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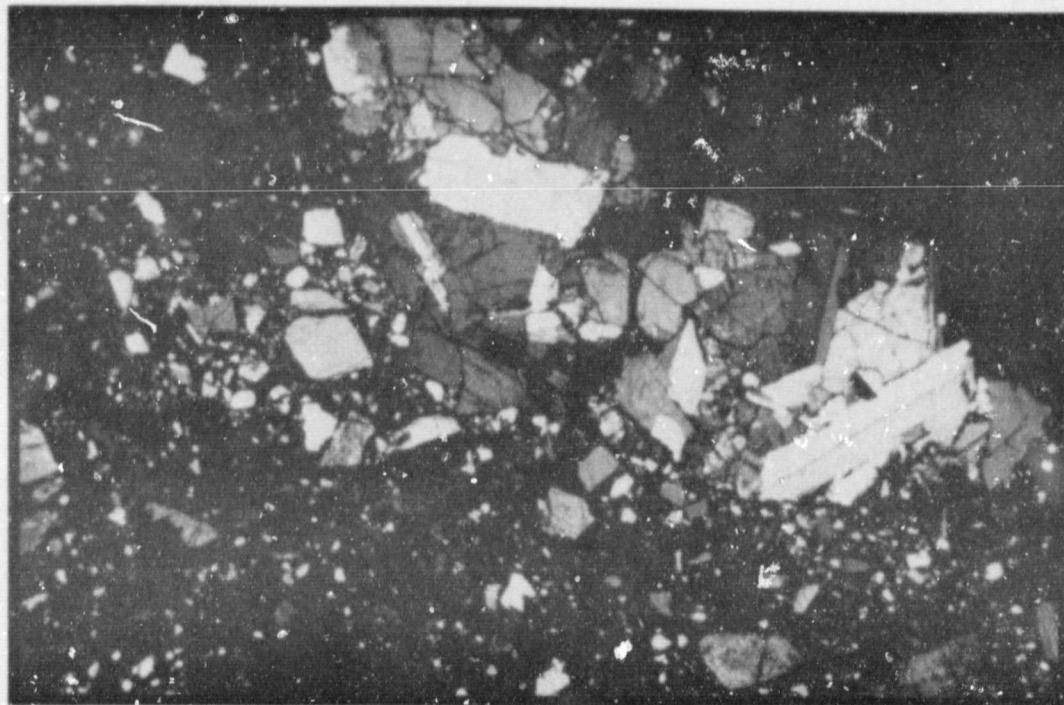
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WORKSHOP ON  
**PRISTINE HIGHLANDS ROCKS**  
 AND THE EARLY HISTORY OF THE MOON



(NASA-CR-174584) WORKSHOP ON PRISTINE  
 HIGHLANDS ROCKS AND THE EARLY HISTORY OF THE  
 MOON (Lunar and Planetary Inst.) 94 p  
 HC A05/MF A01

N84-13059

CSSL 03B

Unclass

G3/91 42504

# LPI Technical Report Number 83-02

LUNAR AND PLANETARY INSTITUTE

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HOUSTON, TEXAS 77058

WORKSHOP ON  
PRISTINE HIGHLANDS ROCKS  
AND THE EARLY HISTORY OF THE MOON

Edited by:  
John Longhi  
and  
Graham Ryder

Sponsored by  
The Lunar and Planetary Institute

A Lunar and Planetary Institute Workshop  
New Orleans, Louisiana  
October 15-17, 1982

Lunar and Planetary Institute 3303 NASA Road 1 Houston, Texas 77058

LPI Technical Report 83-02

Compiled in 1983 by the  
**LUNAR AND PLANETARY INSTITUTE**

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This report may be cited as:

Longhi J. and G. Ryder (1983) *Workshop on Pristine Highlands Rocks and the Early History of the Moon*. LPI Tech. Rpt. 83-02. Lunar and Planetary Institute, Houston. 92 pp.

Papers in this report may be cited as:

Author A. (1983) Title of paper. In *Workshop on Pristine Highlands Rocks and the Early History of the Moon* (J. Longhi and G. Ryder, eds.), p. xx-yy. LPI Tech. Rpt. 83-02. Lunar and Planetary Institute, Houston.

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**Cover:** *Sample 76255,71 in thin section viewed with crossed polarizers—a pristine gabbro clast in a glassy breccia matrix. Long dimension of photo is 9 mm.*

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## Introduction

Historically, the moon has been a source of wonder and mystery. The Apollo missions probed and sampled the moon and took away some of its mystery, but enough of the mystery and most of the wonder still remain to keep planetary scientists interested, active, and arguing. Much of the attraction that the moon holds for planetary scientists lies in the belief that the origins of the Earth and moon are inextricably intertwined. Because the Earth is much more complex than the moon and because geological processes have obliterated the record of the Earth's earliest history, planetary scientists have logically turned to the moon for answers to the questions of planetary origin and early evolution. As is often the case with inquiry in a new field, however, planetary scientists have discovered more new questions than answers to old ones. The answer to the question of lunar origin depends critically upon determination of lunar structure and bulk composition, which in turn are linked to the extent and duration of early differentiation and crustal formation. The most tangible clues to these problems lie in the suite of ancient rocks from the lunar highlands that have survived meteorite bombardment. These "pristine" rocks record an early melting event and appear to be the primary constituents of the lunar crust. Accordingly, lunar investigators have directed a great deal of effort into studying these samples. However, many questions regarding the identity, age, and distribution of these samples still remain.

Addressing these questions formed the basis for the Workshop on Pristine Rocks. The program committee felt that limiting the workshop to discussions of the rocks themselves would be neither desirable nor possible given the interests of the lunar science community, so equal time was allotted to the broader questions of lunar structure, composition, and origin.

The time was well spent. Despite the lack of new missions, the lunar science community remains active and innovative. Not only were new twists added to old arguments, but there were presentations on newly recognized pristine samples, on newly determined partition coefficients, on new dating techniques, and even a "new" model of lunar origin. There was even the



introduction of a new ray of hope for future sampling and geophysical missions—the European SELENE project. It was clear at the end of the workshop that even though a little more of the moon's mystery had been stripped away, the sense of wonder was still there, and so were many unanswered questions.

# Program

The following program of keynote speakers, discussion leaders, and summarizers constituted the presentation and discussion portion of the workshop. This portion of the program was held at the Travel Lodge Motel in New Orleans, Louisiana and lasted three days. Summaries of and discussions on these sessions will be found following the program.

The program committee consisted of J. Longhi and G. Ryder, conveners, and O. B. James, P. Jones, G. Lugmair, G. J. Taylor, D. Walker, and J. Wasson.

**Friday, October 15, a.m.**

## Session I

*A Brief Survey of the Early Lunar Igneous History*

John Longhi

*The Mineralogy and Petrology of the Pristine Rocks*

Odette James

*Chemical Trends in Pristine Rocks (Including KREEP)*

Gordon McKay

*Isotopic Systematics of the Lunar Highlands Rocks*

Richard Carlson

*Major Chemical and Petrological Components of the Lunar Crust*

Randy Korotev

**Friday, October 15, p.m.**

## Session II

### Discussion Topics

- I. What are the original igneous characteristics of the pristine rocks?  
Graham Ryder - Leader/Stephen Huebner - Summarizer
- II. When did the pristine rocks form?  
Gunter Lugmair - Leader/Joe Wooden - Summarizer
- III. How representative of the lunar crust are the pristine rocks?  
John Wasson - Leader/Ursula Marvin - Summarizer

Saturday, October 16, a.m.

**Session III**

*The Relations Between Early Lunar Magmatism and the Formation of the  
Mare Basalt Source Regions*

John Delano

*What can be Learned from Random Sampling of Layered Intrusions?*

Ian Campbell

*Geophysical Constraints on Lunar Structure and Composition*

Roger Buck

*Models of Bulk Moon Composition*

Paul Warren

*Models of Pristine Rock Formation*

John Wood

Saturday, October 16, p.m.

**Session IV**

**Discussion Topics**

- I. What are the relationships among the groups of pristine rocks?  
Stewart McCallum - Leader/Doug Blanchard - Summarizer
- II. What was the nature and extent of early lunar melting?  
Mike Drake - Leader/Claude Herzberg - Summarizer
- III. What is the bulk composition of the moon?  
Robin Brett - Leader/John Morgan - Summarizer

Sunday, October 17, a.m.

**Session V**

Summary Talks  
Research Plans

ADJOURN

# Summaries of Technical Sessions and Discussions

Prepared by: D. P. Blanchard, C. T. Herzberg, S. J. Huebner, J. Longhi, U. B. Marvin, J. W. Morgan, G. Ryder, and J. L. Wooden

## First Day: The Petrography, Chemistry, Age, and Distribution of Pristine Lunar Rocks

The ancient samples from the lunar highlands are part of a more or less random collection of rocky debris strewn about by meteorite impacts. The focus of the first day's session was threefold: (1) to discuss which samples retain a useful record of endogeneous igneous activity; (2) to describe these samples as completely as possible; and (3) to establish the proportions of these samples in the lunar crust.

Table 1. Criteria for determining pristinity.

Cumulate texture
Coarse grain size and pyroxene exsolution lamellae
Intra- and intergranular homogeneity
Low meteorite component (siderophile element abundances $\leq 3 \times 10^{-4}$ CI)
Low KREEP component
Radiometric ages $\geq 4.2$ b.y.

### Pristinity: Criteria and limitations

The term "pristine rock," coined by Paul Warren and John Wasson, has proved to be enormously useful in focusing attention on the concerted effort of lunar investigators to reconstruct the igneous history of the moon despite the pervasive effects of meteorite impacts—brecciation, melting, and mixing—on the lunar crust. The conferees discussed at some length the criteria originally proposed by Warren and Wasson (Table 1). It soon became apparent that the criteria had evolved as a flexible baseline for comparison, that no one criterion could stand alone, and that as a group the criteria were too exclusive if applied rigidly. For example, ancient volcanic rocks are implicitly excluded, yet they might well be the most informative samples. The criterion of coarse grain size was discussed by Mac Rutherford and Larry Taylor. They cited field and experimental evidence to show that coarse grain size is a poor indicator of magma body size: cooling rates  $<10^\circ\text{C}/\text{day}$  do not affect grain size, hence the interiors of slowly cooled impact melt sheets might (and in fact do) develop coarsely crystalline textures, including pyroxene exsolution, similar to those found in larger intrusions. The criterion of mineral homogeneity was questioned by Gordon Nord and Graham Ryder. Nord showed that during subsolidus cooling Na-, Fe-, and Mg-zoning developed at the margins of plagioclase grains in ferroan anorthosites, which are acknowledged to be pristine. Ryder maintained that although sample 60025 consists of four mineralogically distinct types of anorthosites, he found no reason to classify 60025 as non-pristine. Rather, he interpreted 60025 as a mechanical mixture of closely related rocks, perhaps adjacent layers crystallized from the primordial magma ocean. The criterion of low concentration of KREEP component was considered to be the weakest of the criteria, since a number of samples (KREEP basalts, quartz monzodiorite, granites) have KREEPy characteristics, but are judged to be pristine by most workers. Indeed, Paul Warren reminded the conferees that this criterion was never meant to exclude KREEP plutonic samples, but only to exclude those samples for which there is petrographic evidence of KREEP contamination by mixing, brecciation, and melting. Perhaps the strongest criterion listed in Table 1 is low siderophile element abundances. Yet flexibility is required here as well. Warren acknowledged that the upper limit for siderophile element concentration ( $3 \times 10^{-4}$  times CI

concentration) is somewhat arbitrary and that at least one sample thought to be pristine, the dunite from the Apollo 17 site, has subsamples that lie above (72417) and below (72415) the designated limits. Odette James pointed out that there is petrographic evidence for admixture of foreign material into dunite samples. These observations imply that there is a class of highlands samples whose pristinity is partially compromised by meteoritic processes, but which still retain recognizable features of their primary igneous origin.

No explicit revision was made of the Warren and Wasson criteria listed in Table 1. However, there was a general feeling that there was much useful information locked up in partially compromised and even totally mixed highlands samples. Excluding a sample from the family of pristine rocks, therefore, should not be grounds for excluding it from further study, but rather a challenge to petrologists and chemists to unravel the original characteristics of its protolith(s).

### **Petrography and chemistry of pristine rocks**

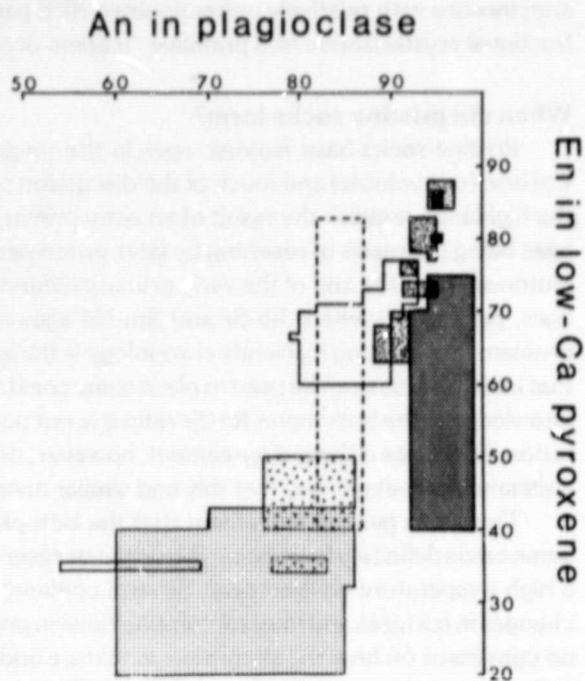
Details of the petrography and chemistry of individual pristine rocks are to be found in the annual *Proceedings of the Lunar and Planetary Science Conference*. The emphasis in the workshop was upon the petrographic and chemical basis of classification and petrogenesis.

Odette James, in her keynote talk (this volume), proposed that the majority of pristine rocks fall into three groups with each group representing the products of a distinct type of parent magma. The groups are: the ferroan anorthosite suite (consisting mostly of anorthosites and troctolitic anorthosites); the Mg-norite group of the Mg-suite (consisting mostly of norites and troctolites, but also including some spinel troctolites, the alkali anorthositics, the granitic rocks, and the quartz monzodiorite); and the Mg-gabbronorite group of the Mg-suite (consisting of gabbronorites, gabbro, feldspathic lherzolite, and sodic ferrogabbro). These groups are distinguishable on the basis of six criteria: An in plagioclase vs. Mg\* in mafic minerals, Ti/Sm ratio, Sc/Sm ratio, REE pattern, assemblages of trace minerals, and age. Figure 1 illustrates how mineral compositions easily separate ferroan anorthositics from Mg-norites and Mg-gabbronorites. There is also a subtle distinction between the Mg-norite and Mg-gabbronorite trends. This distinction is borne out by systematic differences in Ti/Sm and Sc/Sm (both ratios are lower in Mg-norites) and in trace mineral assemblages (ilmenite is the dominant Ti-mineral in Mg-gabbronorites but is less abundant in Mg-norites where armalcolite and rutile are dominant; zircon, baddeleyite, and zirkelite are present in Mg-norites but have not been found in Mg-gabbronorites). A number of samples do not fall simply into these groups and are classified tentatively in the most appropriate group. James considers KREEP basalts with low siderophiles and low Ti/Sm and Sc/Sm to be related to the Mg-norites even though the basalts have younger crystallization ages (~4.0 b.y.), thus implying that KREEP-rich source regions have been tapped more than once during early lunar history.

Perhaps the single most important feature of pristine rocks as a group is that they are random samples. This randomness is both a blessing and a curse. As for the blessing, randomness suggests (but does not guarantee) that similar samples found at different landing sites (ferroan anorthosites, Mg-norites) represent an abundant rock type. Several aspects of the curse were dealt with by Ian Campbell in his keynote talk (this volume). He noted severe problems in trying to reconstruct the composition and history of a terrestrial layered intrusion from a limited set of random samples. Indeed the study of terrestrial layered intrusions begins with careful, stratigraphically controlled sampling in order to determine crystallization order, proportions of rock types, and number of injections of fresh magma. He demonstrated that sample size is also important by showing striking variations in La, Sm, Yb, Th, and Na<sub>2</sub>O in ten small (2 to 5 gram) subsamples of the same two pyroxene gabbros. For example, La/Sm and La/Yb vary by factors of 1.5 and 1.9, respectively. Many pristine rock samples fall within this size range.

John Longhi suggested that some information about the physical nature of lunar magma bodies may eventually be extracted from the slopes of trends on the Mg\* vs. An diagram. He noted that the slopes of

**Fig. 1.** An vs. En for pristine rocks. Boxes enclose areas of overall variation for groups; individual rocks are not plotted where overlap occurs. Dashed line: KREEP basalt. Solid lines: irregular stipple, Mg-norites, and related troctolite. Unpatte... Mg-gabbro-norites; black, Apollo 14 troctolites; v's, alkali anorthosites; fine dots, evolved rocks; gray, ferroan anorthosites and related troctolitic anorthosites. Sources of data are given by James (this volume).



mineral composition trends of cotectic rocks (troctolites, norites, gabbros) vary considerably from intrusion to intrusion. Analysis of fractional crystallization and cumulus processes shows that these variations are due to physical, not compositional, differences. For example, the flattest slopes are associated with spreading-center magma chambers where intrusion of new magma is thought to be nearly continuous; intermediate slopes are typical of multiply-intruded magma bodies (Stillwater, Bushveld), and the steepest trends are typical of chambers formed from a single pulse of magma (Skaergaard). Thus, the difference in slope of the Mg-gabbro-norite and Mg-norite trends might be due to the former crystallizing in more complex intrusions than the latter. Longhi's modeling showed that the near-vertical trend of the ferroan anorthosite series was almost unavoidable, given the composition of the parent magma ( $Mg^* < 0.5$ , low alkali concentration).

In his keynote talk, Gordon McKay reviewed trace element partition coefficients and their application to calculating the parent magma compositions of pristine anorthosites. Trace element partition coefficients have been obtained by measuring phenocryst/matrix concentration ratios in terrestrial volcanic rocks and by analysis of coexisting crystals and quenched liquid produced in melting experiments. Because the experiments involved lunar-like compositions, many investigators have favored the experimental over the natural coefficients. These coefficients have been used in mass balance calculations that yield a parent magma composition of a cumulate rock when the proportions of cumulus crystals and trapped liquid are known. These proportions are usually difficult to establish petrographically, and one must assume that all elements are effectively "trapped," but in the case of lunar anorthosites the low concentration of iron (< 4 wt.%) limits the range of cumulus mafic minerals and trapped liquid. Calculations based upon McKay's early experimentally determined plagioclase/liquid partition coefficients produced light REE enrichment and marked positive Eu anomalies in the model parent magmas of the anorthosites. The origin of such fractionated magmas sparked considerable speculation in previous years. However, newly measured experimental partition coefficients for plagioclase differ considerably from the old coefficients ( $D_{Ce}^{Pl-L}$  is now  $\gg D_{Ce}^{Pl-L}$ ). The result is that the same mass balance calculations now produce model parent magmas of the

anorthosites with relatively unfractionated REE patterns. Such parent magmas could be derived by simple fractional crystallization of a primitive "magma ocean" with a chondritic REE pattern.

### When did pristine rocks form?

Pristine rocks have isotopic ages in the range 3.9—4.5 b.y. (Table 2). The keynote talk by Richard Carlson (this volume) and much of the discussion centered about two fundamental questions: (1) Are all of the highlands samples the result of an early primary differentiation of the moon with the relatively younger ages being the result of resetting by later processes? (2) Do some of the younger ages (4.2 b.y.) represent plutonism after the end of the early primary differentiation? Carlson concluded that in general the inferred ages, particularly where Rb-Sr and Sm-Nd ages are concordant, reflect the time of crystallization. One problem in evaluating highland chronology is the lack of data for the ferroan anorthosites. Carlson showed that initial Sr ratios can be used to place some constraints on the ages of the ferroan anorthosites, but without knowledge of the bulk moon Rb/Sr ratio it is not possible to be confident about ages inferred from  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios. There was a general agreement, however, that the Ar-Ar age of anorthosite 60025 (~3.9 b.y.) records a metamorphic event and that this and similar anorthosites are significantly older.

There was general agreement that the isotopic systematics of the highlands samples could be, and in some cases definitely have been, disturbed or reset by processes associated with impact and slow cooling in a high temperature environment. Several conferees presented petrologic evidence for post-crystallization changes in textures and mineral compositions in pristine rocks due to these processes. However, there was no consensus on how the alterations in texture and mineral compositions could be used to infer changes in radioisotope ratios and distribution in the highlands rocks. Although enough information exists to model the diffusion of Rb, Sr, Sm, Nd, K, Ar, etc., there is still little real information (and it is often conflicting) about how these elements actually behave in various geologic situations. Günter Lugmair raised the very difficult but important question of "if diffusion occurs, where do the elements go?" Because allocated lunar samples are so small and the highlands rocks have a generally coarse grain size, the problem of open and closed isotopic systems is particularly difficult. The data for terrestrial rocks shows that *all* of the isotopic systems examined in detail are susceptible to disturbance and that the conditions necessary for disturbance can vary greatly for each system. In a given sample, one or more isotopic systems may be "open" while the remaining systems are "closed," and in another sample, just the opposite systems will be open or closed. There is still much to learn about the when, how, and why of radioisotope movement.

A second general area of discussion concerned the isotopic characteristics of the source regions of the highlands rocks, and whether or not these characteristics could distinguish among the models for the origin of the highlands samples. There was general agreement among the isotopic specialists that the isotopic differences that do exist in highlands samples are so small that the problems of interlaboratory bias and post-crystallization isotopic disturbance overwhelm them. Lugmair observed that this problem is very sample dependent; i.e., there are usually just a few critical samples that give information about source heterogeneity at a given time (the Apollo 12 ilmenite basalts are an example), and such a critical sample simply has not yet been analyzed for the highlands suite. Carlson drew some inferences concerning the source of the Mg-suite. These inferences are that: (1) it definitely had a low Rb/Sr ratio similar to the source region of the anorthosites; (2) it possibly had a slightly depleted light rare-earth-element pattern; and (3) it could be similar to the source of some of the mare basalts (Apollo 12 olivine-pigeonite, Apollo 15, Apollo 11 high-K). Stuart McCallum commented that most pristine rocks are cumulates, not solidified melts, and are therefore removed one additional step from the characteristics of their source.

The KREEP Mg-suite connection was also discussed in the context of isotopic sources of highlands rocks. Of particular interest was the old question of when the KREEP source formed and what role assimilation of KREEP could have played in the formation of the relatively incompatible element-enriched

Table 2. Rb-Sr and Sm-Nd isotopic results for pristine highlands rocks.

Sample	Type	Age (Sr)	$I_0(\text{Sr})$	Age (Nd)	$\epsilon_{\text{Nd}}(0)$	Reference
15455c	An. Nor	$4.48 \pm 0.12$	$0.69894 \pm 6$	—	—	16
60025	An.	$3.85 \pm 0.02$	$0.70030 \pm 20$	—	—	7
67667	Feld. Lhz.	—	—	$4.18 \pm 0.07$	$+0.8 \pm 0.3$	14
72255c	Nor.	$4.08 \pm 0.05$	$0.69913 \pm 7$	—	—	8
72417	Dun.	$4.45 \pm 0.10$	$0.69900 \pm 7$	—	—	9
73255,27	Nor.	—	—	$4.23 \pm 0.05$	$+0.5 \pm 0.4$	13
76535	Troc.	$4.51 \pm 0.07$	$0.69900 \pm 3$	$4.26 \pm 0.06$	$+0.3 \pm 0.5$	10,11
77215	Nor. Brec.	$4.33 \pm 0.04$	$0.69901 \pm 7$	$4.37 \pm 0.07$	$-3.0 \pm 0.5$	12
78236	Nor.	$4.29 \pm 0.02$	$0.69907 \pm 2$	$4.34 \pm 0.04$	$+1.1 \pm 0.7$	13,15

All ages in units of  $10^9$  years, Rb-Sr ages calculated using  $\lambda_{\text{Rb}} = 1.42 \times 10^{-11} \text{yr}^{-1}$ , Sm-Nd ages with  $\lambda_{\text{Sm}} = 6.54 \times 10^{-12} \text{yr}^{-1}$ . Initial  $^{87}\text{Sr}/^{86}\text{Sr}(I_0)$  and  $\epsilon_{\text{Nd}}$  are as reported in original references. See the paper by Carlson (this volume) for the reference key and additional discussion.

Mg-norite suite. The model Sm-Nd and Rb-Sr ages of KREEP are about 4.3 b.y. and may reflect the time at which the KREEP source formed: If the older isochron ages of some of the Mg-norite suite rocks (Table 2) prove correct, then processes acting on the source of the Mg-norites before 4.3 b.y. may have produced isotopic characteristics similar to those of the KREEP source.

The discussion also included a very interesting presentation of the  $^{39}\text{Ar}$ - $^{40}\text{Ar}$  continuous laser degassing technique by Derek York and Christopher Hall. This technique permits stepwise outgassing of very small samples, including single mineral phases, and has produced very precise and otherwise unobtainable data for terrestrial samples. In particular York pointed out that blocking temperatures of minerals are routinely determined in his lab. By plotting blocking temperature vs. age for several different minerals, the cooling path can be determined. This type of information for highlands samples would help in choosing between the later plutonism vs. slow cooling/resetting models for the relatively young ages of some of the pristine highlands rocks. Unfortunately, this technique has yet to be applied to lunar samples.

### The distribution of pristine rocks

The distribution of pristine rocks in the lunar crust has two aspects: (1) How well can highlands soils and breccias be modeled as mixtures of known pristine rocks?; and (2) What are the areal and vertical variations in lunar crustal composition? Every mission except Apollo 11 has yielded one or more fragments of probable pristine highlands rocks. Yet only one mission (Apollo 16) visited a highlands area; the others landed on mare surfaces, the rims of large basins, or on an ejecta blanket of disputed provenance (Apollo 14). Because orbital gamma-ray and X-ray measurements have shown large areas of the lunar highlands to be compositionally similar to soils at Apollo 16, Randy Korotev has been modeling Apollo 16 soil compositions as mixtures of pristine rocks plus meteoritic and mare basalt components in the hope that the results would apply to large areas of the lunar surface. In his keynote talk, Korotev pointed out that mixtures of ferroan anorthosite, Mg-suite rocks, and KREEP are adequate to explain the major element compositions of Apollo 16 soils but are inadequate in terms of minor and trace elements even when meteoritic and mare basaltic components are included. Korotev has calculated two mixing models that approximate the Apollo 16 soil. The first model requires employing nearly all the individual samples of the Mg-gabbro group that have been found at the Apollo 16 site as mixing end-members along with ferroan anorthosite and KREEP: 60.7% ferroan anorthosite, 10.5% KREEP, 8.9% 61224 (eucritic gabbro), 10.9% 67667 (feldspathic lherzolite), 4.6% 67915 (sodic ferro-gabbro), 0.6% 67435 (spinel troctolite), and 3.8% CI chondrites. This model assumes that Mg-gabbroites constitute all the mafic lithologies at Apollo 16. The second model assumes that "ferroan



anorthositic norite," a previously unrecognized pristine rock type with 20–30% mafic minerals, is the major mafic component of the Apollo 16 soils. The ferroan anorthositic norite lithology has been identified by M. Lindstrom (this volume) in one possibly pristine rock (67215) and in clasts of several breccias from North Ray Crater. Korotev preferred the second model because he felt it seemed less *ad hoc*. A major unresolved problem with the second model is why these samples should be so poorly preserved compared to ferroan anorthosites, despite their being so similar mineralogically. The conferees did not decisively favor one model over the other. But the implications of the two models are clear: if the second proves correct, then Mg-norites and Mg-gabbronorites may be only minor constituents of the upper lunar crust and their abundance in the highlands collections may be the result of biased sampling near the margins of large basins (Mg-norites) or near small, relatively young intrusions (Mg-gabbronorites).

