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CHARACTERIZATION AND DISTRIBUTION  
OF LUNAR MARE BASALT TYPES USING  
REMOTE SENSING TECHNIQUES

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May 1977                              Planetary Sciences  
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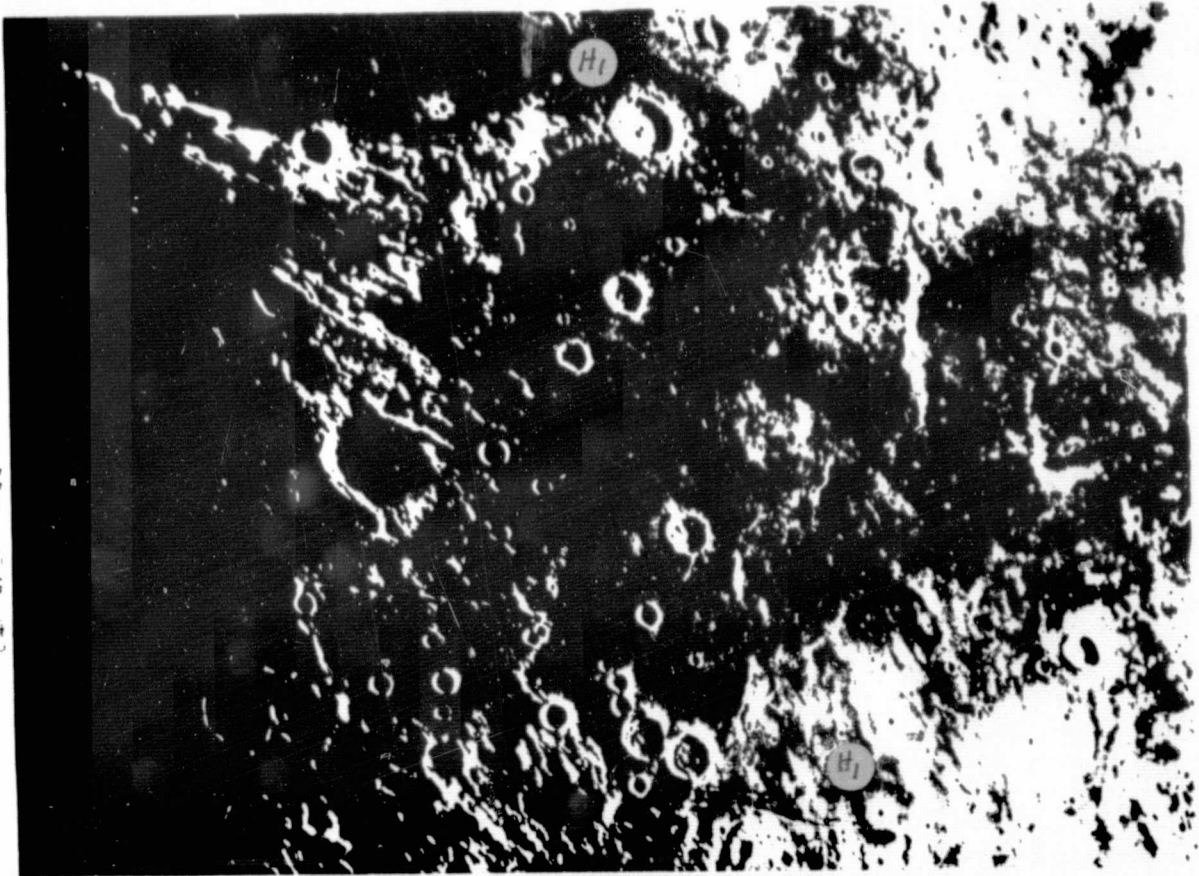
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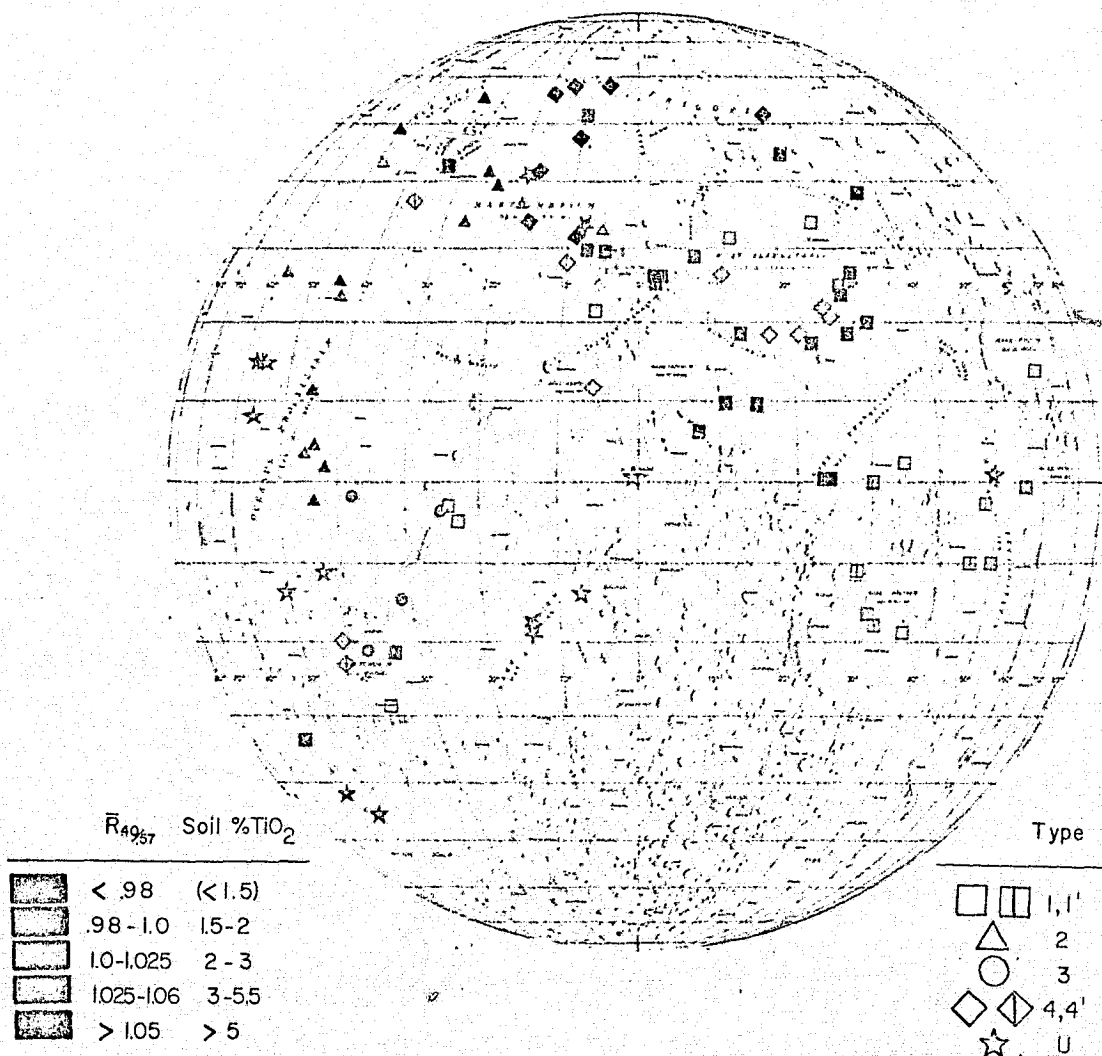
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May 1977

