text{S}$ value for the samples analysed is $-4\%$ and systematic
errors of $\pm 2\%$ in the factor of 0.515 in the $\Delta^{33}\text{S}$ definition, or
in the conversion of mass spectrometer ion current data to $\delta^{33}\text{S}$
values, could easily account for such a deviation. Thus there
appear to be no significant mass independent effects for sulphur
isotopes in the lunar material examined so far.

The absence of significant sulphur concentration variations
between rocks and fines in Apollo 12 and 14 samples makes it hard
to support mechanisms for $\delta^{34}\text{S}$ enrichment in fines involving loss
of sulphur (2,3,9). A preliminary experiment has been performed
Sulphur Contents and Isotope Ratios in Lunar Samples
H. G. Thode

### Table 1

<table>
<thead>
<tr>
<th>Sample</th>
<th>Sulphur Content ppm</th>
<th>δ^{34}S,‰ (a)</th>
<th>Δ^{33}S,‰ (b)</th>
<th>Δ^{36}S,‰ (c)</th>
</tr>
</thead>
<tbody>
<tr>
<td>12002A (d)</td>
<td>632</td>
<td>0.46</td>
<td>-0.06</td>
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<tr>
<td>12002A (e)</td>
<td>---</td>
<td>0.47</td>
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<tr>
<td>12018-33 (d)</td>
<td>549</td>
<td>0.68</td>
<td>+0.05</td>
<td>-0.7</td>
</tr>
<tr>
<td>12021-60 (d)</td>
<td>892</td>
<td>0.38</td>
<td>+0.04</td>
<td>-1.5</td>
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<tr>
<td>12021-60</td>
<td>768</td>
<td>0.56</td>
<td>0.00</td>
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</tr>
<tr>
<td>12021-60</td>
<td>683</td>
<td>0.65</td>
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</tr>
<tr>
<td>12022-50 (d)</td>
<td>914</td>
<td>0.37</td>
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<td>-0.1</td>
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<tr>
<td>12053-71 (d)</td>
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<td>0.68</td>
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<tr>
<td>12053-71</td>
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<td>+0.7</td>
</tr>
<tr>
<td>12070-57 (d)</td>
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<td>14321-220</td>
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<tr>
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<td>14163-132</td>
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<td>14163-132</td>
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<td>14163-132 (f)</td>
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<td>10086 (g)</td>
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<td>---</td>
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<td>-0.18</td>
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</tr>
<tr>
<td>10002-54 (g)</td>
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<td>---</td>
<td>-0.13</td>
<td>-0.9</td>
</tr>
<tr>
<td>10049 (g)</td>
<td>---</td>
<td>---</td>
<td>-0.05</td>
<td>+2.3</td>
</tr>
<tr>
<td>10057-40 (g)</td>
<td>---</td>
<td>---</td>
<td>-0.10</td>
<td>+1.2</td>
</tr>
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</table>

Mean value, ‰: -0.06
Standard deviation of mean, ‰: ±0.02

Footnotes:
(a) δ^{34}S values relative to Canyon Diablo troilite.
(b) Δ^{33}S = δ^{33}S - 0.515 δ^{34}S.
(c) Δ^{36}S = δ^{36}S - 1.90 δ^{34}S.
(d) Previously reported in reference (7)
(e) Repeat determination on Ag_{2}S.
(f) Mean value from fines fractionation experiment, see text.
(g) Determinations on Ag_{2}S separates provided by I.R. Kaplan.

on Apollo 14 fines collecting separately successive sulphur fractions released by acid treatment. Three successive fractions of ~30% each had distinctly different δ^{34}S values (-2.5, +0.1, +2.0 ‰ with respect to total) in a manner which may not be explained on the basis of simple isotope fractionation in the generation of H_{2}S from troilite. The results may be contrasted
with those of Epstein and Taylor (10) which indicate surface correlated enrichments (for 1-2% reaction) of the heavy isotopes of oxygen and silicon in fines. Our data show heavy isotope depletion in the first fraction (~30% reaction) and changing $\delta^{34}$S values throughout the entire reaction. The surface effects for oxygen and silicon isotopes are related to solar wind bombardment but the effect noted with sulphur extends through the bulk of the material and may not be explained by this mechanism.

References and Notes

1. $\delta^{33}$S, $\% = \{ [ (M/32S)_{sample}/(M/32S)_{standard} ] - 1 \} \times 1000$
   
   M = 33, 34, 36. Standard is Canyon Diablo troilite.


11. We are grateful for the opportunity to examine $\Delta^{33}$S and $\Delta^{36}$S effects in Apollo 11 material supplied to us by Prof. I. R. Kaplan.


In previous studies of the magnetic properties of individual glass spherules from the lunar samples (Thorpe et al, 1970; Sullivan et al, 1971), various magnetic components were measured and identified. The temperature dependent magnetism observed in the lunar glass spherules, at temperatures down to 77 K, was shown to be due primarily to the iron ions dissolved in the glass. At lower temperatures the susceptibility is less than that predicted by the Curie law, but can be fitted to a Curie-Weiss relation with a Weiss temperature of about 3-7 K in contrast to 2-3 K for tektite glass.

The Weiss temperature $\theta$ is related to either or both (1) antiferromagnetic interactions in a crystalline phase or (2) a distorted octahedral ligand field about the Fe$^{+2}$ ion in either the glassy or crystalline phases of the spherules.

If the value of $\theta$ is the result of antiferromagnetic interactions in the glass phase, then $\theta$ should depend on the concentration of iron, or, as the Curie constant $C$ is proportional to the amount of iron, then $\theta$ should also be a function of the magnitude of the Curie constant. As shown by the data in Table 1 and the plot in Figure 1 there is essentially no dependence of $\theta$ on the Curie constant (i.e., iron content) in tektites or other glasses, but there appears to be a dependence of $\theta$ on the value of $C$ in the glass spherules. If the observed $\theta$ is due to antiferromagnetism in a single crystalline phase in the glass, then $\theta$ would not be expected to be concentration dependent. However, if there are multiple crystalline antiferromagnetic phases embedded in the glass, then one might expect $\theta$ to vary with the iron depending on the relative crystalline phase concentration.

If the Weiss temperature, $\theta$, is due to the Fe$^{+2}$ ions in a distorted octahedral ligand field in the glass phase then the $\theta$ would not be expected to be concentration dependent, and the situation can be represented by a fine structure spin Hamiltonian, $H = DS_z^2$, where $D$ is the zero field splitting parameter and $S_z$ is the $Z$-component of the total spin in an assumed cylindrical symmetry. $D$ is a measure of the ligand field distortion and is directly related to $\theta$. However, when one attempts to fit the low temperature susceptibility data of the lunar glasses to the equation for the susceptibility derived from the above Hamiltonian, the fit is not good as might be expected. The data in Figure 1 suggests that there is a dependence of $\theta$ on concentration of iron for the lunar glasses, but not for the other glasses. It is possible that the total susceptibility of the lunar glasses may possibly be a function of the sum of two components. Thus, the sum of the antiferromagnetic component in the crystalline phases and a ligand field distorted paramagnetic component in the glassy phase will depend on the concentration of iron.
Table 1

The Weiss temperature, $\theta$, and the Curie constant, $C$, for lunar glass spherules, tektites and other glasses

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>$\theta$ (K)</th>
<th>C</th>
<th>Sample No.</th>
<th>$\theta$ (K)</th>
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<td>3</td>
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<td>7.01</td>
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<td>1.1</td>
<td>1.85</td>
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<td>4.9</td>
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</table>
Assuming the measured susceptibility is a function of both components, one can fit the data to an equation consisting of both components. A typical fit to the experimental data is shown in Figure 2.

Tektites contain no known antiferromagnetic fine particle phases and have a small $\theta$. The lunar glass spherules have a larger $\theta$ and are known to contain small antiferromagnetic mineral grains. Therefore, the two-component model of the temperature-dependent paramagnetism is favored for the lunar glasses.

REFERENCES


The Rayleigh wave properties of Apollo 12, 14, 15 and synthetic rocks were measured by the impulse technique. The lunar samples with the exception of 14321 are igneous rocks of competent, crystalline composition. Microscopic examination of our 14321.211 suggests that it does not belong to any of the clasts in the rock but represents the less coherent matrix material of the breccia boulder. Because of the sample friability, measurements could not be made in the normal manner, but estimates of the velocity were obtained by the time of flight method. The value of \(0.9 \pm 0.5\) km/sec could be significant since geological investigations of the Apollo 14 landing site thus far indicate that rock 14321 may be typical of most boulders in the block field on the rim of Cone Crater. However, it appears that our specimen is a breccia or matrix fragment whose velocity may not necessarily represent those in the Fra Mauro formation. Lunar rocks 14310 \((v_R = 1.20\) km/sec) and 12063 have velocities which are lower by a factor of as much as three from those typically measured in terrestrial basalts. One measurement on a sample of basaltic flow from northern California gave 2.97-3.15 km/sec for comparison. A similar difference was observed with bulk waves and has been explained as being due to the presence in lunar rocks of microfractures unfilled with water. Rayleigh wave velocities calculated for rock 12063 from measurements of the density and shear and longitudinal bulk wave velocities range from 1.27 to 1.64 km/sec and thus tend to be higher than those measured by us directly, \(v_R = 1.16-1.26\) km/sec, although the difference is small compared to the total range of bulk velocities reported thus far.

Our previous measurements of the Rayleigh wave velocity and bulk wave data reported by Wang et al. and Warren et al. suggest changes of velocity with propagation direction. This was checked by measuring the relative changes in signal arrival time systematically as a function of angle at fixed transducer separation on one surface of rock 12063. As the angle \(\theta\) between the propagation direction and one sample edge taken as reference was increased, a systematic rise in velocity was observed for low angles followed by a decrease and leveling out of the velocity at higher angles. Polished section microscopic examination of rock 12063 showed that microfractures, vesicles and ilmenite blades tend to be preferentially elongated in the direction \(10^\circ < \theta < 25^\circ\), which gave higher velocity values. The total change in velocity amounted to 9%. Similar measurements in the augite basalt from northern California which has a strong flow structure gave a total change in velocity of only 6% between values obtained parallel and perpendicular to the flow structure. Measurements of the linear compressibility of Apollo 11 samples by Anderson et al. showed anisotropy at low pressure which they suggested as arising from a preferential orientation of microfractures. We mapped the surface of 12063 by scanning
electron beam microscopy (magnification from about 50 to 500) and found that on a statistical basis the microfractures showed a preferred orientation in the direction in which the velocity was highest. The width of microfractures observed ranged from 0.001 to 0.3 mm with most of them about 0.01 mm in width. A rough estimate of the surface area of microfractures ranged from 5 to 15% for the total surface area. Therefore it would appear that one contribution to the velocity anisotropy may be a result of the material being more compliant for stresses perpendicular to the fractures. Thus, if there is an anisotropy in the distribution of fracture directions, one may expect an anisotropy in the compliance and therefore in the velocity of sound. In an effort to shed more light on the contribution of microfractures to the velocity, the sample was saturated with a low surface-tension fluid, reagent grade ethanol, and the velocity monitored as the alcohol was allowed to leave the sample by evaporation and then by evacuation to $10^{-8}$ mm of Hg. The velocity decreased from 1.61 to 1.26 km/sec with a net change of 25%. This result demonstrates that the presence of microfractures and their degree of filling can have a pronounced effect on the velocity of Rayleigh waves.

Velocity measurements were also performed on a synthetic basalt 10017 analogue which gave values for $v_R = 2.21\text{-}2.26$ km/sec, a factor of more than two times greater than those calculated ($v_R = 0.95\text{-}0.97$ km/sec) for 10017 from bulk wave data, but about 30% lower than on the terrestrial basalt. This result suggests the presence of fractures in the lunar samples, in addition to those introduced during the cooling process. The synthetic rock may be representative of the basalts which have not been fractured by exposure at or near the lunar surface.

The impulse technique was also used to measure relative changes in and approximate absolute values of the Rayleigh wave attenuation by recording the change in the amplitude of the received signal. In previous experiments, substantial relative changes in Rayleigh wave amplitude had been observed in rock 12038 when the absolute air pressure was changed from 1 atmosphere to $6 \times 10^{-7}$ mm of Hg. The amplitude increased by a total of 25% as the pressure was reduced with most of the change occurring between one atmosphere and 1 mm of Hg. In tests where the sample was previously outgassed, then pressurized with dry nitrogen gas, this change between atmospheric pressure and 1 mm of Hg was absent. This result is similar to the results obtained in bulk wave experiments by Warren et al. Adsorption studies have shown that an estimated 10 to 20 monolayers of H$_2$O are adsorbed onto glass and silicate surfaces. On the basis of these results and other considerations, it is believed that the trapping of water molecules in the microfractures is an important factor in decreasing the Q when the lunar return samples are allowed to come in contact with air. The source for the attenuation is believed to arise from water molecules collected in the regions of the fracture tips. During the passage of the sound wave pulse the relative displacements of opposing fracture faces near the tip effectively shorten the fracture length. This causes the flow of small amounts of liquid, giving rise to losses due to the viscosity of the liquid. When our sample was saturated with ethanol and then evacuated to $7 \times 10^{-8}$ mm of Hg, the relative amplitude of the Rayleigh waves increased by a factor of 3.75. This rather substantial change dramatizes the effect of filling of the microfractures on the relative Q of the sample.
The velocity data collected on rock 15555 were obtained on many locations on three different sample faces, one of which was polished. The values fell into the range 0.28-0.34 km/sec at a frequency of 700 kHz and are three to four times lower than those observed on homogeneous rock previously, and less than half those obtained on lunar breccia; i.e., for 14321 $v_R \approx 0.9$, and for 10046 $v_R \approx 0.7$ km/sec as calculated from bulk wave data. Microscopic examination shows that our 15555.90 is an olivine basalt with vesicles and vugs, showing little evidence for severe shock, but being highly fractured. Care was taken to make the measurements on the more massive portions of the rock in which the vugs were few and not exceeding about 0.1 mm in size. While it is hard to estimate the influence of the fractures on the stiffness of the sample, the sample is coherent and competent, and our values probably represent intrinsic properties. The existence of very low elastic wave velocities in this igneous rock indicates that the existence of a thick low velocity zone near the lunar surface does not necessarily imply the existence of a thick layer of fines. Field observation shows that the regolith in Station 9a at which the sample was collected is very thin or absent. The area is devoid of large craters, the surface is relatively flat with mature topography, and it is unlikely that the material of the boulder is derived from a far away source. From the information derived the sample most likely represents the upper layer of bed rock volcanics that are exposed at the Hadley Rille. Fractures aside, these represent unusually low velocities, that may in part account for the low seismic velocities observed in the upper layer of the lunar surface. This work was supported in part by NASA Contract NAS9-11542.

REFERENCES

Lunar seismic data from three Apollo seismometers are interpreted to determine the structure of the moon's interior to a depth of about 100 km. Seismic signals from three SIV-B and three LM-ascent stage impacts on the moon have been recorded over a distance range of 67 to 356 km by the Apollo 12, 14, and 15 seismometers. The travel times of the first (P) arrivals and the reflected phases are determined. The observed travel times and amplitudes are interpreted in terms of a compressional velocity model extending to a depth of 80 km inside the moon. This model is further refined by matching the observed seismograms with theoretical seismograms computed using generalized ray theory. Velocities very near the surface of the moon are determined from LM Thruster test firings and agree with Active Seismic Experiment results.

The seismic compressional velocity profile based on the combined data and interpretation techniques is shown in Figure 1. The most outstanding feature of the model is that the moon has a layered crust and the thickness of the crust at the investigated Fra Mauro region of Oceanus Procellarum is about 65 km. Other features of the velocity profile are: i) Very rapid increases at very shallow depths from a surface value of 0.1 km/sec to about 5 km/sec at the 10 km depth. ii) A sharp increase (discontinuity) at a depth of about 25 km. iii) Near constant value (about 7 km/sec) between 25 and 65 km. iv) A significant and discontinuous increase at the base of the lunar crust (65 km). v) As can be determined from a single data point corresponding to the distance of 356 km, very high velocity (greater than 9 km/sec) below the lunar crust.