The areal distribution of pristine rock types was discussed by Paul Spudis (this volume), who reported on attempts to correlate orbital X-ray analyses of the elemental composition of the lunar highlands with photogeology and sampling results. Spudis and his colleagues used the digital data bases of the La Jolla Consortium to generate color composited images representing the Mg and Al abundances typical of mare basalts plus two pristine rock types, anorthosites and norites. The data were also used in an attempt to distinguish formations of pristine norites from low-K Fra Mauro "basalts" (which were assumed to be impact mixtures). The results showed that pure anorthosite (defined as  $\text{Al}_2\text{O}_3 > 30\%$ ,  $\text{MgO} < 4\%$ ) is rare, but it occurs in several small regions of pre-Nectarian highlands west of Pasteur Crater ( $11^\circ\text{S}$ ,  $94^\circ\text{E}$ ), southwest of Smythii Basin ( $11^\circ\text{S}$ ,  $89^\circ\text{E}$ ), and west of Mendeleev ( $4^\circ\text{N}$ ,  $117^\circ\text{E}$ ). Spudis suggested that these Al-rich zones may well delineate areas of primordial ferroan anorthosites, uncontaminated by later intrusions or basin ejecta. In addition, large regions of the eastern limb, and also the Apollo 16 site, though highly aluminous, are too high in magnesium to be classified as pure anorthosite. These may be regions where the pristine anorthositic crust has been contaminated by mafic ejecta from the Crisium and Smythii basins and pre-Imbrian mare flooding.

Regions rich in norite (defined as  $\text{Al}_2\text{O}_3 = 20\text{--}21.6\%$ ,  $\text{MgO} = 6\text{--}7.5\%$ ) include the Montes Apenninus, Haemus, and Taurus highlands, which are dominated by ejecta from the Imbrium and Serenitatis Basins and which yielded samples of pristine norite in the Apollo 15 and 17 collections. Spudis suggested that this region, which now appears to constitute an Mg-suite petrologic province, may never have had a crust of ferroan anorthosite or, if one originally formed there, it may have been removed by the Procellarum basin impact. Additional small areas of ejecta dominated by norite also occur near Gibbs ( $15^\circ\text{S}$ ,  $82^\circ\text{W}$ ) and close to the Crisium and Nectaris Basins.

Spudis said that a search of highlands areas for mare basalts ( $\text{Al}_2\text{O}_3 < 19.6\%$ ,  $\text{MgO} > 8.3\%$ ) was made in an effort to identify mafic deposits not attributable to the large mare basins. None were found.

Vertical variations in lunar crustal composition were considered from two points of view. The first (Longhi, this volume) is that the highlands surface layer, as typified by Apollo 16 soil ( $\geq 60\%$  ferroan anorthosite) but known to be of global extent from orbital data, extends downward at least as far as the maximum relief (5–8 km) of the highlands. The absence of ferroan anorthosite in melt ejecta from the Imbrium and Serenitatis impacts suggests that the anorthositic layer is thin or absent in these regions.

The second point of view is that some idea about vertical variations in crustal compositions can be obtained by estimating the depth of formation of pristine rocks using various textural, mineralogical, and thermodynamic criteria. On the basis of comparing Fe/Mg homogeneity and development of exsolution lamellae in pyroxenes with zoning and exsolution typical of terrestrial layered intrusions, James argued that most ferroan anorthosites and Mg-suite rocks formed at depths of 5–30 km with anorthosites forming at the shallower end of the range. On the basis of strong zoning in both pyroxene and plagioclase, some Mg-rich rocks (e.g., 67667) formed in near-surface intrusions, whereas other samples (e.g., 76535) with homogeneous mineral compositions, annealed textures, and symplectic intergrowths suggest depths of formation of

10-30 km. Claude Herzberg presented revised thermodynamic calculations showing that all pristine rock mineral assemblages were consistent with subsolidus equilibration at depths < 30 km.

These considerations suggest that the upper half of the 60-km-thick lunar crust is heterogeneous. This heterogeneity may arise either from (1) a primary layering in which the uppermost layer, i.e., one lying above the excavation depth of the basin-forming impacts, is enriched in anorthosite and lower levels consist primarily of Mg-suite rocks (the model of Ryder and John Wood) or (2) large Mg-suite intrusions situated at various depths in an initially anorthositic crust (the models of James and Herzberg). There seems to be no firm petrological evidence as to the constitution of the lower half of the lunar crust.

### **Second Day: Pristine Rocks and Other Clues to Early Lunar Melting, Bulk Composition, and Origin**

The primary igneous nature of pristine rocks, the preponderance of their inferred crystallization ages at times  $\geq 4.2$  b.y., and global extent of an  $\sim 60$  km crust apparently composed primarily of pristine rocks all suggest an early igneous event that led to the formation of a feldspathic to mafic crust overlying a complementary mantle below. This mantle is the source for the mare basalt magmas. Models of lunar structure and composition rely heavily upon inferences of the depth and nature of differentiation caused by the early crust-forming event, and models of lunar origin will be on firm foundations only when they can account for a well-constrained bulk composition and initial physical state. Therefore inquiries into the origin of the moon inevitably lead to consideration of its geophysical and geochemical features, and these in turn are inextricably linked to the early differentiation and the formation of pristine rocks.

#### **Lunar structure and the depth of early lunar melting**

John Delano, in his keynote talk (this volume), presented a variety of experimental and chemical data that indicated that the lunar mantle is markedly heterogeneous. Evidence for this heterogeneity comes from mare volcanic glasses (4.0-3.2 b.y.) that span a wide range of composition (Ti and Al are especially variable), but which on the basis of high-pressure liquidus experiments appear to be derived from similar depths (350-500 km). Those basalts that were apparently derived from depths approaching 500 km carried with them a primordial volatile component. The presence of these volatiles in surface coatings suggests, according to Delano, that early differentiation extended to approximately 500 km and left the deep interior more or less undifferentiated. This interpretation is bolstered by (1) inversions of seismic velocity data that show a transition zone from shear wave velocities of  $4.45 \pm 0.04$  km/sec above 450 km to  $4.25 \pm 0.13$  km/sec below 500 km and (2) thermal stress calculations that suggest the moon originally had a hot exterior and cold interior. Both of these points were challenged by Alan Binder, who asserted that the velocity profile model is heavily dependent upon questionable interpretation of weak reflections and that thrust fault scarps in the highlands ranging from 1 to 30 km long are consistent with a 6 kb modern-day stress field that would have developed if the moon had been completely molten initially.

In his keynote talk (this volume), Roger Buck considered a number of geophysical constraints on the composition of the lunar interior, which in turn hinge upon the depth of lunar differentiation. He assumed that the shear wave transition at 450-500 km was real and was due to the chemical change from  $\text{Al}_2\text{O}_3$ -free differentiated material above to aluminous, undifferentiated material below. He also assumed that the  $\text{Al}_2\text{O}_3$  that was once in the upper mantle was now in the crust. He noted that shear wave velocity for the lower mantle is consistent with only small amounts of garnet. Buck minimized the normative garnet content by

apportioning some  $\text{Al}_2\text{O}_3$  to pyroxenes as  $\text{MgAl}_2\text{SiO}_6$  and  $\text{CaAl}_2\text{SiO}_6$  components. With these assumptions in mind Buck then treated bulk  $\text{Al}_2\text{O}_3$  and Mg/Si as adjustable parameters and calculated high-pressure normative mineral assemblages that were consistent with shear wave velocities. He concluded that an  $\text{Al}_2\text{O}_3$  concentration of 3–5 wt.% and an Mg/Si ratio  $\sim 0.8\%$  were the optimum values for the bulk moon. Herzberg pointed out that the effect of increasing pressure would be to create more garnet at the expense of pyroxene, and hence the low  $\text{Al}_2\text{O}_3$  values favored by Buck might still be too high.

Mike Drake led an informal session on the potential heat sources of early lunar melting. In addition to heat acquired from a hot, fissioning proto-Earth (Binder's model), solar luminosity, electrical induction, short-lived nuclides, accretion, tidal capture, and adiabatic compression of gases were discussed. The relative contributions of these processes remain uncertain, but there was a general feeling that accretional energy must play a major role.

Dave Walker suggested that the effort put into finding heat sources for a globe-encircling magma ocean might prove futile. He pointed out that a major incentive for a magma ocean involving a substantial portion of the moon was the belief that the crust was primarily anorthosite. He cited mixing model studies by Larry Haskins' group that require only a small portion of the crust to be anorthosite. If these studies accurately reflect bulk crustal composition, then the need for a magma ocean disappears and with it the need for exotic heat sources.

### **Lunar bulk composition**

Paul Warren reviewed geochemical models of the moon in his keynote talk (two papers in this volume). He noted that there is general agreement that (1) the low density of the moon implies low abundances of Fe-Ni and probably other siderophile elements, and (2) the moon is greatly depleted in volatile elements relative to the Earth and more so relative to chondrites. Among the uncertain compositional parameters, the abundances and ratios of refractory elements are the most controversial and received the most attention.

The widely accepted value of bulk lunar U, 35 ppb, is inferred from heat flow measurements and corresponds to a 2.8 $\times$  enrichment over abundances in the silicate phases of chondrites. This enrichment serves as a means of estimating abundances of other refractory elements in many geochemical models. Warren suggested, however, that these estimates are too high because (1) the Apollo 15 and 17 heat flow measurements were made at the edges of maria where heat flow may be high due to anomalous thermal conductivity, and (2) the concentrations of U, Th, and K on the nearside of the moon might be higher than average. He also noted that if the offsets of the moon's centers of figure and mass were due to a more Fe-rich near-side mantle, rather than a light, thick anorthositic crust on the farside, as most workers believe, then constraints on the moon's total  $\text{Al}_2\text{O}_3$  content are relaxed and "chondritic" levels, i.e., 3–4 wt.%  $\text{Al}_2\text{O}_3$ , are possible.

Warren also noted a conflict between the results of Roger Buck's geophysical modeling (Mg/Si  $\sim 0.8$ ;  $\text{Mg}^* \approx 0.75$ ) and the value of  $\text{Mg}^*$  ( $\sim 0.85$ ) most likely in the source region of the Mg-suite magmas. The two sets of inferences cannot be accommodated unless the Mg-suite magmas formed from remelted olivine plus orthopyroxene cumulates. A cumulate source might also alleviate the difficulty in deriving the Mg-suite's parent magmas from source regions with chondritic Ca/Al ratio, and would also be consistent with Richard Carlson's suggestion of a light REE-depleted source region for the Mg-suite. No heat source for the remelting of refractory cumulates was tendered, however.

### **Lunar origin**

Perhaps the best constraint on lunar origin would be a firm knowledge of the moon's bulk composition. While a confident determination of lunar bulk composition awaits future sampling of the crust and geophysical probing of the interior, the range of estimates of bulk composition has narrowed considerably in the last

few years to compositions generally similar to the Earth's mantle and the eucrite parent body. A major incentive for this narrowing range is the similarity of lunar and terrestrial oxygen isotopes, which suggest that the compositions of the Earth's mantle and the moon cannot be totally unrelated. Within this range two models still seem possible: (1) derivation from the Earth by fission or spallation, and (2) accretion in Earth orbit.

Ross Taylor, in an informal presentation (this volume), discussed lunar composition and origin in the context of the chemical fractionations that can be demonstrated to exist between terrestrial-type planets. For example, in a plot of  $K/U$  vs.  $K$ , very distinct disparities in ratio can be seen between planets, but the ratio is constant in material from any given planet, thus implying characteristic volatile depletions for the planets. In addition, in the planets and in the chondrites, it is apparent that significant bulk separations of metal and silicate have taken place. Taylor argued that the similarity of crystallization ages of eucrites and H and E chondrites ( $\geq 4.5$  Ga) implied that: (1) fractionation of metal from silicates and volatiles from refractory elements must have occurred at a time indistinguishable from solar system  $T_0$ ; (2) these fractionation events occurred before accretion of the Earth and the moon, so that the planets must have inherited such characteristic ratios as  $K/U$  from precursor planetesimals; (3) the core and mantle of the Earth accreted separately (with FeS as the low density component in the core) and metal silicate equilibration was established in the precursor objects; and (4) the differences between the moon and the terrestrial mantle compositions reflect heterogeneous accretion. Such a model is obviously at odds with a lunar origin by fission. Taylor suggested that a general depletion of volatile elements in the inner solar system may have occurred as a result of processes in the early solar nebula, such as a strong T-Tauri wind, within a few million years of the formation of the sun. Thus depletion of  $K/U$  in the region of the inner planets may have already occurred by the time these bodies accreted from the fractionated planetesimals on a time scale of  $10^8$  years.

As a proponent of the origin of the moon by fission, Alan Binder presented a model (this volume) in which a large, extremely hot proto-moon is spun off the Earth and initially loses mass by flash evaporation through the  $L_2$  Lagrangian point of the close binary system. Subsequently, volatiles are lost from the proto-moon to the Earth through the  $L_1$  point as the moon cools from  $2000-3000^\circ\text{C}$  to  $\sim 1200^\circ\text{C}$  (formation of a thin crust) over a period of perhaps 500 years. The model results in a bulk moon closely resembling "pyrolite" except for marked volatile depletion (e.g., P, Na, K) and a twofold enhancement of FeO in the moon which perhaps reflects depletion of FeO in the Earth's mantle by core growth after Earth-moon fission. The corollaries of this model are high initial temperatures and a totally molten initial state. The low  $\text{Al}_2\text{O}_3$  concentrations (2-4 wt.%) estimated for the bulk moon by Warren are more easily reconcilable with the relatively high  $\text{Al}_2\text{O}_3$  concentrations (19-25 wt.%) observed in the crust if the moon was totally molten initially.

John Wood (this volume) suggested the problems of producing an essentially coreless moon, and providing sufficient heat to provide large-scale (or total) melting of the moon may be resolved by a model of lunar origin involving low velocity coalescence of two smaller hemi-moons. Formation of a magma ocean by accretional heating is not a new suggestion, but accretion of small planetesimals takes too long ( $10^7$  to  $10^8$  years) and implants heat at too shallow a level for a significant proportion of the moon to be molten at any given moment. If the size distribution of planetesimals were skewed toward larger sizes, then heat could be implanted deeply; collision of two bodies of roughly 0.5 lunar masses represents the limit of this condition. In order to preserve a moon (a necessary boundary condition), the two moonlets would need to be in geocentric orbits with low relative velocities; heliocentric orbits would probably provide too much collisional energy. Wood argued that two moonlets could be produced in geocentric orbit possibly by fission or spallation by glancing impact, or by accretion in the Earth's orbit. He believed that the first mechanism probably would not work since most of the fragments would be reaccreted by the Earth, and a large proportion of those remaining would be lost to space. On the other hand, accretion in Earth orbit has some

interesting consequences that might lead to the formation of two moonlets. Two moonlets in similar Earth orbits would coalesce on very short time scales. In order to store two moonlets the  $10^7$  to  $10^8$  years required to accrete them separately, it may be necessary for these bodies to have formed sequentially rather than simultaneously. Thus the first moonlet would have accreted in near-Earth orbit and then would have undergone tidal recession, while continuing to enlarge at the expense of the circumterrestrial swarm. After repopulation of the near-Earth region, the second moonlet would have nucleated and begun its tidal recession. Because the rate of tidal recession decreases with distance from the Earth, the second moonlet would have eventually overtaken the first, causing a collision. Coalescence of moonlets of similar size, therefore, provides sufficient heat to form a magma ocean and may be responsible for the offset of the moon's center of figure and center of mass.

### Acknowledgments

Logistic and administrative support for this workshop has been provided by Pamela Jones and LeBecca Turner (Projects Office, Lunar and Planetary Institute). This Technical Report has been prepared by René Dotson (Technical Editor, Lunar and Planetary Institute).

# ABSTRACTS

## AN ESTIMATE OF THE BULK, MAJOR OXIDE COMPOSITION OF THE MOON.

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An estimate of the bulk, major oxide composition of the moon (neglecting the small core which contains 2% of the lunar mass (1,2)) can be made on the basis of the following assumptions and observations:

1) As is discussed in an accompanying abstract and in earlier papers (3,4), the moon was initially totally molten and thus underwent total differentiation in its earliest phase of development (e.g., 5). As such, the composition of the lower mantle can, in part, be determined from expected differentiation sequences using the compositions of the crust and mare basalt magma source region or upper mantle as boundary conditions. These considerations indicate that the lower mantle consists of olivine-orthopyroxene (6). Internal structure models of a totally differentiated moon (7) and independent seismic data (8) indicate the the  $Mg'$  of the lower mantle is in the range of 72-88. Further, thermal history models of an initially totally molten moon (9) indicate that the lower mantle must contain small amounts of U, Th, and K and hence that the lower mantle originally contained the equivalent of 15±10% trapped liquid (bulk composition). On the basis of the maximum depth of the upper mantle discussed under 2) below, the mass of the lower mantle is 66±12% of the lunar mass.

2) The mare basalt magma source region or upper mantle is olivine dominated (75-80% O1) and at shallow depths (maximum depth 150 to 300 km, probable maximum depth 200 km) as indicated by Binder (10). The composition of the upper mantle is derived from Table 2 of (10) and its mass is 23±10% of the lunar mass.

3) The remnants of the KREEP layer at the crust-mantle boundary have the composition given by (11) and is 4±2 km thick; hence it contains only 0.6±0.3% of the lunar mass.

4) The average composition of the crust is 70% Pl, 20% Pyr, 9% O1, and 1% Ilm (e.g., 12). The uncertainty of the crust's Pl content is 10% with corresponding uncertainties in the other components. The average  $Mg'$  of the crust is 70. The average thickness of the crust is 75±10 km and hence its mass is 10.4±1.3% of the lunar mass.

The bulk composition of the moon (minus core) derived on the basis of the above considerations is given in Table 1 along with estimates of the bulk composition of the terrestrial mantle taken from the literature (13, 14, 15, 16). These data are also shown in Fig. 1, from which it is apparent that, except for the well known depletion of volatile components ( $P_2O_5$ ,  $K_2O$ , and  $Na_2O$ ) and a possible enrichment of FeO, the bulk composition of the moon is very close to that of the terrestrial mantle. This result, when couple with the obvious fact that the moon lacks any significant Fe/FeS core, supports the fission origin of the moon (e.g., 5). As shown earlier (17,18), the depletion of the volatile components in the moon, in terms of the fission model, is due to their having been preferentially concentrated, via fractional vaporization, in the hot atmosphere of the hot (2000-4000°C) fissioned moon and the loss of the atmosphere to the earth through the  $L_1$  point. The possible FeO enrichment in the moon with respect to the earth's mantle, as we know it today, may be due to a decrease in the FeO contents of the terrestrial mantle as the core grew by the reduction and removal of FeO from the mantle during geological time (19). Thus the current "high" FeO content of the moon may reflect an earlier higher FeO content of the terrestrial mantle.

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Table 1. Estimated bulk compositions of the Moon and the Earth's mantle.

Oxide	Bulk Moon	Terrestrial Mantle			
		Pyrolite* (13)	Total Mantle (14)	Upper Mantle (15)	Lower Mantle (16)
SiO <sub>2</sub>	39.8±1.3	44.2±1.0	44.4±0.4	49.6	40
MgO	34.8±1.1	38.5±1.0	40.3±1.2	41.1	28
FeO	18.0±0.7	8.3±0.1	7.1±0.6	9.3	21
CaO	2.3±0.4	2.9±0.2	2.4±0.5		
Al <sub>2</sub> O <sub>3</sub>	3.6±0.6	3.8±0.3	3.4±0.9		
Na <sub>2</sub> O	0.10±.01	0.59±.03	0.28±.05		
K <sub>2</sub> O	0.03±.01	0.18±.05	0.06±.03		
Cr <sub>2</sub> O	0.6±0.3	0.42±.20	0.35±.05		
TiO <sub>2</sub>	0.4±0.2	0.65±.07	0.16±.04		
P <sub>2</sub> O <sub>5</sub>	0.03±.01	0.06±.02	0.05±.01		

\*Average of Pyrolite I and III

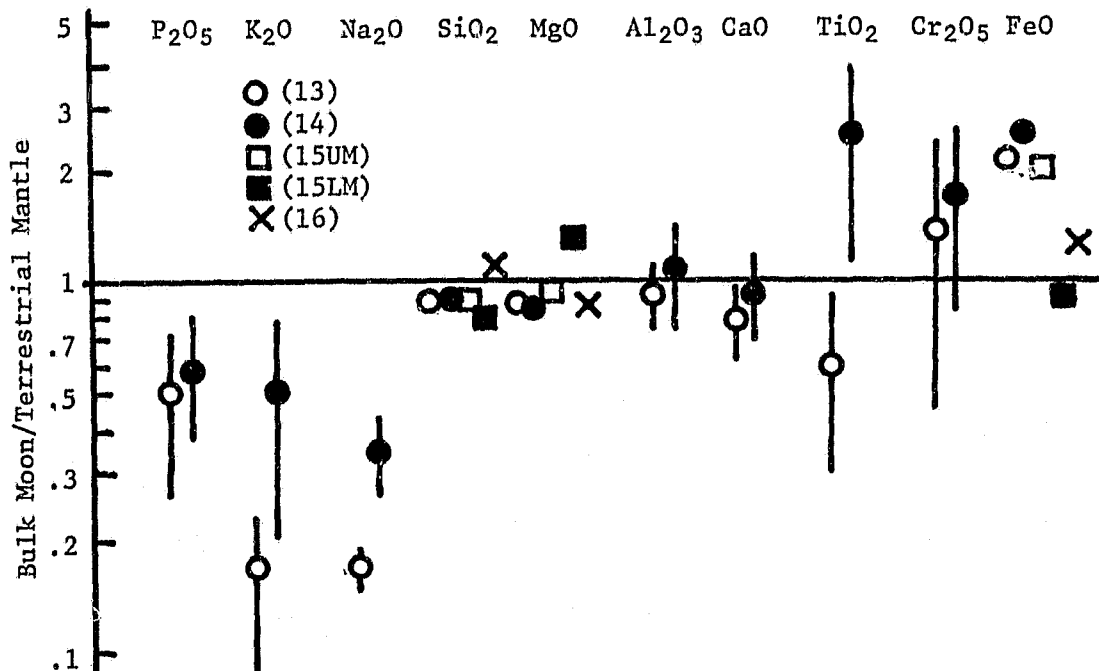


Fig. 1. Ratio of the bulk composition of the moon to that of the terrestrial mantle.

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SELENOLOGIC AND SEISMOLOGIC EVIDENCE FOR AN INITIALLY TOTALLY MOLTEN MOON. A.B. Binder, H.-C. Gunga, and J. Oberst, Erde-Mond Forschergruppe, Institut für Mineralogie, Universität Münster, 4400 Münster, West Germany.

As previously discussed (1,2), thermal history- and thermoelastic stress models (3) indicate that if the moon was initially totally molten, it should have modest scale ( $\approx 10$  km), young ( $\approx 0.5$  to  $1 \times 10^9$  yr old) thrust fault scarps in the highlands. Also, shallow (outer 5 to 10 km of the crust), compressional moonquakes (a subclass of the HFT moonquakes) with maximum stress drops in the 1 to 3 kbar range should be occurring on these faults. In contrast, thermoelastic stress calculations for magma ocean models (lunar models initially molten only in the outer few 100 km (4,5)) indicate that the highlands should be free from any global, post-Imbrian, compressional tectonic features and that the crust should now be seismically quiet.

A cursory review of the Apollo high resolution imagery indicated earlier (1,2) that fault scarps with the expected geometry and ages, assuming that the moon was initially totally molten, are found in the highlands. As an extension of this work, we report here the initial results obtained from an extensive photoselenological study of these highland scarps. We also report the initial results obtained from a program carried out to determine the stress drops of the 35 observed HFT moonquakes.

61 thrust fault scarps have been found in the highlands using the Apollo 15 and 16 panoramic imagery. The scarps are visible on images with solar illumination angles ( $A$ ) between  $5^\circ$  and  $45^\circ$  and are best seen at  $A \approx 20^\circ$ . Correcting these data for incompleteness due to illumination effects, we find that about 95 such scarps probably exist in the areas photographed with  $5^\circ < A < 45^\circ$ . Since these areas cover only about 5.7% of the highlands, there must be  $\approx 1700$  such fault scarps in the highlands. The majority of the observed 61 scarps are members of scarp complexes which are made up of up to 10, but on average 3 to 4 separate scarps. 19 such complexes are observed in the areas investigated, which, from the above, indicates that  $\approx 500$  such complexes are in the highlands.

The individual scarps range in length from 1 to 30 km and have a typical length of 12 km, i.e., they are on the 10 km scale as predicted on the basis of the initially totally molten moon model. The complexes are typically 30 to 50 km in dimensions, with the largest being 120 km from end to end.

Using the crater morphology-age classification of Trask (6) and the calibration data of Moore et al. (7), the ages of 34 of the thrust scarps could be determined on the basis of craters they cut or overrun (maximum ages) and craters which are located on the fronts of the scarps (minimum ages). The ages thus determined range from  $60 \pm 40$  million years to  $620 \pm 280$  million years, with younger scarps being much more frequent. Thus this epoch of recent thrust fault activity began about  $0.6 \times 10^9$  years ago and the level of activity has been increasing with time. These ages and this increase in the level of activity are also in accord with the initially totally molten model which predicts that the faulting should have begun at  $\approx 0.5$  to  $1 \times 10^9$  years ago in response to the ever increasing thermoelastic stress levels (also see Fig. 6 of (2)).

Spectral analysis of the 35 observed HFT or shallow moonquakes shows that their corner frequencies, which are used in the calculation of their stress drops, lie at frequencies higher ( $> 12$  Hz) than are measurable by the Apollo seismometers. Thus only minimum stress drops can be calculated from the available data. Nevertheless, following the theory presented by Goins et al. (8), we find that 4 of the 25 HFT's thus far analysed have minimum stress drops in the range from 1 to 4 kbar. The existence of such high stress moonquakes is consistent only with the initially totally molten moon model in which quakes

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are predicted to be occurring in the outer 5 to 10 km of the crust.

In conclusion, the photoselenologic and seismologic results presented clearly support the concept that the moon was initially totally molten.

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GEOPHYSICAL CONSTRAINTS ON LUNAR COMPOSITION  
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The geophysical data on the moon can be used to constrain the composition of the moon. Assumptions must be also made about the evolution of the moon since that will affect the radial distribution of material within the body. There are uncertainties in the geophysical data pertinent to this study and the path of evolution of the moon is not agreed upon. Therefore, we will investigate the effect of variations in data and assumption on each of several compositional factors.

The data considered in this study are the mass of the moon:  $7.35 \times 10^{25}$ g (1), the upper mantle seismic shear velocity:  $4.45 \pm 0.04$ km/sec (2), the lower mantle seismic shear velocity:  $4.25 \pm 0.13$ km/sec (2), the moment-of-inertia factor:  $0.391 \pm 0.002$  (3). Mineral properties are also input into this work. First the normative mineral assemblages must be calculated for several pressure ranges, then the bulk moduli, thermal expansion coefficients and seismic velocities and their pressure and temperature derivatives must be calculated. For the low pressure in the moon (<40kbars) the bulk moduli and thermal expansion coefficients are well determined from laboratory measurements. Lack of data on the Tschermactic minerals affect the outcome of the modelling and the effect of uncertainty in these data is considered.

The path of evolution of material in the interior is important to several assumptions of the modelling. The distribution of the oxide  $Al_2O_3$  is most critical.  $Al_2O_3$  goes into low density, low seismic velocity minerals at pressures less than 8 kbars (~180 km depth). Below this region it is stable in minerals with high densities and high seismic velocities (spinel, garnet and the Tschermactic minerals). It is generally assumed that the crust was derived from only the outer part of the moon (300-500 km) (4). For testing models with low  $Al_2O_3$  abundances the whole moon is taken to have differentiated to make the crust. Since the seismic velocities in the lower mantle (> 450km) are lower than in the upper mantle, some compositional difference must be assumed. This has been taken to be iron in the silicates ( $Fe^{+2}$ ) or free iron. The free iron is more efficient at lowering the seismic velocities for a given increase in density. The crust is taken to be 75km thick and made of anorthositic gabbro. The temperature profile is uncertain and two profiles are considered here: one from Toksoz, et. al. (5) and one 100°C hotter overall.