Department of Earth and Planetary Sciences  
Massachusetts Institute of Technology



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## MARE BASALT CLASSIFICATION



Classification of nearside lunar basalt types using telescopic reflectance spectra from .3 to 1.1  $\mu\text{m}$  of small (5-20 km diameter) mare areas (Pieters and McCord, Vol. 3). The shape of the oversized symbols indicates a distinction of basalt types based on near-infrared features of the soil spectra. The color of the symbol indicates the percent TiO<sub>2</sub> of surface soil as derived from continuum slope of the spectrum:  $R_{.40/.57}$  = reflectance at .40  $\mu\text{m}$  relative to a standard area in Mare Serenitatis (MS-2) and scaled to unity at .57  $\mu\text{m}$ . The data indicate that there exists a variety of basalt types unsampled by the landing missions.

CHARACTERIZATION AND DISTRIBUTION OF LUNAR MARE  
BASALT TYPES USING REMOTE SENSING TECHNIQUES

by

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CHARACTERIZATION AND DISTRIBUTION OF LUNAR MARE BASALT TYPES  
USING REMOTE SENSING TECHNIQUES

by

Carle Pieters

Submitted to the Department of Earth and Planetary Sciences, MIT  
on May 6, 1977 in partial fulfillment of the requirements  
for the Degree of Doctor of Philosophy

ABSTRACT

Although their composition is narrowly defined, basalts are igneous rocks emplaced on a planet's surface as melts generally derived through partial melting of the planet's interior. Variations in the age, volume and chemistry of basalts from different provinces provide clues to the composition and structure of the interior. The types of basalt that can be identified on the surface of the moon have been studied using spectral reflectance measurements. Reflectance spectra (.3 to 1.1  $\mu\text{m}$ ) have been obtained using earth based telescopes for a wide variety of lunar mare surfaces and craters and have been classified according to their spectral features. Geochemical interpretations are provided from a background of laboratory reflectance measurements of lunar, terrestrial and meteoritic samples. Some of the principal results of this study are: a) There are major (regionally extensive) basaltic units on the moon not returned by Apollo or Luna missions to date. About 2/3 of the lunar maria are not represented in the lunar sample collection. b) Particular extensive units unsampled include: i. medium to high-Ti basalts near Flamsteed in southern Procellarum and in western Imbrium, ii. medium high-Ti basalts of Humorum and SE Procellarum and iii. low-Ti basalts of NE Imbrium and Frigoris. c) The existence of late stage high-titanium basalts is confirmed although these young basalts are distinctly different in mafic composition from the early high-titanium basalts. d) All mare surfaces contain abundant lateral variations of compositionally heterogeneous basalts. e) Some maria are vertically inhomogeneous with subsurface basalts being distinctly different in composition (eg. low-Ti surface, high-Ti subsurface). f) Some basalt types are spectrally gradational suggesting minor variations in geochemistry. Three regional series can be identified. g) The data suggest that most ( $\geq 80\%$ ) of the lunar surface is composed of a finite number of discrete describable compositional units. h) Mineral components of unsampled units can be defined if spectra are obtained with sufficient spectral coverage (.3 to 2.5  $\mu\text{m}$ ) and spatial resolution ( $\sim .5$  km).

Thesis Supervisor: Dr. Thomas B. McCord

Title: Associate Professor of Planetary Physics

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A PhD thesis represents the formal part of a transition in a person's life from clever student to professional. The process that leads to a PhD degree is different for each student. When I obtained my first undergraduate degree eleven years ago, it never occurred to me that a doctor's degree in science was a realistic goal in my life. The process that led to this thesis perhaps first started with a vague dissatisfaction with my previous position as a high school teacher and a non-directed but avid interest in astronomy and planetary science. During the last seven and a half years at MIT (three of which were as a junior staff scientist), the foundation for a strong and confident scientific career was gradually developed. A few key people provided the essential personal and academic support so important in this process. I will always be deeply and warmly grateful to Drs. Tom McCord, John Adams, Jim Head, Tom McGetchin, and Roger Burns. Without their help and encouragement it is unlikely I would ever have developed scientific enthusiasm nor faith in my own contributions. The continuous manner in which Dr. Tom McCord was able to integrate the difficult roles of teacher, advisor, friend and colleague is specially appreciated.

The roots of the thesis content presented here are difficult to define and are imbedded in almost all the work I have been involved with at MIT in association with Tom McCord and John Adams. However, three people have generously acted as principal advisors throughout the period of thesis preparation and contributed significantly to the content and structure of the project: Drs. Tom McCord, Jim Hays, and Roger Burns.

Much of the telescopic data used in this thesis was obtained while I was a guest observer at the following observatories: (1) Mount Wilson Observatory (24-inch telescope) which is operated by the Department of Geological Sciences, California Institute of Technology, Pasadena; (2) Cerro-Tololo Inter-American Observatory, which is operated by the Association of Universities for Research in Astronomy, Inc., under contract with the National Science Foundation; and (3) Mauna Kea Observatory, Institute for Astronomy, University of Hawaii. The research was supported by NASA grants NSG-7048 and NGL-22-009-790 (Tom McCord, principal investigator).

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## EXTENDED OUTLINE

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## PREFACE

The content of this thesis can be considered in three parts: a background section (I) that discusses what basalts are and why they are interesting, a section (II) that provides detailed background of spectral reflectance remote sensing techniques, and three sections (III, IV, and V) that discuss the application of spectral reflectance techniques to understanding the geochemistry of the lunar maria. The length of each section is not necessarily proportional to its contribution to the scientific output of this thesis, but is determined more by an attempt for completeness.