The compositional implications of the velocity model are also illustrated in Figure 1. The high-pressure data of laboratory measurements on lunar and terrestrial rocks are also shown on the same plot. It is clear from the comparison that:

i) Near the surface, to a depth of about 1 to 2 km, self-compacting fines, breccias, and broken rocks can explain the velocities. Below this depth, basaltic rocks similar to those studied from Apollo 11 and 12 missions fit the velocity values to a depth of 25 km. The high
velocity gradients to a depth of 10 km result from the pressure effects on dry rocks having cracks and pores.

ii) The second layer of lunar crust seems to be distinctly different from the basaltic lunar rocks sampled from the surface. Terrestrial equivalents would range from norite to grabbro or pyroxenite or anorthosite.

iii) The mantle velocities, although tentative at this stage, are higher than most earth rocks and are close to Mg-rich olivines.

The evolution of such a lunar crust is discussed in the light of thermal history calculations and the geological properties of lunar rocks.

STRUCTURE AND ULTRASTRUCTURE OF LUNAR GLASSES IN APOLLO XIV SAMPLES

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Lunar glasses in samples n° 14 163 and 14 003 exhibit two different main morphologic facies. In the first one, glass makes up the most important part (sometimes the totality) of fragments without characteristic external shape, mostly angular, more or less scoriaceous or microbrecciated. The second facies is related to the classical glassy spheres described in the lunar regolith from Apollo XI landing. Both facies have been observed in the samples collected during Apollo XI and Apollo XII flights.

The morphologic features of vitreous fragments in Apollo XIV samples are identical to those already noted in previous samples: blebs of adhering materials due to spattering of molten particles -impact craters, symetric or not-dome-shaped blisters or roofs of outgassing bubbles -vesicles inside the fragment- fractures due to impact or thermal effects.

The matter of this paper is related to the microstructural feature which can be observed on the surface of fragments of both facies, angular and spherical. It can be described as microblebs, arranged closely one to another, and approximately homogeneous in size (200 Å to 500 Å in the samples studied).

Their presence gives to the glassy surface a microspherical etched appearance. This can be observed with the scanning electron microscope. The blebs due to the spattering of molten material, of either Fe-Ni or silicate composition, stand out against (and upon) this back-ground.

What is the origin of these microblebs? The first idea flashing through the mind is to consider them as minute grains either of regolith dust or of molten material similar to the big blebs adhering to the surface. These grains would have a somewhat homogeneous size and would cover the surface with some regularity.

But another hypothesis is suggested by the similarity of this surface's feature with what can be observed on terrestrial volcanic glasses surfaces. The microspherical etching is indeed also present. This feature has been studied especially on Hawaiian recent basaltic glass but found again in other glasses of various compositions. The conclusion is that the presence of the microspheres perceptible on the surface can be interpreted "from the interior" of the fragment (Trichet, 1969-1970). It is caused by the emerging of
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Demixed particles existing inside the glass. The existence of phase segregations is known in industrial glasses and important physical properties such as light scattering, thermal expansion, have been related to the size and shape of the demixed phases. It has been found too in volcanic glasses of various compositions and related to the existence of the subspherical particles of allophanes in volcanic soils (Trichet, id.)

Confirmation of the existence of such demixed particles in glasses can be obtained by two means: either on freshly fractured surfaces, using direct observation with scanning electron microscope or replicas techniques or by transparency in very thin chips under the electron microscope. In the first technique (observation of freshly cracked surfaces) artificial etching with acids is often necessary but not systematically. Hydration of the surface is known to be critical for the vanishing of the contrast between adjacent particles.

Both experiments have been conclusive with both lunar glasses types (angular fragments or spheres).

Microparticles exist inside the glasses and their size between 100 Å and 600 Å, what is in agreement with the surface’s microblebs one.

Unfortunately, such a size does not make it possible to get in situ analysis of the particles with the microprobes. Global analysis of 1 µ of glass surfaces (including one or more particles and surrounding matrix) have not shown any anomaly in Fe or Ni. This weakens the meteoritic hypothesis for the origin of these particles. Furthermore abundance of Fe-Ni particles inside the glass fragments is greater in the external portions of the fragments. This is not the case for the observed particles. It is known that all the blebs adhering on the glass fragments are not Fe-Ni in composition. They are then considered as spatters of silicate glasses and the microblebs inside the fragments could then be related to this type of particles having penetrated inside the fragment. But the density of the observed particles is not greater in the margin of the grain than in the heart and the penetration hypothesis is then weakened.

Explanation by emergence of internal in situ particles is then satisfactory.

Natural etching of the surface appears to be responsible of the emergence of the particles. Of course, the surface can become more and more covered by adhering blebs of various origins and sizes. It is then, and there, difficult to make a sure distinction between microblebs of similar size but different origin.

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COSMOGENIC NUCLIDES IN FOOTBALL-SIZED LUNAR ROCKS; M. Wahlen, M. Honda, M. Imamura, J. Fruchter, R. Finkel, C. Kohl, J. Arnold, R. Reedy, Univ. of Calif., San Diego, La Jolla, Calif., 92037

The study of lunar surface samples from the Apollo 11 and Apollo 12 missions has shown that bombardment by two classes of high energy particles can account for the production of cosmogenic radionuclides observed. The solar cosmic rays (SCR) produce significant effects on a depth scale of a few g/cm², while the galactic cosmic rays (GCR) penetrate to greater depths (1).

Rock 14321, a 9 kg fragmental rock, photographically documented on the lunar surface, provided an opportunity for extending our study of the variation of production with depth down to about 40 g/cm², as well as for continuing our investigation of solar effects. We received a series of samples (denoted as FM in Fig. 1) from a vertical column through the center of this rock. The counting data for each of our samples is given in Table 1 along with the various depths. Table 2 gives chemical compositions of rocks 14321 and 14310, another rock we studied.

The samples FM-3 and FM-5 carry us clearly into a depth region where, according to present ideas, only GCR effects should be significant. For the profiles of the various isotopes the observed gradient is zero or even positive between 18 g/cm² (FM-3) and about 40 g/cm² (FM-5), indicating that by these depths SCR's have been almost completely attenuated. Because the irregular shape of the rock puts FM-3 relatively close to the surface in some directions, there may be a small solar contribution at this depth. There is some indication, nevertheless, for Fe⁵⁵ and Mn⁵⁴, of a positive gradient between FM-3 and FM-5. This gradient is what one would expect to observe from the buildup of GCR secondaries at intermediate depths. We have used the FM-5 values, in which there is certainly no solar contribution, as normalization points for the theoretical galactic depth profiles (2). These normalized profiles are used in determining the net solar induced activity in the surface samples as discussed below.

The surface samples, FM-1 and FM-2, show the now familiar effects due to SCR bombardment. Co⁵⁶ in these samples is essentially produced only by SCR bombardment. Because of its short half-life, Co⁵⁶ production is dominated by recent solar flare events. We know of only one flare of any importance, that of January 24, 1971, that seems to have occurred close enough to the time of the Apollo 14 mission to have caused a significant production of Co⁵⁶. Comparison of the measured activities with those predicted from preliminary satellite data for the flux and shape of the January 24 flare (3) immediately implies that an attrition, i.e. loss of surface material due to handling of the rock, has occurred. This conclusion is well confirmed by the results of the other short-lived nuclides, Fe⁵⁵ and Na²², which, along with the Co⁵⁶, imply an attrition of (0.4±0.2) mm. Adopting this attrition value we obtain, for the January 24 flare, a spectral shape, R₀, of 70-100 MV and an equivalent steady state intensity of J (>10 MeV, 4π) ~ 170 protons/cm²sec, in agreement
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with preliminary data obtained from satellite measurements (3).

From the depth gradients of Fe$^{55}$ and Na$^{22}$ ($\tau_{1/2} = 2.6$ y), we derive simultaneously an $R_0 = 85$-$100$ MV and a $J \approx 100$ p/cm$^2$ sec for the solar flare particle flux responsible for the production of these nuclides, in good agreement with the data obtained from rock 12002.

Evaluation of the Al$^{26}$ and Mn$^{53}$ measurements requires some consideration of rock history. An erosion rate of about 1.5-2.0 mm/10$^6$ y is implied by these profiles. Using this erosion rate the Al$^{26}$ and Mn$^{53}$ profiles are mutually consistent and imply an $R_0$ of about 100 MV and a $J$ ($\omega 10$ MeV, $4\pi$) $= 70$-$100$ p/cm$^2$ sec. There is no evidence for undersaturation of Mn$^{53}$ due to a short exposure age for this rock.

A few radionuclides were measured in a five gram sample of the comprehensive fines--sample 14259--material reported to have been collected largely from the top centimeter of the lunar soil. Our Co$^{56}$, Al$^{26}$, and Na$^{22}$ data are all in good agreement with values obtained in the same sample by other workers (4, 5). The results look puzzling at first. The low Co$^{56}$ activity (although the error is quite large) together with the high Al$^{26}$ and Mn$^{53}$ activities could imply a rather complicated history for this sample. However, a simple explanation might be possible. The surface values of rock 14321, corrected for the slightly different chemistry, can be used to predict activity levels for soil layers of various thicknesses. Using such a calculation the observed net solar activities for Co$^{56}$ (if this is taken at the upper edge of its error range), Mn$^{53}$, Al$^{26}$, and Na$^{22}$ in the soil are just consistent with an actual sampling depth of 0 cm to about 2 cm (0-$3.6$ g/cm$^2$).

We also obtained a sample of rock 14310, a basaltic rock weighing 3.4 kg and containing 66% plagioclase ($\rho = 2.8$ g/cm$^3$) and 31% clinopyroxene ($\rho = 3.3-3.5$ g/cm$^3$). Our sample was a rectangular slab extending from the bottom toward the center of the rock, and had a mean depth of about 24 g/cm$^2$. 27 g of this sample were taken for mineral separation. The sample was ground to finer than 200 mesh and separated into 3 density fractions using thallium malonate-thallium formate heavy liquid. The heavy density separate, FIH ($\rho > 3.22$ g/cm$^3$), made up 36% of the total sample. The light density separate, FIL ($\rho < 3.00$ g/cm$^3$) made up 50%. The separation of clinopyroxene and plagioclase indicated in Table 2 allowed us to observe target effects in the induced activity of Al$^{26}$, Na$^{22}$, and Cl$^{36}$. Counting data for FIH and FIL are shown in Table 1. The production of Al$^{26}$ from different target elements was calculated, after applying a small correction for production from Ca and Fe to be 170±50 dpm Al$^{26}$/kg Si and 270±60 dpm Al$^{26}$/kg Al. These numbers are in good agreement with the data of Fuse and Anders (6) and are consistent with the Reedy-Arnold model calculations. Using these Al$^{26}$ production rates we calculate the activity in FM-5 to be 58 dpm/kg, in good agreement with the measured value of 56±8 dpm/kg. In the case of Na$^{22}$ and Cl$^{36}$, it is more difficult to derive production rates from different target elements because of the large number of targets involved. Using in addition the result from a deep sample of rock 10017, we obtain 115±25 dpm/kg Ca and 80±35 dpm/kg Ti for the production of Cl$^{36}$.

References:
Cosmogenic Nuclides in Gootball-Sized Lunar Rocks

M. Wahlen

5. Clark, R. S., private communication.

Table 1: Counting results (dpm/µg), corrected to February 6, 1971. The quoted errors include a 5-10% chemical yield error, a 5% self-absorption error where applicable, a 5-10% standardization error and a 10% counting error.

<table>
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<th>Sample</th>
<th>Wt. (µg)</th>
<th>Depth (µg/cm²)</th>
<th>C136 (77d)</th>
<th>Na56 (363d)</th>
<th>Fe55 (2.6µy)</th>
<th>Al26 (7.4x10⁶µy)</th>
<th>Mn54 (2.2x10⁶µy)</th>
<th>Cl36 (3.1x10⁶µy)</th>
<th>Be10 (2.5x10⁶µy)</th>
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<td>6.9</td>
<td>0.0-0.63</td>
<td>220±45</td>
<td>41±11</td>
<td>310±40</td>
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<td>176±22</td>
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Table 2: Chemical composition determined by atomic absorption spectrometry and colorimetry.

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<th>Sample</th>
<th>Na ppm</th>
<th>Mg ppm</th>
<th>Al ppm</th>
<th>Si ppm</th>
<th>K ppm</th>
<th>Ca ppm</th>
<th>Ti ppm</th>
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Fig. 1: Location of the samples from rock 14321. A and B denote samples received by Fireman and coworkers for Ar37-Ne33 studies.
BULK, REE AND OTHER ABUNDANCES IN APOLLO 14 SOILS (3), CLASTIC (1) AND IGNEOUS (1) ROCKS

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Abundances of 34 major, minor and trace elements, including the 14 REE (rare earth elements) plus Y, have been determined by INAA (instrumental neutron activation analysis) and RNAA (radiochemical neutron activation analysis) in two Ap. 14 soils, 14003-31 and 14163-55. These soil samples were scooped from the smooth Fra Mauro terrain near the LM (lunar module). Abundances of 21 major, minor and trace elements, including six representative REE have been measured by INAA in one Ap. 14 soil, 14240-10 (SESC sample), one clastic rock, 14047-30 and one igneous basaltic rock, 14073-3. Soil 240 and rock 73 were obtained at station G, ~100 m E of North Crater and these samples could represent material ejected from North and Center Craters. Rock 47 was collected at station B, ~300 m ENE of the LM on the smooth terrain.

In general, the abundances of our work agree well with those obtained by Schnetzler and Nava (1971), Brunfelt et al. (1971) and Taylor et al. (1971). The Ce abundances in soil 14163 agree well with those reported by the latter two groups and these Ce values are ~20% higher than the Ce obtained by Schnetzler and Nava (1971) and Hubbard and Gast (1972) for the same 14163 soil. Our Na₂O abundance for 14163 agrees well with that measured by Brunfelt et al. (1971); our Na₂O abundance is higher than the Schnetzler and Nava (1971) value by ~10%.

Ap. 14 soils are characterized by significant enrichments of Al₂O₃, Na₂O and K₂O and depletions of TiO₂, FeO, MnO and Cr₂O₃ relative to Ap. 11 and 12 and Luna 16 soils. Chemical compositions, including trace elements, of these three Ap. 14 soil samples, 3, 163 and 240, are essentially indistinguishable. If soil 14240 represents material ejected from the North and Center Craters, then the regolith must be relatively well-mixed within ~200 meters from the Triplet Craters.

Scandium abundances in Ap. 14 soils are lower than in other lunar soils (Ap. 11 and 12 and Luna 16); Sc correlates directly with the FeO content and inversely with REE abundances. Vanadium abundances are generally less in Ap. 14 soils than other lunar soils. Cobalt abundances, indicative of metallic

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meteoritic influx, do not vary significantly among all lunar soils measured to date. Abundances of Hf, Th, Ba and the 14 REE greatly exceed those observed in Ap. 11 and Luna 16 soils by a factor of ~5, and by a factor of ~2 in some Ap. 12 soils. However, abundances of these trace elements in Ap. 14 soils are nearly equal to those measured in Ap. 12033 soil and Ap. 12010 and 12034 breccias (Goles et al., 1971 and Wakita et al., 1971). Chondritic normalized REE + Y abundances and the Sm/Eu ratios in 3 Ap. 14 soils are nearly identical to those obtained for 12033 soil and 12010 and 12034 breccias. These close relationships in both bulk and trace elements suggest that either Fra Mauro breccia rocks were ejected onto the Ap. 12 landing site (Schnetzler and Nava, 1971) or that some of the Mare Imbrium ejecta blanket extended to the Ap. 12 site.

The bulk and trace element composition and Sm/Eu ratio of this work and the volatile and trace elemental abundances (Anders et al. 1972) for the 14047 fine-grained clastic rock are nearly identical to the composition of the Ap. 14 soils. Such a close chemical similarity suggests that 14047 is merely a compacted soil sample. Similarities in noble gas contents (LSPET, 1971) between 14047 and 14259 soil support this conclusion. Within the estimated sampling error, the bulk and trace chemical compositions and the Sm/Eu ratio of 12.0 ± 1.2 of the 14073 igneous basaltic rock are also nearly identical to the average Ap. 14 soil composition and Sm/Eu ratio of 11.6 ± 0.2. The bulk and the REE compositions and the Sm/Eu ratio of 14073 are very similar to those of 14310 (LSPET, 1971 and Hubbard and Gast, 1972). These compositional similarities and the same Rb-Sr ages for these two rocks (Papanastassiou and Wasserburg, 1971) suggest a common lava flow for these two rocks. Assuming a two component mixing model for soil derivation from KREEP and pulverized igneous rocks, we calculate that Ap. 14 soils are composed of ~20% KREEP and ~80% comminuted basaltic rocks like 14073 or 14310. This is consistent with the findings of the Ap. Soil Survey (1971) which suggested that the progenitor rock source for Ap. 14 soil would have a chemical composition similar to 14310.