The effect of variations in the data and assumptions on bulk compositional parameters have been explicitly tested through inversion of the data. The parameters considered were Mg/Si weight ratio, the Mg/(Mg+Fe) molar ratio, the core radius and the amount of Fe (free or in silicates) in the lower mantle. The  $Al_2O_3$  composition and the other assumptions are changed outside the inversion, since there are only four pieces of data to invert. Those data are the mass, the moment-of-inertia factor ( $I/MR^2$ ) and the upper and lower seismic shear velocities. The compressional wave velocities are not used because they are more uncertain than the shear velocities and contain similar information. The inversion is necessary because the compositional parameters are interrelated through the model so that a change in one produces a change in several of the geophysical observables. The forward problem part of this work (the model) is similar to that outlined in Buck and Toksöz (6). A starting bulk composition is assumed and the crust is removed from the upper

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mantle portion of it. The minor components of the bulk composition, which have little effect on the geophysical data were taken from Morgan, *et. al.* (7). The  $Al_2O_3$  content was varied as were the different temperature profiles, mechanisms for lowering the lower mantle seismic velocities and the inclusion of Tschermactic minerals in the normative calculations.

Unique compositions are found for a given set of assumptions and data values. Several of the factors which have a strong effect on the compositional parameters are shown in figures 1-3. Figure 1 shows the effect that uncertainty in the moment of inertia factor has on the Mg/Si ratio and the core size. Figures 2a and b show the influence of the normative schemes on composition. 2a is done without allowing for Tschermactic minerals so only garnet and spinel are the sites for  $Al_2O_3$  for pressures greater than about 10 kbars. 2b is for a norm allowing for Tschermactic minerals. Both show a strong dependence of lower mantle iron content on  $Al_2O_3$  content. Lower  $Al_2O_3$  abundances are compatible with no difference in upper and lower mantle iron contents, allowing for variation of seismic velocities within the range of uncertainty. Lower  $Al_2O_3$  contents are also preferred on the grounds of the Mg/Si ratio. The earth's mantle is often taken to have a value of 0.95 (8) for the ratio but this may be only representative of the upper mantle. The value for chondritic meteorites is about 0.80 (9). Changing the temperature profile  $100^\circ C$  changes the Mg/Si ratio by 0.1. The variation of upper mantle seismic velocities by 0.05 km/sec changes Mg/Si about 0.1, Mg/(Mg+Fe) by 0.03 and the free Fe in the lower mantle by 2.0 weight percent. Changing the lower mantle velocity by 0.1 changes the Mg/Si ratio by 0.08 and the lower mantle free iron content by 4.0 weight percent. Finally, changing the velocity lowering factor from free iron to  $Fe^{+2}$  decreases the Mg/Si ratio by about 0.1.

These results allow for a range of compositional parameters. The Mg/Si ratio is likely to be closer to the value for the chondritic meteorites than that for the upper mantle of the earth. If the  $Al_2O_3$  content is in the low range (3-5 weight percent) it is more consistent with the data.

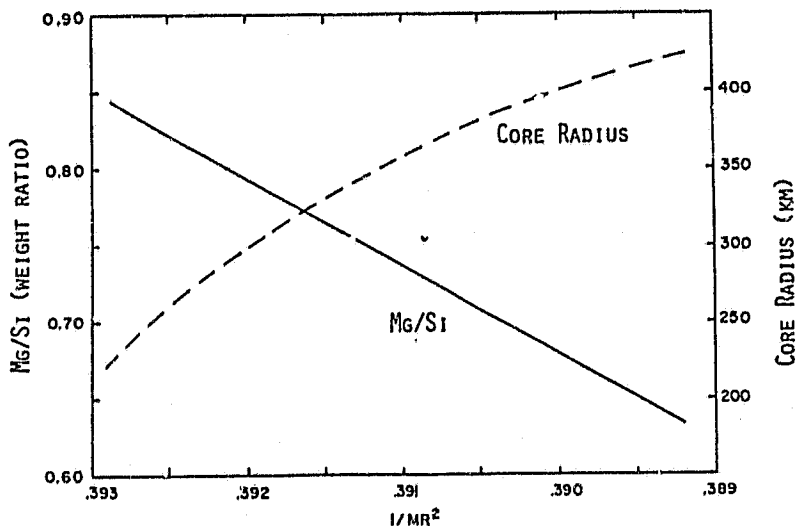


Figure 1. The effect of the variation in the model  $I/MR^2$  on the Mg/Si weight ratio and the core radius is shown for a core density of  $6.5 \text{ g/cm}^3$ . In this series of tests the upper and lower mantle seismic shear velocities were 4.5 and 4.3 km/sec respectively and the temperature profile was  $100^\circ$  hotter than that of Toksöz *et al.* (5)

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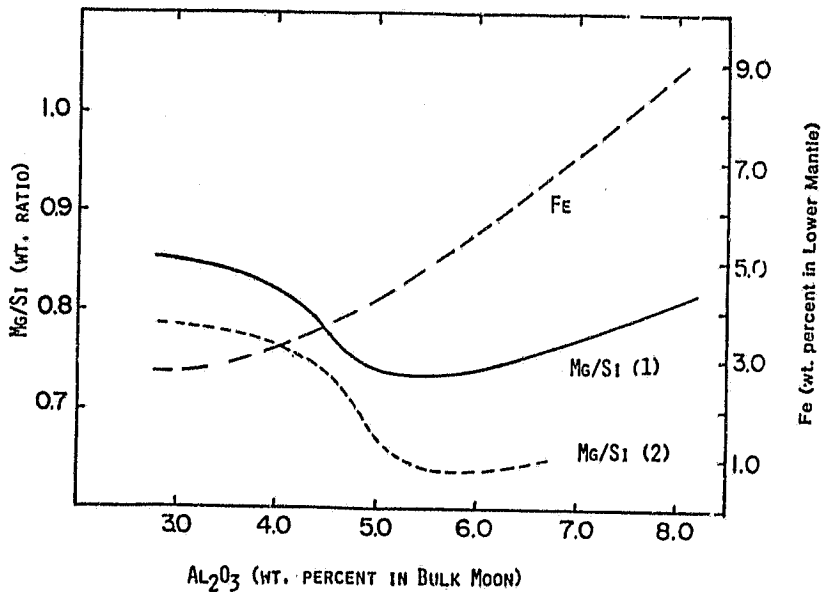
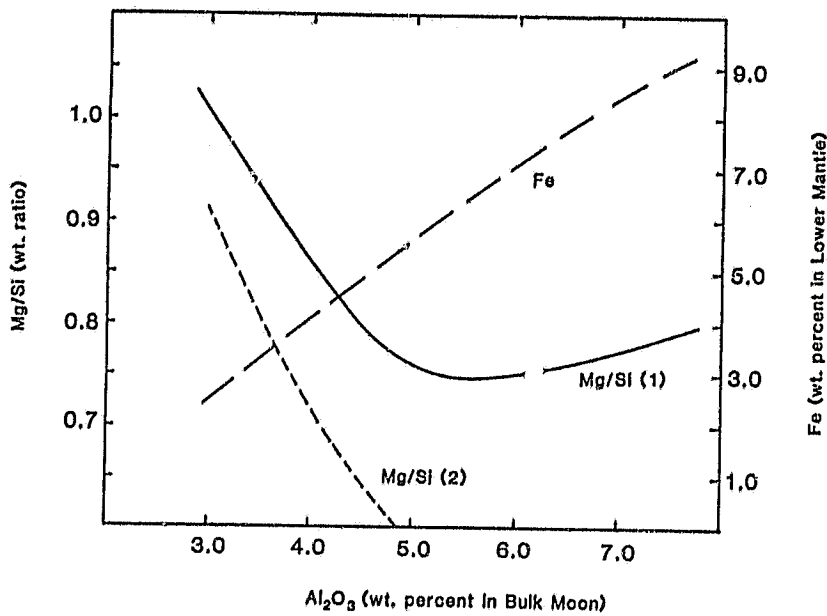


Figure 2. The effect of variation in  $Al_2O_3$  content for two normative schemes: (a) with no Tschermaclitic minerals and (b) with inclusion of Tschermaclitic minerals. Case (1) on each figure is for free iron as the velocity reducing factor in the lower mantle and (2) is for  $Fe^{+2}$  as the factor. The other geophysical variables are set as in figure 1.

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### WHAT CAN BE LEARNED FROM RANDOM SAMPLING OF A LAYERED INTRUSION?

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Introduction: Interpretation of the data obtained from lunar cumulates suffers from two important limitations: (i) the field relationships between samples is uncertain (ii) the samples available for study are small in size. How much can be learned from a limited number of small samples collected at random from a layered intrusion? The best way to approach this question is to pose it first for a terrestrial layered intrusion and then, by analogy, to the moon.

Problems introduced by small sized samples: Consider first the problems presented by small samples of cumulate rocks. The chemistry of a cumulate rock has two components; a mineral component controlled by the chemistry and modal abundance of the cumulus minerals and a liquid component controlled by the percentage of intercumulus liquid trapped between the cumulus grains. The major element and compatible trace element geochemistry are controlled by the mineral component and are thus susceptible to changes in the modal percentage of the cumulus minerals. The problems of obtaining a representative analysis of these elements from a small sample of a medium- to coarse-grained cumulate rock are obvious, especially if the rock is layered. The inhomogeneous distribution of the compatible elements develops at the liquidus temperature and is due to the preferential partitioning of the element into certain phases. The incompatible elements enter a cumulate as a liquid and initially are homogeneously distributed throughout the rock, but they eventually partition into the late crystallising accessory minerals during the final stages of crystallisation. The abundance of an incompatible element in a cumulate is a measure of the amount of trapped liquid contained in the cumulate and its ratio to another incompatible element should be the same as that in the melt from which it crystallised, provided the sample taken is not too small. If the sample taken is small, biased sampling of the accessory phases becomes a problem. This is illustrated in Table 1, which gives the results of analyses of nine 2-5 gm samples of a two-pyroxene gabbro. The sampling error, expressed as the  $(S.D \times 100) \div \text{mean}$ , is appreciable for the incompatible elements and for incompatible element ratios, but is far greater for compatible elements controlled by the modal abundance of the cumulus minerals.

A more serious problem is encountered when sampling cumulate rocks with only one cumulate phase. Consider a bronzite cumulate which crystallises intercumulus augite and plagioclase. Crystallisation of the intercumulus liquid starts at selected points within the cumulate pile, probably where the texture is locally the most adcumulate (see below). The first intercumulus plagioclase crystals to form are An-rich and the fractionating intercumulus liquid is pushed in front of an advancing crystal-liquid interface. It eventually crystallises in pockets characterised by a high incompatible element content and by plagioclase crystals rich in Ab. As a consequence very large samples are required when sampling monocumulate rocks for incompatible elements.

Random sampling of a layered intrusion: The second major problem with lunar cumulates is that the stratigraphic relationship between the samples is unknown. The most valuable studies of terrestrial layered intrusions have been made on bodies which are complete, undeformed and unmetamorphosed. It is important that the samples collected are large and that their stratigraphic position is well-known. The bulk composition of the intrusion can then be calculated from a weighted average of the analysed samples.



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Successive liquids can also be calculated by excluding the layers one at a time from the bottom up (1,2), provided the chamber has remained a closed system.

Can this approach be applied to samples collected at random? The first requirement is to be able to place samples in their correct stratigraphic sequence. It is well-known that as a magma fractionates its  $Mg/(Mg+Fe)$  and normative  $An/(An+Ab)$  ratios decrease, resulting in the crystallisation of olivines and pyroxenes with a decreasing  $Mg$  number (where  $Mg$  = molar  $Mg/(Mg+Fe)$ ) and plagioclase with a decreasing  $An$  content. The compatible trace element content of minerals is also a useful guide. Here the  $Ni$  content of olivines and the  $Cr$  content of pyroxenes are especially important.

If randomly collected samples can be ranked in stratigraphic order, the bulk composition of the magma can be calculated from the mean of the analyses. Successive liquids can then be calculated by omitting the samples one at a time.

There are a number of potential problems in using this approach to rank samples from a terrestrial layered intrusion in stratigraphic order. Firstly, cyclic layering is a common feature of layered intrusions. Each cycle is believed to represent the entry of a new pulse of magma into the chamber which resets the fractionation of the magma, producing a reversal in the mineral variation trend. A sample collected above the level at which a new pulse enters the chamber may have olivines with a higher  $Mg$  number and  $Ni$  content than a sample from well below this level. Secondly, if the chamber has acted as an open system by emitting lavas at the surface, the calculations will yield early liquid compositions which are more primitive than the parent magma (3). Thirdly, some layered intrusions (3,4) have a zone of reversed fractionation at their base. Here the most fractionated samples are at the bottom of the zone and the most primitive at the top. Fourthly, the  $Ni$  content of olivines may increase with increased fractionation. This is because  $D_{Ni}^{Ox}$  increases with decreasing temperature and with increased fractionation of the silicate liquid. Under some circumstances, the increase in  $D_{Ni}^{Ox}$  may more than offset the decreasing  $Ni$  content of the melt, producing a reversal in the fractionation trend. Because of these difficulties, especially the problems introduced by multiple injection and the extrusion of magma at the surface, it is impossible to place a random set of samples from a layered intrusion in their correct stratigraphic sequence with a high degree of confidence. It is, however, possible to rank samples according to their degree of chemical fractionation.

The use of mineral chemistry: The problem connected with small samples can be avoided by concentrating on the chemistry of the cumulus minerals. In principle, many of the properties of a magma can be deduced from the chemistry of the phases which precipitate from it. For example the  $MgO/FeO$  ratio in a silicate can be calculated from the  $MgO/FeO$  ratio in olivines, using the Roeder and Emslie (5) relationship

$$K_D^{Ox} = \frac{x_{FeO}^{Ox} \cdot x_{MgO}^L}{x_{FeO}^L \cdot x_{MgO}^{Ox}} \quad (1)$$

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Similar equations can be written for pyroxene and plagioclase (6,7). The equation for plagioclase is

$$K_D^{P\&L} = \frac{(1-An)^{P\&L} \times An^L}{(1-An)^L \times An^{P\&L}} \quad (2)$$

where An is the composition of normative plagioclase in the liquid. The  $Al_2O_3$  content of pyroxenes also gives information about  $as_{SiO_2}$  of the magma.

An important property of cumulate rocks is that their chemistry is modified by sub-liquidus re-equilibrium reactions. Two types of intercumulus crystallisation can be recognised. In the first, equilibrium is maintained between the cumulus crystal and the intercumulus liquid. The principle involved can be illustrated using Fig. 1. Consider a liquid of composition L. Cumulate crystals of composition S will form in equilibrium with this liquid at temperature t. A cumulate rock formed under these conditions with 50% trapped liquid will have a bulk composition  $B_1$  and a solidus temperature  $t_1$ , the final solid composition being  $S_1$ . If only 10% liquid is trapped between the cumulus grains, the bulk composition of the system will be  $B_2$ , the solidus temperature  $t_2$  and the composition of the final solid  $S_2$ . If equilibrium is not maintained (i.e. perfect fractional crystallisation), crystallisation will cease at  $t_3$ , the rims of the cumulus grains being zoned from S to  $S_3$ . The bulk composition of the cumulus grains at any temperature lies on the curve  $SP_1$  for 50% trapped intercumulus liquid and  $SP_2$  for 10% trapped liquid.

Cumulate pyroxenes show little evidence of zoning and follow an equilibrium intercumulus trend whereas plagioclase crystals are zoned in rocks with an orthocumulate texture and follow a fractional crystallisation path. In practice, plagioclase crystals may maintain partial equilibrium with the intercumulus liquid so that the solidus temperature is at some intermediate temperature  $t_4$ , the final solid to crystallise having the composition  $S_4$ .

Grain size: It has been known from the early days of geology that the grain size of a rock which crystallises from a basaltic magma is a function of the cooling rate. Basalt which cools rapidly has a finer grain size than diabase which, in turn, has a finer grain size than a slow cooling gabbro. Following this line of argument it might be reasonable to expect that the grain size of cumulate rocks is also a function of cooling rate and therefore of the size of the magma chamber. Empirical observations suggest that this is not the case. There is little difference between the grain size of cumulates from the giant Bushveld Complex and the smaller Skaergaard Intrusion. Cumulates from the intermediate sized Stillwater Complex are coarser grained than either of the former. These observations suggest that factors which affect the balance between nucleation and crystal growth, such as the viscosity of the melt, are probably more important than the cooling rate in controlling grain size in cumulates. Studies of kinetic processes such as exsolution in pyroxenes (9) or subsolidus re-equilibration of Mg and Fe between chromite and olivine (10) offer a better potential for obtaining the size of an intrusion from a single sample than the grain size of the minerals.

An:Mg diagrams: An obvious method of comparing a random set of cumulate rocks is to plot the major fractionating constituents of one type

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of cumulate minerals against those of another. Plots of the An constant of cumulus plagioclase against the Mg content of coexisting cumulus olivines and pyroxenes were introduced by Wager as a means of comparing different terrestrial layered intrusions. Two factors produce deviations from the normal trend of decreasing An with decreasing Mg: (i) postcumulate reactions (ii) the extent to which An and Mg are fractionated between the liquid and cumulus crystals.

The influence of postcumulus reactions on An:Mg trends has been discussed by Longhi (11) and Raedeke and McCallum (12). The trapped liquid has a lower An and Mg content than the cumulus crystal. If equilibrium is maintained between liquid and crystals, the effect of postcumulus re-equilibration is to lower both the An content of the plagioclase and the Mg content of coexisting olivines or pyroxenes. The resultant trend subparallels the liquidus fractionation trend provided the minerals are present in approximately their cotectic proportions. If the minerals are not present in their cotectic proportions, deviations from this trend become possible.

Consider the case of a plagioclase cumulate with a small amount of liquid trapped between the cumulus grains. The intercumulus pyroxene which crystallises from this liquid will have the Mg content of the liquid which will be lower than that of cumulus pyroxenes from adjacent layers. Raedeke and McCallum (12) have shown that varying the proportion of cumulus plagioclase to pyroxene from the cotectic proportions to pure plagioclase, results in a variation in Mg at constant An, producing a vertical trend on an An:Mg diagram. Similarly, it is theoretically possible to produce a horizontal trend in olivine or pyroxene cumulates by varying the amount of cumulate plagioclase.

In terrestrial layered intrusions, equilibrium between the cumulus grains and the intercumulus liquid is maintained for pyroxenes and olivines but not for plagioclase. This has two important effects:

- (i) Re-equilibration in cumulate rocks with cotectic mineral proportions decreases the Mg content of the olivines and pyroxenes but produces no change in the An content of the cores of the plagioclase crystals.
- (ii) Horizontal trends in ultramafic rocks cannot be the result of re-equilibration between cumulus and intercumulus plagioclase. Such trends are seen in bronzite cumulates from the Jimberlana Intrusion (Fig. 2) and are the product of fractional crystallisation of the plagioclase component of the intercumulus liquid.

Longhi (7) has shown that variations in  $K_D$  in equations (1) and (2) can also influence the fractionation trends on an An:Mg plot. As  $K_D$  approaches 1.0 fractionation tends to zero. Thus the vertical trend for the lunar anorthosites could be produced by  $K_D$  in equation (2) approaching 1.0 provided there is no corresponding change in the  $K_D^{Ox}$ . Increasing the SiO<sub>2</sub> content of the melt or lowering the temperature of crystallisation, increases the value of  $K_D^{Px}$  and reduces the compositional contrast between plagioclase and liquid. The effects are most pronounced near the end-members where fractionation is least effective. Significantly, the lunar anorthosites are believed to have crystallized from liquids with a low normative Ab content. Longhi's conclusions are supported by plots of An:Mg for cotectic gabbros from the Jimberlana Intrusion which follow a near-vertical trend during the final stages of fractionation (Fig. 2).

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Table 1. Analyses of nine samples of the same two pyroxene gabbro

Sample	wf. (gm)	La	Sm	Yb	Th	Na <sub>2</sub> O	La/Th
1	4.6	5.31	0.91	0.53	.70	2.84	7.59
2	3.4	3.92	0.81	0.57	.62	1.21	6.36
3	5.3	5.33	1.14	0.68	.73	2.51	7.30
4	3.2	5.56	1.17	0.63	.72	2.69	7.70
5	3.3	4.13	1.07	0.78	.80	1.17	5.17
6	4.3	5.62	1.13	0.67	.71	2.83	7.94
8	3.3	5.85	1.11	0.64	.72	2.87	8.11
9	2.3	5.95	1.34	0.65	.76	2.89	7.84
10	3.4	4.04	0.93	0.65	.67	1.28	6.05
<u>S.D. × 100</u> mean		16%	15%	11%	7%	35%	14%

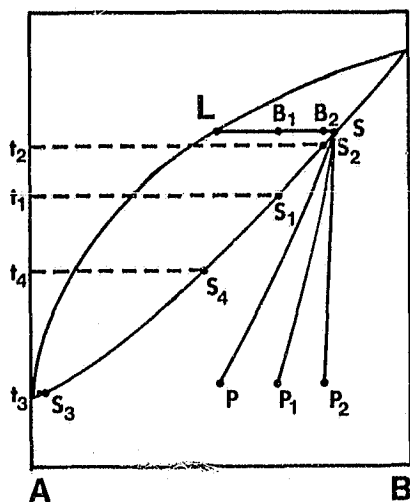


Fig. 1

Fig. 1. A phase diagram for a hypothetical two-component system AB in which there is complete solid solution between A and B.

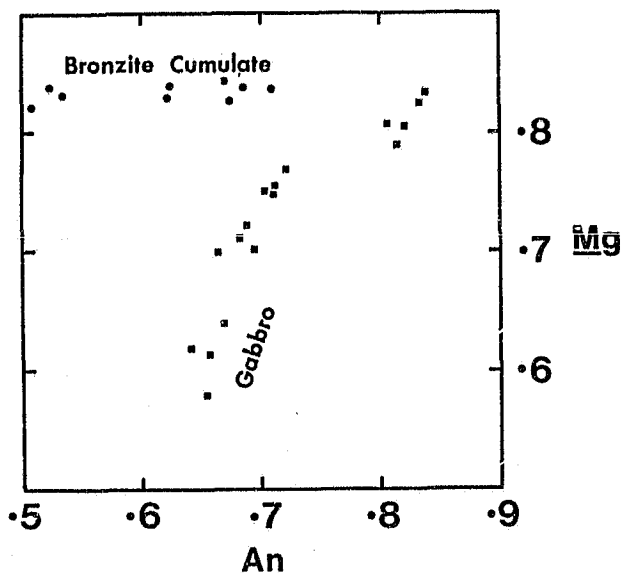


Fig. 2

Fig. 2. A plot of An against Mg for cotectic gabbros and bronzite cumulate from the lower layered series of the Jimberlana Intrusion.

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CHRONOLOGIC AND ISOTOPIC SYSTEMATICS OF LUNAR HIGHLAND ROCKS. Richard W. Carlson, Department of Terrestrial Magnetism, 5241 Broad Branch Road N.W., Washington, D.C. 20015.

Recorded in the chemical and isotopic characteristics of the pristine lunar highland rocks are the clues necessary for understanding the character of the earliest igneous activity on the moon. Deciphering these clues is a difficult process hindered by several factors such as, a limited sample collection, poor stratigraphic control on the samples that are available, and the fact that many of the pristine rocks have been affected by impact induced metamorphic overprinting of their original characteristics. In spite of these weaknesses, several fundamental conclusions regarding the nature of the early igneous evolution of the moon can be drawn from this data set.

First, and perhaps foremost, of these conclusions is that the outer regions of the moon experienced a major differentiation episode within the first 100-150 Ma of the moon's formation. Evidence for this primary differentiation event is contained in the Rb-Sr, Sm-Nd and U-Pb isotopic systems of a wide variety of lunar rock types including high-Ti mare basalts (1,2), KREEP (3-6) and the pristine crustal rocks (7-16). In the case of the mare basalts and KREEP, despite their relatively young (i.e. <3.9 Ga) crystallization ages, model age data suggest that the respective incompatible element depletion and enrichment of their source materials originated during events occurring 4.36 to 4.55 Ga ago (1-6). For pristine crustal rocks the evidence for a major early lunar differentiation is shown not only by greater than 4.3 Ga crystallization ages for some samples (Table 1) but also by their alignment along "total rock" Sm-Nd (14) and Rb-Sr (figure 1) isochrons of 4.33 to 4.38 Ga age. Though caution should be used in assigning precise chronologic significance to these "total rock" isochrons, the correspondence of these ages and the model ages obtained for a highly fractionated material such as KREEP (3-6), strongly suggest that the first major episode of crust-mantle differentiation on the moon was complete by 4.35 to 4.40 Ga ago. Association of this primary differentiation event with a "magma ocean" episode is an obvious choice although other models for lunar crustal evolution involving a series of huge impact melt pools (17) or more limited melting of the lunar exterior (18) cannot be ruled out by the isotopic data alone. Regardless of the exact physical nature of this event(s) the major episode of chemical segregation of the plagioclase and incompatible element enriched lunar crust and the more mafic, incompatible element depleted mantle took place during the first 150 Ma of the moon's existence.

Determining the precise details of the igneous history of the lunar crust during and following the primary differentiation episode is more difficult and several important points concerning early lunar crustal evolution remain to be answered (e.g. 16). These include the questions of whether some pristine samples formed during the primary differentiation episode or whether a more prolonged period of igneous intrusion into the lunar crust occurred (e.g. 19, 20, 21).

The available Rb-Sr and Sm-Nd internal isochron ages for pristine highland rocks are given in table 1. Notably rare in

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this list are age data for samples of the ferroan anorthosite group. Though several ferroan anorthosite group rocks have been dated by the Ar-Ar method (22, 23) the relatively young ages obtained ( $\sim 3.9$  Ga) conflict with the extremely low initial  $^{87}\text{Sr}/^{86}\text{Sr}$  of these samples (e.g. 7) suggesting that the young Ar-Ar ages record the time of metamorphic, but not crystallization, events. The lack of detailed crystallization age information for this group of highland rocks is truly unfortunate because the chemical characteristics of the ferroan anorthosites suggest that they may be the most likely representatives of the earliest phase of crust formation on the moon (e.g. 24). The relatively old Rb-Sr internal isochron age of  $4.48 \pm 0.12$  Ga determined for anorthositic norite 15455c (16) is consistent with this idea though the relatively large uncertainty associated with the age makes a definitive interpretation impossible.

Rocks of the high-Mg group are enriched generally in incompatible trace elements, especially when compared to the ferroan anorthosites. While this enrichment has made the high-Mg rocks more suitable for radiometric dating, it also indicates that these rocks did not form directly from magmas of "bulk-moon" composition, but are the products of chemically highly evolved liquids. Whether these evolved liquids represent late stage residuums of a "magma ocean" episode or are partial melts of preexisting crustal cumulates (16, 19-21) remains to be answered.

The exact chronologic significance of the ages obtained for the high-Mg rocks (Table 1) is not certain and at least three interpretations must be considered: 1) Ages represent the time at which the various samples crystallized from the "magma ocean". 2) Ages record a prolonged period (4.18-4.51 Ga) of partial melting and plutonism into preexisting lunar crust. 3) Ages represent "cooling ages" indicating the time when the temperature of the rock cooled below the particular radiometric system's blocking temperature. As discussed by (16) the initial  $\epsilon_{\text{Nd}}$  values determined for the majority of high-Mg samples tend to be marginally greater than zero. While this may suggest the parental magmas of these rocks formed by melting of preexisting "depleted" mantle materials (16) sufficient uncertainty exists both in the "bulk-moon" Nd reference parameters and the initial isotopic values to allow all of samples to be products of a magma ocean with approximately chondritic Nd isotopic composition. For example, at 4.3 Ga ago the maximum difference expected in  $\epsilon_{\text{Nd}}$  for lunar materials ranges only from a high of +2 for the source region of high-Ti mare basalts (25) to a low value of -0.6 for the proto-KREEP materials (6).

The question of whether the ages obtained for lunar crustal rocks represent "cooling ages" or the time of their crystallization becomes more important as more relatively young (i.e.  $\sim 4.2$  Ga) ages are reported. For troctolite 76535, the evidence suggests that the olivines in this rock retain memory of a much older event ( $\sim 4.5$  Ga, 10, 26) than is recorded by other minerals in the sample (11, 27). In light of the data for 76535, the meaning of the  $\sim 4.2$  Ga ages reported for 73255 (13) and 67667 (14) becomes more ambiguous. This is especially true for 73255 where other, non-pristine, clasts within the same host impact

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breccia also give ~4.2 Ga Ar-Ar ages (28). Resolving this question must wait until a more complete isotopic and petrologic data base is obtained for these samples. The task of determining the true chronological significance of these ~4.2 Ga ages is of great importance because the various possible interpretations carry distinct implications for the nature of the moon's early thermal history.