Section I is a brief description of basalts in the solar system and could easily be skipped by geochemists and petrologists familiar with terrestrial and lunar basalts. It is clear this section is too short for the detailed discussion that the topic merits. On the other hand, I realize the length of this background material is perhaps out of proportion with its direct relevance to the content of the thesis. I hope that sufficient detail has been presented so that readers unfamiliar with basalts are able to understand what they are and why they are interesting.

Section II provides detailed information concerning spectral reflectance studies of lunar material. Although much of this material has appeared in the literature during the last ten years, some of the basic concepts presented have not been discussed fully or are currently being prepared for publica-

tion by' the 5 people involved in the field. Emphasis is placed in this section on establishing a basis for interpretation of remotely-obtained spectral reflectance data.

Sections III, IV, and V contain most of the scientific contributions of this thesis. This material is largely in the form of self-contained manuscripts that were prepared as the thesis progressed. The content concerns the classification of remotely-obtained lunar reflectance spectra and the applications of spectral reflectance data to the characterization and distribution of lunar mare basalt types. The material presented in Section I and II provides the detailed background on the nature of basalts and spectral reflectance techniques that cannot be included in a (page-limited) published manuscript. The major conclusions that can be drawn from the currently available spectral reflectance data are summarized in Section V.

I. BASALT TYPES: AN OVERVIEW

## I. BASALT TYPES: AN OVERVIEW

The formal definition of a basalt concerns its chemistry, mineralogy, and mode of emplacement. Basaltic composition is generally 45-52%  $\text{SiO}_2$ ,  $\text{MgO} + \text{FeO} + \text{Fe}_2\text{O}_3 > 15\%$ , and  $\text{CaO}/\text{Al}_2\text{O}_3 > .60$ . The mineralogy of common basalts is dominated by calcic plagioclase ( $\text{An}_{>50}$ ), augite,  $\pm$ Ca-poor pyroxenes,  $\pm$ olivine, and a minor amount of opaques. Basalts are extrusive igneous rocks, i.e., melts which have cooled in a low pressure environment on the surface.

Basalts are suspected to have erupted on the earth's surface for at least the last 2.7 billion years. They were emplaced on the moon's surface between about 3.9 and 3.0 billion years ago (AE) and account for 17% of the moon's surface area. From photogeologic studies, basalts are expected to also have occurred in particular regions on Mars and Mercury. Laboratory studies of basalt samples from the earth and moon show they have crystallized from a melt derived from sub-crustal material and as such provide an opportunity to study the composition of the interior. The composition and age of basalts put limits on the thermal evolution of the planet. The basaltic achondrite class of meteorites fits the above chemical definition, and texturally are often considered igneous. Since there are asteroids

that have similar geochemical properties, asteroids are often considered to be the parent bodies for the achondrites.

More loosely, basalts are considered the major igneous rock types derived more or less directly by partial melting of a mantle source. As such, basalts provide important geochemical constraints on mantle composition. (The relation of basalts to the mantle at times involves a circular definition.) It is this more general description of basalts that makes them interesting as a planetary phenomenon. However, since the nature of the mantle, or source region of 'basalts', is not likely to be the same for all terrestrial<sup>1</sup> planets, the prevalent igneous rock type derived by partial melting of the interior may not necessarily be a basalt. For example, the lunar high-titanium 'basalts' commonly contain <40% SiO<sub>2</sub>, primarily because of their high TiO<sub>2</sub> content. The likely igneous rock type derived from the mantle of Mars would be extremely ultramafic (McGetchin and Smyth, 1977) and well beyond the range of terrestrial<sup>1</sup> basalts.

It is tempting to postulate a common period of 'basaltic' volcanism for all solid surface planets. Volcanism

---

<sup>1</sup>The adjective 'terrestrial' usually means pertaining to earth. However, in planetary science, it is often used to describe a whole class of solid surface objects. The 'terrestrial' planets include Mercury, Venus, Earth, Mars, and often the moon and asteroids.

is certainly a common occurrence on the terrestrial planets, but each planet (and the asteroids) has evolved in a unique manner. It is hoped that a detailed study of the basalts on each planet will help define some of the general principles of planetary evolution.