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BULK, REE ... APOLLO 14
H. Wakita

Elemental abundances in three Apollo 14 soils, and one plastic and one igneous rock

<table>
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<th>Element</th>
<th>Soil 14003-31</th>
<th>Soil 14163-55</th>
<th>Soil 14249-10 (SPSC)</th>
<th>Basalt 14047-34</th>
<th>Igneous rock 14073-3</th>
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<td>Na₂O (%)</td>
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a One standard deviation due to counting statistics and other errors for single determinations are approximately +2-3% for Al, Na, Mn and Cr; +5% for Ti, Fe, Mg, Ca, Sc, Co, Cd, In, 14 rare earth elements and Y; +10% for K, Rb, Cs and Hf; +15% for V, Zr, Th and Ba, Na and +10% for Ce and Eu in 14073 analysis. Cd and In abundances are probably high due to LLNL contamination.

b Abundances were obtained via instrumental neutron activation analysis.

c Abundances were obtained via radiochemical neutron activation analysis.

d This sample was prepared by D. S. Burnett as a member of the G. J. Wasserburg consortium for analysis of the 14073 rock.
EXPERIMENTAL PETROLOGY AND ORIGIN OF FRA MAURO ROCKS AND SOIL

David Walker, John Longhi, and James Fred Hays
Department of Geological Sciences, Harvard University

Melting experiments over the pressure range 0 to 20 kilobars on Apollo 14 igneous rocks 14310 and 14072 and on comprehensive fines 14259 demonstrate the following:

1) Low pressure crystallization of rocks 14310 and 14072 proceeds as predicted from the textural relationships displayed by thin sections of these rocks. The mineralogy and textures of these rocks are the result of near surface crystallization.

2) Material having the chemical compositions of 14310 or 14259 cannot be representative of the average lunar composition since each of these compositions crystallizes to a dense (>3.5 g/cm³) clinopyroxene-garnet mineral assemblage at pressures greater than about 18 kilobars. Such a dense mineralogy at depth would contradict the mean density of the moon.

3) Partial melting of a lunar crust consisting largely of plagioclase, orthopyroxene, and olivine, followed by crystal fractionation at the lunar surface is a satisfactory mechanism for the production of the igneous rocks and soil glasses sampled by Apollo 14. The KREEP component of other lunar soils may have a similar origin.

In rock 14310 at low pressures, plagioclase (An 92) crystallizes first from the melt at 1310°C and precipitates alone until 1203°C where it is joined by orthopyroxene (En 88, Wo 3) and a trace of olivine (Fo 89). Plagioclase and orthopyroxene co-precipitate until about 1190°C where the liquid reacts with the orthopyroxene to produce pigeonite (En 86, Wo 5). At 1100°C a silica phase, probably tridymite, is present in the residuum. With increasing pressure this crystallization sequence is modified so that at 5kb orthopyroxene no longer crystallizes before pigeonite. It can be concluded from this and the vesicular nature of 14310 that the mineralogy and texture of this rock are the result of near surface crystallization.

In rock 14072 the low pressure liquidus phase is olivine (Fo 70) appearing at 1275°C, followed by chrome spinel at 1200°C. Pigeonite and plagioclase appear simultaneously at 1180°C. Olivine is consumed with falling temperature and is replaced as
the liquidus phase above 10 kb by aluminous orthopyroxene. We again infer near surface crystallization for the rock. The occurrence of early pigeonite rather than orthopyroxene in this rock is consistent with an Fe/Mg ratio greater than that of 14310.

Consider now the liquidus phase diagram for the system anorthite-forsterite-silica and its extension to iron bearing compositions (1). For all bulk compositions in the pyroxene-olivine-anorthite volume, the first liquid produced on partial melting has a composition buffered by the triple-saturation curve: anorthite-olivine-pyroxene-liquid. This triple saturation curve is a reaction curve such that olivine is consumed with falling temperature. Liquids generated along this curve, once removed from their olivine bearing parent rocks, follow crystallization paths on the plagioclase-pyroxene saturation surface.

We note with interest that the preferred glass compositions B, C, and D reported by the Apollo 14 Soil Survey appear to be controlled by these equilibria, as do the residual liquids in our experimental runs. Furthermore, the anorthositic gabbro compositions reported by Wood et al (2), the Surveyor 7 experiment (3), and by the Apollo 14 Soil Survey, upon partial melting yield liquids of composition near 14259 and Soil Survey glass B.

Our experiments show that rock 14072 and soil 14259 compositions both melt to liquids triply-saturated with olivine, plagioclase and pyroxene at 1180°C, and that rock 14310 is triply-saturated at 1200°C.

In the light of these observations, we propose the following model for the origin of Fra Mauro igneous rocks and soil. A lunar crust consisting largely of plagioclase, olivine, and orthopyroxene is partially melted to yield liquid of composition near B or 14259. (The process is quite insensitive to details of crustal chemistry as long as the bulk composition falls in the anorthite-pyroxene-olivine volume.) Liquid produced would be in equilibrium with plagioclase, pyroxene, and olivine and normal petrologic processes could act on those crystals. In our experimental charges, olivine has a tendency to sink whereas plagioclase tends to remain suspended. Local accumulation of plagioclase and a small loss of olivine by settling in a liquid near B would yield rock 14310. The complementary rocks enriched in olivine are represented by 14072. There is textural evidence that such a process has indeed occurred. 14310 contains some large, blocky, compositionally distinctive plagioclase phenocrysts and 14072 has large anhedral olivine remnants.

Unmodified liquids follow crystallization paths on the pyroxene-plagioclase saturation surface toward compositions such
as C and D. Extreme fractionation might lead to silica-saturated liquids of high K₂O content as also reported by the Soil Survey.

The KREEP component recognized in Apollo 11 and Apollo 12 soils (4) and in rock 12013 has a composition similar to that of the Apollo 14 fines. We suggest that it may have a similar origin.

If we are mistaken, and rock 14310 is indeed a primary magma as suggested by the Soil Survey authors and not a plagioclase cumulate as we believe, then our experimental data preclude an origin by direct partial melting at depths less than 100 km or greater than 400 km. An origin by partial melting at depths of 200-300 km is possible and would imply an orthopyroxene-clinopyroxene-spinel residue similar to that proposed by Ringwood and Essene (5) on the basis of their Apollo 11 work.

REFERENCES
Inclusions in two Apollo 14 breccias (samples 14171 and 14305) provide a more diverse petrologic suite than that observed in previous lunar samples. A major aim of this study was to detect petrogenetic relationships between the different rock types represented by the clasts. Petrogenetic relations are complicated by uncertainties in fragment source area and by the multiple brecciation events recorded by clasts composed of multiple generations of breccia.

In sample 14171, the two thin sections studied (-7, -13) are similar and consist of larger clasts composed of diverse breccias, together with smaller monomineralic clasts, contained in a finely devitrified matrix. Several fragments of hercynitic spinel were observed; a partial electron microprobe analysis gave (in weight per cent): total FeO 21.3, MgO 12.5, Al2O3 54.6, TiO2 1.22, CaO 0.5. In the thin section of sample 14305 studied (-92, -93, -104), the sample differs from 14171 in that a significant amount (about 25 per cent) of the clasts are rock fragments. Clasts composed chiefly of plagioclase are common, and many of the clasts are distinctively shock-metamorphosed.

Breccia mineralogy. Pyroxene occurs chiefly as separate monomineralic fragments within breccia clasts. Two types of pyroxenes are distinguished by chemical analyses: (1) "normal" pigeonites (low Ca) and subcalcic augites (high Ca) similar to those found in Apollo 11 and Apollo 12 mare basalts; (2) "unusual" (non-mare?) Fe-bearing diopsides (~Di90Fs10) and low-Ca, high-Mg orthopyroxene (~En95Fs15) associated with plagioclase-rich rocks of noritic to anorthositic composition, which occur particularly as clasts in 14305. The chemistry of the "normal" pyroxenes resembles that of pyroxenes in Apollo 11 and Apollo 12 mare basalts, for example in the TiO2-FeO and TiO2-Al2O3 relations (Fig. 1), although the Al2O3 level seems lower than that observed in Apollo 11 and Apollo 12 pyroxenes. There is a direct correlation of TiO2 and Al2O3 in both the "normal" and "unusual" pyroxenes (Fig. 1), but the "unusual" diopside and enstatite pyroxenes concentrate at the low end of this diagram (Fig. 1; crosses) with 0.75 weight per cent of both oxides. Furthermore, the "unusual" pyroxenes do not exhibit the negative correlation of TiO2 and FeO observed in "normal" mare pyroxenes, but instead form a separate group with 0.75 weight per cent TiO2 and 0.18 weight per cent FeO.

Olivine is generally found as isolated monomineralic clasts in breccia clasts. It may be associated with low-Ca orthopyroxene and was not observed in clasts containing diopside pyroxene. The olivine crystals are similar in composition to those found in Apollo 11 and Apollo 12 mare basalts in terms of CaO (~0.7), TiO2 (0.7), and Al2O3 (~0.4) contents (in weight per cent).
Plagioclase feldspar. Feldspar in clasts in breccias 14171 and 14305 is variable in An content, even within individual clasts. One clast in sample 14171 contains both the most anorthitic (An97) and albite (An74) plagioclase compositions analyzed. FeO contents (∼0.5 weight per cent) are at the low end of the range of previously reported values for lunar plagioclases. The feldspathic rock fragments (norite to anorthosite) in 14305 generally contain highly calcic plagioclase with a slightly lower FeO content (∼0.3 weight per cent).

Bulk chemistry of breccia clasts. Microprobe analyses were made on the matrix of larger (>0.3 mm) clasts and on the fine-grained breccia matrices to obtain chemical characteristics that might indicate petrogenetic relations. Analyses were made simultaneously for the elements K2O, CaO, Al2O3, FeO, MgO, and SiO2, using an electron beam in scan mode on areas about 20 μm on a side. Compositions of the breccia clast matrices, plotted on a normalized triangular diagram with the components SiO2, (FeO + MgO), and (CaO + Al2O3) (Fig. 2), can generally be expressed in terms of olivine, pyroxene, and a plagioclase with An contents between An70 and An100. The results indicate a poorly expressed trend toward An80. Similarly, on a plot of CaO vs. Al2O3 (Fig. 3), the breccia clast matrices are significantly higher in Al2O3 than are the Apollo 11 and Apollo 12 samples and one of the Apollo 14 crystalline rocks (14053). (The other Apollo 14 crystalline rock, 14310, plots among the matrix analyses.) The trend of analyses plots close to the ratio for anorthitic feldspar, suggesting...
either: (1) that fractionation of anorthitic feldspar during crystallization has been involved in the formation of the source rocks for the clasts, or (2) that the clasts have originated by dilution of anorthite-rich material with other phases that are not significantly enriched in CaO or Al2O3. If the trend is real, then the source rocks making up the clast matrices could have developed by removal of plagioclase of about An80 composition from a starting material with an Al2O3 content similar to, and a CaO/Al2O3 ratio slightly lower than, some Apollo 12 samples (e.g., 12004, 12018, and 12040).

Sample 14310 (crystalline melt rock). Specimens of this rock (section -182 and a fragment from -131) contain small crystalline inclusions with virtually the same mineralogy as the host rock. A sharp interface occurs between one inclusion (characterized by finely-crystalline plagioclase about 0.02 mm long) and the host rock (plagioclase about 0.3 mm long). The inclusion also contains a small angular grain of metal.

Compositions of plagioclase in both the inclusion and host rock range from An05 to An06. FeO contents are similar (~0.6 weight per cent), but the Or content of fine-grained plagioclase in the inclusion may be slightly higher (2 per cent) than that in plagioclase in the host rock (1 per cent). In the host rock, two types of pyroxene, one pigeonitic and the other augitic, were identified by chemical analysis. The pigeonites are consistently lower in FeO / FeO + MgO than the augites, whose FeO contents range upward to pyroxferroite compositions. Similar pyroxenes in the inclusion have compositions intermediate between the extremes observed in the host rock, i.e., the pigeonites in the inclusion are more FeO-rich, while augitic pyroxenes are more MgO-rich than in the host rock. The variation of FeO and Al2O3 contents as a function of FeO is similar for pyroxenes in both the fine-grained inclusion and in the coarser-grained host rock. The close chemical similarities of component minerals in both inclusion and host rock suggest that the inclusion is not exotic and may be a cognate inclusion.

Physical property measurements on lunar samples emphasize the roles of microcracks, absence of water, vacuum, texture, and prehistory. These parameters are different for lunar and earth rocks. Systematic variation of one parameter while the others are held constant is a near impossible task so that interpretation of experimental results often contains ambiguities. Some preliminary experiments on terrestrial rocks with thermally induced microcracks and new data on Apollo 14 samples are reported within this abstract.

Thermally cycled Fairfax diabase. If microcracks reduce velocities, can we try to create them in a systematic manner? Our experiment was to thermally cycle Fairfax diabase cores in vacuum to maximum temperatures between 100°C and 1000°C. The heating and cooling was gradual over a 24 hour period with the maximum temperature held constant for about 12 hours. The compressional velocity \( V_P \) was then measured to 5 kb at room temperature and the quality factor \( Q \) was measured at a frequency of a few Hertz at room conditions.

The velocity results are shown in Fig. 1. Note that the 5 kb values for all the Fairfax diabase cores are about 6.9 km/sec. For a core cycled to 600°C, \( V_P \) at \( P=0 \) is reduced to 4.2 km/sec. For a core cycled to 1000°C, \( V_P \) at \( P=0 \) is reduced even more to 2.2 km/sec. Thus we have simulated the low \( V_P \) values found in lunar rocks.

The Fairfax diabase contained less than 3% mica and no mineral alteration was seen in thin sections even for the core cycled to 1000°C. Externally, however, the core did become lighter in color. Perhaps the mica breakdown is responsible for the color change and also for some of the cracking. The plagioclase grains from the thermally cycled cores were observed to contain significantly more cracks than the grains before cycling. Evidently the cracking is caused by stresses between grains due to the mismatching of thermal expansions across the grain boundaries.

We hope to make a quantitative correlation between velocities and microcracks. Then we could also test theories which relate the pressure dependence of elastic moduli with the aspect ratio and number density of cracks.

The \( Q \) results on the thermally cycled cores do not match the seismically determined lunar \( Q \). The result might be expected
Velocities and thermal expansion
H. Wang

since the measurements were made at room conditions. The Q rises slightly from 450 to 550 for cores cycled between maximum temperatures of 100°C and 600°C. We interpret the behavior in terms of water finally being driven out at 600°C. For higher maximum temperatures the Q drops to 250 when the friction on cracks dominates.

Thermal expansion. Linear thermal expansions were measured between -100°C and +200°C for three samples: 12022,95; 14318,30; and Fairfax diabase. 12022,95, an igneous rock, had a thermal expansion less than that of Fairfax diabase while 14318,30, a breccia, had an expansion almost identical with that of the Fairfax diabase. The relative length change \( \Delta L/L_0 \) versus temperature is plotted in Fig. 2.

The thermal expansion values are 1/3 to 1/2 the intrinsic value calculated from mineral content. We interpret the low value at P=0 to be due to microcracks. The zero pressure thermal expansion is less than the intrinsic value while the zero pressure bulk modulus is also less than the intrinsic value. For most minerals and ceramic materials the thermal expansion and bulk modulus are inversely proportional. Again this anomaly is probably related to the microcracks. We hope to develop a firmer theoretical basis for the thermal expansion of rock by treating microcracks as a second phase in an isotropic composite.

Discussion. Laboratory physical property experiments combined with in situ ALSEP measurements have two primary purposes in lunar science: (1) To describe the physical state of the lunar interior and (2) To characterize possible lunar geologic processes.