Table 1: Rb-Sr and Sm-Nd isotopic results for pristine highland rocks

Sample	Type	Age (Sr)	$I_0$ (Sr)	Age (Nd)	$\epsilon_{Nd}(0)$	Reference
15455c	An.Nor.	4.48±0.12	0.69894±6	-	-	16
60025	An.	3.85±0.02	0.70030±20	-	-	7
67667	Feld.Lhz.	-	-	4.18±0.07	+0.8±0.3	14
72255c	Nor.	4.08±0.05	0.69913±7	-	-	8
72417	Dun.	4.45±0.10	0.69900±7	-	-	9
73255,27	Nor.	-	-	4.23±0.05	+0.5±0.4	13
76535	Troc.	4.51±0.07	0.69900±3	4.26±0.06	+0.3±0.5	10,11
77215	Nor. Brec.	4.33±0.04	0.69901±7	4.37±0.07	-3.0±0.5	12
78236	Nor.	4.29±0.02	0.69907±2	4.34±0.04	+1.1±0.7	13,15

All ages in Ga. Rb-Sr ages calculated using  $\lambda_{Rb}=1.42 \times 10^{-11} \text{yr}^{-1}$ , Sm-Nd ages with  $\lambda_{Sm}=6.54 \times 10^{-12} \text{yr}^{-1}$ . Initial  $^{87}\text{Sr}/^{86}\text{Sr}(I_0)$  and  $\epsilon_{Nd}$  values are as reported in original reference. At 4.2 Ga ago the "chondritic" Nd reference values used by La Jolla (6), JSC (15), and USGS (12) differ by a maximum of 0.3  $\epsilon$  units.

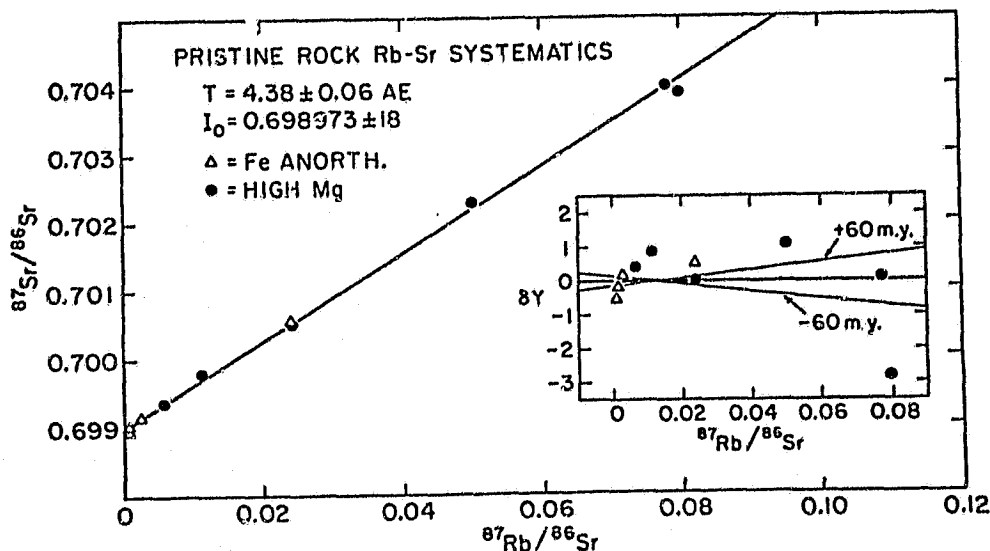


Figure 1: Rb-Sr "total rock" isochron for ferroan anorthosite and high-Mg highland rocks. Data for samples 60025 (7), 61016c (29), 15415 (30), 76535 (10), 67667 (16), 15455c (15), 78236 (15), 72417 (9), 77215 (12), and 72255c (8) are plotted in order of increasing  $^{87}\text{Rb}/^{86}\text{Sr}$ . All Sr isotopic values are bias corrected for comparison with data from CIT.

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THE EARLY EVOLUTION OF THE MOON AS INFERRED FROM MARE BASALTIC LIQUIDS. J. W. Delano, Dept. of Geological Sciences, State University of New York, Albany, N. Y. 12222

#### INTRODUCTION

Information concerning the depth and residual mineralogy of source regions for basaltic magmas can be less than unique if the basalts have had any history of crystal/liquid fractionation prior to solidification. Melts that have remained closed systems since leaving their source regions are termed *primary* and are much sought after by lunar and terrestrial petrologists/geochemists.

Many compositional varieties of mare volcanic glasses have been analyzed and found to display relatively simple inter-relationships (i.e., two linear arrays; 1), since these trends (Fig. 1) would have been eliminated if only  $\sim 5$  wt.% olivine fractionation had occurred during emplacement, the mare volcanic glasses may be samples of primary melts from the deep lunar interior (1-5). If that view is correct and those liquids were multiply saturated in the lunar mantle, then the liquidus phase relations of the volcanic glasses should furnish reliable information about the depths and residual mineralogies of the mare source regions.

The most recent additions (Table 1) to the growing list of volcanic glasses have brought the present total to twenty-three chemical varieties.

#### BOUNDARY CONDITIONS

Investigations of mare basalts and volcanic glasses have provided valuable insights into the Moon's early evolution. Eight constraints that are considered by the author to be of particular significance are listed below:

[A] The source regions of most mare liquids became isotopically closed at 4.3-4.4 AE (e.g., 6-11) and had some chemical characteristics that were complementary to the highlands crust (e.g., 12).

[B] Mare basalts and volcanic glasses were extruded within the time interval from 3.2 to 3.9 AE (7,9-11,13,14). Additional mare volcanism may have occurred outside that period (15-18).

[C] Contrary to earlier views, mare basaltic liquids do not exhibit any simple correlations among composition of melt, age of eruption, and inferred depth of source region (1,3,4,19,20).

[D] Mare source regions were chemically and isotopically heterogeneous on small scales (1,7,21,22).

[E] The Mg/(Mg+Fe) ratio of the *residual* olivine (and low-Ca pyroxene) in the mare source regions was nearly uniform at  $\sim 0.80 \pm 0.02$ , regardless of the abundances of titanium and incompatible trace elements in the mare liquid (2-5,23,34,36).

[F] Mare volcanic glasses suggest that their differentiated source regions were located at depths ranging from 350 to 500 km (2-5,20). The validity of this, however, hinges on confirmation of a key assumption (i.e., multiple saturation).

[G] Chemical systematics occur among mare volcanic glasses (Fig. 1) collected from widely separated landing sites (1).

[H] Volatiles associated with mare volcanic glasses (Table 2) indicate that volatile-enriched, undifferentiated regions existed within the Moon at the time of mare petrogenesis (24-30).

The bearing of these factors on two prominent questions in lunar science are examined in the next two sections of this abstract.

#### HOW DEEP WAS THE MAGMA OCEAN?

Estimates for the initial depth of the magma ocean have ranged from 200-300 km (31-33), to  $\sim 400$  km (34-36), to  $\sim 500$  km (37), to 1738 km (38). While constraints [A,F] imply that the magma ocean had a thickness of  $\gtrsim 500$  km,

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constraint [H] indicates that the whole Moon was not involved. A seismic discontinuity apparently occurs at  $\sim 400$ -500 km (39,40). Some investigators have speculated that that feature marks the limit of the magma ocean (20,21, 35). However, until a sample of primary basaltic liquid derived from undifferentiated Moon is discovered and examined by experimental petrologists, a bracket on the depth of the magma ocean will be difficult to constrain observationally.

*DID THE MOON ACCRETED HOMOGENEOUSLY?*

Although lunar samples are depleted in volatiles relative to chondrites and terrestrial rocks, Pb-isotopes in mare volcanic glasses (24-29) and volatile-rich breccia 66095 (41) suggest that this may not be the whole story. It seems that two possibilities exist: if the Moon accreted homogeneously, then it is not as depleted in volatiles as commonly believed (26,29,41); alternatively, the Moon may have accreted heterogeneously with respect to the volatile elements (42,43). This issue may be resolved by isotopic analyses of noble gases trapped within vesicles in mare volcanic glasses (44).

Table 1. Major-element compositions ( $\pm 1\sigma$ ) of lunar volcanic glasses from Apollo 17 that have been investigated subsequent to the listing that occurs in [1]. Within any group where a crystal/liquid fractionation trend was observed, the least fractionated composition has been tabulated.

	Apollo 17 yellow (II)	Apollo 17 orange (II)*	Apollo 17 orange (I)*	Apollo 15 green (B)
SiO <sub>2</sub>	40.5 $\pm$ 0.1	38.8 $\pm$ 0.3	39.4 $\pm$ 0.1	46.0 $\pm$ 0.1
TiO <sub>2</sub>	6.90 $\pm$ 0.06	9.30 $\pm$ 0.09	8.63 $\pm$ 0.13	0.40 $\pm$ 0.02
Al <sub>2</sub> O <sub>3</sub>	8.05 $\pm$ 0.07	7.62 $\pm$ 0.03	6.21 $\pm$ 0.07	7.92 $\pm$ 0.06
Cr <sub>2</sub> O <sub>3</sub>	0.63 $\pm$ 0.01	0.66 $\pm$ 0.05	0.67 $\pm$ 0.03	0.55 $\pm$ 0.03
FeO	22.3 $\pm$ 0.2	22.9 $\pm$ 0.1	22.2 $\pm$ 0.1	19.1 $\pm$ 0.1
MnO	0.25 $\pm$ 0.02	0.29 $\pm$ 0.03	0.28 $\pm$ 0.01	n.a.
MgO	12.6 $\pm$ 0.1	11.6 $\pm$ 0.04	14.7 $\pm$ 0.06	17.2 $\pm$ 0.1
CaO	8.64 $\pm$ 0.02	8.55 $\pm$ 0.07	7.53 $\pm$ 0.04	8.75 $\pm$ 0.04
Na <sub>2</sub> O	0.39 $\pm$ 0.01	0.39 $\pm$ 0.02	0.41 $\pm$ 0.03	n.d.
K <sub>2</sub> O	n.d.	n.d.	0.04 $\pm$ 0.01	n.d.
$\Sigma$	100.3	100.1	100.1	99.9
Mg#	50.2	47.4	54.1	61.6

\* Chemically distinct from the well-known orange glass in 74220, which was used as a working standard during these microprobe analyses. (Analysis of Apollo 17 glasses was still in progress at the time of writing.)

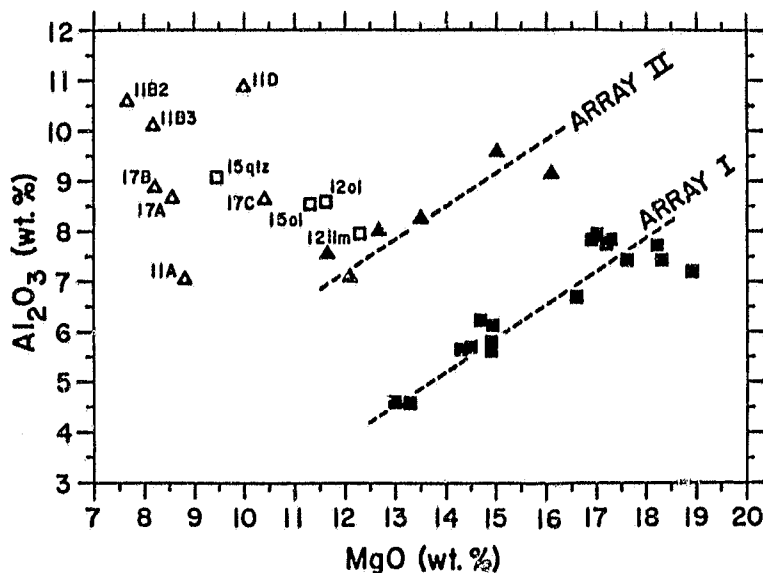


Figure 1. The mare volcanic glasses (solid symbols) form two subparallel, linear arrays. Mare basalts (open symbols) generally have lower abundances of MgO than the volcanic glasses. The least-fractionated compositions of the twenty-three varieties of glass plotted here are listed in (1) and in Table 1 of this abstract.

Table 2. Analyses of lunar volcanic glasses have indicated that the 24 elements listed below were constituents in the gases associated with lunar fire-fountains. The sources of data are given in parentheses.

B	(45)	Ag	(50,66,72)
C	(73-76)	Cd	(50-52,66,68,71,72)
F	(24,46-49)	In	(50,51,68,72)
Na	(50-52,71,77)	Sb	(50)
S	(24,52-57,71,77)	Te	(68)
Cl	(24,46,71,77)	I	(69)
Ar	(58-65)	Xe	(69)
Cu	(24,49)	Au	(50,51,58)
Zn	(24,49-54,66-68,71,72,77)	Hg	(68,70)
Co	(24,49,51)	Tl	(24,66,72,78)
Ge	(50,51,68,72)	Pb	(24-29,52,67,78)
Br	(46,66,69)	Bi	(50)

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MINERALOGY AND PETROLOGY OF THE PRISTINE ROCKS: Odette B. James, 959 National Center, U.S. Geological Survey, Reston, VA, 22092.

### Introduction

The ultimate goal of the studies of pristine highland rocks is to determine the constitution, origin and history of the lunar crust. Because pristine rocks are igneous rocks that crystallized from internally generated melts, they are the basic building blocks of the crust. Their compositions give information on the compositions of their parent magmas and the sources of these magmas.

### Processes That Alter Original Igneous Characteristics

Determining the original igneous characteristics of pristine rocks is not usually straightforward, because all pristine rocks have been modified, to some extent, by impact or metamorphism. For this reason, the igneous characteristics must be inferred after subtracting out the inferred modifications.

Granulation and mixing: Many pristine rocks have been granulated by impact. In granulated rocks, there is always the possibility of mixing. If unrelated rocks are mixed to form a polymict breccia, the effects are usually easy to detect. If rocks from the same or related plutons are mixed, however, there may be almost no evidence for mixing, but the bulk composition of the resultant breccia may not be simply related to the composition of the parent magma and its mineralogy may not be simply related to igneous processes.

The Apollo 16 ferroan anorthosites provide an example of a way in which mixing can obscure original igneous characteristics, in this case, the compositional variability of the minerals. The variability of minerals in an igneous rock provides clues to its depth of origin. All other conditions being equal, rocks formed deeper will cool more slowly so that subsolidus homogenization of mineral compositions by diffusion will be more effective. In many anorthosites, the mafic minerals have wide compositional ranges; if the mafic minerals of the original anorthosites were as variable, then the rocks probably formed near the lunar surface. The anorthosites in question, however, are probably mixed rocks that consist of clasts from a single pluton [17,21,24,33]. This mixing limits the information that can be obtained on depth of origin.

Reheating: Common processes that involve heating are shock, metamorphism, and incorporation in impact melt. While the bulk compositions of rocks reheated by these processes may remain pristine, their isotopic systematics and mineral compositions are likely to be altered.

Contamination: Another process that can affect the rocks is the addition of contaminant material of nonmeteoritic origin. A granulated clast included in impact melt is susceptible to contamination by the surrounding melt because of its high porosity. Because many impact melts are much more KREEP rich than most pristine rocks, addition of only a small amount of melt-derived material can greatly affect bulk composition. The 72255 Civet Cat norite [34] has probably been modified by such contamination. This norite is mineralogically nearly identical to the other Apollo 17 Mg-suite norites, but it has much more abundant REE, with a KREEP pattern, than the other norites [3]. It also has lower Ti/Sm and Sc/Sm ratios [3,26] and a younger Rb-Sr age [7]. These anomalies can be explained if Rb [7] and REE were added to the clast from the surrounding KREEPy melt. The texture of the clast indicates that such contamination is likely; the norite was shocked and granulated prior to being incorporated in the 72255 melt and subsequently recrystallized.

Another example of possible addition of nonmeteoritic contaminant is dunite 72415-7. This rock had a complex history, which included granulation and subsequent permeation by plagioclase-rich melt [14,18]. The source of this melt is uncertain. Most workers have assumed it represents remobilized interstitial material from the original dunite, so that the bulk composition of the rock was unchanged, but there is no real justification for this assumption.

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Characteristics of the Major Suites of Pristine Rocks

Highland plutonic rocks can be divided into two suites that differ in modes and mineral compositions [38,39] and also in trace-element ratios [26]. The two suites have been termed the ferroan-anorthosite suite and the Mg-suite [13,38,39].

Modal mineralogy: Rocks of the ferroan-anorthosite suite are mostly anorthosites and troctolitic anorthosites. Plagioclase is by far the dominant mineral, making up >90% of most of the rocks. Olivine is the next most abundant mineral. Low-calcium pyroxene and/or high-calcium pyroxene are present in minor amounts in most rocks. Other phases are rare and present in trace amounts; these are ilmenite, chrome spinel, a silica mineral, troilite and iron metal.

The Mg-suite has a far greater range of rock types than the ferroan-anorthosite suite. Most samples are spinel-bearing troctolites, troctolites, and norites; but there are also several gabbronorites, several gabbros, a feldspathic lherzolite and a dunite. Mafic minerals are more abundant than in the anorthosites, and, in most rocks, low-calcium pyroxene is dominant over high-calcium pyroxene. Trace minerals are common and include K-feldspar, chrome spinel, a silica mineral, apatite, whitlockite, baddeleyite, armalcolite, rutile, troilite and iron metal.

Mineral compositions: In the ferroan-anorthosite suite, plagioclase is very calcic and is about the same composition in all the rocks (Fig. 1). In most rocks of this suite, the mafic minerals have compositional ranges of ~10-25% En content [17,21,22,33]. As mentioned above, some of this variation may be due to mixing because the rocks are probably not monomict [17,21,33]. The trend of mineralogic variation in the suite, as shown on Fig. 1, is steep but not quite vertical [17,25,33].

The Mg-suite shows an overall trend of mineralogic variation having a shallower slope on Fig. 1 than that of the ferroan-anorthosite suite. The Mg-suite also shows modal variations correlated with the mineralogic variations, e.g., in the magnesian part of the trend, olivine is the dominant mafic mineral and in the more iron-rich part of the trend, pyroxenes are dominant. None of the samples has as variable mafic minerals as most of the anorthosites.

Trace-element ratios: The two suites show distinctive patterns in Ti/Sm and Sc/Sm ratios [26]. Most of the Mg-suite rocks have lower Ti/Sm than the ferroan anorthosites (Fig. 2). A difference in Ti/Sm ratio as great as that between the ferroan anorthosites and the Mg-suite rocks usually indicates a difference in parent-magma composition, because the mineral fractionations occurring during crystallization of magmas like the inferred parent magmas of the two suites do not appreciably change Ti/Sm ratios in the melts [16,19,26].

The Mg-suite rocks also have lower Sc/Sm ratios than the ferroan anorthosites (Fig. 3). The Sc/Sm ratios can be more easily changed by fractionation than Ti/Sm ratios, however, especially by fractionation of pyroxene [16]. Removal of pyroxene from a magma decreases Sc/Sm of the remaining melt, and concentration of pyroxene in cumulates increases Sc/Sm of the cumulates. Pyroxene fractionation of this sort probably explains some of the trends on Fig. 3 [16]. First, the decreasing Sc/Sm correlated with decreasing Mg' shown by the ferroan anorthosites could be the result of pyroxene separation from the parent magma. Second, the fact that most Mg-suite norites have higher Sc/Sm than Mg-suite troctolites could be the result of pyroxene accumulation in the norites.

Depths of origin: Some data are available for depths of origin of the Mg-suite rocks. Depths have been estimated using thermodynamic data for troctolites and spinel-bearing troctolites in which high-pressure mineral assemblages are present. In the troctolites, a high-pressure reaction between olivine and plagioclase has formed symplectic intergrowths of low-Ca pyroxene, high-Ca pyroxene and aluminous spinel [9]; these rocks probably equilibrated between 10 and 30 km below the surface [9]. In the spinel troctolites, alumi-

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nous low-Ca pyroxene coexists stably with aluminous spinel [1,10]; these rocks probably equilibrated at depths >26 km [11].

Depth of origin of the anorthosites is more difficult to estimate, largely because of the absence of suitable mineral assemblages. The scale of pyroxene exsolution, the grain sizes of the rocks, and the scale of the mineralogic variations all suggest that the Apollo 16 anorthosites did not form especially deep in the lunar crust [13]. The troctolitic anorthosites of this suite do not show the high-pressure reaction between olivine and plagioclase observed in some Mg-suite troctolites, so they may have formed at shallower levels.

#### Characteristics of Other Types of Pristine Rocks

Recently, the interpretation that there are two suites of pristine rocks, although basically correct, has been shown to be oversimplified. The Mg-suite rocks especially seem to be more diverse than previously thought. Mg-suite noritic rocks have been divided into two groups, termed Mg-norites and Mg-gabbro-norites [16]. A previously unknown type of Mg-suite rock, termed alkali anorthosite, has been discovered [40,41]. Apollo 14 Mg-suite troctolites have been found to differ from other Mg-suite troctolites in trace-element contents and ratios [41].

Mg-norites and Mg-gabbro-norites: Mg-norites and Mg-gabbro-norites differ in several critical parameters [16]: 1) a higher proportion of the total pyroxene is high-Ca pyroxene in the Mg-gabbro-norites than in the Mg-norites; 2) the Mg-gabbro-norites have a more sodic mineralogic variation trend than the Mg-norites, with a shallower slope on Fig. 1; 3) the Mg-gabbro-norites have higher Ti/Sm (Fig. 2) and Sc/Sm (Fig. 3) ratios than the Mg-norites; 4) assemblages and compositions of trace minerals differ (e.g., trace minerals rich in incompatible elements are more abundant in the Mg-norites than in the Mg-gabbro-norites); and 5) the Mg-gabbro-norites are younger than the Mg-norites, ~4.2 b.y. old versus ~4.4 b.y. old [5,6,7,23,28,29,30].

Alkali anorthosites: Most alkali anorthosites are cataclasites and recrystallized cataclasites, with some evidence for an original coarse grain size but poor preservation of igneous texture [41]. Because most samples are very small, their modes are not necessarily representative of their parent rocks. Plagioclase is by far the dominant mineral; the largest sample contains 84% plagioclase and 16% pigeonite. Minor minerals are augite, olivine, K-feldspar, ilmenite and iron metal. These rocks fall roughly on the extension of the Mg-norite mineralogic variation trend (Fig. 1). The Ti/Sm ratios (Fig. 2) and Sc/Sm ratios (Fig. 3) are similar to those of the Mg-norites, though slightly lower.

Apollo 14 troctolites: The Apollo 14 troctolites [41] consist mostly of plagioclase and olivine; accessory ilmenite, chrome spinel, low-Ca pyroxene, high-Ca pyroxene, and K-feldspar have been reported. The mineral compositions (Fig. 1) plot along the Mg-suite trend, but incompatible-element contents and Eu/Sm ratios are higher than in other Mg-suite troctolites. Because Ti and Sc contents are about the same as in the other troctolites, the Ti/Sm and Sc/Sm ratios are lower (Figs. 2, 3).

Evolved rocks: These are the quartz monzodiorite, the felsites and the granites. Most of these rocks have relict igneous textures. The quartz monzodiorite [32,37] is rare and consists mostly of plagioclase, low-Ca pyroxene, high-Ca pyroxene, K-feldspar, and a silica mineral; minor and trace minerals are ilmenite, zircon, whitlockite, chromite and iron metal. The granites and felsites are more common and widespread [15,35,42] but typically are found only as very small clasts. These rocks consist mostly of quartz and K-feldspar; variable amounts of plagioclase, iron-rich pyroxenes, iron-rich olivine, ilmenite, iron metal, zircon, and phosphate minerals are present. The mineral compositions (Fig. 1) fall close to those of the alkali anorthosites, roughly on the extension of the Mg-norite trend. The Ti/Sm (Fig. 2) and Sc/Sm ratios

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(Fig. 3) also fall along the same trend as the alkali anorthosites.

KREEP basalt: Only one type of highlands pristine volcanic rock has been recognized, the Apollo 15 KREEP basalt. (The Apollo 17 KREEPy basalt may be a KREEP-mare basalt hybrid, because in many characteristics it is intermediate between Apollo 15 KREEP basalt and mare basalt [36].) The textures of KREEP basalt indicate that its parent magma was probably endogenous and crystallized rapidly at the lunar surface [8,12]. The major minerals (Fig. 1) are relatively sodic plagioclase and relatively iron-rich orthopyroxene and pigeonite, with compositions that fall roughly along the Mg-suite trend. Ti/Sm ratios (Fig. 2) plot close to the Mg-norite trend.

#### Interrelations of the Various Groups of Pristine Rocks

On the basis of the parameters discussed herein, I propose the following scheme for interrelations of the various groups of pristine rocks. The ferroan anorthosites and associated troctolitic anorthosites constitute one distinct suite, the ferroan-anorthosite suite; all other pristine rocks constitute a second distinct suite, the Mg-suite. The Mg-suite is divided into three groups: first, the Mg-gabbronorite group, which includes the sodic ferrogabbro; second, the Mg-norite group, which includes most of the Mg-suite troctolites and spinel troctolites, the alkali anorthosites, and the evolved rocks; and third, the Apollo 14 troctolites (grouped separately because they have no clear affinity with any other group). The affinity of the Mg-norites with most Mg-suite troctolites and spinel troctolites, alkali anorthosites, and evolved rocks is established by the An vs. En, Ti/Sm, and Sc/Sm trends and by the presence of similar assemblages of trace minerals.

KREEP basalt is probably related to the Mg-suite, specifically to the Mg-norite group, as indicated by the following evidence. 1) The pyroxene-plagioclase compositions fall roughly along the Mg-suite trend. 2) Ti/Sm and Sc/Sm ratios plot along the Mg-norite trend (Fig. 2,3). 3) Calculated parent liquids for several rocks of the Mg-norite group have KREEP rare-earth patterns [2,43]. 4) The Sm/Nd data suggest a relationship between troctolite 76535, which is a member of the Mg-norite group, and KREEP basalt [4].

#### Genesis

The rocks of the ferroan-anorthosite suite are generally considered to have formed as igneous plagioclase cumulates. Many theories have been advanced to explain the near-vertical trend of mineralogic variation of these rocks as shown on Fig. 1. One theory [31] is based on analogy with the Stillwater complex, where the anorthosites show a similar trend. This hypothesis considers the anorthosites to be mixtures of three components: floated plagioclase crystals; trapped, relatively magnesian cumulus mafic crystals; and trapped, relatively iron-rich late-stage liquid. The wide range in mafic-mineral composition is explained as the result of local reequilibration between the trapped liquid and cumulus mafic minerals, with the relative proportions of these two constituents as well as their relative compositions determining the final composition of the mafic minerals in any given area. The other theory [33] considers the anorthosites to be the product of nearly perfect adcumulus growth during strong fractional crystallization of a parent magma that had a very low sodium content. Two lines of evidence strongly favor the adcumulus hypothesis over the trapped-liquid hypothesis: the very low trace-element contents of the rocks suggest very low contents of trapped liquid; and the pronounced decrease of Sc/Sm ratio with decreasing Mg' suggests extensive fractionation of pyroxene during crystallization.

The relict textures and compositional variations of the minerals of the Mg-suite plutonic rocks indicate that they are cotectic igneous cumulates. Cumulus minerals were olivine and plagioclase in the troctolites and low-Ca pyroxene and plagioclase in the norites and gabbronorites. Most of the rocks

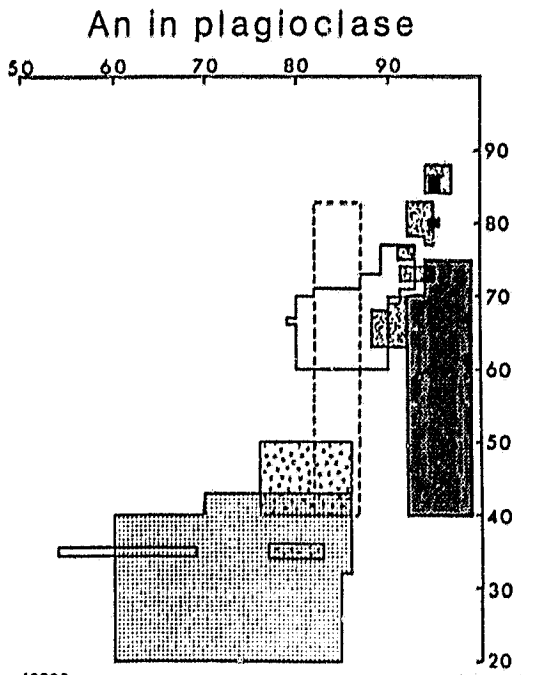


Figure 1. An vs. En for pristine rocks. Boxes enclose areas of overall variation for groups; individual rocks are not plotted where overlap occurs. Dashed line: KREEP basalt. Solid lines: irregular stipple, Mg-norites and related troctolite; unpatterned, Mg-gabbro-norites; black, Apollo 14 troctolites; v's, alkali anorthosites; fine dots, evolved rocks; gray, ferroan anorthosites and related troctolitic anorthosites. Most sources of data are given in [11,14,38]; additional data in [39,40,41].