## I. BASALT TYPES: AN OVERVIEW

### A. Basalts as a planetary phenomenon

#### 1. Brief description of known and implied basalts

##### a. Earth

There are about 800 known active volcanoes on the surface of the earth today (Verhoogen et al., 1970, p. 262) and an undetermined number of dormant or extinct volcanoes. The composition, mineralogy and mode of occurrence of material from these sources are a function of the specific environment and planetary processes associated with the region (see Section I-B). It is clear that the earth is a dynamic planet and has been for most likely all its 4.5 AE history. The current surface of the planet is constantly undergoing severe alteration and rejuvenation from both tectonic and meteorological processes. The effects of plate tectonics which dominate the geology of the earth may be unique in the solar system. Nevertheless, there does exist stable regions in the continental plates that still contain a record of early volcanic rocks (see Section IB-3), implying that some form of basaltic volcanism occurred early in the earth's history, perhaps before 3.0 AE.

Of recent terrestrial basalts, the largest surface area and perhaps the largest volume of basalts occur

on the ocean floor and are generated by the midocean spreading centers. Additional forms of volcanism occur near the region of a subduction zone as oceanic and continental lithospheric plates collide: (1) the island arc volcanics which include a variety of basalts and more silicious material, and (2) the continental volcanics which are generally more silica-rich (e.g., andesite). Major volumes of basalts can also occur within a lithospheric plate such as is observed with the Hawaiian series of shield volcanoes and the extensive continental flood basalts (e.g., Columbia River, Deccan). Alkali-rich basalts occur both on ocean islands and also well within a continental plate. Additional continental volcanics associated with rift zones are often extremely complex and may include such 'exotic' compositions as carbonatites as well as more familiar compositions such as basalt-rhyolite associations. In short, terrestrial volcanism is not only currently active in a sometimes perplexing variety of forms, but it also has a complex history as a planetary phenomenon, much of which has been erased.

b. The Moon

There are two major geochemical families of material on the lunar surface: the mare basalts and the feldspathic highland crust. The composition, mineralogy

and texture of returned mare rocks are clearly basaltic (Section I-C). Although a component of highland material (perhaps 20%, Taylor, 1975, p. 255) can be considered basaltic in composition, many highland rocks with igneous textures are considered to be the result of impact melts (Irving, 1975) and the highlands in general are not basaltic.

Mare basalts account for about 17% of the lunar surface area and less than 1% of the total volume for an estimated 60 km thick lunar crust. [This and much of the details that follow can be found in a review by Head (1976)]. Basaltic volcanism occurred on the front side of the moon filling the lowlands and forming the maria from about 3.8 to 2.5 AE (estimate). Some form of volcanism may also have occurred in the highlands prior to the filling of the lowlands by mare basalts. Lunar mare lavas were very fluid and voluminous and formed extensive flows. The lunar crust formed early in the evolution of the planet (<4.1 AE). The mare basalts were derived from partial melts of sub-crustal materials and were later emplaced in the low-lying regions of the fractured surface. The total volume of basalt is estimated to be about  $10 \times 10^6 \text{ km}^3$ . The process of filling the maria took place over about 1.3 billion years and produced complex deposits of smooth plains. About 2.5 billion years ago, lunar volcanism ended. The

evolution of the atmosphereless surface was then essentially frozen with only meteoroid bombardment to alter its form. Unlike the earth, the moon provides an opportunity to study early stages of the evolution of a planet.

c. Mars

The largest known volcano in the solar system is the Martian Olympus Mons [500 km diameter, 23 km high,  $2.12 \times 10^6$  km<sup>3</sup> volume (Blasius and Cutts, 1976)]. From Mariner imagery many such volcanoes (but smaller) can be identified (Carr, 1973) with a variety of ages. Although much of the highly cratered surface of Mars could have an age comparable to the lunar highland crust, the Tharsis volcanoes could be younger than 750 m.y. old (Soderblom et al., 1974). Some of the volcanic plains (similar to the lunar maria) could be as old as 2.4 AE. The composition of Martian soil as determined by the x-ray fluorescence experiments of Viking could be consistent with many mineral assemblages which range from ultramafic igneous rock to an iron-rich extensively weathered product (Toulmin et al., 1976; Baird et al., 1976). Earth-based spectral studies of the Martian dark regions are consistent with oxidized basalt (Adams and McCord, 1969); laboratory and theoretical studies show photostimulated oxidation is the likely chemical alteration process (Huguenin, 1974). Although