These two purposes are illustrated by the often discussed paradox of lunar Q and velocities. The Apollo 11 seismic experiment gave low velocities for the subsurface layer yet extremely high Q. Laboratory measurements on returned samples were in good agreement with the velocity-depth curves but not with Q. Despite experiments by many investigators on the effect of water and vacuum conditions, in our opinion no satisfactory answer has been given to the physical conditions responsible for the seismic result.

But there is some certainty that the low velocities themselves are due to the presence of a greater number of microcracks than in earth rocks. What geologic process is responsible for their occurrence? Repeated shock impacts has been suggested as a means of producing microcracks. Another possibility is that the diurnal temperature variation may create cracks because of fatigue induced by a low level cyclic stress. From our experiment on Fairfax diabase we would like to add the possibility that cracks in lunar rocks are created in relatively high temperature subsolidus thermal cycling. Temperatures of 600°C to 800°C in a dry, vacuum environment would not induce chemical changes while
it does reduce velocities to very low values. We do not believe that such a heating would be inconsistent with natural remanent magnetism (NRM) studies on lunar samples. Some investigators suggest the possibility that the NRM was acquired subsequent to the initial cooling at temperatures slightly higher than the maximum diurnal temperature. If such a heating can be established as an actual previous geologic process, then the low velocities would be consistent with our experiment.

Fig. 1. Velocity vs. pressure. Solid curves from top to bottom are \( V_p \) for Fairfax diabase cores cycled to maximum temperatures of 200°C, 600°C, and 1000°C, respectively. Dashed curves are for sample 14310 (\( p=2.88 \text{ g/cm}^3 \)).

Fig. 2. Relative length change \( \Delta L/L_0 \) vs. temperature.
MULTIELEMENT ANALYSES OF LUNAR SAMPLES.

H. Wänke, H. Baddenhausen, A. Balacescu, F. Teschke, B. Spettel, G. Dreibus, M. Quijano, H. Kruse, F. Wlotzka and F. Begemann
Max-Planck-Institut für Chemie (Otto-Hahn-Institut), Mainz, Germany

The major, minor and many geochemically important trace elements have been determined in 7 Apollo 14 and 2 Apollo 15 samples. Nearly all measurements were carried out using neutron activation methods. About half of all 49 elements were determined via instrumental activation analyses before radiochemical neutron activation analyses using both fast and thermal neutrons. For the samples 14066-31;14321-184-1E and 14321-184-25 INAA was applied prior to destructive analysis by Wasson and co-workers.

Table 1 shows the results for a part of the elements only. From the three samples of 14321 analysed two are igneous fragments with identical composition. The third is a micro-breccia with a composition similar to that of the soil samples.

Negative europium anomalies have been found in all our lunar samples (Fig. 1). In two mesosiderites a rather substantial positive europium anomaly has been detected.

From the concentrations of noble metals in the Apollo 14 and 15 soils a meteoritic component of about 3.9% resp. 0.9% is estimated.

The results of the measurements have been used to compare the Apollo 14 material with the two components found in the Apollo 12 soil according to our mixing model. The soil samples 14163 and 14259 fit well into this mixing diagram indicating that the composition of the foreign component in the Apollo 12 soils is identical to that of the Fra Mauro region.

Our computer program used for the mixing model of the Apollo 12 soils has been extended for the three-component mixtures of the Apollo 15 soils. From the data\(^1\) of 13 elements of Apollo 15 soils and breccias alone three components were estimated. The first component being the local basalts, the second KREEP (norite) and the third anorthosite (Fig. 2). With these three components the average deviations of the computed concentrations from the experimental data is only about 6%. The anorthosite component is nearly 0% at the Rille station and increases going towards the mountain front (23% at St. George station).

1) PET - Report Apollo 15.
### Title
Multielement Analyses

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Other elements determined: Cu, Rb, Sr, In, Cs, Ba, Ce, Pr, Nd, Sm, Gd, Tb, Ho, Er
Fig. 1: Distribution of rare earth elements in several Apollo 12 and 14 and 15 samples relative to chondrites. The three lower curves were obtained on the silicate phase of mesosiderites.

Fig. 2: Three component mixing diagram for Apollo 12 (left diagram) and 15 (right diagram) soils and breccias. In both cases only the soils and breccias (x) were used to establish the mixing ratios. All entries with two digits in the left diagram refer to Apollo 12 samples. All entries with three digits in the right diagram refer to Apollo 15 samples. Note that in both cases the basalts (□) cluster in the right corner. For the upper corners the fitted average of KREEP and rock 12013 was chosen rather arbitrarily. This choice has no bearing as to the distribution of the mixtures (soils and breccias).
All Apollo 14 breccias for which thin sections are available (27 rocks) were studied petrographically. The samples display a range in matrix mineralogy and texture from glassy breccias with a detrital matrix, through intermediate members with little glass and an annealed matrix, to glass-free breccias with an interlocking, euhedral matrix. Based on the abundance of matrix glass, the abundance of glass clasts, and the matrix texture, the 27 samples are classified into eight groups fitting into three metamorphic facies (Table 1). Sketches of a typical matrix texture for six groups are presented in Figure 1.

Electron probe compositional data was obtained on one sample in each of six grades. Analyses were performed of matrix (less than 25 micron) plagioclase, pyroxene, and olivine and of rims and cores of plagioclase and pyroxene clasts (50-300 microns) (Table 1).

Matrix plagioclase and pyroxene in the low grade rock display a wide range of compositions, suggesting a detrital or accumulative origin. However, matrix plagioclase and pyroxene in the high grade rocks display a narrow range of compositions (about An 75-80 and Wo 5-10 En 65-70 Fs 20-25) suggesting a metamorphic origin. The nearly constant composition in the high grade matrix silicates suggests that the high grade breccias are equilibrated in the sense that equilibrated condrites are equilibrated. Low grade rocks are unequilibrated. Partially equilibrated states are observed in medium grade samples (Table 1).

Essentially all pyroxene and plagioclase clasts in high grade rocks display a homogeneous core and a 1-5 micron rim. The rim has the "equilibrated composition" regardless of the core composition. About half of the clasts in medium grade display rims and no rims were found in the low grade rock.

Apollo 11 and 12 breccias (excluding rock 12013) are surface breccias that were probably formed as relatively local base surge deposits. Low grade Apollo 14 breccias are texturally and mineralogically similar to Apollo 11 and 12 breccias. The Apollo 14 site lies in the Fra Mauro Formation which is a large ejecta blanket from the Imbrium Basin. Since the known base surge breccias are all low grade, and ejecta breccias are to be expected at the Fra Mauro site, it is concluded that at least the medium and high grade Apollo 14 breccias formed as part of the ejecta blanket that is the Fra Mauro Formation.

However, the continuous nature of the Apollo 14 breccia series suggests a common origin for the samples. Medium and high grade ejecta breccias may be derived from low grade equivalents by a metamorphic process - probably an autometamorphism in the thick, hot, Fra Mauro ejecta blanket. This does not deny the presence of any later surface breccias among the Apollo 14 suite: the low grade breccias may be either later surface breccias or unmetamorphosed ejecta breccias. It is concluded that surface breccias grade into and are indistinguishable from low grade ejecta breccias.
### Table 1

<table>
<thead>
<tr>
<th>FACIES</th>
<th>GROUP</th>
<th>MATRIX AMOUNT</th>
<th>GLASS AMOUNT</th>
<th>GLASS CLASTS AMOUNT</th>
<th>MATRIX TEXTURE</th>
<th>SAMPLE NUMBERS (14XXX)</th>
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<tr>
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<td>DETRITAL</td>
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<td>305 320</td>
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<td></td>
<td>7</td>
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<td>312 314 319</td>
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<td>NONE</td>
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<th>FACIES</th>
<th>GROUP</th>
<th>RIMS ON PLAG. AND PYROX. CLASTS</th>
<th>EQUILIBRATED MATRIX:</th>
<th>SAMPLE NUMBERS (14XXX)</th>
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<td>NO</td>
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<tr>
<td></td>
<td>2</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>MEDIUM</td>
<td>3</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td></td>
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<td>PARTIAL</td>
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</tr>
<tr>
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<td>PRESENT</td>
<td>YES</td>
<td>PARTIAL</td>
</tr>
<tr>
<td></td>
<td>6</td>
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<td>YES</td>
</tr>
<tr>
<td></td>
<td>8</td>
<td>PRESENT</td>
<td>YES</td>
<td>YES</td>
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</tbody>
</table>

1. **IN MATRIX GLASS AMOUNT:**
   - PRESENT = DARK BROWN, >50% ISOTROPIC MATERIAL,
   - INTERMEDIATE = LIGHT BROWN, <50% ISOTROPIC MATERIAL,
   - LOW = GREY, <2% ISOTROPIC MATERIAL,
   - NONE = GREY, NO ISOTROPIC MATERIAL.

2. **IN GLASS CLAST AMOUNT:**
   - PRESENT = >10% GLASS CLASTS, LITTLE DEVITRIFICATION,
   - INTERMEDIATE = <10% GLASS CLAST, ABOUT HALF ARE DEVITRIFIED,
   - LOW = <2% GLASS CLASTS, ALL SHOW DEVITRIFICATION,
   - NONE = NO ISOTROPIC MATERIAL PRESENT IN CLASTS.

*MINERAL CHEMISTRY NOT STUDIED

@ UNDERLINED SAMPLES WERE ANALYZED.
Figure 1. Photomicrographs of breccia textures in reflected light. Black areas are opaque minerals; light areas are pyroxene; dotted areas are plagioclase and glass; and dashed areas are either vesicles or polishing artifacts.

SCALE - 50 MICRONS

Sample 14042 - Groups 1, 2, 3
Sample 14171 - Group 4
Sample 14311 - Group 5
Sample 14303 - Group 6
Sample 14319 - Group 7
Sample 14068 - Group 8
LUNAR ROCKS, GLASSES, AND ARTIFICIAL LUNAR GLASSES: APPLICATIONS TO LUNAR GEOPHYSICAL MODELS. N. Warren, Institute of Geophysics & Planetary Physics, UCLA, Los Angeles 90024; N. Soga, Department of Industrial Chemistry, Faculty of Engineering, Kyoto University, Kyoto, Japan; and O. L. Anderson, Institute of Geophysics & Planetary Physics, UCLA, Los Angeles 90024.

Geophysical modeling of a mare region is considered. Recent data on artificial glasses, earlier velocity models (1), and soil data (2) have been compared to observed mare velocity profiles (3).

Glasses: Elastic properties were determined for a number of artificial glasses with compositions similar to 12009, 14052, 14049, 14305, and 14065. The range of chemistries covers both mare (m) and nonmare (f--feldspathic) basalt types. The glasses were melted at 1450-1550°C for 1 to 2 hours in reducing atmosphere, to prevent oxidation of FeO during melting. Shear (V_s) and compressional (V_p) velocities were measured at room temperature using pulse superposition method (4). The results (in the table below) are very similar to Apollo 11 glass data.

<table>
<thead>
<tr>
<th>Glass of Rock Type</th>
<th>Density (gm/cc)</th>
<th>Longitudinal Velocity (km/sec)</th>
<th>Shear Velocity (km/sec)</th>
<th>Poisson's Ratio</th>
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<tr>
<td>12009</td>
<td>3.066</td>
<td>6.693</td>
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<tr>
<td>14052</td>
<td>2.907</td>
<td>6.494</td>
<td>3.612</td>
<td>0.276</td>
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<tr>
<td>14049</td>
<td>2.802</td>
<td>6.519</td>
<td>3.670</td>
<td>0.268</td>
</tr>
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<td>14305</td>
<td>2.798</td>
<td>6.633</td>
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<td>0.272</td>
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<td>14065</td>
<td>2.748</td>
<td>6.553</td>
<td>3.698</td>
<td>0.266</td>
</tr>
</tbody>
</table>

The density appears to increase linearly with increasing FeO content.

Velocities of m and f basalts can be estimated from these data using the fourth power law (5). V_p = 7.4 km/sec [in agreement with observed V_p in the 40 to 70 km region (3)] are expected for m basalts with density 3.3 and f basalts with density 3.0. Enrichment of garnet or FeO would increase V_p for f basalts to
Lunar Rocks, Glasses, & Artificial Lunar Glasses

N. Warren

8.5 km/sec at $\rho = 3.3$. If $f$ basalts underly $m$ basalts, a density inversion with depth is possible.

Rocks and Soil: Lunar mare velocity at shallower depths ($\leq 30$ km) can be quantitatively accounted for by increasing structure (e.g., microfracturing, jointing, faulting) in parent $m$- or $f$-type basalts (6). The original floor of the (unfilled) mare basin is suggested to be at a depth of about 20 km.

Scattering of seismic energy and rock structure in the mare upper 10 to 20 km has been discussed (7). It is suggested that the gross elastic scattering properties of the region can be satisfied assuming a very low-porosity, jointed (or fractured), dry bedrock model in which the parent material has an intrinsic $Q$ of a few thousand.

Bulk porosities as low as $10^{-3}$ to $10^{-4}$ are sufficient if the voids are mainly flat cracks and joints. (A small partial porosity of low aspect ratio voids will not strongly affect the elastic moduli.) Density of the mare fill may be 3.0 to 3.2, with extensive breakings of the bedrock restricted to the near surface ($\leq 0.5$ to $1$ km) ($\rho = 1.6$ at the surface).

A distribution of fractures or joints with major dimensions on the order of hundreds of meters appears to be sufficient to account for seismic energy scattering. This model implies that extensive complete fracturing and brecciation of bedrock to depths of several kilometers by meteoritic impacts is not necessary.

References


Lunar Rocks, Glasses, & Artificial Lunar Glasses

N. Warren

5. N. Soga, R. Ota, and M. Kunugi; in PROCEEDINGS OF THE 1971 INTERNATIONAL CONFERENCE ON MECHANICAL BEHAVIOR OF MATTER (in press)

6. N. Soga, N. Warren, and O. L. Anderson; to be submitted to Proceedings of the Third Lunar Science Conference (1972)

COMPARISON OF Rb-Sr, K-Ar and U-Th-Pb AGES; LUNAR CHRONOLOGY AND EVOLUTION


A summary of the Rb-Sr internal isochron ages of lunar rocks determined by the authors is shown in Fig.1. The ages for rocks from five missions lie between 4.00 and 3.16 AE. No rocks have yielded ages younger than 3.16 AE. If we consider mare basalts only, we obtain a rather narrow time interval of 0.55 AE for the filling of the maria. Alternatively, we may be sampling only the last major lava flooding at each site.

K-Ar ages using the 4°Ar-39Ar technique have been obtained on crystalline rocks from Apollo 14 and 15 using carefully controlled standards which have been calibrated absolutely: 14310, 3.89 AE; 14053, whole rock 3.94 AE, plagioclase 3.95 AE; 14073, whole rock 3.90 AE, plagioclase 3.88 AE; 15555, whole rock 3.22 AE, plagioclase 3.31 AE. Fragments from the soil have yielded 14001.7, 3.90 AE; 14167.8, 3.94 AE. Uncertainties in these results are ± 0.05 AE. The good agreement between the Rb-Sr and the K-Ar ages indicates that there are no serious discrepancies between the two methods when the 4°Ar-39Ar release pattern allows a precise age determination.

Replicate 4°Ar-39Ar analyses of the anorthosite 15415 show a pattern of monotonically increasing 4°Ar/39Ar with high temperature values corresponding to 4.05 ± 0.15 AE. The ratio of spallogenic 38Ar to Ca determines a cosmic ray exposure age of 118 m.y.

Analyses of Pb-U-Th were carried out on total rock samples and mineral separates using small samples (2-50 mg) with typical blanks ranging from 0.6-2.0 x 10-9 g Pb. Total rock model ages are shown in Table 1. Separates from both 14310 and 14053 showed a wide range in 207Pb/206Pb. Internal isochrons were obtained for two rocks and yielded initial 207Pb/206Pb as shown in Table 1. These isochrons are completely compatible with the Rb-Sr internal isochrons and K-Ar ages indicating no major discrepancies between the methods. The initial 207Pb/206Pb corresponds to the integrated production of these isotopes in a high uranium reservoir between 4.50 and the time of formation of these rocks (∼ 3.90 AE). The data substantiate the model involving original radiogenic Pb proposed by Tatsumoto. Fig.2 shows the data for rock 14053, plotted on a new representation which is analogous to the Rb-Sr evolution diagram.