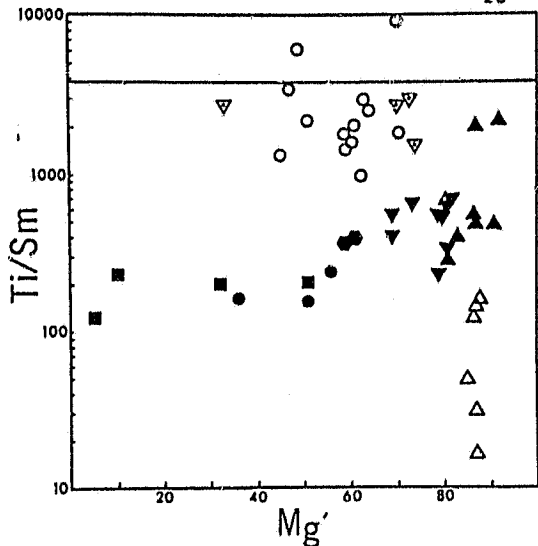


Figure 2. Ti/Sm vs. Mg' in pristine rocks. The horizontal reference line indicates the chondritic ratio. Symbols are as follows: open circles, ferroan anorthosites; filled circles, alkali anorthosites; filled upward-pointing triangles, troctolites and spinel troctolites of Mg-norite group; open upward-pointing triangles, Apollo 14 troctolites; filled downward-pointing triangles, Mg-norites; downward-pointing triangles with centered dot, Mg-gabbro-norites; filled squares, evolved rocks; filled hexagons, KREEP basalts. Most sources of data are given in [14,24]; additional data in [39,40,41].

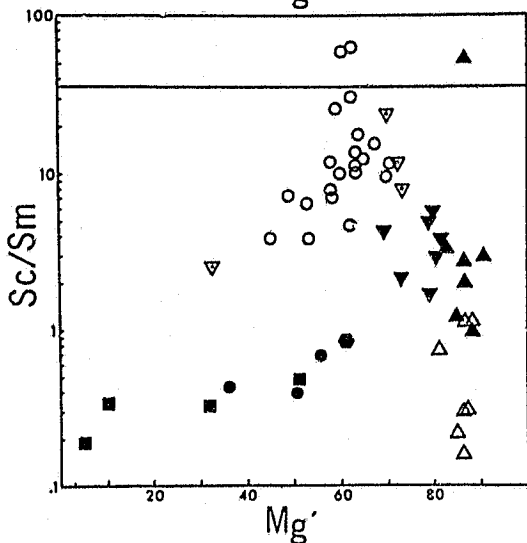


Figure 3. Sc/Sm vs. Mg' in pristine rocks. Symbols and sources of data are the same as for Fig. 2.

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contain interstitial minerals crystallized from trapped late-stage liquids.

As suggested by [13,20], KREEP-basalt magma probably formed by remelting of Mg-suite rocks. The evidence is as follows. As summarized above, the bulk-compositional and mineralogic data indicate that KREEP basalt is related to the Mg-norite group of the Mg-suite. Although KREEP basalt crystallized ~3.9 b.y. ago [4,27,30], the Sm-Nd and Rb-Sr data suggest that its parent magma formed by extensive remelting of a distinctive source that was produced earlier, roughly 4.3 b.y. ago [4]; this age is about the same as, or slightly younger than, the crystallization age of the rocks of the Mg-norite group [5,6,7,23,28,29,30].

A corollary of the proposed grouping of pristine rocks is that at least three distinct types of parent magma are required to have formed the observed rocks: a ferroan-anorthosite magma, a Mg-norite magma, and a Mg-gabbro-norite magma. (A fourth magma may be required for the Apollo 14 troctolites.) Hypotheses can probably be formulated to derive all these rocks from a single magma ocean, but such hypotheses would be extremely complex. Simpler hypotheses, involving distinct parent magmas, are more likely. Determination of the compositions of these magmas and the processes by which they formed are major research directions for the future.

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COMPOSITIONAL RELATIONSHIPS OF THE PRISTINE NONMARE ROCKS TO THE HIGHLANDS SOILS AND BRECCIAS. Randy L. Korotev, Dept. of Earth and Planetary Sciences and the McDonnell Center for the Space Sciences, Washington University, St. Louis, MO 63130.

The small subset of nonmare lunar samples which are recognized as "pristine" [19] can be divided into three general groups 1) ferroan anorthosites and norites, 2) magnesian norites, troctolites, and related materials (the "Mg-rich" suite), and 3) LIL-element rich norites/basalts and other materials of similarly evolved compositions. Most attempts at modeling mixed crustal materials - the soils and breccias that comprise the bulk of the highlands sample collection - as mixtures of more primary rock types have concluded that components with chemical characteristics similar to each of these three groups of pristine rocks are the minimum required (e.g., [13,18,16,1,23,11,6]). (A meteorite component and, for samples like Apollo 15 and 17 soils, a mare basalt component are also required, but these may be regarded as contaminants for the purpose of understanding the highlands crust.)

Since the advent of the pristine rock concept it has usually been assumed that the suite of pristine rocks currently recognized is sufficient to explain the bulk composition of the crust. Warren and Wasson [20,21] note that among the breccias and soils, bulk compositions corresponding to anorthositic norite (gabbro) and LKFM basalt (e.g., [10]) predominate, yet these compositions are not represented among the pristine rocks but are intermediate to them. They explain that the compositions "are intermediate because they are preponderantly mechanical mixtures of abundant pristine rock types" [20]. Ryder [11] suggests that "most highland breccias can probably be modelled using anorthosite, [pristine Mg-] norite, dunite, a meteoritic component, and variable KREEP as end members. An ultramafic component dissimilar to the known pristine samples is not required." Warren and Wasson [21] have noted the similarities between pristine 67667 ("feldspathic lherzolite") and their SCCRV component, a hypothetical ultramafic composition derived from mafic-rich breccia compositions and used with KREEP and anorthosite to model these breccias [23].

We have argued [5,6], however, that although the minimum number of major components necessary to model the highlands crust is three and that anorthosite and some KREEP-like component are certainly two of these, that the third component - the major carrier of Fe, Mg, and related elements - is compositionally dissimilar to the Mg-rich pristine rocks. These geochemical arguments have been directed primarily against the dunite, troctolites, and Mg-norites and are reviewed below.

Mg and Fe. Average  $mg'$  values (molar Mg/(Mg+Fe)) for sampled highlands soils from four sites (Apollo 15, 16, 17 and Luna 20) fall into the remarkably narrow range of 0.68 to 0.72 when corrected for the effects of mare basalt and meteorite contamination (Table 1). Values for averages of LKFM glasses from these same sites average from 0.66 to 0.73 (Table 1). Apollo 14 soil values are lower at 0.62 (no basalt correction). Values of  $mg'$  for bulk breccias from Apollos 16 and 17 average near the soil values. Of "mixed" samples, only VHA-type impact melts from Apollo 16 tend to be higher (range 0.65-0.80, typically 0.75).

These data are difficult to reconcile in terms of mixtures of pristine rocks alone. The ferroan anorthosites ( $mg' = 0.4-0.7$ , typically 0.5 to 0.65) contain too little Fe and Mg to affect significantly the Mg/Fe ratio of any mixture. Most of the pristine norites and troctolites are too magnesian ( $mg' > 0.8$ ) to account for the Mg/Fe ratio in the soils and breccias. Although mixtures of KREEP ( $mg' < 0.65$ ) and Mg-rich rocks can yield values in the 0.7 range, all such mixtures contain far greater LIL element concentrations than are found in the soils and breccias. Some pristine Mg-norites do have low values of  $mg'$ : 77075/7 (0.695/0.68) and possibly the norite clast from 76255 (0.67 from modal recombination data in [12]). These samples have Sc/Sm and Ti/Sm ratios similar to other Mg-norites and thus cannot account for the ratios of these elements in the mixed samples (below).

The similarity of Mg/Fe ratios in the soils from these four sites (which range over a factor of 2 and 3 in Al and LIL element concentrations) suggests but does not require that a single component with an  $mg'$  of about 0.7 is a major carrier of Fe and Mg in these samples. Alternatively, Fe and Mg are carried by a mixture of components with some more ferroan and others more magnesian than 0.7 which recur in approximately the same proportions from site to site.

TABLE 1. Average FeO and MgO concentrations and molar Mg/(Mg+Fe) ratios ( $mg'$ ) in soils from highland sites, raw and corrected for mare basalt and meteorite contamination (data from Korotev et al., 1980). LKFM data from Reid et al. (1977).

	raw mean			minus		corrected mean			LKFM $mg'$
	%FeO	%MgO	$mg'$	%MB	%CC1	%FeO	%MgO	$mg'$	
A14	10.45	9.4	0.615	0	2.7	10.1	9.25	0.62	---
A15-AF	12.25	11.1	0.62	27.4	1.67	8.2	11.2	0.71	0.67
A16	5.38	6.0	0.665	0	3.2	4.77	5.67	0.68	0.73
A16-NRC	3.93	4.35	0.665	0	0.85	3.76	4.25	0.67	---
A17-SM	8.6	10.05	0.675	7.5	1.79	7.3	9.9	0.705	0.70
A17-NM	10.55	10.95	0.65	21.2	1.29	7.95	11.2	0.715	
L20	7.5	9.15	0.685	0	1.35	7.3	9.05	0.69	0.66

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Sc, Ti, Sm. On a plot of  $Ti/Sm$  vs.  $Sc/Sm$  (not shown, but see Fig. 18, in [21] and Fig. 11, in [22]) soils and breccias plot along a line between KREEP (low ratios) and anorthosite (high ratios). Warren and Wasson [22] observe that the trend is due to varying KREEP content which increases Sm much more than Sc and Ti. Upon casual inspection the trend appears to be the result of mixtures of KREEP and anorthosite. Although pure KREEP-anorthosite mixtures do plot along a line connecting the two components (actually, a curve slightly concave downward on this log-log plot), KREEP and anorthosite alone are not responsible for the soil/breccia trend. Apollo 15 soils, for example, fall on a region of such a curve corresponding to 0.2-0.7% KREEP, far less than must be present to account for the LIL element concentrations in the soils. In other words, anorthosite does not contain enough Fe and Ti to alter the overall ratio and thus account for the trend. The plot implies the importance of a component with  $Sc/Sm$  and  $Ti/Sm$  ratios similar to ferroan anorthosite (or perhaps even higher) but with significantly greater concentrations of all three elements. This component is clearly not the typical pristine troctolite or Mg-norite. All of these plot together in a field off the trend, with  $Ti/Sm$  ratios too low to account for the trend. The small amount of dispersion about the line again suggests that the component in question may be a single rock type, or perhaps a mixture of rock types with similar  $Sc/Sm$  and  $Ti/Sm$  ratios.

We are thus left with two extreme possibilities: (1) The samples that comprise the majority of the Mg-rich suite of pristine rocks are volumetrically unimportant to the mixed highlands materials and instead another chemical component with higher Fe/Mg,  $Sc/Sm$ , and  $Ti/Sm$  carries the bulk of the Fe, Mg, Sc, and Ti, or (2) the Mg-rich rocks are important but there is in addition another component which is at least equally important and which has a composition even more dissimilar to the Mg-rich rocks (i.e., still higher Fe/Mg,  $Sc/Sm$ , and  $Ti/Sm$ ) than that of alternative (1). Intermediate cases are also possible (i.e., two components more ferroan than the Mg-rich rocks). In any case, another component is needed. Several lunar rock types have characteristics similar to the implied missing component(s). These are discussed below in the context that each of them alone might account for the compositions of Apollo 16 and 17 soils and breccias as mixtures with KREEP and ferroan anorthosite.

Mare Basalt. These all have high  $Sc/Sm$  and  $Ti/Sm$  ratios. The Apollo 12 and 15 mare basalts and the KREEP-poor Apollo 14 basalts (e.g., 14053) are similar to the ferroan anorthosites in this regard and have the necessarily high concentrations of the three elements. Mg/Fe ratios are so low ( $mg' = 0.4-0.5$ ), however, that high Mg/Fe components like dunite or troctolite are then also necessary to account for the intermediate values in the soils and breccias. No known mare basalt when mixed only with dunite (troctolite, or Mg-norite); anorthosite, and KREEP can explain, for example, Apollo 16 soil and breccia compositions when other elements are taken into account. Kempa et al. [4] have claimed success with Apollo 16 soils using a modified version of this model. Apollo 15 pigeonite basalt was used with an unusually troctolitic impact melt (62295,  $mg' = 0.81$ ), anorthosite (67075), and a poikilitic impact melt (60315,  $mg' = 0.71$ ) as the KREEP component. The model, however, consistently and significantly underestimates Sr and Eu concentrations and the  $mg'$  value, overestimates Si, Al, and Ca concentrations, and predicts 7-8% basalt component even in the North Ray Crater (NRC) soils.

Basalt/dunite or basalt/troctolite combinations are not a satisfactory geochemical explanation for the overall Mg/Fe,  $Sc/Sm$ , and  $Ti/Sm$  ratios in mixed highlands materials. They would fortuitously have to occur in similar proportions in a wide variety of materials. The quantity of mare basalt required is just too high to explain materials like NRC soils and breccias and the anorthositic norites from Apollo 17. Finally, unless a mare basalt with a yet unobserved composition is invoked, some other relatively Fe and Mg-rich component is also required to account for the concentrations of all elements.

Ferroan Anorthositic Norites. The bulk compositions of many highlands soils and breccias correspond to anorthositic norite/gabbro. These have been interpreted as mixtures of pristine rocks, as no pristine rock of this general composition had been found [20,21]. Recently, however, samples of anorthositic norite have been found in breccias from North Ray Crater. These are more ferroan than previously known samples ( $mg' = 0.58 - 0.60$ ) and are uncontaminated by KREEP ( $REE = 3-5 \times CCI$ ) [9].  $Sc/Sm$  and (presumably)  $Ti/Sm$  ratios are similar to those in ferroan anorthosites, but absolute concentrations are 10-30 times greater. One of these samples, 67215, is a 273g cataclastic breccia which Taylor [14] suggests is probably pristine. Lindstrom [8] argues that these samples represent an important cumulate rock. Only one ferroan anorthositic norite has been positively identified as pristine, a small clast from 66035 which [21] thought to be an unusually mafic chip from a ferroan anorthosite ( $mg' = 0.59$ ). The compositions of the NRC ferroan anorthositic norites correspond well to a 2:1 mixture of the 66035 clast and ferroan anorthosite. No other mixture of pristine rocks can account for the composition.

As a mixing component ferroan anorthositic norite is appealing.  $Sc/Sm$  and  $Ti/Sm$  ratios as well as absolute concentrations are approximately those expected for the "missing" component. The Mg/Fe ratio is not as dissimilar to that of the soils and breccias as are other ferroan components (e.g., mare basalt, sodic ferrogabbro) needed to offset the high ratios in troctolites and Mg-norites. The large proportion of a ferroan anorthositic norite component predicted by any model using it to model highlands mixtures is not as petrologically unappealing as an equivalent proportion of some mixture of troctolite and basalt or various Mg-gabbro-norites with extreme compositions (below). No other component accounts more reasonably for the composition of those polymict, anorthositic samples with several percent FeO and MgO,  $mg' < 0.7$ , and low LIL-element concentrations [e.g., NRC soils and breccias; 15418 ( $mg' = 0.63$ , [2,17]), and 70017 ( $mg' = 0.63$ , [7])].

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However, ferroan anorthositic norite alone (with KREEP and anorthosite) cannot account for the mixed samples. A small amount of a magnesian component and a Ti-bearing component is still required. For example, a mixture of 50% anorthosite, 10.3% KREEP, 26% ferroan anorthositic norite, 5.4% dunite, 6.3% sodic ferrogabbro (below), and 3.7% meteorite component accounts as well for the mean Apollo 16 soil composition as any model yet suggested. The problems with Ba and Eu discussed below are significant, however.

Pristine Mg-Gabbronorites. This small subset of the Mg-rich suite was recently recognized and characterized by James [3]. It is composed of samples to which the preceding generalizations about Mg-rich pristine rocks do not apply. Although there is considerable compositional dissimilarity among the six samples classified into this subgroup, the four for which compositional data are available are characterized by high Ti/Sm, high Sc/Sm (except for the sodic ferrogabbro), and  $mg'$  values on the low end of the range for Mg-rich samples.

It is difficult to generalize about this group of samples. The compositions are too diverse to average as a single "Mg-gabbronorite component." Models using each of the chemically analyzed samples (61224 eucritic gabbro,  $mg' = 0.70$ ; 67667 feldspathic lherzolite,  $mg' = 0.73$ ; 67915 sodic ferrogabbro,  $mg' = 0.33$ ; and 73255, 25, 47 norite,  $mg' = 0.74$ ) together along with anorthosite and KREEP go a long way toward accounting for the compositions of diverse mixed samples such as Apollo 16 soils, the non-mare portion of Apollo 17 soils, VHA impact melts, and Apollo 17 noritic breccias. Including the other Mg-rich pristine rocks and ferroan anorthositic norites mathematically improves the fits even further, but no such fit approaches perfection.

Unfortunately, compositional mixing models are not well enough constrained to prove whether the Mg-gabbronorites are or are not important constituents of the highlands crust or are more important than ferroan anorthositic norites. Each of the four samples above is the unique representative of its particular rock type. We have no way of knowing how volumetrically important or compositionally diverse the rock types are (were). Experiments using the Mg-gabbronorites, ferroan anorthositic norites, and other pristine rock types to model various mixed samples are nearly inconclusive as a result of the combined effect of three problems: 1) the uncertainty associated with the variability in composition of the Mg-gabbronorites and KREEP, 2) the large number of components with similar compositions under consideration (all or none of which may be important), and 3) the small number of elements for which data in common are available. Consequently, there is no way of assessing the importance of poor fits for certain elements or model predictions of negative proportions of certain components.

A few generalizations can be made on the basis of such experimentation, however. (1) No mixture of known pristine rocks which does not include the sodic ferrogabbro (SFG, 67915, [15]) can account for the high Ti concentrations in most Apollo 16 and 17 material. Ryder [11] has discussed the problem of excess Ti in the mixed samples and has suggested that the Apollo 16 and 17 KREEP component is more Ti-rich than that at Apollo 14 and 15. This may be the preferred explanation as models including the SFG to account for the excess significantly overestimate Ba and sometimes Na (both are high in 67915). Lindstrom and Salpas [9], however, have found other SFG clasts in NRC breccias (which are generally low-KREEP materials) and have successfully used such a component to model the bulk breccias. (2) Models including only pristine rocks to account for the compositions of mixed samples always underestimate the Eu concentration by a significant amount (5-10%). This problem occurs for both low- and high-KREEP samples from Apollo 16 and 17. The excess is probably too great to attribute to variability in anorthosite or KREEP alone. Another component may be indicated which may or may not be related to the Na- and Eu-rich samples from 67711/2 [5]. (3) If the Mg-gabbronorites are invoked as the principle component responsible for the bulk Mg/Fe, Sc/Sm, and Ti/Sm ratios in the mixed samples, then the proportion of these components may be too large to explain on any reasonable petrologic grounds. For example, the average Apollo 16 soil composition is, in fact, surprisingly well-modeled as a mixture of 62% anorthosite, 10.5% KREEP, 8.9% 61224, 10.9% 67667, 4.6% 67915, 0.6% 67435 (spinel troctolite), and 3.8% CC1. Despite the success of the model on compositional grounds, it is difficult to rationalize on petrologic grounds (at least to a geochemist).

Conclusions. Pristine rocks are rare and atypical lunar samples. They are also difficult to recognize. There is no particular reason to believe that the suite of samples currently recognized as pristine should (or should not) account for the compositions of the mixed samples. There is no reason to believe that even with thousands of pristine samples the relative distribution of pristine rock types would correlate with their relative volumetric importance as components of the crust. It is conceivable that an important primary rock type might be underrepresented among the pristine rocks, or vice versa.

Evidence presented here argues that:

- (1) The Mg-norites and troctolites are probably not as volumetrically important to the mixed crustal materials as generally assumed and as might be expected based on their prevalence among the pristine rocks. A more ferroan component(s) with high Sc/Sm and Ti/Sm must be an important carrier of Fe, Mg, Ti, Sc, etc.
- (2) The Mg-gabbronorites may include that component, but the compositional diversity of the few known samples is so great that no definitive argument for or against them can be made.
- (3) Ferroan anorthositic norite may be that component. It is probably an important primary component of the crust which has been overlooked as such because of its current lack of representation among the pristine rocks. If it is and if it is a cumulate composed of roughly equal proportions of anorthosite and trapped liquid [8], then its importance to the evolution of the lunar crust and the understanding of the other pristine rocks should not also be overlooked.

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FERROAN ANORTHOSITIC NORITE: AN IMPORTANT HIGHLAND CUMULATE; M.M. Lindstrom, Dept. of Earth & Planetary Sciences & McDonnell Center for Space Sciences, Washington University, St. Louis, MO 63130.

Previous studies have shown that lunar breccias and soils are not simple mixtures of pristine highland rocks because the magnesian suite norites and troctolites do not have the right mg' value to be the only major KREEP-poor mafic-rich components in these polymict samples (1). Even for the well constrained case of North Ray Crater (NRC) feldspathic fragmental breccias, where many components were analyzed as breccia clasts, an unidentified ferroan mafic-rich component was required to fit the data (2). Candidates for this component were observed petrographically in 67455 by Minkin et al. (3) and in 67016 by Norman (4) as noritic (gabbroic) anorthosites. Analyses of separated 67016 clasts uncovered two such samples. Discussions with M. Norman and G.J. Taylor suggested that bulk rock 67215 might fall into this classification, and our analyses confirmed this.

Data for these three samples are given in Table 1 (5) with that of a plagioclase separate from 67215. These three samples are very similar to each other and represent the most mafic ferroan rock type yet analyzed. They have 20-30% mafic minerals with mg' the same as in the anorthosites (.55-.60). 67215 is described by Taylor (6) as a 273 g cataclastic monomict breccia that may be pristine. Its clasts have either cumulate or granulitic texture with the same mineral compositions and an average of 75% plagioclase. The two clasts selected from feldspathic fragmental breccia 67016 are plutonic rocks, but 67016,275 is partially metamorphosed with small regions of granular texture. Other ferroan mafic-mineral bearing anorthosites have been found (62236 and 62237) but they contain only 15% mafic minerals. Warren and Wasson (7) report on a small clast from breccia 66035 which has 40% mafic minerals, but due to its small size consider the clast to be a mafic clump from an anorthosite, and hence be due to poor sampling. The two clasts from 67016 might be explained in this way, but 67215 is a large rock with essentially the same composition and petrologic characteristics. This indicates that even the analyses of our small clasts are reasonably representative of a distinct highland rock type. In addition, such ferroan mafic rocks are required to account for compositions of bulk NRC breccias and melt rocks.

Returning to Table 1 and the REE plotted in Fig. 1 we see that these samples have quite low incompatible element concentrations with REE patterns more similar to those of anorthosites than to the KREEP patterns which dominate highland rocks. What, if any, relationship might these plutonic rocks have to the ferroan anorthosites? Both anorthosites and anorthositic norites have similar mineral compositions and relict cumulate textures. Might the anorthositic norites represent cumulate plagioclase with a larger percentage of trapped liquid than the anorthosites? These rocks contain 25% modal mafic phases which, if they crystallized from trapped liquid would have crystallized with an approximately equal amount of plagioclase. Thus, as a simple approximation, the samples may represent about 50% cumulate plagioclase and 50% trapped liquid. Since the plagioclase contributes little of the REE (except Eu), the concentrations in the trapped liquid would be approximately twice that in the bulk rocks, i.e., REE concentrations in trapped liquid would range from 6-13 times chondrites with a nearly flat pattern. Eu has approximately the same concentration in both plagioclase and liquid and the liquid has little Eu if any anomaly. Finally, one can calculate the REE pattern of the cumulate plagioclase in

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equilibrium with this trapped liquid using the distribution coefficients of McKay (pers. comm., 1982). These plagioclase patterns fall just below the range observed for ferroan anorthosites suggesting that the ferroan anorthosite norites could be cumulates related to the anorthosites but with more trapped liquids. The analyzed plagioclase fraction is REE-enriched compared to FAN and the equilibrium plagioclase. This could be due to trapped liquid contamination or crystallization from a late stage liquid. The probable relationship to anorthosites and the larger proportion of trapped liquid make these ferroan anorthositic norites important samples in the study of the nature of the anorthositic parent liquids and the evolution of the lunar highlands.

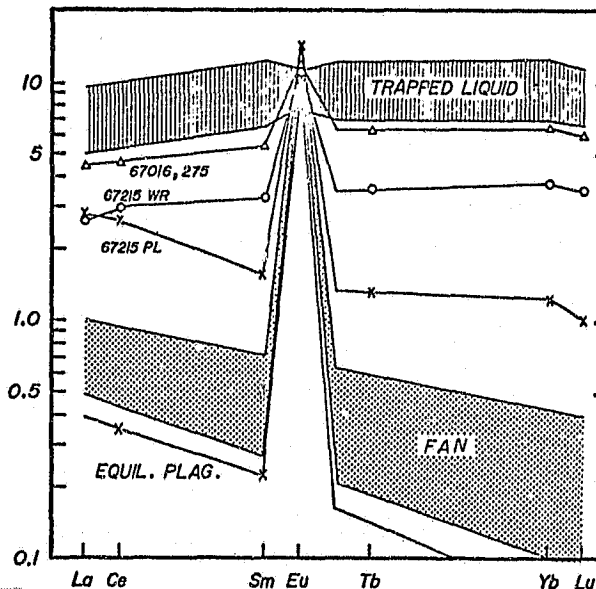
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Fig. 1. REE in Ferroan Anorthositic Norites and Anorthosites. Chondrite normalized REE patterns for ferroan anorthositic norites (67215 WR and 67016, 275) and separated plagioclase (67215 PL) are plotted. The range for calculated trapped liquid values is shown and the plagioclase in equilibrium with that liquid is compared to the observed range for ferroan anorthosites (FAN).

Table 1. Compositions of Ferroan Anorthositic Norites

	67215 WR	67215 PLAG	67016 275	67016 285
Al <sub>2</sub> O <sub>3</sub>	24.3	----	24.2	21.6
FeO	7.95	1.72	8.00	8.95
MgO	6.8	----	6.2	7.5
CaO	14.9	19.6	15.3	14.2
Na <sub>2</sub> O	0.302	0.417	0.292	0.272
Sc	17.3	4.7	17.8	23.8
Co	13.6	1.06	6.57	6.60
Ni	45	9±8	34	----
La	0.867	0.906	1.53	1.35
Sm	0.600	0.285	0.932	1.00
Eu	0.74	1.00	0.77	0.74
Yb	0.77	0.25	1.02	1.33



EARLY LUNAR IGNEOUS HISTORY. John Longhi, Dept. of Geology and Geophysics, P.O. Box 6666, Yale University, New Haven, CT, 06511.

The early history of the moon is recorded in ancient igneous rocks that have survived meteorite contamination and reworking. Decoding the message recorded in these "pristine" rocks is far from trivial, however, because sampling is incomplete, primary field relationships are poorly known, isotopic systems are commonly disturbed, and the volumes of the major rock types are not known. Pursuit of early lunar history through the study of pristine rocks is far from hopeless, though, rather it is a challenge that requires sharper analysis and interpretation than studies of terrestrial magmatism. The rewards for meeting the challenge are more than worth the effort for the history recorded in pristine lunar rocks--early planetary evolution and crustal formation--has been obliterated on earth.

Interpreting early lunar history through the study of pristine rocks requires data from five general disciplines: 1) petrology and petrography, 2) geochemistry, 3) isotopic dating, 4) photogeology and remote sensing, 5) geophysics. The following overview of lunar igneous history will draw discriminately on all five disciplines.

### I. The Anorthosite-Forming Event

Ferroan anorthosites apparently record the oldest and, perhaps, the most voluminous episode of lunar magmatism. This series of plagioclase-rich rocks is (1) characterized by a relatively narrow range of plagioclase composition ( $An \equiv Ca/(Ca + Na + K) = 0.94 - 0.98$ ) and a fairly wide range of olivine and pyroxene composition ( $Mg \equiv Mg/(Mg + Fe) = 0.7 - 0.4$ ). Although some samples are nearly pure plagioclase, others (2,3) contain significant amounts of olivine crystals with grain size similar to that of the plagioclase. The preferred interpretation (2,3) is that these rocks crystallized from a magma co-saturated with olivine and plagioclase (and possibly pyroxene as well). Experimental data and modeling show that the limited range of plagioclase composition coupled with the extensive range of Mg in mafic minerals in the anorthosite series is a normal consequence of fractional crystallization of a cotectic magma with low alkali concentrations and intermediate Mg (4). What is most unusual about ferroan anorthosites is the absence of mafic rocks (troctolite, norite, gabbro) or ultramafic rocks (dunite, pyroxenite) among the highland samples with mineral compositions similar to those in the anorthosites. Although mineral isochron ages have not been obtained, extremely low measured values of  $^{87}Sr/^{86}Sr$  (5) are consistent with these rocks having crystallized 4.5 to 4.6 G.y. ago.

Mixing models of highland soils and breccias (6) show that anorthosite comprises 60 to 70% of the highland surface near the Apollo 16 landing site. Orbital XRF data show soils with Mg/Al similar to that in the Descartes region to be typical of the lunar farside (7). Thus the anorthosite-forming event (AFE) was igneous, apparently moon-wide, and most probably represents the first stage of crustal formation.

How deep the anorthosite-rich surficial regolith persists is not known, but the depth of this layer is perhaps the most important constraint on the scale of the AFE. A minimum estimate of the depth of anorthositic regolith is the relief of the farside terrain, 5 to 8 km, across which the regolith persists. These depths imply a minimum 3 to 6 km column of anorthosite. If the cotectic magma from which the anorthosites crystallized separated completely into anorthositic and ultramafic fractions, then a 5 to 10 km column of cotectic magma is required. If intermediate rocks (subequal proportions of plagioclase and mafic minerals) formed from this magma, then the magma column could easily have been several times as thick as the column of anorthosites. This situation is readily observed in terrestrial layered intrusions such as the Stillwater complex where massive anorthosite makes up only about 1/6 of the exposed section (8). Furthermore, given the reported compositions of cumulus olivine in anorthosites (2,3), the cotectic parent magma of the anorthosites must have had a Mg value of about 0.4, which is far too low to be a primary melt of any reasonable whole moon composition, hence the parent magma of the anorthosites must have evolved by fractional crystallization of a much more voluminous primary magma.

These considerations suggest a moon-wide magma column tens of kilometers thick. The simplest conceptualization of the AFE is thus a "magma ocean", tens if not hundreds of kilometers deep, with anorthosites forming a stable buoyant crust (9, 10). Thermal modeling shows that a magma ocean 200-400 km deep with a floating crust that acts as a thermal boundary layer is consistent with the formation at 4.4 G.y. of a widespread incompatible element-enriched residuum (KREEP) and the source regions of the mare basalts (11). An initially hot exterior and cold interior is also consistent with the absence of large scale compressional or tensional features throughout lunar history (12).

Consideration of post-accretionary meteorite flux poses complications for magma ocean models, however, because plausible fluxes could easily have disrupted the floating crust, thus accelerating the cooling rate and decreasing the solidification time (13). Furthermore, slow accretion by relatively large projectiles (14) would tend to produce a heterogeneous array of overlapping magma seas rather than a global ocean. So the actual anorthosite-forming event may have been considerably more complex than magma ocean models imply. Valuable constraints could be placed on the AFE if an isopach map of the anorthositic layer of the lunar highlands were available.

Other possible explanations for the AFE include an outward zone-refinement of plagioclase



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during accretional growth (15) and growth of an anorthositic crust above a partially-molten convecting mantle (16). One possibility that seems to be excluded by absence of mafic rocks complementary to the anorthosites is that anorthosites crystallized as the upper portions of basaltic intrusions similar in scale to those found on earth.

## II. KREEP Genesis

KREEP is an ephemeral incompatible element-rich component readily discernible in soils and breccias, but hard to identify as a rock type. The component defined from Apollo 14 breccias by (17) has a negative slope on its chondrite-normalized REE abundance pattern with light REE approaching 500X Cl and heavy REE in excess of 200X Cl. This component is also characterized by spectacular relative depletions of certain elements with a dual character which are compatible only for certain minerals: there are negative anomalies relative to the REE for Sr and Eu, which indicate plagioclase fractionation, for Ti which indicates ilmenite fractionation, and for Sc which indicates mafic mineral (probably clinopyroxene) fractionation. By virtue of the high incompatible element concentrations a small amount of KREEP assimilated into an unfractionated magma or admixed into a breccia is sufficient to dominate the incompatible element abundance pattern.

The similarity of chondrite-normalized incompatible element abundance patterns of KREEP-bearing breccias at several near-side landing sites (17) and the fact that KREEP-breccias from various sites also tightly define a whole-rock Sm-Nd isochron of  $4.36 \pm .06$  G.y. (18) suggests a large-scale, very extensive, and relatively homogeneous differentiation at  $\sim 4.4$  G.y. This age is in general agreement with the inference of a large scale lunar differentiation at 4.42 G.y. suggested by U-Th-Pb systematics (19).

KREEP genesis occurred at a pivotal time in lunar history because various isotopic studies (20, 21) suggest that the mare basalt source regions (MBSR) were established at this time. There may be more than a coincidental relationship because the KREEP parent obviously lost ilmenite during its evolution (22), while the source regions of the high-Ti mare basalts obviously gained ilmenite (23). A popular concept is that the MBSR were cumulates of an evolving magma ocean that produced a floating anorthositic crust (20) and that KREEP was the ocean's ultimate residual liquid formed at the base of the existing crust (17). Convective overturn of the ocean's basal cumulate pile may be responsible for the depths and chemical heterogeneities inferred for the MBSR (924), while the more buoyant residual liquid infiltrated into the crust (25) or was assimilated and carried upwards by magmas intruding the crust (1). If one long-lived magma body was responsible for the AFE, KREEP, and the MBSR, then more rapid crustal growth and a lower meteorite flux than suggested by (13) are required.

## III. Mg-rich Plutonic Rocks

Mg-rich plutonic rocks are pristine igneous cumulates whose mafic minerals have significantly higher Mg than those of ferroan anorthosites (26). Most of these rocks have cotectic proportions of mafic minerals and plagioclase (troctolite, norite, gabbro). Most of the Mg-rich rocks also have unexpectedly low Ti/Sm and Sc/Sm ratios (22, 27, 28). These low ratios ( $<1/6$  of the chondritic values) together with the absence of ilmenite or clinopyroxene as cumulus phases in these rocks indicates that their parent magmas assimilated some material high in Sm and low in Ti and Sc. The obvious candidate is KREEP. This interpretation is strengthened by the fact that several of the Mg-rich rocks tightly define a Nd-Sm whole rock isochron that is colinear with KREEP and yields an age of  $4.33 \pm .08$  G.y. (29). Some Mg-rich samples, however, have apparent isochron ages of  $\sim 4.5$  G.y. and have initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios equivalent to those of the anorthosites (31), yet still have low Ti/Sm and Sc/Sm ratios characteristic of KREEP. These older ages suggest that KREEP-like material may have been produced many times over a long span of early lunar history and that the ubiquitous KREEP found in soils and breccias today represents the culmination of a prolonged, repetitious differentiation process (32).

Studies of fragment-laden melt-rocks thought to represent the melt sheets of the Imbrian and Serenitatis basin-forming impacts indicate that no ferroan anorthosite was incorporated into the melt sheets (33). One interpretation is that the lunar crust is layered with anorthositic layers, of a variable thickness nearer the surface and less aluminous layers comprised of Mg-rich rocks and KREEP below (33). These observations plus radiometric ages of Mg-rich rocks extending from 4.5 to 4.2 G.y. suggest that the anorthositic crust was intruded by magma throughout this time. The younger ages probably represent intrusions of magma entirely within the existing crust (1), while older ages may represent crystallization of magma at the base of the crust as well (32).

What is not clear are the proportions of Mg-rich rocks and anorthosites averaged over the entire crust. Is the crustal structure suggested by (33) under the near-side basins moon-wide? Or have the Imbrium and Serenitatis impacts sampled large intrusions contained within more anorthositic crust (1)? The answer is important because the lunar crust, apparently  $\sim 60$  km deep (34), is a non-trivial fraction (1/10) of the moon's volume and thus knowledge of the crustal composition can constrain the lunar bulk composition and models of evolution.

Regardless of the proportions of the various pristine rocks, these rocks do tell us that there was extensive melting of the moon early in its history: crustal formation was an amalgam of igneous events whose records have been altered by meteorite bombardment. Determining the scale and duration of these events is the goal of the study of the pristine lunar rocks.

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COMPOSITIONAL VARIATION IN FERROAN ANORTHOSITES; CRYSTALLIZATION  
 RELIC OR SUBSOLIDUS OVERPRINT? Gordon L. Nord Jr., 959 National Center,  
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Compositional zoning in anorthite is shown by electron microprobe analysis in six lunar ferroan anorthosites. The zoning is best developed at plagioclase-pyroxene grain boundaries (Fig.1), but is also weakly developed at plagioclase-plagioclase grain boundaries and internal boundaries such as twin boundaries (Fig. 2) and low-angle subboundaries. Spot analyses and profiles indicate that the zoning involves a decrease in albite content (reverse zoning) and in Mg and Fe content as the boundaries are approached. The width of the profiles measured at plagioclase-pyroxene grain boundaries are in the range 100-200  $\mu\text{m}$ . One such profile measured in ferroan anorthosite 61015 shows a change in  $\text{Ca}/(\text{Ca}+\text{Na}+\text{K})$  content from 0.980 at the contact to 0.945 at a point 200  $\mu\text{m}$  away. Other ferroan anorthosites that show reverse zoning in anorthites are:

60025 [0.985  $\rightarrow$  0.956]; 60055 [0.981  $\rightarrow$  0.964]; 61015 [0.985  $\rightarrow$  0.956];  
 65035 [0.981  $\rightarrow$  0.964]; 65315 [0.978  $\rightarrow$  0.967]; 62255 [0.987  $\rightarrow$  0.954].

Two lines of evidence suggest that compositional zoning is a subsolidus overprint rather than a relic of crystallization: (1) Zoning is reverse zoning the opposite of the normal zoning (albite enrichment), expected for an evolving fluid; (2) reverse zoning also occurs, although weakly, at twin boundaries and low-angle subboundaries. Zoning that took place during crystallization would be expected to cut these boundaries, not to be coincident with them. Twin and low-angle grain boundaries, however, would be expected to act as rapid conduits for diffusion under subsolidus conditions, especially for large ions such as sodium.

In light of the reverse zoning overprinted on the ferroan anorthosites, any use of compositional trends in evaluating the crystallization history of the anorthosites must be carefully examined. Zoning of even a few percent albite content will effectively mask any differentiation trend deduced from plagioclase endmember abundances because of the restricted compositional range of lunar plagioclase in the ferroan anorthosites. The effects of the zoning can be eliminated by considering only analyses taken from areas several hundred micrometers from any type of structural boundary. Figures 3 and 4 show such analyses for  $\text{Mg}/(\text{Mg}+\text{Fe})$  in anorthite and coexisting orthopyroxene, respectively, versus  $\text{Ca}/(\text{Ca}+\text{Na}+\text{K})$  in anorthite. The positive correlations shown by both figures suggest that the six ferroan anorthosites crystallized from an evolving liquid.

Fe-Mg

The distribution coefficient  $K_{\text{pl-L}}$ , which describes the partitioning of Fe and Mg between basaltic liquid and anorthite during crystallization was determined by Longhi et al. (1976) to be 0.49. The data from this present study yield a coefficient of approximately 0.39. This lower coefficient may reflect a change in the exchange sites, from mostly tetrahedral in rapidly crystallized samples, to mostly large coordinated sites in slowly cooled samples. In support of this suggestion, most microprobe analyses of the six anorthosites studied here indicate that the tetrahedral sites in anorthite are completely filled with Al and Si, leaving Fe and Mg to occupy larger sites.

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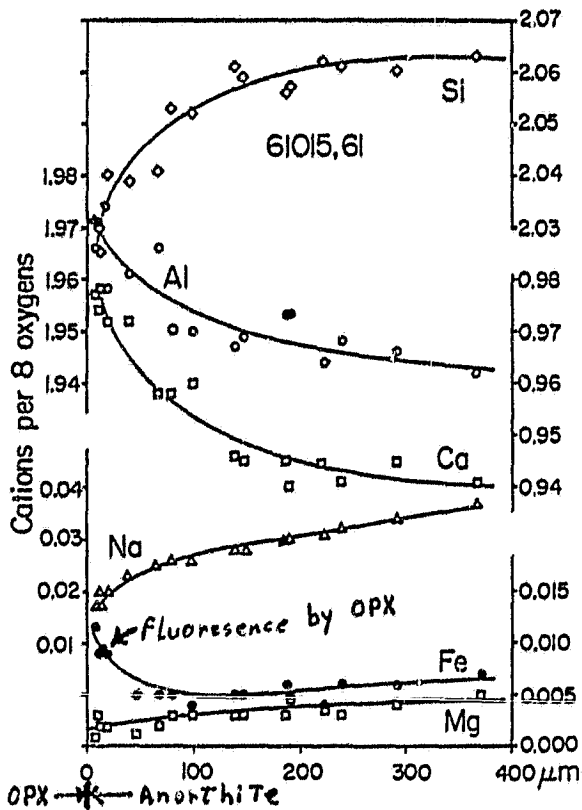


Fig. 1 Traverse from orthopyroxene-anorthite contact in anorthite. Sample 61015.

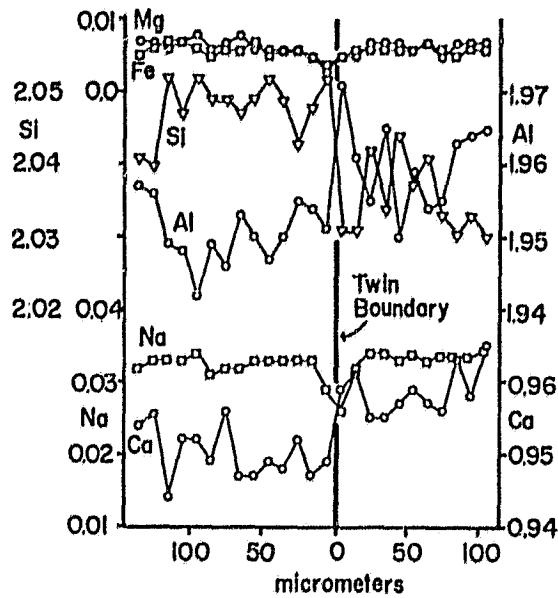


Fig. 2 Traverse across twin boundary in anorthite. Sample 60055.

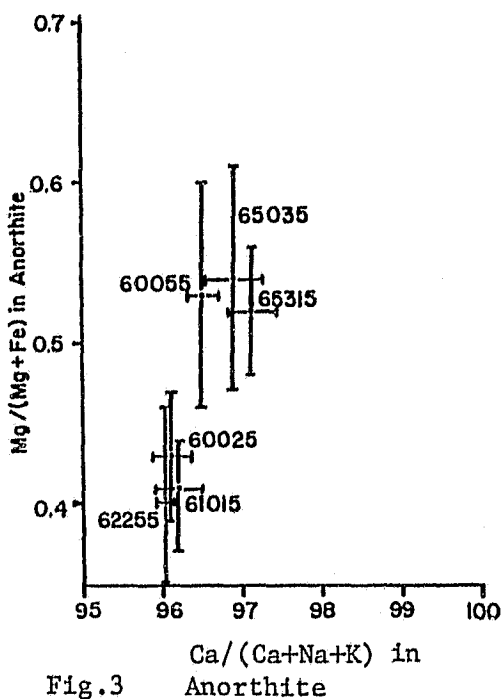


Fig. 3

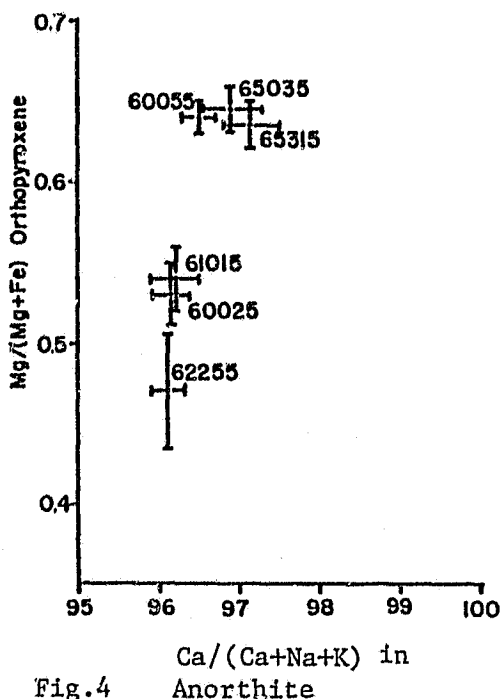


Fig. 4

## Macro and Microsegregation in a Solidifying Lunar Magma Ocean

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A solidification history of a magma ocean is developed utilizing industrial experience such as "carbon (i.e., CO<sub>2</sub>) boil" arising from volatile supersaturation (eg > x15) of melt due to quench cooling. Resultant vesiculated slag provides a floating structure from whence solidification may proceed downward from the surface of the melt through which most of the heat is transferred out. The governing equations are given by

$$h_r T_s = h_l (T_{bl} - T_f) + V \rho \Delta H_f \quad \text{where the radiative heat transfer coefficient is}$$

$$h_r = \sigma \epsilon F_R T_{SR}^4 / T_{SR} \quad \text{and the convective film coefficient is, for a Prandtl}$$

$$\text{number } Pr \ll Sc, \text{ the Schmidt number, } h_l = B Sc^{1/2} [Gr + (Pr/Sc)^{1/2} Gr']^{1/4} (k_l/R) \times$$

$$(R/l)^a / \{Pr^{1/4} [0.625/Pr + 1]^{1/4}\}. \quad T_s = \text{temperature of solidified melt, } T_{bl} = \text{temp-}$$

erature of bulk melt,  $T_f$  = equilibrium freezing point,  $V$  = velocity of solidification,  $\rho$  = density,  $\Delta H_f$  = heat of fusion,  $\sigma$  = Stefan-Boltzmann constant,

$\epsilon$  = emissivity,  $F_R$  = shape factor,  $k_l$  = thermal conductivity of melt,  $R = \frac{1}{2}$  the

breadth of the surface of the magma ocean,  $l$  = depth of magma ocean,  $Gr'$  = mass

transport Grashof number, i.e.  $Gr' = g \alpha (w_1 - w_1') R^3 / \nu^2$  where  $g$  is gravity,

$\alpha$  coeff. of expansion,  $\nu$  = kinematic viscosity,  $w_1$ ,  $w_1'$  are weight fractions

at solid-liquid interface, and bulk liquid respectively.  $Gr$  is the familiar Grashof number.  $B$  and  $a$  are empirical constants. We invoke Chvorinov's

$$\text{Rule, i.e. } S = 2 \frac{T_l}{\sqrt{\pi} \rho_s H} \sqrt{k_s \rho_s c_s t} \quad (\text{where subscripts } l \text{ and } s$$

designate liquid and solid respectively.  $c$  is heat capacity and  $t$  is time,  $s$  = thickness solidified) to estimate the time at which the insulation of the crust becomes significant. Volatile exsolution follows

$$(C_1 - C_1^*) dw_s \rho_1 v = (1 - w_s) dC_1 \rho_1 v - J A dt \quad \text{where } v \text{ is solid volume, } C's \text{ are bulk}$$

$$\text{compositions, } J \text{ the flux across the solid-liquid interface } = -\alpha' \rho_1 (w_1^e - w_1),$$

$\alpha'$  the evaporation coeff. and superscript  $e$  implies gas-liquid equilibrium conditions. Distributions for binary compositions (eg, olivine, plagioclase) are ascertained from Schiel's (non-equilibrium) eq. for convecting melts

$$C_s^* = k' C_o (1 - w_s) (k' - 1)$$

$$C_l = C_o w_1 (k' - 1)$$

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$k'$ , the effective partition ratio, is given by  $k' = k / \{k + (1-k)e^{-R\delta/D_1}\}$ ,

$k = C_s^*/C_1$ , Subscript 0 implies initial conditions on melt,  $\delta$  is boundary layer thickness,  $D_1$  is mass diffusion coeff.

Theoretical considerations provided by these descriptive equations provide initial concentration gradients for numerical assessment of Soret convection in the cooling magma ocean, which would manifest itself as diffusive layers of considerable depth underlying an upper region of solute banding. Both industrial and astrophysical work will be cited indicating Soret and double diffusive convection are one and the same.

## NICKEL IN OLIVINES AND PARENT MAGMAS OF LUNAR PRISTINE ROCKS.

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Mainly to determine the nickel content of their parent magmas, and hence to constrain how the parent magmas evolved, I have made analyses for Ni in the olivines of some of the pristine rocks. Figures 1 and 2 are plots of the determined Ni against mol % Fo for the olivine.

**Methods:** The data collection so far is really a feasibility study to ascertain whether attempting to analyze at the  $\sim 10$  ppm level is productive; the results are positive. Previous analyses on such olivines (e.g., 1, 2, 3) have had detectability levels of  $\sim 100$  ppm Ni and serve to demonstrate only that the olivines contain little Ni. The Fo is taken from previous analyses, published data, or from an Mg-count made simultaneously with the Ni count and calibrated against known olivines. Thus they are not necessarily precise for the grain analyzed, except in the case of 60025. The probe was operated at 30 kv with sample currents of 0.1-0.5  $\mu$ amps (rarely 1.0  $\mu$ amps), and the standard was a synthetic diopside glass containing 1.00 wt.% Ni. Backgrounds were determined by one of two methods: (i) simultaneously measuring an off-peak background on one spectrometer while measuring the Ni peak on another (G. Mackay, pers. comm.) and calibrating one spectrometer against the other with Ni-free phases or by step scans; (ii) successively measuring high background, peak, and low background, and repeating to check consistency of peak height. Much of the data in Fig. 1 were obtained by the first method, most of that in Fig. 2 by the second method. Counting times on peaks and on backgrounds were 100-300 seconds, and counting statistics alone lead to uncertainties of 6-12 ppm for each point in the Figures. The extra uncertainty in background location leads to total estimated uncertainties of 10-15 ppm Ni, and this is consistent with repeated analyses on a single grain. Data were reduced making an allowance for dead-time; matrix corrections were negligible and not made. There is a systematic difference of unknown origin between Figs. 1 and 2 in data for similar materials. Fig. 2 shows 10-15 ppm more Ni, but the general relations among samples are similar. In Fig. 1 the data for Marjalahti are similar to that accurately measured by (4), so Fig. 1 can perhaps be considered the more accurate. As a general guide, two duplicate analyses of Apollo 15 Green Glass balls gave 173, 179 and 141, 161 ppm Ni respectively, placing them in the main cluster found by (5). All the data for pristine rocks presented here are consistent with published analyses for Ni for bulk rock specimens (see 6).

**Mg-suite rocks:** The olivines in spinel troctolite 15445 have a fairly restricted range of Ni contents, but the measured values are higher than those obtained by (2) for the same sample (however, their precision was  $\sim 60$  ppm). The magma in equilibrium with these olivines would have had only 15-30 ppm Ni. The olivines in dunite 72415 have a wide range of Ni contents, from  $\sim 220$  ppm down to  $\sim 70$  ppm (compared with terrestrial dunites, 1000-3000 ppm Ni). Analyses for Fo were not made simultaneously, so it is not known whether there is a correlation of Ni abundance and olivine compositions. However, this range of Ni contents is much greater than for other single specimens, and is greater than appears to be the case for dunites from terrestrial intrusions (e.g., Muskox, (7, 8)). Hence there is the real possibility that dunite 72415 is genomict rather than strictly monomict. Alternatively, there has been subsolidus exchange of Ni between Fe-metal and some of the olivine in the sample.) A silicate liquid in equilibrium with these olivines would have had the same low Ni contents as the parent of the spinel troctolite. In contrast, the Ni content of troctolite 76535 olivines is much lower than that of 72415, and the silicate liquid would have had only a few ppm Ni, similar to that for the anorthosites. Assuming 72415 and 76535 are related by a few % fractional crystallization (1) (their mineral phases are very similar), the precipitous drop in the Ni abundance of the magma suggests the separation of a metal phase, such as is found in 72415. A similar conclusion was reached by (9) on the basis of bulk rock siderophile abundances. The data show that the Mg-suite parent magma contained much less Ni than low-Ti mare basalts; thus the similarity in Ni abundance for pristine rocks and low-Ti mare basalts (10) is merely a coincidence, an artifact of comparing cumulate rock with silicate liquid compositions.

**Anorthosites:** All the anorthosites have very low Ni in their olivines, corresponding to a parent magma with 3-6 ppm Ni. The abundances are similar to those in the Mg-suite troctolites. There is some evidence for a slight increase in Ni with evolution at first, consistent with the interpretation that olivine was not a crystallizing phase during most of anorthosite genesis, and only crystallized sporadically (11). The steady or rising values also preclude the separation of a metal phase during the crystallization of anorthosites, unlike the Mg-suite. Pyroxenes analyzed have Ni abundances at the detectability limit, e.g.,  $12 \text{ ppm} \pm 12 \text{ ppm}$ .

**Mare basalt sources:** It is evident that cumulates complementary to anorthosites would contain far too little Ni to produce the Ni abundance of the low-Ti mare basalts ( $\sim 60$  ppm) by partial melting. In fact, if Ni is carried in olivine in the source of these basalts (9) then the origin of these sources is obscure vis-a-vis any pristine rocks, as these source olivines are required to carry about 300 ppm or more Ni. High-Ti mare basalt sources are not required to carry such high Ni contents and could be complementary on the basis of Ni alone.





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IDENTIFICATION OF REGIONAL DEPOSITS OF LUNAR PRISTINE ROCKS FROM ORBITAL GEOCHEMICAL DATA. Paul D. Spudis<sup>1,2</sup> and Philip A. Davis<sup>2</sup> 1. Dept. of Geology, Ariz. State Univ., Tempe AZ 85287 2. U.S. Geological Survey, 2255 N. Gemini Dr., Flagstaff AZ 86001.

**Introduction.** Lunar pristine rocks are identified by a number of petrographic and geochemical criteria (1,2) that may only be determined in terrestrial laboratories. The need remains to understand the geologic context of these important samples and to relate them to regional areas of the lunar highlands. The only source of regional chemical information for the Moon is the data provided by the Apollo orbiting geochemical instruments (3). These experiments provide limited major element chemistry for large regions of the Moon and although not of global coverage, provide a geologic setting for pristine rocks and information about regional geochemical provinces (4). The limited resolution of these data (5) suggest that only statistical distributions of rock types may be inferred. Thus, although not definitive, the orbital data may be used to locate highland regions that are likely to contain large quantities of pristine rocks.

**Method.** This study utilizes the digital lunar data bases of the La Jolla Consortium (3,6). For this initial effort, the consortium X-ray data were used to generate images of two variables: Mg and Al. Images were produced by color compositing the Mg and Al abundances representative of three pristine rock types: anorthosite, norite and mare basalt (in highlands only). The data were also used to distinguish regions of noritic composition (possibly pristine norites) from low-K Fra Mauro basalt (probably an impact mixture; 7). No regions are found on the Moon to be composed of these pure rock types; usually, a mixture is present. This study attempts to identify areas consistent with a given component abundance of greater than 90 percent. Moreover, the results of Apollo sampling provide ground truth on the petrology of limited areas (e.g., 4,8). Combined with photogeologic analysis of the geologic history in the regions studied, these color maps provide a reasonable estimate of the geographic distribution of pristine rock types.