Rb-Sr and K-Ar ages on a 0.06 g basalt boulder from Luna 16 yield ages 3.42 ± 0.17 AE and 3.45 ± 0.04 AE, in excellent agreement. The total range in 87Sr/86Sr measured is 0.15% and results in the high uncertainty of the Rb-Sr age. We obtain a precise I value for this basalt, I = 0.69906 ± 4.

In contradistinction to the ages of lunar rocks, we show in Fig. 3 data on soils. All soils with widely different Rb/Sr and chemistry lie on the
Comparison of Rb-Sr

G. J. Wasserburg et al.

same isochron through BABI corresponding to 4.6 AE. Deep drill stem samples from Apollo 15 show a wide range in Rb/Sr indicating considerable heterogeneity in the soils at this site. The typical model ages of soils are somewhat less than 4.6 AE and range down to 4.3 AE. Model Rb-Sr whole rock ages for a large number of basalts are close to 4.6 AE and within the range 4.6 ± 0.3 AE. The Luna 16 soil is the least radiogenic sample in the soil array. It is observed at all sites that the soils have Rb/Sr higher than the average Rb/Sr of the local basaltic rocks. We infer, therefore, the universal presence of a high Rb/Sr component in the soil with a model age of 4.6 AE. Such a component was clear at the Apollo 11 site.

From the T-I diagram (Fig.1), all mare basalts have very low I values and demonstrate that the source regions of these basalts have extremely low Rb/Sr, incompatible with chondritic composition. For a given age we obtain a range in I values; this effect is largest for the Apollo 14 rocks. It is also clear that the I values are not a monotonic function of the age of the rock or of the time of extraction of these basalts from their respective sources. In addition the I values are correlated with Rb/Sr in the whole rock. This correlation is suggestive of a process whereby upwelling low Rb/Sr basaltic magmas are contaminated with small but variable amounts of high Rb/Sr crustal material. The evidence from the I values, from the soils and from rock 12013 indicates the existence of a lunar crust high in Rb, K, U and Th formed close to 4.6 AE ago. The location and means of formation of this crust as well as its detailed interaction with the basaltic lavas are not clear at this time. In particular the crust may be the result of early lunar magmatic activity superimposed on a chemically non-uniform moon due to layering during the accretional process which produced that planet with a high (200-300 ppb U, today) concentration of U and Th in the outer 200 km.

From constraints on the thermal evolution of the moon, and, in particular, from the apparent cutoff of significant magmatic activity after 3 AE ago, we have also concluded that the mantle of the moon must be chemically heterogeneous, with the radioactive heat sources increasing towards the surface. It is necessary that the inhomogeneities or layering in terms of the abundance of specific key elements not involve significant density contrasts in order to satisfy the observed moment of inertia of the moon.

The key questions which remain are: 1) Is there significant lunar magmatic activity for times much younger than 3 AE ago? 2) Was there significant magmatic activity subsequent to the initial production of a radioactive crust and prior to the onset of mare flooding at 4.0 AE? 3) Must there have been an iron core produced by deep differentiation in order to explain the observed remanent magnetization?

<table>
<thead>
<tr>
<th>Sample</th>
<th>(^{206}\text{Pb})</th>
<th>(^{207}\text{Pb})</th>
<th>(^{207}\text{Pb})</th>
<th>(^{208}\text{Pb})</th>
<th>(^{208}\text{Pb})</th>
<th>(^{207}\text{Pb})</th>
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<td></td>
<td>(^{238}\text{U})</td>
<td>(^{235}\text{U})</td>
<td>(^{232}\text{U})</td>
<td>(^{232}\text{Th})</td>
<td>(^{208}\text{Pb})</td>
<td>(^{207}\text{Pb})</td>
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<td>4.28</td>
<td>4.10</td>
<td>1.45</td>
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<tr>
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<td>4.25</td>
<td>4.26</td>
<td>4.06</td>
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<tr>
<td>14053</td>
<td>5.54</td>
<td>5.16</td>
<td>5.01</td>
<td>5.36</td>
<td>1.46</td>
<td>----</td>
</tr>
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</table>
Fig. 1. T-I diagram for lunar rocks, including 15555 and a Luna 16 basalt. Apollo 14 rocks fall in two distinct groups: high I, lower T (14310; 14276; 14073; 14001, 7, 3); low I, higher T (14053; basaltic clast from breccia 14321).

Fig. 2. $^{207}\text{Pb}/^{206}\text{Pb}$ evolution diagram for total rock and mineral separates for 14053. A similar result was obtained for 14310. Points B are corrected for blank but not for primordial Pb. Points C are corrected for blank and primordial. Initial $^{206}\text{Pb}/^{204}\text{Pb}$ is $80^{+40}_{-20}$.

Fig. 3. Rb-Sr evolution diagram showing the remarkable linear array formed by all analyzed soils from five lunar missions. The line drawn is a 4.6 AE isochron through BABI. Note the wide range in Rb/Sr of Apollo 15 soils and the primitive nature of Luna 16 soils.

Electron magnetic resonance spectra of Apollo 14 fines (14003-60, 14148-31, 14149-47, 14156-31, 14163-68), fragmental rocks (14301-66, 14303-42, 14311-35, 14318-36, 14321-166), and a crystalline (igneous) rock (14310-68) have been measured at 9 and 35 GHz and at temperatures ranging from 40 to 300°K. The spectrum of a plagioclase fraction (99% plagioclase) of 14053 has been investigated as a function of °Cs gamma-ray irradiations, electron (2 MeV) irradiations, and subsequent heat treatments. Measurements have also been made on various fractions of the fines, fragmental rocks, and the crystalline rock.

The spectrum of the fines is dominated by a resonance absorption that has a width of $\Delta H = 607^{\pm 15}$ oersteds at 9 GHz and $817^{\pm 9}$ oersteds at 35 GHz where these values of $\Delta H$ are the average of all samples and the + and - values are the extremal values of $\Delta H$. The line shapes are asymmetric and the $g$-value of the absorption maxima is $g = 2.1$. This resonance is similar to the one which was observed in Apollo 11 and 12 fines with respect to the intensity of the absorption, $g$-values, and the temperature dependence of the peak intensity and $\Delta H$. The line width, $\Delta H$, of the resonance in the Apollo 14 fines is less at both 9 and 35 GHz, 607 as compared with 770 oersteds at 9 GHz and 817 as compared with 950 oersteds at 35 GHz. The rate of increase in $\Delta H$ with increasing frequency is greater for fines from Apollo 14 as compared with fines from Apollo 12, 817/607 as compared with 950/850, respectively. A weak absorption peak is observed at $g = 4.2$ when $\nu = 35$ GHz in the Apollo 14 fines. Although the intensity of this peak relative to the $g = 2.1$ peak is $< 2 \times 10^{-3}$, it has been observed in all the samples on which measurements have been made with the exception of 12033-15.

Neither of these absorption peaks is observed in any of the pyroxene, plagioclase, colorless, and transparent fractions of the fines whose diameters are $> 0.05$ mm. (This diameter is the minimum that has been separated.) They are observed in that fraction of particles whose diameters are $< 0.05$ mm, in black basaltic particles, and in scoriaceous glassy particles. In the case of these latter two types of particles, absorption at $g > 2$ (magnetic fields smaller than that field necessary for the $g = 2.1$ absorption) is of the same order of intensity as the $g = 2.1$ absorption.1

†Deceased.
It has been suggested that the $g = 2.1$ absorption is due to spherical iron particles. The $g$-value, asymmetry of the absorption, and the width $\Delta H$ are in reasonable agreement with values calculated from simple theory.\textsuperscript{3} $\omega = - \gamma (H - K_1/2M_s)$, where $\omega$ = spectrometer frequency, $\gamma$ = gyromagnetic ratio, $H$ = magnetic field of resonance maximum, $K_1$ = anisotropy constant of cubic iron, and $M_s$ = saturation magnetization. The line width, $\Delta H$, is given by $\Delta H = [K_1/M_s]$, and for iron $|K_1/M_s|$ = 540 oersteds.\textsuperscript{4} If it is assumed that $H$ at resonance at 9 GHz is sufficient to produce saturation, then $\Delta H$ is independent of an increase in $\omega$. If it is not sufficient and $M < M_s$, then $\Delta H$ should decrease with increase of $\omega$. As noted above $\Delta H(9 \text{ GHz}) < \Delta H(35 \text{ GHz})$. The increase may be due to the contributions from other magnetic phases present in the fines. The weak absorption at $g = 4.3$ may be due to hematite, perhaps derived from a meteoritic component.\textsuperscript{1} An absorption peak at this $g$-value and at $g = 2.1$ is characteristic of several hematites which have been measured.

The spectra of the fragmental rocks either at 9 GHz or 35 GHz are not similar and all differ markedly from the spectrum characteristic of the fines. There are at least two distinctive magnetic phases in 14301-66, 14311-35 and 14318-36, while the spectra of 14303-42 and 14321-166 are apparently due primarily to one phase. These spectra are also dissimilar to the spectra of crystalline rocks (14310-68, 12021-55, 12075-19, and 10047-49). The temperature dependence of the width and intensity of these components is characteristic of ferromagnetic phases. A lustrous metallic phase is visible in 14318 which is presumably iron and iron-nickel alloys. Resonance absorption from these phases does contribute to the total absorption. Paramagnetic resonance absorptions due to Fe$^{3+}$ and Mn$^{2+}$ were detected in samples of 14311-35, 14321-66, 14303-42, 14310-68, 14318-36, and 14321-166. The concentrations of these two ions are of the same order as that found in crystalline rocks in which they have been detected.\textsuperscript{1} A third paramagnetic absorption was also detected in 14321-66 with $g = 2.001$ and a width $\Delta H = 3$ oersteds which saturated at high microwave power levels. The concentration of the paramagnetic state producing this absorption is of the order of $10^{16}$ gm$^{-1}$. All of these properties are in accord with those of the paramagnetic state of oxygen vacancies in silicate minerals.\textsuperscript{5,6} This paramagnetic state has not been detected in samples from Apollo 11 or 12, although it was produced in 12021-55-5 by irradiation with 1-MeV electrons.\textsuperscript{7} Lacking data on the times these samples were exposed to energetic solar radiation and cosmic rays, it is not possible to determine if the oxygen vacancies are due to radiation damage. Spectra of the paramagnetic states of Fe$^{3+}$, Mn$^{2+}$ and Ti$^{3+}$ have been detected in a plagioclase fraction of 14053-47. It is estimated that the sample has less than 1% mineral impurities.\textsuperscript{8} The concentrations of Fe$^{3+}$ and Ti$^{3+}$ are of the order of $10^{16}$ g$^{-1}$. The primary impurity is ilmenite\textsuperscript{8} and the fraction of Ti$^{3+}$ in it is below the limits of detection; hence it is assumed that these two ions are contained in the plagioclase. Approximately 1% by weight iron was detected in plagioclases of other samples.\textsuperscript{9} Irradiation of this sample with $^{137}$Cs gamma rays at 78ºK and measurement without warmup produced an increase in the intensity of the resonance absorptions due to Fe$^{3+}$ and Ti$^{3+}$ and introduced two new absorptions at $g = 2.01$ and $g = 2.001$. The Fe$^{3+}$ absorption increased by a factor of $\sim 1.5$ and the Ti$^{3+}$ absorption by a factor of $\sim 2$. Heating the sample at various
temperatures up to 400°C reduced the intensities to their original values. These results indicate that the electron transfer reactions previously proposed are not sufficient. Alterations in the valence state of other ions occur during irradiation, two of which produce the absorptions at \( g = 2.01 \) and \( g = 2.001 \). Changes in the spectra of a plagioclase fraction of 10047-49 which were in part similar to these, were produced by a similar irradiation.\(^7\)\(^9\)

The most intense component of the spectra of Apollo 14 fines on which measurements have been made is similar with respect to intensity and temperature dependence of \( \Delta H \) and peak absorption intensity to the characteristic spectrum of Apollo 11 and 12 fines. However, \( \Delta H \) is smaller and has a larger increase with increase in spectrometer frequency as compared with Apollo 11 and 12 fines. This ubiquitous spectrum of the fines cannot be produced by mechanical comminution of crystalline rocks, also an association with glassy particles has not been established. There is abundant data that is inconsistent with spherical metal particle hypothesis, but which is consistent with the presence of a ferromagnetic iron oxide phase.\(^10\)\(^7\)\(^1\) Fragmental rock specimens have spectra that differ markedly from the spectra of fines and from rock specimen to rock specimen. Paramagnetic \( \text{Fe}^{3+} \), \( \text{Mn}^{2+} \) and \( \text{Ti}^{3+} \) are present in two fragmental rock specimens but have not been detected in three other specimens. The paramagnetic state of oxygen vacancies was also detected in these two specimens. A large fraction of the \( \text{Fe}^{3+} \) and \( \text{Ti}^{3+} \) in Apollo 14 fragmental rocks appears to be in the plagioclase fraction, as was the case for one Apollo 12 crystalline rock specimen.

REFERENCES

4. Ibid., p. 352.
OPTICAL ORIENTATION, COMPOSITION AND TWIN-LAWS OF PLAGIOCLASES
FROM ROCKS 12051, 14053 AND 14310

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From thin sections 12051,57, 14053,11, 14053,19, 14053,61 and 14310,14 plagioclases were selected for this study. Their compositional range determined by microprobe is An 86 to 94 for the wide cores, in the narrow marginal zones the An content can drop to 75%. The broad centers of the crystals to which our optical data refer are not homogeneous, as shown by the following microprobe point analyses in 4 traverses across one and the same crystal of 14310,14 (data in brackets refer to marginal zones):

90, 91, 93, 92, 93, 92, 94, 91, 93
(82, 79, 88), 89, 90, 93, 92, 93, 92, 92
94, 92, (84, 92)
91, 93, 90, 86

This inhomogeneity shows up in the imperfect optical extinction of the crystals and renders the U-stage study rather difficult.

The attention of petrologists is drawn to the occurrence of bytownites An 71-80, with the exceptional Or-content of 3%, in slide 14053,11 which are surrounded by silica minerals and iron ore and which can hardly be distinguished morphologically from the predominant anorthites embedded in pyroxene. The nuclei of the bytownites show compositions similar to the marginal zones of the anorthites.

Nearly all feldspars of these rocks are twinned, in the great majority with (010) or vicinal faces as composition planes. Simple or polysynthetic albite twins predominate (90% of all) and none of the other laws represents more than about 5% of the total twins. The general sequence in frequency appears to be: albite > Roc Tourné, Carlsbad, pericline-acline, Baveno > Ala, albite-Ala > Manebach which is very rare. But there is some variation from rock to rock; in sample 14310 simple, never polysynthetic pericline twins form conspicuous squares and are second in importance after albite, third being Baveno. In samples 12051 and 14053 the laws Ala B and albite Ala - which are easily overlooked unless a full U-stage study is made - are combined with albite and Roc Tourné twins, while pericline and Baveno are the least common. Thus, on the moon a set of twin
laws known from igneous rocks of the earth is found. No law other than albite forms polysynthetic lamellae. In composite (010) twin groups Carlsbad as well as Roc Tourné laws may occasionally be repeated, but never in adjacent lamellae.

At An 80 high and at An 82 low the directions [†] of Carlsbad-twins coincide with each other and occupy the position of pole (I10). This plane contains the optical axes and becomes an additional plane of symmetry of the twin, as does very nearly also (130) which is virtually perpendicular to (I10). Therefore, at this very special composition a Carlsbad-twin can be interpreted also as a twin according to the prism-law (130) as far as the symmetry of the indicatrices is concerned. However, in all cases examined (010) was determined as composition plane and also the combination with albite and Roc Tourné proves that the interpretation as Carlsbad-twin is correct. So far, (110) and (130) were ascertained in moon-feldspars only as faces of interpenetrant crystals not as twin-planes.

Interpenetrant twin-groups show regular and irregular mutual orientation. Apart from pseudo-symmetry which appears to be quite common, some examples of interpenetrant twins were found in which individuals of the two intergrown groups show the symmetry of the Baveno law.