**Results.** Anorthosite was defined in the data system as  $Al_2O_3$  greater than 30 percent and MgO less than 4 percent. The resulting image shows that large regions of the eastern limb are sufficiently aluminous, but typical Mg values are too high for pure anorthosite. This probably results from the combined effects of mafic ejecta from the Crisium and Smythii basins and copious pre-Imbrian mare flooding (9,10) in this region. Nevertheless, these large aluminous regions may contain some quantity of pristine anorthosite because the Apollo 16 Descartes area is displayed in the color image of this category. More interesting are several small regions located west of Pasteur crater (11 S, 94 E) that appear to consist of pure anorthosite. This area consists of rugged pre-Nectarian highlands (11) that may be in situ ferroan anorthosite from the primordial lunar crust. Additional regions of pure anorthosite occur in rugged terra SW of Smythii basin (11 S, 89 E) and west of Mendeleev (4 N, 117 E).

For this study, norite was defined with  $Al_2O_3$  between 20 and 21.6 percent and MgO of 6-7.5 percent. This composition is actually closer to anorthositic norite, but mixing effects are acute for intermediate rock compositions and the resulting map closely corresponds to known regions of norite-rich crust (e.g., Taurus-Littrow; 4,8). The norite map shows that large regions of the Montes Apenninus, Haemus and Taurus highlands are rich in norite. These areas are dominated by ejecta from the Imbrium and Serenitatis basins. Moreover, large quantities of pristine norite were returned by the Apollo 15 and 17 missions (12). These results support earlier suggestions that norite was a

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major component of the Imbrium/Serenitatis impact targets (4,13). The proposed Procellarum basin (14) may have removed most of the ferroan anorthosite crust in this region while subsequent Mg-suite intrusions formed a petrologically complex crust. Alternatively, anorthosite may never have been formed in this region and Mg-suite cumulates represent the original lunar crust in this area. Other norite provinces include a small region near Gibbs (15 S, 82 W) that may be ejecta from the Balmer or Smythii basins. Smaller norite regions are also associated with the Crisium and Nectaris basins as minor ejecta components.

Mare basalt was defined as  $Al_2O_3$  less than 19.6 percent and MgO greater than 8.3 percent. This component was mapped to look for occurrences of mafic materials in the highlands that are not attributable to subsequent mare volcanism. The largest mafic regions are found in the Montes Apenninus and Haemus. These areas are covered by Imbrian age dark mantle (15) that are probably pyroclastics associated with mare volcanism. Other than this explainable anomaly, no major regions of mafic highlands are identified. A small mafic region near Firsov (4 N, 115 E) may correlate with an ancient mare basalt-filled basin Al-Khwarizmi-King (11) whose pre-Nectarian basalt fill has been gardened into the local megaregolith. This concept is further supported by anomalously high regional Ti values in the terra comprising this proposed basin (16). Thus, all highland mafic anomalies may be related to mare volcanism of widely variable age.

Conclusions. Regions of the lunar highlands with major element chemistry consistent with large quantities of pristine rocks have been identified in the orbital geochemical data. Areas with composition corresponding to pure anorthosite are rare, but may occur in isolated areas of pre-Nectarian terra on the eastern limb. Moreover, large regions of the eastern highlands may contain significant amounts of pristine anorthosites included within brecciated basin and terra deposits. Extensive norite provinces occur within the rim units of the Imbrium and Serenitatis basins (sampled by Apollo 15 and 17) and this area may represent a regional Mg-suite petrologic province. Ferroan anorthosite is not an important component in this region and may have been removed by the Procellarum basin impact or was never present. Mafic regions in the lunar highlands are rare; most can be related to mare volcanism that has occurred throughout lunar history. The results of this attempt to locate regional deposits of pristine rocks are not definitive but are very suggestive. Global geochemical coverage of the Moon by a future orbiting spacecraft would enable a more detailed analysis of the regional distribution of pristine lunar rocks and a greatly improved knowledge of lunar geologic history.

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LUNAR AND PLANETARY COMPOSITIONS AND EARLY FRACTIONATION IN THE SOLAR NEBULA. S.R. Taylor, Research School of Earth Sciences, Australian National University, Canberra, Australia.

Estimates for bulk planetary compositions for Earth, Moon, Mars, Venus and Mercury (1,2) show substantial differences in total Fe, Fe/Mg and K/U as well as density. Although considerable uncertainties exist it is clear that these differences reflect variations in metal/silicate and refractory/volatile element ratios. Table 1 lists K/U ratios, representing volatile/involatile element ratios, and densities, providing information on the proportions of metal to silicate. K/U ratios remain relatively unchanged during planetary fractionation processes and accordingly provide an estimate for the bulk planetary ratios. The K/U ratios for CI chondrites are about 60,000 (2). Other chondrites have variable ratios (Fig. 1). The eucrites have ratios of about 3000 ( $3100 \pm 375$ ) (3). The eucrites have low contents of other volatile elements, resembling the Moon in this respect. Although our estimates of the density of the eucrite parent body are uncertain (It is not clear whether it has a core or substantial metal content), the eucrites themselves are depleted in siderophile trace elements (Ni is typically much lower than 10 ppm) to an extent similar to the levels in the lunar high-Ti basalts (4,5). Two modes of formation of the eucrite parent body appear possible (A) Melting and separation of a metallic core from a silicate mantle, from which the eucrites are subsequently derived by partial melting. (The W/La and U/W studies show that metal was absent in the source region of the eucrites (6,7). (B) Accretion of metal-free silicate planetesimals to form the eucrite parent body. In the first case, the eucrite parent body becomes one of the sites where metal-silicate fractionation occurs: in the second scenario, this event has already occurred. Eucrites give internal mineral isochrons around 4.54 Aeons (8). Accordingly the separation of metal from silicate in the parent body occurred at or before that date, which represents the crystallisation of the eucrite magma, derived by partial melting from the silicate metal-free mantle of the eucrite parent body. Although we cannot date the metal-silicate fractionation (it occurs either during or before accretion of the eucrite parent body) we can date the depletion of the volatile elements (eg K) as occurring before accretion, since there is no evidence that they could be lost following accretion. The eucrites have chondritic levels of refractory trace elements (4,5). Accordingly, the volatile/involatile element fractionation occurred before 4.54 Ae, at a time not clearly resolvable from 4.57 Ae generally assigned as the age of formation of the Solar System. (The eucrite parent body is small with pressures not exceeding 2 kbars (9)).

The considerable implications of these data are

(a) Metal-silicate and volatile-refractory element fractionation occurred very early in Solar System history at a time indistinguishable from 4.57 Aeons.

(b) These events occurred before the accretion of the Earth and the Moon which inherited their K/U ratios from precursors.

(c) The core and mantle of the Earth accreted as separate phases. Metal-silicate equilibration was established in precursor events, and was only locally established during core separation in the Earth. Accordingly, only local reaction between metallic core and silicate mantle is predicted during core-mantle separation.

(d) Comparisons between bulk Moon and terrestrial mantle compositions provide evidence for heterogeneous accretion of a diverse suite of planetesimals with differing K/U and metal/silicate ratios (10).

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Implications for the fission hypothesis of lunar origin are

- (a) The differences in total Fe, siderophile element content, Fe/Mg ratios and refractory/volatile elements between the Earth and the Moon (2) are inherited from the accreting planetesimals.
- (b) No simple mechanism can produce these differences following core separation.
- (c) The mantle Fe/Mg ratio is fixed for the Earth before accretion, not by processes during core separation.
- (d) Although some material knocked off from the Earth by collisions may be incorporated into the Moon, its compositional signature is swamped by independently accreted planetesimals.
- (e) The W/La correlations in the Moon, Earth and eucrite parent body are due to processes occurring before accretion and have no local genetic significance (11).

Some other constraints may be noted: Astrophysical observations indicate that during star formation, nebulae clear on time scales of  $10^6$  years (12). Strong T Tauri Solar winds also occur on these short times scales (13). Calculations for accretion of the inner planets by accretion from a swarm of planetesimals indicate time scales of  $10^8$  years for this process (14), two orders of magnitude longer. Jupiter and Saturn clearly grew in a gas-rich environment and accordingly probably formed early. Earth and inner planets accrete from planetesimals in a gas-free environment (except for gas trapped in planetesimals) up to 100 million years later. Jupiter was thus able to prevent formation of a planet in the asteroid belt and stunt the growth of Mars by sweeping up planetesimals. The depletion of volatile elements in the inner Solar System, leading to low K/U ratios may have resulted from loss of volatiles during early nebula/clearing stage within a few million years of formation of the Sun. This process is likely to have been less effective in the region of the giant planets. Accordingly it may be predicted that K/U ratios for the Galilean and Saturnian satellites are about chondritic, with implications for radiogenic heat production, melting and differentiation.

Table 1. Planet	Mercury	Venus	Earth	Moon	Mars	
Density(2)	5.435	5.245	5.514	3.344	3.934	
K/U(2)	-	7-20x10 <sup>3</sup>	10 <sup>4</sup>	2500	2200	
Meteorites:	C1(2)	H(15)	L(15)	E4(15)	Eucrites(3)	Nakhla(3)
K/U	6.3x10 <sup>4</sup>	6.2x10 <sup>4</sup>	5.8x10 <sup>4</sup>	8.6x10 <sup>4</sup>	3100	2.1x10 <sup>4</sup>

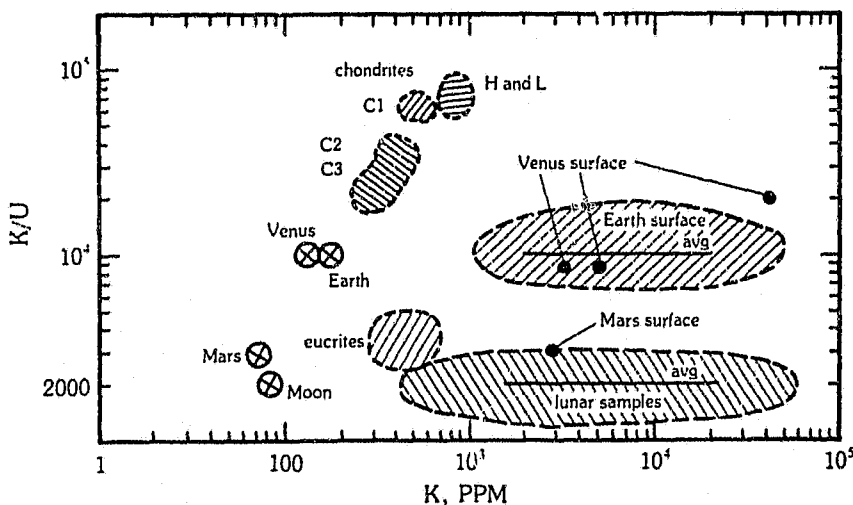


Figure 1.  
Variations in K/U ratios in meteorites and planets (Reproduced with permission from Ref. 2).

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## MODELS OF BULK MOON COMPOSITION: A REVIEW

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### INTRODUCTION

#### Why investigate the Moon's bulk composition?

There are basically two reasons why it is important to constrain the Moon's bulk composition (BC) as best we can. First, there is probably no more fundamental constraint on the origin of a planetary body than its BC, and the origin of the Moon is of particular importance due to its possible relationship with Earth. Second, the Moon's BC (along with its size, which is known, and its thermal status at birth, which is practically unknown) was a key factor governing its evolution as a planet.

#### The chondrite connection

A starting point common to all models of the Moon's BC is an assumption that it formed out of solids fractionated from the solar nebula, and its BC must therefore be derivable from components known from undifferentiated meteorites (chondrites), which, for all but the most highly volatile elements, have BCs that closely resemble the sun. The range of possible BCs is still very great, however, because among chondrites as many as seven separate components may be recognized [15]. The latest estimates of the Moon's silicate fraction BC from each of the various research groups who have worked at the problem since the heat flow data were revised [24] are listed in Table 1, and the "average" is compared with the silicate fractions of H chondrites in Fig. 1.

Another point of agreement among all BC models is that the Moon's low bulk density proves that it is depleted in the denser phases found in chondrites, namely Fe-Ni and FeS: the Moon has only a small (perhaps ~2 wt.% of the Moon [41]) or nonexistent core. Siderophile trace elements are commensurably low, everyone agrees. A third major point of agreement is that the Moon is greatly depleted in volatile elements, most notably H (water), C, O, Na and K, but also numerous trace elements, e.g., Zn, Rb, Cd, In, Cs, Hg, Tl and Pb. These depletions are extremely important aspects of the Moon's BC. However, since (1) they are not highly controversial, (2) they may be reconciled about equally well with all three of the basic models for lunar origin (double planet, fission or capture), and (3) the Fe-Ni and FeS depletions are better assessed via geophysics [5] than via geochemistry, I will not discuss them, per se, further.

The most controversial aspect of the Moon's BC involves refractory or "early condensate" (out of the solar nebula) lithophile elements: Al, Ca, Sc, Ti, V, Sr, Y, Zr, Nb, rare earths, Hf, Ta, Th and U (I will henceforth refer to these as RL elements). Almost simultaneously with the return of the first lunar samples, revealing that the Moon's crust is greatly enriched in RL elements, the Allende carbonaceous chondrite fell. Unlike most chondrites, ~5 wt.% of Allende (and other, less well-studied meteorites of its class) consists of inclusions that are greatly enriched in RL elements. There are still no whole meteorites that consist exclusively of refractory materials, and the Mg-normalized BC of Allende is only enriched in RL elements by ~38% [20]. Nevertheless, Allende showed that the "early condensate" might have been segregated from the rest of the chondritic condensates. Early post-Apollo models suggested that the Moon accreted a refractory shell in its final stages of formation [e.g., 16], or even that the whole Moon is as refractory as the Allende inclusions [2]. Detailed analysis of lunar geochemical trends revealed that the heterogeneous accretion model "presents more problems than it solves" [4], and estimations of the Moon's BC have tended to have smaller and smaller proportions of early condensate (see, e.g., the series of Wanke et al. models [45-48]). But most workers still believe the Moon's BC is at least modestly enriched in RL elements, relative to chondrites.

It is generally acknowledged that all the RL elements are in approximately chondritic proportions to each other (as they are in the Allende BC), even if as a group they are enriched in the Moon. Thus, the ratio of the trace element U to Al is much more constant in Table 1 than Al/Si. Longhi [27, 28] has argued that the Moon may have a subchondritic Ca/Al. I address this issue in a contributed abstract [50].

There is general agreement about the SiO<sub>2</sub> and MgO contents of the Moon, but not about its FeO content (Fig. 1). This reflects the fact that chondrites vary much more in FeO than they do in MgO (the elemental abundance of Fe is relatively constant, but there are vast differences in Fe-metal/FeO ratios, i.e., oxidation states). Despite the general agreement about MgO and SiO<sub>2</sub>, geophysical evidence [6] strongly indicates that the Moon's Mg/Si is less than 0.84, whereas the average of the estimates in Table 1 is 0.92.

In summary, the Moon's BC is generally modeled as a mixture of chondritic materials, greatly depleted in Fe-Ni, FeS and volatiles. The two main sources of controversy are (1) how large is the RL component?, and (2) what is the nature of the "regular" chondritic component?; in particular, what are its FeO content and Mg/Si?

#### MAGNITUDE OF THE REFRACTORY LITHOPHILE COMPONENT

The volatile depletions are dramatic, but they should never be confused with significant refractory element enrichments. The volatiles that are depleted in the Moon amount to only about 2% of an ordinary chondrite; they may have been lost with little effect upon the major element BC. Ordinary chondrites and the Earth are depleted in most volatile elements, relative to carbonaceous chondrites, yet no one argues that these objects are greatly enriched with RL elements.

#### Heat flow

One of the arguments for large RL enrichments is based on heat flow (HF). Assuming we know the global avg. HF, it is very simple to derive global concentrations of Th and U (and by implication, all the other RL elements). In the view of Langseth et al. [23, 24], the global avg. HF is virtually identical to the avg. of the HFs at the two sites where HF measurements were obtained (Ap 15 and

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17), which would imply RL in the Moon are  $\sim 2.8 \times$  RL in the silicate portions of H chondrites. This interpretation has been uncritically accepted by most geochemists. However, other geophysicists have criticized it, because the HF measurements were both made right on the edges of maria, which are regions of shallow regolith depth, and hence, anomalously high thermal conductivity. Mizutani and Osako [30] stated that "the [average lunar surface heat flow at the highland should be substantially smaller than that observed at Apollo 15 and 17, ... for any reasonable lunar crust model." Conel and Morton [9] considered 15 different models, and concluded that the HF data were sufficiently low to match all but one under the assumption that the Moon's RL are enriched to  $1.5 \times$  chondritic silicates (and this was before the HF data were revised downward by an average of 38% [24]). Another reason to doubt the representativeness of the two HF data is that the general area of the AP 15 and 17 sites is richer in K, Th and U than the bulk of the crust [49]. Langseth et al. [24] admitted that "uncertainty in the global average heat flow cannot be estimated until further data become available."

#### Mass balance

Another argument for RL enrichments [43] assumes that we know the average composition of the entire upper 60-100 km (i.e., 10-13 vol. %) of the Moon from averaging data for soils. Assuming this entire layer has 25%  $Al_2O_3$  (as an example), either an implausibly large percentage of the Moon's Al must be bound up in this layer, or the Moon must be at least a little enriched in Al, over the  $\sim 3.0\%$  contained in the silicates of ordinary chondrites. However, remote sensing data clearly indicate that there are huge lateral variations in the crust's composition, and there may well be equally great variations with depth [43]. In fact, there is some evidence that the crust grows poorer in Al with depth [e.g., 36, 53]. The seismic data are about equally consistent with an Al-poor lower crust as with an Al-rich one [26]. One way of explaining the Moon's center-of-figure/center-of-mass offset [21] is to invoke a thick, low-density (i.e., Al-rich) crust, which is thicker beneath the farside than the nearside. However, the offset can also be explained by assuming that the mantle (and crust) are slightly more Fe-rich beneath the nearside than beneath the farside [54]. This model is no less consistent with more Fe in nearside soils than in farside ones [e.g., 10], than models invoking a thicker farside crust.

#### The Mainz "primary component" model

H. Wanke and coworkers [45-48, 12] estimate the Moon's BC by assuming that the highlands samples consist of mixtures of a "primary" component that was added to the Moon via meteoritic bombardment, and never underwent magmatic fractionation, except for removal of enough (31%) olivine to make for a CI-chondritic Mg/Si, plus a trace-element poor anorthosite component, plus KREEP. They subtract out KREEP and anorthosite from the highlands compositions, and assume that the composition of the primary component is equivalent to the Moon's BC. Their latest model indicates a RL enrichment of only  $\sim 50\%$  (their first model of this basic type indicated  $\sim 500\%$  [46]). This model involves two questionable assumptions: (1) that none of the Mg in the highlands comes from pristine cumulates such as norites and troctolites (surely an oversimplification [35]), and (2) that the Moon's Mg/Si is the same as the CI chondrite value, which is ruled out by geophysical evidence [6].

#### Mare basalt compositions

Ringwood [32] argues that the low Al (etc.) content of most mare basalts constrains the Moon's Al to be at least nearly as low as that of pyrolite (Earth's upper mantle).

#### Constraints (?) from phase equilibria

One argument for RL enrichments has been effectively overthrown. Several groups [44, 17] argued that the existence of cumulates with cumulus olivine together with cumulus plagioclase means that the Moon's BC must have very high Al and Ca (or else implausibly low Mg/Si), because under 1-atm phase equilibria primitive lunar magmas would otherwise have ceased crystallizing olivine, and begun crystallizing pyroxene instead, before crystallizing plagioclase. This overlooks the probability that the lunar magma "ocean" was deep enough for pressure to completely alter the olivine/pyroxene phase equilibria [52].

In the view of Ringwood [32], the preponderance of low-Ca pyroxene compared to high-Ca pyroxene in the crust shows that the Moon is low in Ca. The same phenomenon is explained by Longhi [28, 29] by invoking a subchondritic Ca/Al ratio, however. Using the phase diagrams of [37], norms may be calculated for model BCs at lower mantle pressures [6]. Results indicate that the geophysical constraints of density and moment-of-inertia factor are best met by BCs with only 4-5%  $Al_2O_3$ ; otherwise, too much dense garnet is formed. This difficulty may be circumvented by postulating only 3.0% FeO [11]; this seems implausibly low, however (see below).

#### SILICON, MAGNESIUM AND IRON

The strong geophysical constraint that Mg/Si be  $< 0.84$  [6] is only met by 2 of the 8 model BCs in Table 1, and one of those is the new one of [6]. Of the remaining 6, only one even comes close. The two satisfactory models have only slightly lower MgO, but considerably higher  $SiO_2$ , than the "average" BC. Both yield norms with much lower olivine/pyroxene ratios than the "average" BC.

An important constraint on FeO comes from magnetic permeability data [13]. Assuming FeO occurs roughly half in low-Ca pyroxene, half in olivine, the Moon's FeO must be very close to 9%. With a lower ol/px ratio, the FeO might be a few % higher, but 3 of the 4 models in Table 1 that have  $FeO \gg 9\%$  yield norms with ol/px  $\gg 1$ , not  $< 1$ . This model is somewhat oversimplified, because other FeO-bearing phases, e.g. high-Ca pyroxene, garnet and spinel, have the same effect on the FeO value as a lower ol/px ratio. Seismic velocity profiles indicate that the mg ratio in the lunar mantle is only  $\sim 0.75$  [8] but see also [8], considerably lower than that of pyrolite (0.90). The mg inferred for the mare basalt source region is also  $\sim 0.75$  [33]. However, the mare basalt sources were previously differentiated, so there is no reason to expect their mg to precisely equal that of the Moon's BC. The well-known dichotomy among pristine nonmare rocks probably arose because the ferroan anorthosites formed by flotation of plagioclase over the primordial magma "ocean," whereas the Mg-rich suite formed

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in numerous later intrusions [52, 19]. If this is so, the  $mg$  of the magma ocean was probably roughly 0.85 [52], and the  $mg$  of the source region of the magmas parental to the Mg-rich intrusives was probably at least 0.80 [50]. The  $mg$  ratios of KREEP basalts are surprisingly high (generally 0.5-0.7 [18, 51]), considering their extremely high incompatible element contents. Incidentally, whatever the correct value for FeO is, the value for MnO ought to be precisely 80 x lower, because FeO/MnO is extremely constant among all types of lunar samples [25].

## IMPLICATIONS FOR AND FROM MODELS FOR THE ORIGIN OF THE MOON

According to Ringwood [e.g., 33], if meteoritic contamination is theoretically removed from highlands samples, the residual siderophile element patterns are evidence of a "close relationship" between Earth and Moon, but this model has been sharply criticized [1, 53]. However, one genetically important aspect of the Moon's BC is known with great precision: Most meteorite groups, particularly the only known primitive solar system objects that are greatly enriched in RL elements relative to chondrites (the Allende-type inclusions), feature oxygen isotope ratios that make it clear that they are genetically distinct from Earth; but lunar samples could not be distinguished from Earth rocks, if our only data were for oxygen isotopes [7]. As Taylor [39] puts it, this "may have dealt a fatal blow to the capture hypothesis, by requiring formation of the Earth and Moon in close association." If so, the BC of the Moon's silicate fraction should not be totally unrelated to the BC of Earth's silicates. It is much easier to explain why the much smaller Moon accreted less metal and volatiles, than why it should have greatly different RL contents, Mg/Si,  $mg$ , etc.

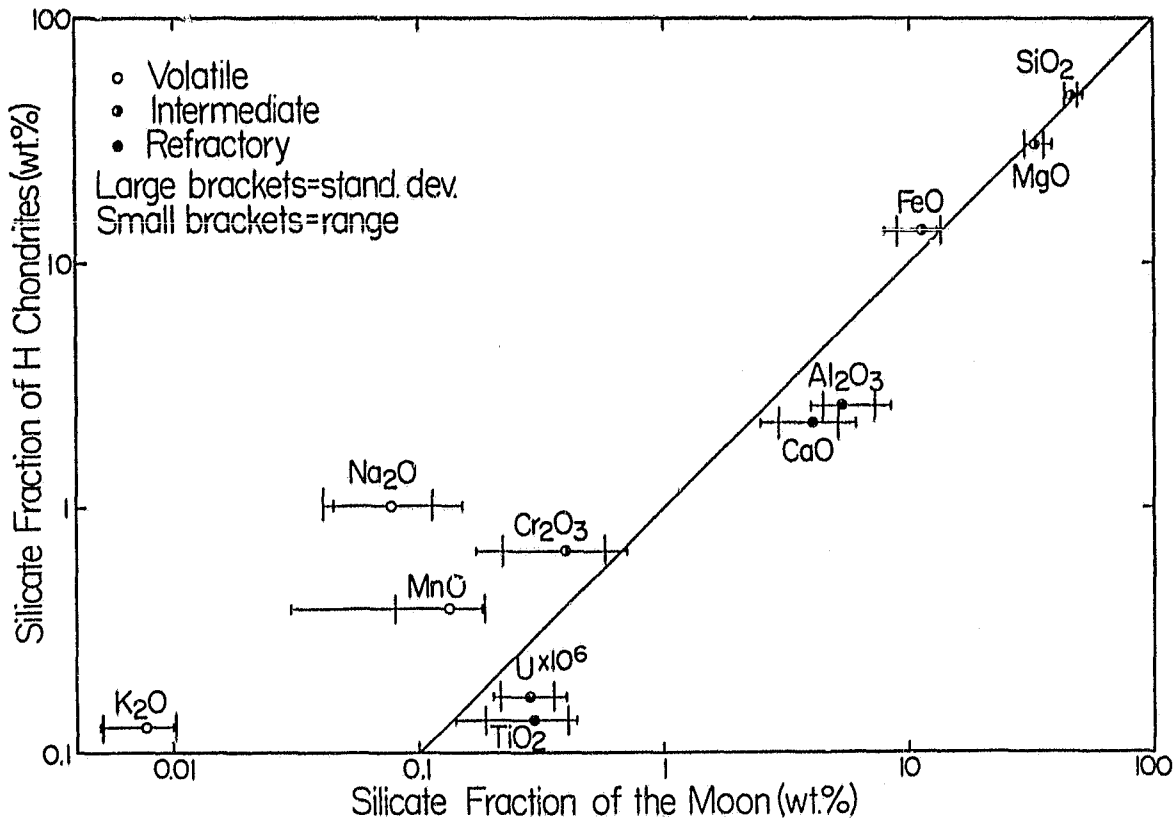


Fig. 1

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Table 1

Model	W77	M78	LB79	WW79	B/WW79	R79	BT80	T82	"Average"	SD
SiO <sub>2</sub>	45.6	43.3	42.8	51	45.1	44.8	48.4	44.4	45.68	2.73 (6.0)
MgO	32.3	29.1	33.6	32.6	38.1	32.7	29.0	32.8	32.53	2.84 (8.7)
FeO	13.0	13.0	9.4	8.5	7.9	13.9	12.9	11.0	11.20	2.33 (21)
Al <sub>2</sub> O <sub>3</sub>	4.6	7.6	8.38	3.0	3.9	4.2	5.0	6.1	5.35	1.86 (35)
CaO	3.8	0.1	4.88	2.5	3.1	3.7	3.83	4.6	4.06	1.12 (27)
Cr <sub>2</sub> O <sub>3</sub>	0.39	0.3	0.17	0.7	0.3	0.4	0.3	0.61	0.40	0.18 (44)
MnO	0.18	0.15	0.03	0.11	0.10	-	-	0.15	0.12	0.05 (44)
TiO <sub>2</sub>	0.20	0.39	0.44	0.14	0.2	0.3	0.4	0.3	0.30	0.11 (37)
Na <sub>2</sub> O	0.061	0.10	0.07	0.045	0.045	0.05	0.15	0.09	0.076	0.036 (47)
K <sub>2</sub> O	0.006	0.01	0.01	0.005	0.005	-	-	0.01	0.0077	0.0026 (34)
Sum	100.14	100.65	99.78	98.60	98.75	100.05	99.98	100.06	99.68	
U (ng/g)	27	40	-	20	25	25	-	34	28.5	7.2 (25)
Ca/Al	1.12	1.08	0.79	1.13	1.07	1.19	1.03	1.02	1.05	0.12 (11)
U/Al	11.1	9.9	-	12.6	12.1	11.2	-	10.5	11.25	0.98 (8.7)
Al/Si	0.114	0.20	0.22	0.067	0.098	0.106	0.117	0.156	0.135	0.053 (39)
Mg/Si	0.914	0.867	1.01	0.825	1.09	0.942	0.773	0.953	0.92	0.10 (11)
mg	0.816	0.800	0.864	0.872	0.896	0.807	0.800	0.842	0.837	0.037 (4.4)
FeO/MnO	71.1	85.3	308.3	77.3	79.0	-	-	72.2	116	95. (82)
CIPW pl	12.8	21.1	23.1	8.4	10.8	11.7	14.2	16.8	14.9	5.1 (34)
CIPW ol	51.6	55.9	62.7	22.5	58.8	55.5	33.2	54.4	49.3	13.9 (28)
CIPW px	34.8	18.5	12.9	66.4	28.4	31.8	51.4	25.2	33.7	17.5 (52)
ol/px	1.48	3.02	4.86	0.34	2.07	1.75	0.85	2.16	1.51	0.23 (16)

References: W77 [48], M77 [31], LB79 [29], WW79 [52] (this is a 1:1 mixture H and IAB chondritic mixtures), B/WW79 [52, 3] (this is "pyrolite," according to [34], with volatiles Na and K adjusted by [52] to suit the lunar BC), R79 [33], BT80 [6], T82 [41].

Note: SD = standard deviation. the numbers in parentheses give the SD in %.

## PETROGENESIS IN A MOON WITH A CHONDRITIC REFRACTORY LITHOPHILE PATTERN

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Among models of the Moon's composition [35] there is considerable controversy about refractory lithophile (RL) elements: Al, Ca, Sc, Ti, V, Sr, Y, Zr, Nb, rare earths, Hf, Ta, Th and U. Most models assume that RL elements are enriched in the Moon relative to chondrites. The extent of the enrichment is controversial. However, most models assume that even if the RL elements are enriched as a group, they are still in chondritic proportions to each other. Thus, the ratio of the trace element U to Al is much more constant in Table 1 of [35] than Al/Si. This is a crucial assumption, because: a) U and Th are the dominant sources of heat in the Moon; b) isotopes of Sr, Sm, Nd, Th and U are used to calculate model rock ages; and c) the Moon's crust is extremely RL-rich (how?-- why??).

Longhi [15-18] has made the intriguing suggestion that the Moon has a Ca/Al ratio considerably lower than that of chondrites. He points out that in the crystallization sequence inferred for the Mg-rich pristine rocks, orthopyroxene came before clinopyroxene. Moreover, opx predominates over cpx in most highlands breccias. Whether modeling these rocks as products of a magma ocean [15, 16, 18], or younger magma(s) formed by partial melting [17], it is difficult to explain their high opx/cpx ratios. Ryder [28] concurs. Of course, if the Ca/Al ratio is not chondritic, most of the other RL element ratios are probably not chondritic, either.

This constraint is model-dependent. First, let us assume that the noritic rocks are products of the magma ocean. Longhi tends to favor high (Al + Ca) in his magma ocean models. Consequently, his models generate relatively large amounts of plagioclase cumulates (troctolites), before any norites can form. As noted by Longhi [16], plagioclase crystallization increases Ca/Al. Thus, a model that assumes high (Al + Ca) in the initial composition requires a lower Ca/Al than a model that assumes low (Al + Ca). If the magma ocean is modeled assuming low initial (Al + Ca) [36], the composition of the ocean at the point that it begins (copious) plagioclase crystallization yields a CIPW norm with a diopside/hypersthene ratio of only 0.26 (this is for the model assuming a 1:1 hybrid of H and IAB silicates as the initial composition, and 360 km as the initial depth). For comparison, the estimated average composition of the highland crust [31] yields a CIPW norm with a di/hy ratio of 1.45.

Now let us assume that the noritic members of the Mg-rich suite are products of younger magma(s) formed by partial melting [36, 10]. The source region(s) under this assumption were not necessarily primitive, undifferentiated materials of whole-moon composition [37]; they may have lost Ca relative to Al during the formation of the "ocean" (e.g., at very high pressures the residuum from the formation of the ocean may have been rich in low-Ca, but high-Al pyroxene). But let us assume for the sake of argument that the sources were completely undifferentiated. Models of high pressure partial melting [17] indicate that, assuming a source with chondritic Ca/Al, only large degrees of partial melting, >40%, produce melts that would crystallize low-Ca pyroxene before high-Ca pyroxene at low pressure. This seems excessively high, so again there is a suggestion that the Moon has a subchondritic Ca/Al.

However, high percentage partial melting is required to explain the high mg of many Mg-rich suite samples, regardless of constraints involving Ca/Al. According to seismic evidence [4] (but see also [5]), the mg of the lunar mantle is probably <80, and almost certainly <83 (data are shown in [4] up to a depth of 1000 km; only 16 vol. % of the Moon is deeper than 1000 km, so we may assume that this holds for essentially the entire mantle. Of course, partial melts always have lower mg than their source compositions. Yet, a high percentage of the pristine Mg-rich suite samples have olivines with mg >87 [e.g., 33]; several even have mg-92 olivines. The value of K (= (Fe/Mg<sub>crystal</sub>)/ (Fe/Mg<sub>liquid</sub>)) for olivine ranges from 0.26 to 0.36 as a function of Si content [19]; in Longhi's [17] models the mole fractions of Si in the partial melts range from 0.46 to 0.51, in which cases the most likely value for K would be ~0.34. If so, the parent magma(s) of the mg-87 olivines had to have mg ~69 (assuming, conservatively, that the troctolites contain no trapped liquid, and no ferroan anorthosite material was assimilated). That the magma(s) parental to the Mg-rich suite had high mg values is also manifested by the mg values of highlands soils. Among the Apollo-16 soils mg is very constant, and averages 66 [11]. The mg of the Luna-20 soil is ~69 [2, 8, 12, 25]. The soils' other major components, non-Mg-rich suite pristine rocks (mostly ferroan anorthosites and KREEPy rocks), and chondritic materials, as a rule have mg <65, so the average mg of the Mg-rich component in the soils is probably at least 68.

Table 1 lists a few of the possible combinations of K, liquid mg, bulk system mg, and % melting, that might have yielded the Mg-rich suite. Note that the K values in these cases are composites, because part of the solids would be pyroxene, which probably has a lower K than olivine: 0.30 was used by [17] for low-Ca pyroxene. Literature data [3, 27, 29, 39] for the K of low-Ca pyroxene scatter somewhat, but pyroxene should easily offset the small effect of pressure to increase the olivine K [3, 19]. Models with composite K >0.33, or liquid mg <69, are probably not very realistic. Whatever composite K is assumed during melting, the liquid must have a high enough mg to subsequently precipitate mg-88 olivine, where K is virtually the same, possibly even higher. As may be seen from the Table, there are no conceivable combinations of assumed K plus assumed liquid mg that are capable of yielding mg-88 olivine, unless the degree of melting is very large, or the mg of the bulk system is >>80. Indeed, in view of the low probability of melting far beyond 40%, the Mg-rich suite data suggest that the lunar mantle's mg is not quite as low (<80) as the seismic data (as interpreted by [4]) indicate. Note that the residuum from the melting has a much higher mg than the bulk system, so if a significant amount of the mantle were involved, the mg of the present-day mantle should be higher than the bulk system value.

Crustal abundances of Eu and Sr support the hypothesis that RL elements in the Moon's bulk composition are in at least nearly chondritic proportions to each other. At extremely low oxygen

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fugacities, such as prevailed on the Moon, the Eu plagioclase/liquid distribution coefficient,  $D(Eu,pl) \sim 1.1$  [24, 39]. Likewise,  $D(Sr,pl) \sim 1.7$  [6, 23, 24, 30; review: 9]. With all other important lunar minerals, these two RL elements act as incompatible elements. Hence, Ca and Al (two other RL elements), which determine plagioclase-liquid equilibria, also determine Eu and Sr fractionations.

The proportion of plagioclase precipitated along mafic silicate-plagioclase cotectics varies somewhat as a function of Mg, but probably averages very close to 50 wt. % for lunar magmas [see 1, 14, 29, 32]. Therefore, the point at which plagioclase is exhausted during partial melting (and commences crystallizing during fractional crystallization) comes when the % liquid in the system  $\sim 2.0 \times$  the % plagioclase in the norm for the system. Barring complex, multi-stage differentiations, a lunar liquid's contents of Eu and Sr can never be higher than they are at this point, without acquiring "anomalies" due to plagioclase fractionation. Also, the more "primitive" lunar plagioclase compositions all contain  $\sim 13 \times$  as much Al as the silicates of an H chondrite. Therefore, for the liquid to be 50% "plagioclase" means that its Al content must be  $\sim 7 \times$  H chondrite silicates, assuming the system's Ca/Al is not much lower than chondritic (Ca/Al is much lower in plagioclase than in chondrites). Thus, there are two predictions that follow from the hypothesis that RL in the Moon are in chondritic proportions to each other: 1) Rocks of the feldspathic lunar crust, obviously formed from liquids "evolved" past the point where plagioclase was a liquidus phase, should have Eu and Sr contents that are seldom  $< 7 \times$  chondritic silicates, and seldom  $>> 7 \times$  (because  $D(Eu,pl)$  and  $D(Sr,pl)$  are high enough to result in almost zero fractionation, when plagioclase is a liquidus phase). 2) The ferroan anorthosites are believed to have formed from the primordial magma "ocean" [36, 37, 10], and many of them are nearly pure plagioclase. These samples should have Eu almost precisely =  $7 \times 1.1$  (i.e.,  $D(Eu,pl)$ )  $\times$  chondritic silicates, and likewise Sr =  $7 \times 1.6 \times$ .

Fig. 1 and 2 show Eu and Sr contents of pristine lunar rocks (the unlabeled noncumulates are averaged data for mare basalt groups, taken from [31]), plotted vs. Sm (a REE adjacent to Eu). The numbers along the ( $\sim$  diagonal) lines of chondritic ratios give enrichment factors relative to the silicate fractions of H chondrites. The nearly vertical lines in the upper right of each figure trace fractional crystallization, with various  $D(pl)$  values being assumed. Cumulates other than the pure plagioclase anorthosites are difficult to interpret on an individual basis, because they may contain trapped liquid (which has the effect of displacing them almost vertically upwards) or cumulus mafics (which displace them almost horizontally to the left). Fortunately, almost none of the pristine cumulates contains more cumulus mafics than cumulus plagioclase (two exceptions are 72417 [7, 13] and 67435c [20, 26]). Western cumulates seem to be atypical of the overall crust [38], so they should almost be ignored. Nevertheless, the general distribution of the rocks is well in accord with prediction (1) above. As for prediction (2), the pure plagioclase anorthosites are seen to plot precisely as predicted (the nearly vertical lines in the lower right of each figure), assuming plagioclase fractionation started when the magma ocean was enriched in Al, Sm, Eu and Sr to  $7.3 \times$  H-chondritic silicates. The value for  $D(Sm,pl)$  is from [21], lowered slightly to be commensurate with the most likely T of the parental magmas [22].) Note that the anorthosites that plot farthest to the left of the lines are those which contain relatively large amounts of cumulus mafics [e.g., 37].

These tests are fairly precise. If, for instance, Eu/Al were  $2 \times$  chondritic, then the pure plagioclase anorthosites would contain  $2 \times$  as much Eu as they do; if Eu/Al were  $0.5 \times$  chondritic, then they would contain  $0.5 \times$  as much Eu. Incidentally, the same types of predictions can be tested for the eucrite parent body [34], with greater precision, because the eucrites underwent less complex differentiation than most lunar rocks. The eucritic data can even be utilized to check the accuracy of the experimental  $D(Eu,pl)$  and  $D(Sr,pl)$ . The indicated values are  $\sim 1.1$  for  $D(Eu,pl)$  (this is based on a better model than that of [34], which was oversimplified, and indicated  $\sim 1.4$ ), and  $\sim 1.5$  for  $D(Sr,pl)$ . The bulk composition of the eucrite parent body appears to have RL elements in precisely chondritic proportions to each other.

In summary, the evidence for the Moon having a subchondritic Ca/Al is at best ambiguous. The evidence for chondritic Eu/Al and Sr/Al is stronger. It is hard to imagine how RL trace elements could be in chondritic proportions to Al, if Ca is not, as well.

Table 1

	0.25	0.25	0.30	0.25	0.30	0.35	0.25	0.30	0.35	0.30	0.35	0.40	0.30	0.35	0.40
K	66	68	68	70	70	70	72	72	72	74	74	74	76	76	76
Liquid mg	88.6	89.5	87.6	90.3	88.6	87.0	91.1	90.0	88.0	90.5	89.0	87.7	91.3	90.0	88.0
solids mg	64	70	66	75	73	69	81	79	77	86	85	83	93	92	92
m% <sub>m</sub> , if mg = 73	55	61	56	67	63	58	74	70	67	78	75	72	85	83	81
= 80	45	52	45	58	52	45	65	60	55	68	64	59	76	73	69
= 82	34	41	31	47	40	30	54	48	40	56	50	42	65	60	54
= 84															

Note: "m%<sub>m</sub>, if mg = " stands for mole % melting, and is calculated for various assumed bulk system mg values, based on the (conservative) assumption that mole % (Mg + Fe) in the solids =  $2 \times$  mole % (Mg + Fe) in the liquid.





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A TERRESTRIAL ANALOG OF LUNAR ILMENITE BEARING CUMULATES. P.W. Weiblen and J.D. Miller, Jr., Dept. of Geology and Geophysics, Univ. of Minn., Mpls, MN 55455; Pekka Lestinen, Geological Survey of Finland, Box 237, SF-70101 KUOPIO 10 Finland; and Patrick J. Ryan, Mineral Resources Research Center, Univ. of Minn., Mpls, MN 55455.

Current interpretation of differences between ferroan anorthosites and the Mg-rich plutonic suite rocks of the lunar highlands crust suggests (on the basis of Ti/Sm data) that the Mg-suite reflects an episode of ilmenite fractionation prior to crystallization of norites and troctolites (1). Similarly, the present consensus for high-Ti mare basalt petrogenesis appears to require ilmenite bearing source rocks (2-4). Despite this indirect evidence for ilmenite-bearing cumulates in the lunar crust, such rocks are conspicuous by their absence in the catalogue of identified rock types in highlands samples. This may be a sampling problem related to levels of excavation during early meteorite bombardment of the highlands crust; a reflection of the limited sampling of the Luna and Apollo programs, or merely due to the incomplete characterization of the collections.

Two ilmenite-rich rock types have been described from the Apollo collection: sodic ferrogabbro clasts in Apollo 16 polymict breccia 67915 (5 and references therein) and just recently an olivine-spinel cumulate clast with 4 modal % intercumulus ilmenite in Apollo 14 breccia 14305 (Fig. 1, Table 1) (6). In light of present data, the genetic relationship of both of these two types of ilmenite bearing clasts to the two highlands suites is equivocal (1, p. 323 and 6). However additional data on their origin can place critical constraints on fractionation models of early evolution of the lunar crust and/or partial melting models of mare basalt petrogenesis.

The texture and mineralogy of an ilmenite peridotite from the Duluth Complex (Fig. 2, Tables 1,2) bears a striking resemblance to the olivine-spinel ilmenite bearing clast in 14305 (Fig. 1). The significance of the ilmenite peridotite as an analog lies in the fact that data on its origin can constrain fractionation models. It appears to represent the mafic counterpart of the anorthositic series rocks in the Duluth Complex (7). Mass balance calculations on major elements indicated that an early high-Mg lava composition could be the parent for the peridotite, anorthositic series rocks and late stage granophyre in the proportions 65:25:10. Recent REE data further substantiate this interpretation (8). Analysis of the processes of crystal melt segregation which resulted in efficient separation of plagioclase from mafics have been hampered by the limited exposures of peridotite at Duluth (Fig. 3) and the fact that the peridotite is probably in fault contact with overlying anorthositic rocks. However recent drilling has intersected the peridotite north of Duluth and analysis of layering in the drill core will aid in interpretation of segregation models. Geochemical studies of major elements and REE of the ilmenite peridotite and anorthositic series rocks in the Duluth Complex may shed light on the processes by which the Ti/Sm and Sc/Sm depletions in the Mg-suite lunar magma systems occurred (1).

The lower An and mg values of the Duluth Complex and the rift environment in which it formed does not make it at first sight a likely source for terrestrial analogs of lunar petrogenetic processes. However, in addition to the relevance of the ilmenite peridotite to lunar petrogenesis, it is intriguing that one major element analog of high-Ti Apollo 11 basalt is a recrystallized lava from the Duluth area (Table 2). It is possible that

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this flow tapped the magma chamber that produced the ilmenite peridotite analogous to the model proposed for the 14305 Fra Mauro clast (6).

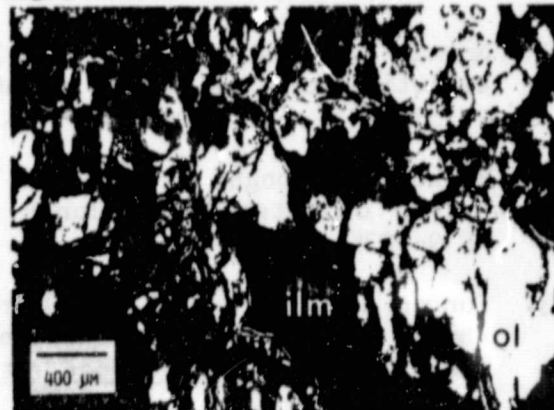
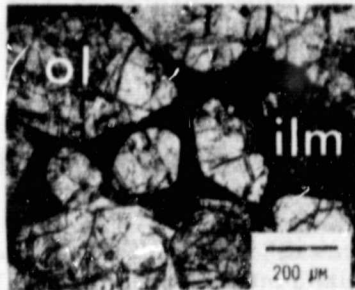


Fig. 1. Olivine-spinel cumulate clast in 14305,92 (6). Ol and Sp are poikil-  
enclosed in pyroxene and Ilm. Fig. 2. Ilmenite peridotite from the  
Duluth Complex. Cumulus Ol is enclosed by intercumulus augite and Ilm.

Table 1. Modal Proportions of Ilmenite Peridotites

Mineral/Mode	1	2
Plagioclase	15.2	10.
Spinel	2.5	1.
Olivine	39.7	50.
Ca-rich pyroxene	9.8	25.
Ca-poor pyroxene	28.8	?
Ilmenite	4.2	14.

1. Mode of clast in 14305,92 (6).

2. Typical mode of ilmenite rich layer in ilm-peridotite in the Duluth Complex.

Table 2. Mineral and rock analyses from the Duluth Complex

Number	1	2	3	4	5	6
SiO <sub>2</sub>	51.70	0.10	35.10	0.09	41.31	47.65
TiO <sub>2</sub>	1.32	50.20	0.11	11.47	7.04	1.93
Al <sub>2</sub> O <sub>3</sub>	2.09	0.10	0.05	7.42	12.12	8.31
FeO	10.20	45.70	38.40	51.04	17.72	13.14
MgO	14.70	2.43	26.30	2.52	6.58	10.99
MnO	0.19	0.54	0.32	0.46	0.21	0.19
CaO	19.50	0.10	0.16	0.00	11.07	11.84
Na <sub>2</sub> O	0.23	-	-	0.00	2.06	1.98
K <sub>2</sub> O	-	-	-	-	0.16	0.88
Cr <sub>2</sub> O <sub>3</sub>	0.24	0.37	0.12	19.10	-	-
NiO	0.02	-	0.09	0.00	-	-
Total	100.29	99.54	100.65	92.14	98.32	96.91
Atoms						
O	6.00	6.00	4.00	4.00		
Si	1.925	-	0.986	0.004		
Ti	0.037	1.916	0.002			
Al	0.091	-	-	0.350		
Cr	0.007	0.015	0.003	0.604		
Fe	0.311	1.939	0.903	1.707		
Mn	0.006	0.023	0.008	0.016		
Mg	0.816	0.184	1.102	0.150		
Ca	0.778	-	0.005	0.00		
K	-	-	-	-		
Ni	0.001	-	0.002	0.001		
Total	3.988	4.077	3.010	3.176		

Notes: 1-4 Mineral Compositions from ilmenite bearing cumulate (Fig. 1, Table 1)

1. augite; 2. ilmenite; 3. olivine; 4. chrome spinel; 5. basalt hornfels at Duluth (9) resembles Apollo 11 High-Ti; 6. Keweenaw basalt flow which is a possible parent basalt magma composition for ilmenite peridotite, anorthositic gabbro and granophyre.

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## THE LUNAR MAGMA OCEAN, THIRTEEN YEARS AFTER APOLLO 11

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Was there a lunar magma ocean? A better way to couch the question is, first, was there a major plagioclase-fractionating event early in the history of the moon? And second, did this require a magma ocean?

That a major early fractionation occurred is proven many times over by the trace element patterns and isotopic systematics of lunar samples. That the fractionation involved plagioclase is clear from the mineralogy of highlands samples. The amount of plagioclase that had to be fractionated is dictated by the gravity data. The nearside highlands regions, which are close to a state of isostatic compensation, stand an average  $\sim 2.6$  km above the ringed maria. The latter display  $\sim 300$  mgal positive gravity anomalies, relative to  $\sim 0$  mgal over the highlands. This situation requires that the highlands are underlain by less-dense material than the maria. The only abundant low-density mineral to be held responsible for the underdense state of the highlands crust is plagioclase, which of course squares with the observed mineralogy of highlands rocks. The amount of plagioclase needed to create the gravity/topography contrast sketched above can be calculated straightforwardly. If the densities of plagioclase (anorthite assumed) and the lunar mantle are 2.76 and 3.35 g/cm<sup>3</sup> respectively, 29 km of plagioclase must underlie the nearside highlands (1). It doesn't matter whether this plagioclase is concentrated as monomineralic anorthosite or dispersed in a variety of plagioclase-bearing rocks, but one way or another the equivalent of 29 km of plagioclase has to be present in the uppermost 100 km or so under the nearside highlands.

A moon with bulk composition equal to the silicate phase of ordinary chondrites (minus alkalis) contains 11.1 vol. % normative anorthite. 43% of such a moon would have to be melted to obtain a 29-km global layer of pure plagioclase, assuming perfect extraction. Thus there is no question about the need for a major plagioclase-fractionating event early in lunar history. But did this require a single, colossal magma ocean, as was proposed in 1970 (2, 3)? Or could the plagioclase simply reside in a  $\sim 60$ -km near-surface zone of basaltic flows and gabbroic intrusions, as Walker (4) has more recently argued?

The evidence remains compelling that a single magma system of great vertical extent produced the lunar highland rocks. That a single system was involved rather than many small ones is strongly indicated, though not unequivocally proven, by the fact that pristine highland and KREEP samples from Apollo sites across the moon align on "total rock" Sm-Nd and Sr-Rb isochrons corresponding to fractionation  $\sim 4.4$  Ga ago (5,6). That the system was deep follows from the absence of mafic cumulate samples, complementary to the abundant plagioclase cumulates, in the Apollo collection. If the crystal fractionation that produced lunar anorthosites occurred in mere flow units or plutons, many impacts-- certainly the basin-forming impacts-- would have exhumed their ultramafic lower layers and mixed these lithologies into the regolith.

On the other hand, the 1970 concept of a magma ocean of bulk-moon composition, from which a great amount of olivine had to crystallize before plagioclase saturation occurred, is almost certainly wrong (unless the moon was totally melted). Whatever melted the magma ocean, the separation of melted and unmelted components would have taken long enough to allow the components to equilibrate with one another (approximately), leaving a liquid of partial-melt

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composition in the magma ocean.

The most serious doubts about the magma ocean have come from the difficulty in understanding what could have melted it. The energy source had to be adequate in amount, and available on the right time scale. Only accretional energy seems even potentially capable of meeting these requirements. Even accretional energy would be ineffectual if the accretion was too slow ( $> \sim 1000$  yr; 7) and/or the material accreting too small in dimension, because then most of the heat generated would be lost immediately by radiation from the lunar surface. In 1970 I (and others) considered the accretion rate to be a free parameter, and the existence of the magma ocean to be evidence enough that accretion occurred rapidly. However, extensive studies of the dynamics of planetary accretion (8) have shown that this process extends over a period of  $\sim 10^8$  yr. This time scale must also apply to lunar accretion, since new material would continue to be fed into the protolunar swarm, and ultimately accreted by the moon, throughout the  $\sim 10^8$  yr. Thus rapid lunar accretion cannot be invoked as a means of conserving the energy of accretion.

Accretion over such a protracted period is still capable of effecting a substantial amount of melting, if the planetesimals being accreted are large enough to deposit energy deep below the lunar surface (8,9). However, the melting would occur piecemeal, producing a series of magma bodies, each of which was capable of differentiating and solidifying on its own time scale. It is not at first apparent that such a system could effect a net transfer of plagioclase upward to the present lunar surface; seemingly it could rise only to the top of each magma chamber. But in fact, there would be some tendency for new melt-zones to involve and remelt only the tops of older magma chambers, since the moon is growing while this succession of melting events occurs; so the magmas would become increasingly felsic. The net effect, operating at an unknown degree of efficiency, would be to skim the plagioclase component of the moon to its surface by a form of zone refining.

However, it seems unlikely that such a system of discrete fractionations would produce rocks in widely separated regions that align on isochrons, or that it would leave its mafic cumulates so deeply buried that their debris is not found on the lunar surface. The need for a single, deep magma ocean remains.

There is one model of lunar formation that would inject a large amount of accretional energy in the moon in such a short time that practically all of it would be conserved as buried heat. This involves the accretion of two or more discrete submoons, which do not coalesce until they have achieved substantial size. Ruskol, who has proposed this mode of origin, shows that 37% of the total energy of accretion of the moon would be released when two 0.5-lunar mass submoons coalesced, in the space of  $\sim 1$  hr (7).

Multiple submoons cannot stably coexist at similar orbital radii; they will coalesce in  $< 100$  yr (10). To survive through most of  $10^8$  yr of accretion before coalescence (so that they incorporate most of the material now in the moon), they have to spend most of that time at substantially different orbital distances. The most plausible scenario is one in which submoon 1 nucleates from the protolunar swarm close to Earth, then is worked out to a substantially larger orbital radius by tidal friction with the earth, sweeping up material as it goes; then after the swept-up inner zone of the proto-

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lunar swarm becomes repopulated, submoon II nucleates there and is in turn worked outward by tidal friction. The rate of tidal recession decreases with distance, so eventually submoon II approaches submoon I and coalescence occurs.

Would such an event provide enough heat to melt a magma ocean? Several energy inputs have to be considered. Ruskol (7) estimates that coalescence of two 0.5-lunar mass submoons at the escape velocity would produce a mean rise in temperature of  $\sim 600^\circ\text{K}$ . Allowing for the fact that the submoons would have perturbed one another's orbits, so that coalescence occurred at somewhat greater than the escape velocity, adds  $\sim 75^\circ\text{K}$  to this value. Much depends upon the internal temperatures of the submoons before coalescence. Kaula (11) estimates that energy released during the prior accretion (from many planetesimals) of a 0.5-lunar mass moon would have heated its interior to  $\sim 400^\circ\text{K}$  (center) -  $\sim 1000^\circ\text{K}$  (near-surface). His calculation assumes an initial temperature of  $\sim 250^\circ\text{K}$  for the material accreting to the moon; to the extent that the submoons accreted from planetesimals that were already hot from  $^{26}\text{Al}$ -decay, or whatever else heated the parent meteorite planets, the pre-coalescence temperatures of the submoons may have been higher than this. Finally, if more than two submoons coalesced to form the moon, more accretional energy is conserved as heat than in the case of two submoons (7). Combining these effects, we arrive at temperatures of  $>1075^\circ\text{K}$  -  $>1675^\circ\text{K}$  in lunar material after coalescence of submoons, clearly enough to promote major amounts of melting.

The coalescence would have been a very complicated event. Heat generated during the coalescence would not have been uniformly distributed throughout the moon, as the above analysis assumes, but concentrated in the most-strained zones. In these, temperatures much in excess of  $1675^\circ\text{K}$ , perhaps high enough to cause total melting, would have been reached. Other zones may not have been heated enough to promote even partial melting. The resulting melange of unmelted, partly-melted, and totally-melted materials would have required years to relax to a mechanically, chemically, and thermally stable configuration, which would have consisted of a magma ocean of more or less gabbroic composition surrounding an interior zone of unmelted material.

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