The optical orientation was derived from 59 complex (010) twin-groups, by constructing the twin axes and symmetry planes between the indicatrices. Similar to the experience with feldspars of terrestrial igneous rocks the composition-planes of twins are highly irregular. Therefore, we did not rely upon morphological data. Cleavage is poor, though the crystals are sometimes bent or broken.

In triplets there is little uncertainty in locating the three mutually perpendicular twin axes of the Carlsbad (Z-axis), albite (Y-axis) and Roc Tourné (X-axis) laws within one degree, and reading the Eulerian angles from the stereogram (diameter 40 cm). In this way two sets of optical data were obtained, one referring to individual crystals, the other giving the average for each thin section. Subsequently, the crystals selected for optical study were analysed by electron microprobe.

The optical analysis was done with special care considering the restricted migration of the indicatrix of basic plagioclases in response to change in composition, thermal history and ordering. In several instances the same crystal was examined by three operators with different microscopes and U-stages. In homogeneous parts the angles measured by various observers agreed very well (within ±0.5°). Only when influenced by zonal structure the angles differed more than 1°. The determination of 2V by measurement of both optical axes requires high tilting angles, at which birefringence of the cement used for the moon slides has considerable influence. Accordingly 2V data show more scatter.
In the bytownite range the relations between the optical orientation and the chemical composition are in line with those for plutonic plagioclases compiled by Burri, Parker, Wenk (1967). This is true not only for selected twin groups, but also for the averages of sample 12051,57 (mean of 14 series of Eulerian angles $\phi$ 25.2, $\Theta$ 36.5, $\Psi$ -4.0, 2$\nu$ 99.5; mean value of 15 An-determinations 88.2).

In the anorthite range, however, conditions are different: Here the numerous new data on the optical orientation of feldspars from the moon do not agree neither with the meagre data for plutonic anorthites from the earth, nor with the more ample evidence on terrestrial feldspars from very young volcanoes (Vesuvius, Etna, Soufrière, Japan). We are struck especially by the fact that lunar anorthites of almost one and the same composition (An 90-92) can show such different and reproducably distinct optical orientations. This indicates that basic plagioclases of similar chemical composition but of different structural state can be present in one and the same lunar rock. The rapidly increasing evidence on lunar anorthites may change our point of view. We are urged to augment our modest knowledge on anorthites from the earth. The Eulerian angles of anorthites, collected 1971 in Iceland and analysed by the same operators, follow very well the high curves for terrestrial material.
Research Project - Carmichael and Wenk

LUNAR PLAGIOCLASE (A MINERALOGICAL STUDY); H. R. Wenk, M. Ulbrich, Space Sciences Lab. and Dept. of Geology & Geophysics, and W. Muller, Dept. of Materials Science, U. C., Berkeley, Calif. 94720.

It was the purpose of this study to describe the properties of lunar plagioclase and to investigate - by comparison with appropriate terrestrial feldspars - if the optics and structure of anorthite is indicative for the thermal history of the crystals.

Techniques which have been used include determination of optical properties on the Universal stage, single crystal X-ray photography with the precession method, chemical microprobe analyses and 650KV transmission electron microscopy. Most of our data were obtained on the only two available thin sections 14310 (basalt) and 14319 (breccia).

Petrography. 14310 is a basalt consisting mainly of clear, inclusion-free crystals of calcic plagioclase (An 85-95), clinopyroxene (2V = 12-20°), orthopyroxene (2V = 72°) and opaque minerals. The texture is subophitic. Plagioclase varies greatly in size. The largest crystals in the thin section are 2 mm long, but most of them range between 0.3 and 0.5 mm. Chemical analyses indicate that there are two groups of plagioclase which, however, are not apparent by grain size, shape or texture. An average of microprobe analyses on many crystals gives for one group a composition An 86.9, Or 1.44 and for the other group An 94.0, Or 0.42. Twin laws which have been identified on the U-stage are Albite, Carlsbad, Albite-Carlsbad (common), Pericline (less common), Baveno-(one crystal only). No Ala B twins have been found (E. Wenk, 1972). Quite common is cruciform intergrowth. There are several cases of pseudotwins where indicatrixes are related by an irrational twin axis. A fragment from 14162 coarse fines has very similar properties to the 14310 basalt and we assume that it is a fragment of the same rock. 14319 is a breccia with a large amount of plagioclase fragments. Plagioclase is heterogeneous. Many crystals are twinned, some are undeformed, some are fractured, some have patchy extinction and bent lamellae. One crystal of plagioclase shows very thin platelets of an opaque mineral on (010). Other components of the breccia are ortho- and clinopyroxene and perovskite. Apart from crystal fragments the breccia contains many lithic fragments (gabbroic anorthositic rocks, basalt similar to 14310 and an older breccia). Brown glass inclusions show beginning crystallization. Noteworthy are spherical aggregates of pyroxene with radial crystallites resembling meteoritic chondrules. Many lithic and crystallite fragments in the breccia are rounded, others, however, have sharp corners. 12021 is a basalt of ophitic texture. Some of the large plagioclase crystals are skeletal with pyroxene and opaque inclusions. Calcic plagioclase from the terrestrial rocks and from eucrite meteorites with similar mineralogical composition have been analyzed and compared with lunar crystals.

U-stage analysis of plagioclase. Euler angles are used to describe the orientation of the indicatrix in the crystal. The results are listed in Table
In Fig. 1 they are compared with data for anorthite available in the literature. Looking at all anorthites $\phi$ and $\psi$ appear to be slightly larger for plutonic than for volcanic feldspars and the rather large angles in lunar and meteoritic plagioclase agree with plutonic optics (Wenk & Nord, 1971; Ulbrich, 1971).

Lattice constants and structure. Lattice constants for four crystals have been measured on precession photographs. The variations are within the standard deviation and no conclusions can be drawn. Except for crystals from 10017 all analysed crystals showed strong $b$ reflections and diffuse $c$ reflections streaking along $b^\circ$. If the diffuseness of $c$-reflections reflects the thermal history, then these lunar anorthites have all been cooled rapidly, and comparing them with terrestrial and eucrite anorthites, the structure of lunar plagioclase is similar to that of plagioclase from volcanic rocks. All metamorphic, plutonic and meteoritic feldspars showed much stronger and sharper $c$ reflections (cf. Wenk & Nord, 1971). Rapid cooling for rock 14310 has also been suggested from order-disorder studies in pyroxene (Ghose, 1972). Of special interest is a crystal from 10017 which shows $b$-split reflections and sharp $b$ reflections in the same crystals.

Electron microscopy. In 14310 selected area diffraction patterns show a sharp $b$, diffuse and elongated $c$ reflections. Diffuse $d$ reflections may be due to double diffraction. Dark field images using $b$-reflections revealed smoothly curved antiphase boundaries. The size of these domains (500-1000 Å) is distinctly smaller than those found by Christie et al. (1971) in rock 10029. $c$-domains, which were imaged in comparative terrestrial anorthites (cf. Muller, Wenk & Thomas, 1972) have not been observed. Diffraction contrast similar to peristerite has been observed in a few crystals. So far it cannot be explained.

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| Euler I angles and chemical composition of plagioclase from lunar rocks and eucrite meteorites. The accuracy of the U-stage measurements is 1°. |
|---|---|---|---|---|---|---|---|---|
| LUNAR ROCKS | Twin Laws | $2\beta$ | $\phi$ | $\psi$ | $\theta$ | An | Or | Ab |
| 12021, 118C | Ab, Ca | 77° | 25° | -6 | 36 | 92.9 | 0.34 | 6.7 |
| basalt frag. | Ab-Ca; Pe | 25° | -2 | 38 | 85.3 | 1.72 | 12.9 |
| 14310, 23 & 95 | Ab, Ca | 25° | -3 | 38 | 87.6 | 1.22 | 11.0 |
| Plag.-pyroxene basalt | Per. (rare) | 81° | 23° | -5° | 38 | 87.6 | 1.38 | 11.2 |
| (only one) | 78° | 24° | -6 | 34 | 94.3 | 0.39 | 5.2 |
| 14319, 6 | Ab, Ca | 23° | -6 | 38 | 93.4 | 0.45 | 6.1 |
| Breccia | | | | | | | |
| A) plag.-pyr. | Ab-Ca | 79° | 22° | -4 | 38 | 93.8 | 0.37 | 5.7 |
| aggr. (anor.) | Pe | 78° | 21° | -7 | 37 | 94.6 | 0.54 | 4.8 |
| B) plag.-pyr. | Ab, Ca | - | -6 | 37 | 92.6 | 0.63 | 6.6 |
| aggr. (gab.) | Ab-Ca, Pe | 20 | -6 | 37 | 92.6 | 0.63 | 6.6 |
| C) plag. aggregate | Ab, Ca | - | -6 | 37 | 92.6 | 0.63 | 6.6 |
| Ab-Ca | Ab-Ca | 26 | -1 | 34 | 84.2 | 1.56 | 14.2 |
LUNAR PLAGIOCLASE (A MINERALOGICAL STUDY)
H. R. Wenk

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<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>Ab, Ca, Pe</th>
<th>Ca, Ab-Ca</th>
<th>Ab-Ca, Ca</th>
<th>Ab-Ca, Pe</th>
<th>Ab-Ca, Ca, Pe</th>
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<td>Ab</td>
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<td>0.39</td>
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<td>0.44</td>
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<td>Pe (rate)</td>
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<td>375 - 625</td>
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**EUCRITE METEORITES**

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<th>94.83</th>
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<td>23 - 6</td>
<td>38</td>
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<td>Ibitira</td>
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<td>37</td>
<td>95.30</td>
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<td>Ab, Ca</td>
<td>28 - 6</td>
<td>42</td>
<td>88.29</td>
<td>0.45</td>
<td>11.25</td>
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</table>

Ab - Albite, Ca - Carlsbad, Pe - Pericline

**VARIATIONS OF EULER ANGLES Φ and Ψ IN BASIC PLAGIOCLASE**

Figure 1. Orientation of the optical indicatrix in calcic plagioclase.
Lunar photogeologic mapping has shown that the multi-ringed circular basins are the dominant geologic features of the Moon's near side (Wilhelms and McCauley, 1971). These basins are the large members (larger than 250 km) of a size series of circular lunar features or craters believed to be produced by impact. Circular features in this series less than about 250 km diameter have one dominant raised encompassing ring or crater rim; larger circular structures are surrounded by two or more raised, roughly circular structures or rings.

In order to sample representatively, and thereby better understand lunar basins, one of the objectives of Apollo missions 14 and 15 was to investigate different parts of the most accessible, largest, and second youngest basin - Imbrium. Apollo 14 sampled radially textured basin ejecta (Fra Mauro Formation) at a point 500 km outside the main ring of mountains bordering the basin just north of the partly mantled pre-Imbrian crater Fra Mauro. Numerous studies of
naturally terrestrial impact craters and experimentally produced impact and explosion craters show that the outer radial ejecta consists of thoroughly broken pre-basin bed rock most of which was excavated from deep within the crater and deposited ballistically along low trajectories upon pre-basin terrain. The rugged raised rim, on the other hand, consists mainly of pre-crater rock, fractured and uplifted, but not grossly disaggregated and redistributed, and overlain by only a thin layer of ejecta. The bulk of the material returned by Apollo 14 should consist of complex finely fragmental debris produced directly during the Imbrium event.

Apollo 15 sampled materials from the foot of the steep inward facing scarp of the Apennine Mountains, which constitute the main mountain ring surrounding the Imbrium basin. Included in the Apollo 15 samples should be ejecta from the nearby Serenitatis basin and an indistinct basin, centered near the crater Copernicus, whose vaguely expressed rings trend toward Imbrium, but are truncated by it. In addition, volcanic rocks like those observed in all large lunar depressions as light-colored plains materials probably occupied parts of these old basins and these pre-mare but post-basin rocks may also be included in the samples from both missions.

Apollo 15 also sampled basalts of Palus Putredinus in which Hadley Rille is incised. These basalts flood the Imbrium basin, but clear stratigraphic relations in the Archimedes region west of the site show that they were emplaced in a considerable time after the basin formed and are not the direct product of the impact that formed the basin. These rocks should be similar in character to those returned by Apollo 11 and 12.
A complex record of multiple brecciation is, therefore, expected in both sample suites - some pre-Imbrian volcanic samples that have undergone only Imbrium-related brecciation and shock metamorphism may be present and if they can be identified, they should help establish the age of the Imbrium event. The samples from the base of the Apennines may be of deep-seated origin and show less fragmentation and shock metamorphism than the more far flung Apollo 14 materials. They may also better retain evidence of earlier basin-forming events. Some ejecta may have covered the uplifted Apennine ring and have found its way into the debris sampled at the base of the scarp, thereby making distinctions between specific basin-forming events difficult, if not impossible.

References
PETROLOGY OF THE FRA MAURO FORMATION AT THE APOLLO 14 LANDING SITE.

The kinds and distributions of simple clasts in the Fra Mauro subunits
have been discussed in the previous paper (Jackson and Wilshire, this volume).
However, virtually every rock examined contains some compound clasts and in
many, such clasts are abundant (LSPET, 1971). The first column of Table 1
gives the sample number and the main types of simple, noncomposite clasts
observed; the following columns list the rock types of compound clasts (the
symbols are those of Jackson and Wilshire, this volume).

Table 1. Clasts and Matrix Types in Compound Clasts

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Main types of simple clasts</th>
<th>Host Clast</th>
<th>Clasts in (1)</th>
<th>Clasts in (2)</th>
<th>Clasts in (3)</th>
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<tr>
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<td>D1, D2</td>
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<td>D2</td>
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It is interesting and significant that older generations of clasts are of metaclastic rocks, showing clearly a history of multiple brecciation and metamorphism in each of the Fra Mauro subunits. The matrices of the breccias range from glassy to crystalline (Warner, 1971); the more coherent rocks are generally the more crystalline ones. In addition, partial to complete re-equilibration reactions have taken place between matrices and some clasts in a number of the rocks, particularly in F4 rocks and to a lesser extent in F2 rocks. Olivine reacted to form pyroxene and opaque oxide, plagioclase and pyroxenes are compositionally zoned, translucent spinels developed colorless mineral coronas, and quartz reacted to form pyroxene. Similar reactions are observed within clasts in all types of fragmental rocks whether glassy or crystalline. These relations indicate that more than one generation of re-equilibration reactions is represented in the Fra Mauro rocks, and that older breccias had thermal histories similar to that of the Fra Mauro Formation itself. The distribution of rocks at the Apollo 14 site (Fig. 1; Swann and others, 1971a) suggests that the Fra Mauro Formation is stratified, that dark clast-dominant rocks (F3, F4) are deepest, and that light clast-dominant rocks (F2) are shallowest.

Poorly consolidated, glass-rich rocks (F1) are considered to be weakly lithified regolith, derived either from disaggregation of F2 type breccias or from originally unconsolidated material at the top of the Fra Mauro deposit. This stratigraphic sequence is consistent with the comparative abundance of glass and the weakly developed re-equilibration reactions in F2 rocks that might be expected to have chilled somewhat more rapidly near the surface, and with the dominance of light colored hornfels among the lithic fragments of soils (Carr and Meyer, this volume). It is presumed that thermal blanketing of the lower, dark clast-dominant breccias allowed time for reactions to proceed farther and for annealing and devitrification of glass. Thus, while the Fra Mauro Formation appears to be composed of three or four distinct units based on clast populations, its internal thermal metamorphism is gradational.

The compound clasts and great variety of lithic fragments indicate that the pre-Imbrium terrane, presumed to have existed prior to about 3.9 to 4.2 b.y. ago, was already exceedingly complex. However, the similarity of older breccias to the Fra Mauro breccias themselves indicates that processes, involving extrusion of basalt, intrusion and differentiation of basaltic magma, impact brecciation, and metamorphism prior to 3.9 to 4.2 b.y. were not unlike later ones. The complex clasts of the Fra Mauro breccias therefore reveal a long history of impact and ejection that probably record a half billion years of planetary accretion, of which the Imbrium was one of the last large events.
Fig. 1. Map showing distribution of fragmental rock types in the Apollo 14 traverse area. Cc4, continuous ejecta blanket of Cone crater; lined areas, bouldery ray deposits of Cone crater; Is, smooth terrain material of the Fra Mauro Formation; Ifr, ridge material of the Fra Mauro Formation. After Swann and others, 1971.

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Jackson, E.D. and Wilshire, H.G., this volume, Classification of the samples returned from the Apollo 14 landing site.


ON LUNAR METALLIC PARTICLES AND THEIR CONTRIBUTION TO THE TRACE ELEMENT CONTENT OF THE APOLLO 14 AND 15 SOILS.

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Lunar soil 14163 contains 0.5% of metallic iron, i.e. three times as much as soil 12001 (0.12%). We separated this metal (grain size < 100 microns) and studied its main and trace element chemistry. Fig. 1 shows the Ni and Co content (microprobe measurements) of individual particles. Their composition falls into a rather narrow range which is typical meteoritic: ~0.5% Co, 5 to 10% Ni. Soil 15601 contains 0.35% metal, which shows quite a different Ni-Co distribution pattern (Fig. 1), apparently the meteoritic component is much smaller here. Included are measurements on metal from soil samples 10084 and 12001 for comparison. These particles also show a cluster around meteoritic Ni-Co values, but they contain a higher proportion of grains similar to those from the local lunar rocks. It is noteworthy that in all four soil samples only two grains were found having a Ni-content of 30 to 50% typical of chondritic taenite.

Table 1 gives the content of several trace elements measured in a 3 mg-sample of metal from 14163. In column 3 and 4 it is shown that the elements Ge, As, Pd, Ir, and Au agree quite well with a carbonaceous chondrite abundance pattern which is normalized to Ni, whereas no such agreement is found for the metal from rock 14321 or from fines 15601. The same is true if one compares these elements with a normal chondrite abundance, except for Ge which is off by a factor 3. This indicates that the meteoritic component in soil 14163 seems to be derived from an unfractonated or carbonaceous meteorite material, in agreement with the conclusions reached by Laul et al. (1971) and Baedecker et al. (1971) on Apollo XI and XII bulk fines.

This meteoritic component, however, is mainly contained in the metallic grains: Table 1 gives a calculation of the contribution of the metal particles to the bulk chemistry of the fines 14163. Apparently the incoming meteoritic material was equilibrated with the lunar soil, i.e. oxidized iron reduced to metal which took up the siderophile elements.
In this way also the high amount of W in the metal can be derived from the lunar silicates. This would imply that W did not behave as a siderophile element in the early history of the moon, when these elements were depleted from the lunar surface rocks. It enters the metal phase now, however, probably because of more reducing conditions.


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<tr>
<th>Metal from rock 14321</th>
<th>Metal from fines 14163</th>
<th>Contribution of 0.5% metal to bulk fines</th>
<th>Metal from rock 15601</th>
<th>Contribution of 0.35% metal to bulk fines</th>
<th>C1-chondrite</th>
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<tr>
<td>Co 1800</td>
<td>5500</td>
<td>65%</td>
<td>10350</td>
<td>71%</td>
<td>505</td>
<td>4100</td>
<td>950</td>
<td>2120</td>
<td></td>
</tr>
<tr>
<td>Cu 134</td>
<td>220</td>
<td>7%</td>
<td>863</td>
<td>37%</td>
<td>200</td>
<td>290</td>
<td>35</td>
<td>165</td>
<td></td>
</tr>
<tr>
<td>Ga 8.6</td>
<td>14</td>
<td>1%</td>
<td>9</td>
<td>0.9%</td>
<td>10.0</td>
<td>17</td>
<td>2.0</td>
<td>1.56</td>
<td></td>
</tr>
<tr>
<td>Ge n.d.</td>
<td>138</td>
<td>117%</td>
<td>n.d.</td>
<td>-</td>
<td>2.7</td>
<td>-</td>
<td>19.5</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>As n.d.</td>
<td>11</td>
<td>63%</td>
<td>n.d.</td>
<td>-</td>
<td>1.6</td>
<td>-</td>
<td>1.5</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Pd n.d.</td>
<td>2.8</td>
<td>50%</td>
<td>n.d.</td>
<td>-</td>
<td>0.32</td>
<td>-</td>
<td>0.27</td>
<td>-</td>
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</tr>
<tr>
<td>W 46</td>
<td>223</td>
<td>56%</td>
<td>25</td>
<td>31%</td>
<td>0.035</td>
<td>34</td>
<td>12</td>
<td>1.65</td>
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<tr>
<td>Ir 1.3</td>
<td>2.0</td>
<td>53%</td>
<td>n.d.</td>
<td>-</td>
<td>0.094</td>
<td>0.9</td>
<td>0.10</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Au 0.20</td>
<td>1.1</td>
<td>92%</td>
<td>0.35</td>
<td>55%</td>
<td>0.044</td>
<td>0.14</td>
<td>0.057</td>
<td>0.021</td>
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</tr>
</tbody>
</table>

Table 1: Composition of metal from an igneous fragment of rock 14321 and of metal from fines 14163 and 15601. In column 3 and 5 the contribution of the metal contained in the fines to the bulk chemical composition is given.
Fig. 1 Co versus Ni for metal particles from fines 10084, 12001, 14163 and 15601 and from rock 12053, respectively.
The Apennine front, bordering the Imbrium Basin on the southwest, largely comprises two major mountains at the Apollo 15 site--Mount Hadley and Hadley Delta. These two mountains rise steeply 3-5 km above the mare surface. According to pre-mission interpretations (Ref. 1), the front massifs are considered to consist of impact breccias from the Serenitatis basin that overlie a complex of impact breccias from still older basins and craters. Faulting related to the Imbrium impact occurred along lines both radial and concentric to the Imbrium Basin and produced the present-day arc of block faulted mountains.

Surface material on the front at stations 2, 6, 6a, and 7 consists of relatively fine-grained cratered regolith that probably contains debris derived from deeper layers and from the nearby mare, as well as some exotic material from distant meteorite impacts. It may include not only pre-Imbrian massif material, but also Imbrium ejecta, either deposited directly in the sampled areas by the Imbrium impact or subsequently transported downslope by mass wasting.

About 80 percent of the fragments collected are breccias. The remainder are crystalline rocks including some of the distinctive mare type basalts, and glass fragments. The breccias contain variable amounts of feldspathic and basaltic clasts and mineral grains, and glass fragments. The indurated breccias, unlike the Apollo 14 breccias, are notable for the scarcity of clasts of older breccias. They may represent materials from relatively deep levels, either ejecta derived from some large craters and basins that excavated bedrock prior to the Imbrium impact or bedrock brecciated and uplifted by the Imbrium event itself.

Well developed systems of lineaments, suggestive of fractures and/or compositional layers were observed by the crew and recorded by surface and orbital photographs of Mount Hadley and Hadley Delta. Previous orbital and surface photographs have also shown sets of lineaments with preferred orientations--predominantly northwest, north, and northeast. This so-called lunar grid has been observed under very restricted lighting conditions, generally low sun with illumination from either the east or the west, which suggests that some of the lineaments may be illusions created by low angle illumination of randomly irregular surfaces. Experiments with small scale models (Howard and Larsen, 1972) show that oblique illumination on randomly irregular surfaces produces systematic sets of apparent lineaments that resemble some of those seen at the Hadley site.
Statistical analyses of lineaments recorded from orbit over a range of sun positions at the Hadley site show similar predominant trends on slopes with different orientations. On most of the studied slopes, major north and northeast lineament trends intersect in obtuse angles that are approximately bisected by the sun line. The lineaments have no consistent relationship to the slope such as gravitational effects might be expected to produce, nor do they vary systematically as slope attitudes vary relative to the sun line. Apparently the lineaments represent either real directional features of the lunar surface, or, if they are illusions of oblique lighting on an irregular surface, they are sensitive only to the lighting direction and not to the orientation of the slope on which they occur. If real, they may represent closely spaced repetitive geologic structures, such as layering or regional fracture patterns projected through the regolith to the surface.

Locally, lineament trends are related to distinct topographic features and are less equivocally interpreted as geologic structures. At Silver Spur prominent northeast and northwest trending lineaments may be directly related to steeply dipping geologic structures with distinctive topographic expression.

If either of the prominent north or northeast trends represents bedrock structure, that structure is more likely to be regional fracturing propagated through the regolith than to be a surface expression of compositional layering. The extensive areal distribution of the pattern and its azimuthal consistency regardless of slope orientation would suggest, if real, that it represents a regional set of nearly vertical conjugate fractures.

References and notes


3. This work done under NASA contract number T-65253-G.
Preliminary observations of the glass portion of a 2 mm wide glass-filled fracture of breccia rock sample #14306.50 indicate the presence of numerous metallic spherules ranging from about 30 Å to 100 μm in size. Metallic inclusions in lunar glasses have been noted previously by several authors in both the Apollo 11(1) and 12(2) mission samples. The inclusions in the 0.1 to 1 μm range are abundant and evenly distributed throughout the glass phase while the particles at the larger and smaller ends of the size range are more sparsely distributed. X-ray energy dispersion, X-ray diffraction, electron microprobe and magnetic susceptibility analyses suggest that the spherical bodies are mainly iron-nickel with segregated zones of iron-nickel sulfide (troilite) and iron-nickel phosphide (schreibersite) although some of the inclusions appear to be nearly all sulfide and/or phosphide. As many as five magnetic phases may be present and there is evidence of single crystal formation. The magnetic analysis indicates magnetic centers as small as 40 to 50 Å.

The glass-breccia interface was sharply defined with the transition band being about 150 μm wide. The transition zone contained a few irregular shaped metallic inclusions, ample evidence of flow lines and moderate numbers of breccia grains with rounded edges. X-ray energy dispersion analysis of breccia grains in the vicinity of the glass-breccia interface indicate at least four phases of different chemical composition. No indication of crystallization from the glass was found at the interface or in the main body of the glass region. Electron microprobe isometric concentration maps of glass-breccia grain interfaces indicated no diffusion of breccia constituents into the glass within the resolution of the microprobe. These observations suggest rapid flow and cooling of the glass melt.
Quantitative chemical analysis of two general areas of glass, one near the center of the 2 mm glassy vein and the other near the glass-breccia interface, with the electron microprobe indicated some variation in the iron concentration in the two zones. The average concentration of the glass constituents in the glass phase expressed in weight percent oxides are: SiO$_2$, 46.63 ± 0.24; FeO, 11.60 ± 0.47; Na$_2$O, 0.72 ± 0.04; TiO$_2$, 1.32 ± 0.03; CaO, 8.67 ± 0.35; MgO, 13.53 ± 0.75; Al$_2$O$_3$, 15.87 ± 0.36; K$_2$O, 0.61 ± 0.04; Total, 98.95.

A synthetic glass batch of sample 14306.50 was melted in a resistance fired furnace in an inert (N$_2$) atmosphere at 1400°C for two hours. The batch melted rapidly and the resulting glass was free of solid and gaseous inclusions. The following physical properties were determined: softening point, 823°C; annealing point, 669°C; strain point, 633°C; expansion, 53.9 x 10$^{-7}$/°C (average expansion 0 - 300°C); density, 2.767 gm/cc; liquidus, 1205°C, phases pigeonite and magnetite. Gradient furnace and microscope furnace tests on the synthetic batch and glass also indicated both melt rapidly and at relatively low temperatures, as low as 1035°C depending upon exposure time at temperature.


Inclusions and Interface Relationships
J. F. Wosinski

Fig. 1. Inclusion and Interface Detail. (A) Scanning electron micrograph of iron-nickel phosphide particle with troilite inclusions and rim. (B) Scanning electron micrograph of metallic particles, one with pronounced nodule. (C) Electron microprobe isometric concentration maps for aluminum, calcium, and magnesium at a glass-breccia grain interface. Diffusion interface very sharp. Resolution 1 to 2 μm. (D) Magnetization vs. temperature curve showing four magnetic phases and indicating a fifth with Curie temperature > 670°C.
Examination of a limited number of Apollo 14 samples by nondestructive multi-dimensional gamma ray spectrometry for natural radioactivities of the thorium and uranium decay chains showed the rocks and soils to be very similar to one another in the concentrations of these elements which are important in lunar history. Whereas the thorium and uranium concentration in Apollo 11 soil appeared to be an average of thorium and uranium in Apollo 11 rocks (and therefore indicated the soil derived from similar rocks), the Apollo 12 soil required some exotic component to account for thorium and uranium levels in the soil that were approximately eight times those in the rocks. The Apollo 14 rocks and soil appear to fit neither previous lunar case but rather would seem to be identical material as far as thorium and uranium concentrations on a macroscopic scale are concerned. For instance, thorium and uranium concentrations for the bulk soil sample (14163) collected at the lunar module and for sawdust derived from cutting a large rock (14321) collected high on the rim of Cone Crater are, respectively, 13.5 ± 0.2 ppm Th/3.8 ± 0.1 ppm U and 11.3 ± 0.2 ppm Th/3.0 ± 0.1 ppm U. Thus the heterogeneity noted by the crew and LSPET* in clastic rocks would not seem to apply to thorium and uranium. One or two exceptions are notable in the LSPET* gamma ray data, primarily a sample from White Rock (14082) collected at the same station as 14321 (Station C1) on the Cone Crater rim. It has thorium and uranium levels roughly a third of those of 14321. The difference is probably due to the obvious inhomogeneity of White Rock with respect to the more abundant surrounding rocks with many dark clasts such as 14321. The possibility exists that White Rock is unique and may be from material underlying the Fra Mauro Formation.

Cosmogenic radioactivities $^{26}$Al and $^{22}$Na were measured in the same samples by the same technique. The $^{26}$Al and $^{22}$Na concentrations, 79 ± 5 dpm/kg and 40 ± 5 dpm/kg, respectively, in the sawdust from 14321 exhibits the 2:1 ratio found for $^{26}$Al and $^{22}$Na in previous sets of Apollo samples. The low concentrations compared to other clastic rocks reported by LSPET* would be due to the fact the saw cuts intercepted relatively little surface so that solar generated activity would be minimized. Comparison with other gamma ray work on 14321 will be useful since the sawdust should be a reasonably good sample of the whole rock. $^{26}$Al concentration in the bulk fines (14163,0) is 79 ± 6 dpm/kg or a factor of two or more less than Apollo 11 or 12 fines or other fines reported by LSPET* for Apollo 14. A deeper average sampling depth could be the explanation; comparison to 14321 with an identical $^{26}$Al concentration would indicate an average sampling depth of ~10 cm.
Radionuclides at Fra Mauro
Robert C. Wrigley

The thorium and uranium concentrations reported above are very high with respect to solar and chondritic concentrations and are still high in comparison to Apollo 11 material and most of Apollo 12 material. Thus, they represent highly fractionated material much as granites on the earth do. Due to the homogeneity in thorium and uranium noted above for the Apollo 14 site, these high concentrations may be taken as representative of, at least, this area of the Fra Mauro Formation. This Formation is interpreted to be ejecta from the Imbrium impact event (550 km north) so that it represents material down to tens of kilometers deep. Unless the area of the Formation visited by Apollo 14 is unrepresentative of Imbrium ejecta with respect to thorium and uranium, the pre-impact crust at Imbrium contained highly fractionated thorium and uranium to depths of tens of kilometers, a depth of high thorium and uranium only approached on the earth. It is possible the Fra Mauro Formation at Apollo 14 derived from near-surface deposits but the similarity noted by LSPET* between Apollo 14 samples and the exotic components from Apollo 12 would rather indicate wide distribution of similar material.

*LSPET, Preliminary Examination of Lunar Samples from Apollo 14, Science 173, 681 (1971).
RADON EMANATION FROM APOLLO 11, 12, AND 14 FINES, A. Yaniv* and D. Heymann, Departments of Geology and Space Science, Rice University, Houston, Texas, 77001. *On leave of absence from Department of Physics and Astronomy, Tel Aviv University, Ramat Aviv, Israel.

The formation of α-radioactive deposits on the lunar surface involves: 1) the emanation of Rn out of regolith particles, 2) the diffusion of Rn through the void spaces of the regolith into the lunar atmosphere, and 3) the implantation of polonium into the top of the regolith. We report here on the emanation of Rn out of lunar fines. Since our experimental techniques involve a novel approach to Rn measurements we will describe these briefly (Fig. 1). The samples were sealed under vacuum (10⁻² mm Hg) in a Pyrex container of about 10 cm³ volume. The container was directly connected with the detector, a counting bottle of about 100 cm³ volume, coated on the inside with ZnS. The lightflashes from the counter were detected with a photomultiplier. Temperatures during the experiments ranged from 27-35°C. Radon diffused from the container into the counting bottle, hence we could follow the buildup of Rn²²² and Rn²²⁰, and by closing the stopcock their decay. The background counting rate of the system was monitored extensively before and after each measurement. For long counting periods (one week) we determined background rates of 4.1 ± 0.2 cph; the error represents counting statistics only. However, we observed nonrandom fluctuations during successive 8 hour periods; i.e. drifts in background upward or downward. The smallest measured background rate (8 hr period) was 2.3 ± 0.7 cph; the largest 5.8 ± 1.0 cph. However, since more than 70% of the individual 8 hr background rates fell within 4 ± 1 cph we have adopted this value to subtract it from counting rates obtained from samples.

Apollo 14 The only clear-cut evidence for Rn emanation was obtained with the combined samples 14003 (1.9003 g) and 14259 (1.7579 g). In this case Rn atoms escaped from 3 cm thickness of fines. The activity (counting rate-background) reached a value of 11.6 ± 1.2 cph after one week. When the stopcock was closed the activity decreased more rapidly than one would have expected from Rn²²² (t½ = 3.825 days) alone, hence we concluded
that Rn$^{220}$ ($t_1 = 54.5$ sec) was also present. The analysis of the decay curve in terms of Rn$^{222}$ and Rn$^{220}$ and their progeny was not entirely satisfactory, probably because of slight day-to-day variations in the background rates. The total Rn activity just before the closing of the stopcock was $11.2 \pm 1.2$ cph. The activity during the first 8 hours decreased to $(65 \pm 10)$% of the initial value, which corresponds to Rn$^{222}$ (Rn$^{222}$ + Rn$^{220}$) activity ratio $= 0.6$ before the closing of the stopcock. In the following 16 hours the activity dropped to $(37 \pm 6)$% which corresponds to an initial Rn$^{222}$ fraction of 0.35. In the final 24 hours the activity dropped to $(20 \pm 5)$%, indicating an initial Rn$^{222}$ fraction of only 0.25. From this we calculate that the Rn$^{222}$ equilibrium activity in the counter was at least $3.7 \pm 0.4$ cph, at most $8.9 \pm 0.9$ cph. The corresponding numbers for Rn$^{220}$ are $4.5 \pm 0.5$ cph and $8.4 \pm 0.9$ cph, but these are obviously minimum values because a substantial portion of Rn$^{220}$ atoms could have decayed in the sample, or on their way to the counter.

LSPET has reported U contents of about 4 ppm and Th contents of about 13.5 ppm in five Apollo 14 fines, including 14259. The over-all efficiency for counting Rn$^{222}$ was 71% (this figure includes counter efficiency and counter/(counter + reservoir) volume ratio). Using these numbers we calculate a minimum emanation-to-production ratio (EPR) for Rn$^{222}$ of about $2 \times 10^{-3}$; a maximum EPR of about $5 \times 10^{-3}$. The corresponding numbers for Rn$^{220}$ are: $3 \times 10^{-3}$ and $5 \times 10^{-3}$ respectively.

Apollo 11 Two experiments were carried out with 6.7 g of 10084. There was no detectable increase in counting rate. The U content of 10084 is $\sim 0.5$ ppm. Hence the EPR of Rn$^{222}$ for this sample is $< 1 \times 10^{-3}$, assuming that 1 cph above background could have been detected.

Apollo 12 One experiment was carried out with 0.889 g of 12070. No significant activity was detected above background. The U content of this sample is about 1.6 ppm. From this we calculate a Rn$^{222}$ ERP of $< 3 \times 10^{-3}$.

Our result for 12070 disagrees with a previous report by Adams, Barretto, Clark, and Duval (Nature, 231 (1971) 174) who have reported an EPR value of about 0.4. Their experiment; however, was carried out in air, ours in a modest vacuum.

The small rate of radon emanation from lunar fines agrees with the observation by Lindstrom, Evans, Finkel, and Arnold (Earth Planet Sci. Lett. 11 (1971) 254) who found that Pb$^{210}$ in surface samples of lunar soil and rock is in equilibrium with the parent uranium to within 5%. These authors have suggested
that the low rate of emanation could be due to the implantation of recoiling Rn atoms in grain surfaces under the conditions prevailing on the lunar surface (no moisture, low gas pressure). However, an alternative explanation is that much of the U and Th is contained in minerals which are known or suspected to be poor Rn emanators even in the terrestrial environment. The prime candidate among the known minerals is zircon (P. Barretto, private communication). We suspect that the U, Th rich lunar mineral zircolite is another good candidate.

The emanation rate of the Apollo 14 samples means that all the Rn$^{222}$ made available for diffusion in a layer of several meters of regolith down from the surface should escape into the lunar atmosphere in order to account for the $\alpha$-activity from Rn$^{222}$ and its progeny observed by Turkevich et al (Science 167 (1970) 1722) in Mare Tranquillitatis (also assuming an average U content of the lunar regolith of a few ppm). However, a more realistic estimation of the rate of Rn escape into the atmosphere must await measurements of the emanation rates for a wide range of temperatures and different types of soil.

![Diagram of measurement setup](image)
COSMONUCLIDES IN LUNAR ROCKS, Y. Yokoyama, R. Auger, R. Bibron, R. Chesselet, F. Guichard, C. Léger, H. Mabuchi, J. L. Reyss and J. Sato, Centre des Faibles Radioactivités, C.N.R.S., Gif-sur-Yvette, 91, France

Lunar samples bear fossil records of cosmic radiation. From studies of Apollo 11 samples, clear evidence was obtained for the long term bombardment by solar particles in the region of tens of MeV per nucleon (1). Studies of Apollo 12 samples (2) showed a similarity of the depth profile of 7.4 x 10^5 yr Al-26 to that of 2.6 yr Na-22. Thus the flux of the solar particles averaged over 10^6 years has been similar to that observed recently. The profile for 3.7 x 10^6 yr Mn-53 suggests a possibility of either an erosion rate of 0.2 mm per million years or the variation of the solar cosmic flux (2).

The aim of the work reported here comprises experimental measurement of the depth variation of radionuclides, theoretical calculations of the expected profiles and comparison of these results. Unfortunately we have received samples late (on 21 Sept. and 5 Nov. 71) and radiochemical analyses are yet in process. We report here therefore mainly the results of the theoretical calculation. The calculation of the expected production by solar particles is carried out as follows. We consider here only the interactions of the primary protons. The production by α-particles will be reported later. The spectrum of incident solar protons is assumed to decrease exponentially with their rigidity:

\[ N(R) \, dR = k \, e^{-R/R_0} \, dR \]  \hspace{1cm} (1)

where \( N(R) \, dR \) is the number of particles having rigidities between \( R \) and \( R + dR \), and \( R_0 \) is the characteristic slope (MV). By using the relation between the rigidity and the energy,

\[ R^2 = E^2 + 1876 \, E \]  \hspace{1cm} (2)

the rigidity spectrum is converted in the energy spectrum:

\[ N(E) \, dE = k \, (E + 938) \, e^{-R/E_0} \, dE \]  \hspace{1cm} (3)

where \( N(E) \) is the number of particles having energies between \( E \) and \( E + dE \) MeV, and 938 is the rest mass of proton in MeV. To calculate the depth variation of the proton flux, we have used a range-energy relation expressed by

\[ P = \frac{2.21}{S} \, E^{1.8} \]  \hspace{1cm} (4)

where \( P \) is the range (mg/cm^2) of a proton of energy \( E \) (MeV) and \( S \) is the relative mass stopping power which depends on the chemical composition of the material and is 1 for air, 0.91 for the rock 14305, and 0.88 for the rock 12002. Then the incident energy \( E \) of the particle coming with a zenith
angle $\Theta$ and having an energy $E_D$ at a given depth $D \text{ (mg/cm}^2\text{)}$ is calculated by

$$E = \frac{S.D.}{2.21 \cos \Theta + E_D^{1.8}} \frac{1}{1.8}$$

(5)

The incident flux is assumed to be isotropic. Incident angles are divided in ten parts, of which $\cos \Theta$s are comprised between 0 to 0.1, 0.1 to 0.2 and so on, in order that each part has equal solid angle. Then each part is represented by its mean $\cos \Theta$, namely $\cos \Theta = 0.05, 0.15$ and so on. The differential flux of particles having $E_D$ at a given depth $D$ is then obtained by integrating the flux of particles having initial energy $E$ ($E$ being a function of $\cos \Theta$) over all angles:

$$N(E_D) \, dE_D = \frac{k}{2\pi} \sum_{\cos \Theta} \frac{E + 938}{R} \cos \Theta \cdot \frac{R}{R_0} \left(\frac{E_D}{E}\right)^{0.8} \, dE_D$$

(6)

where $E$ should be determined by the equation (5) as the function of $E_D, D, \cos \Theta$; $R$ does by the equation (2) from $E$, and summation should be done for $\cos \Theta = 0.05, 0.15, \ldots$ to 0.95. The results are shown in figure 1.

The production rate is calculated in integrating $N(E_D) \, \mathcal{E}(E) \, dE_D$ for all energies, where $\mathcal{E}(E)$ is the excitation function of a given reaction. We have used the excitation functions selected from experiment, corrected for the adopted values of the monitor reactions, of decay schemes, and synthesized by Toballem and de Lassus St-Genies (3). The calculation of the expected production by galactic cosmic rays is carried out by a method described in detail in a separate paper. Figure 2 shows the calculated profiles for $^{26}$Al and observed value obtained preliminary by us with non-destructive $\gamma$-$\gamma$ coincidence spectrometry on the rock 14305 (Table 1). This rock was thought to be turned recently, from the photographic studies of its situation on the lunar surface and from the pits counting. This was confirmed by our results, showing more activities of $^{26}$Al at the bottom than at the top. In Fig. 3, a preliminary result of the measurement of $^{55}$Fe of the rock 14305 is shown. The result of our calculation for the rock 1206Z is in good agreement with that of Finkel et al. (2), figure 4.

Table 1 - Non-destructive $\gamma$-ray measurements of the rocks 14305 and 14302

<table>
<thead>
<tr>
<th>SAMPLES</th>
<th>Weight (g)</th>
<th>Th (ppm)</th>
<th>U (ppm)</th>
<th>Al-26 (dpm/Kg)</th>
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<tr>
<td>this work</td>
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<tr>
<td>14305 top</td>
<td>2.7</td>
<td>14.6 ± 0.9</td>
<td>3.8 ± 0.4</td>
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<tr>
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<td>13.3 ± 0.9</td>
<td>3.8 ± 0.5</td>
<td>160 ± 60</td>
</tr>
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<td>LSPET (4)</td>
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<td>14.3 ± 1.4</td>
<td>3.8 ± 0.6</td>
<td>85 ± 17</td>
</tr>
<tr>
<td>14302 *</td>
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<td></td>
</tr>
</tbody>
</table>

* 14302 was a part of 14305 at the lunar surface.
Cosmonuclides in lunar rocks
Y. Yokoyama et al.

1. See the references cited in the reference (2).
2. R. C. Finkel et al., Proceedings of the Apollo 12 Lunar Science Conference (1971)
3. J. Tobailem and C. H. de Lassus St-Genies, to be published
4. Apollo 14 preliminary science report, NASA SP-272 (1971)

Fig. 1. Differential flux of solar protons as a function of depth.

Fig. 2. Al-26 in rock 14305.

Fig. 3. Fe-55 in lunar rocks.

Fig. 4. Solar production of Al-26 in rock 12002.
40Ar-39Ar AGES OF APOLLO XIV AND XV SAMPLES, Derek York, W. John Kenyon and Roy J. Doyle, Department of Physics, University of Toronto, Toronto, Canada.

Introduction Nine controlled heating analyses have been performed on neutron-irradiated samples of Apollo XIV and XV materials. The results are summarized in the table and the figure.

Apollo XIV Analyses 14072,6 Controlled heating runs were carried out on two neutron-irradiated fragments of igneous rock 14072,6. Low 40Ar/39Ar ratios in the low temperature fractions indicate slight argon loss. The remaining 95 per cent of the gas released, however, has 40Ar/39Ar ratios corresponding to ages in the range 3.83-4.07 b.y. Evidently less than 5 per cent of the radiogenic argon has been lost from this sample and total gas ages of 3.94 and 4.02 b.y. were found. Both analyses appear to show a very slight drop in 40Ar/39Ar ratios in the mid-upper temperature range. It seems that this rock crystallized on the moon about 4.0 b.y. ago and its Ar/K ratio has not been seriously disturbed since. 14167,9 Two basaltic fragments of the 2-4 mm fines were analyzed. The 40Ar/39Ar evolution patterns are clearly reproduced between runs. The pattern, in fact, is apparently an exaggerated version of that of 14072,6. Low 40Ar/39Ar ratios are apparent in low temperature fractions while the largest volumes are evolved in the middle temperature range with equivalent ages of 3.84-3.97 b.y. There is again a diminution in 40Ar/39Ar ratios at higher temperatures, but in this case much more pronounced. As yet, there is no definitive explanation of this effect. We tentatively take the plateau ages of 3.91 and 3.95 b.y. as indicating the age of crystallization of these fragments.

14321 Three basaltic clasts from this breccia have been examined (184,1D;184,12B;184,17B). The results for 184,1D are shown in detail in the figure. The tendency for the 40Ar/39Ar ratios to sag in the mid-high temperature range is again apparent. Despite this, only about 5 per cent argon loss is apparent for this sample which has a plateau age of 3.90 b.y. and a total gas age of 3.84 b.y. The evolution spectra of the other two clasts 184,12B and 184,17B are very similar having plateau ages of 3.99 and 3.95 b.y. respectively. These fragments evidently crystallized 3.9-4.0 b.y. ago. They are also notably low in trapped
argon component. In summary, three different Apollo XIV source materials (basalt, coarse soil, breccia) all yield apparent $^{40}\text{Ar}/^{39}\text{Ar}$ ages in the range $3.95 \pm 0.15$ b.y. indicating a significant event occurred on the moon at this time. Its relationship to the excavation of the Imbrium basin is still not clear, however. Approximate $^{38}\text{Ar}$ exposure ages of between 20 and 30 m.y. probably date the formation of cone crater.

Apollo XV Analyses Two controlled heating analyses were performed on basalt 15,555,26. The sample appears to have lost a little over 20 per cent of its argon. 50 per cent of the $^{39}\text{Ar}$ is released in the temperature range 860-1175°C with essentially constant $^{40}\text{Ar}/^{39}\text{Ar}$ ratios (after corrections). Plateau ages thus found for the two runs are 3.29 and 3.32 b.y. This sample, therefore, probably crystallized on the moon $3.31 \pm 0.05$ b.y. ago. An approximate exposure age of 75 m.y. is found. Both analyses give evidence of high temperature diminution of $^{40}\text{Ar}/^{39}\text{Ar}$ ratios, this time in the last 5 per cent or so of the gas released.

<table>
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<tr>
<th>Sample</th>
<th>Plateau Age (b.y.)</th>
<th>Total Gas Age (b.y.)</th>
<th>Cosmic Ray Exposure Age (m.y.)</th>
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<tr>
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<tr>
<td>14167,9 2-4mm Fines, I</td>
<td>3.91</td>
<td>3.66</td>
<td>26</td>
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<tr>
<td>14167,9 2-4mm Fines, II</td>
<td>3.95</td>
<td>3.63</td>
<td>27</td>
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<tr>
<td>14072,6 Igneous Rock I</td>
<td>4.01</td>
<td>3.94</td>
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<tr>
<td>14072,6 Igneous Rock II</td>
<td>4.06</td>
<td>4.02</td>
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</tr>
<tr>
<td>14321,184,1D Basalt Clast</td>
<td>3.90</td>
<td>3.84</td>
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<td>14321,184,12B Basalt Clast</td>
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<tr>
<td>14321,184,17B Basalt Clast</td>
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<td>Apollo XV</td>
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<tr>
<td>15555,26,1 Basalt</td>
<td>3.29</td>
<td></td>
<td>81</td>
</tr>
<tr>
<td>15555,26,II Basalt</td>
<td>3.32</td>
<td>2.91</td>
<td>72</td>
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