Martian volcanism probably extended over a period longer than that demonstrated by the moon, Mars is clearly not as active as the earth today and may even have been dormant for the last billion years. The martian surface has undergone extensive mechanical and chemical weathering causing some of the record to be blurred.

d. Meteorites and asteroids

Prior to the Apollo missions, the only samples of extraterrestrial material available were fragments of unknown origin that fall to earth as meteorites. (See Was-son, 1974, for a detailed discussion.) Most (>70%) of these falls are ordinary chondrites, or chondrule-bearing silicate assemblages with textures that indicate they have not experienced a severe heating process. About 8% of the meteorite falls are classified as differentiated silicate-rich achondrites, or mineral assemblages (without chondrules) that possess mineralogical and textural properties which imply crystallization from a melt and perhaps fractional crystallization in a gravity field (Duke and Silver, 1967). The composition of the achondritic eucrites (monomict pigeonite breccias), which constitute <3% of falls, is essentially basaltic. The extrusive origin of some such eucrites is demonstrated by the existence of vesicles presumably formed by the expansion of gas evolved from a magma as the

result of pressure drop during extrusion onto a parent body surface.

The return of lunar samples allowed geochemical studies that effectively eliminated the moon as a possible source for the meteorites. Comets, belt asteroids (main body of asteroids in orbit 2.2 - 3.2 AU from the sun), and the Apollo asteroids (earth-orbit crossing) are the remaining possible meteorite sources. Spectrophotometry of over 100 asteroids (e.g., McCord and Chapman, 1975a, b; Chapman et al., 1975) and the mineralogical interpretation of the spectra (McCord and Gaffey, 1974; Gaffey and McCord, 1976) indicate that most asteroids can be understood in terms of mineral assemblages common to meteorite samples. The earth, however, receives a biased distribution of samples. Basaltic achondrite-like material is extremely uncommon among the asteroids with the main belt being dominated by material much like the relatively 'primitive' carbonaceous chondrites. The asteroid 4 Vesta is the singular example (McCord et al., 1970) of possible basaltic material; the spectrum of this asteroid indicates a surface with a pigeonite basalt assemblage much like the eucrites. Somehow it seems a small body with a radius 245-300 km having a density around  $3.0 \text{ g cm}^{-3}$  (4 Vesta) was able to develop a basaltic surface, while in the same part of the solar

system a larger object with a radius 465-590 km having a density around  $2.1 \text{ g cm}^{-3}$  (1 Ceres) remains undifferentiated (Matson et al., 1976).

e. Mercury

Although Mercury was strongly suspected to have a lunar-like surface, little was known about the planet's surface until the Mariner mission in 1974. Earthbased spectrophotometric measurements of the whole planet (McCord and Adams, 1972) indicated that the composition of the surface was comparable to the lunar highlands--being rich in the dark impact generated silicate glasses. Polarization measurements indicated that the surface microstructure was also lunar-like (Dollfus and Auriere, 1974). Meticulous low resolution visual observations implied there were at least albedo differences across the surface (Dollfus, 1961). There is, however, no direct evidence for basalts from the astronomical data.

The Mariner images showed the surface morphology of Mercury to be remarkably similar to the moon (Murray et al., 1974, 1975). Much of the surface is heavily cratered but there exist many level and less cratered surfaces (plains) including the floors and surrounding terrain of many craters and basins. The smooth plains are somewhat darker than the heavily cratered terrain, but the albedo

contrast is less than that observed for the moon. Although no direct evidence for volcanism such as cones or domes is observed, the volume and areal extent of the Mercurian plains material suggest a volcanic origin. If the lunar analogy holds for Mercurian early differentiation, the plains material could be basaltic.

f. Venus

The size and density of the planet Venus is not much different from that of the earth, but little is known about the surface of the planet due to the extensive and opaque atmosphere (~100 bars). Images of the surface were obtained in 1975 by the Soviet spacecrafts Venera 9 and 10 and described by Florensky et al. (1977). The surface contains both fine material and rocks. Some rocks are flat and show evidence of layering. One region contains sharp-edged "slabs" of rock whereas the other region shows rocks with smooth edges. Earth-based radar measurements (e.g., Goldstein et al., 1976) reveal major surface features reminiscent of the large basins on the moon and Mars. There is currently no evidence for or against the existence of volcanic material on the surface of Venus.

2. Limits of planetary chemical composition set by Solar System models

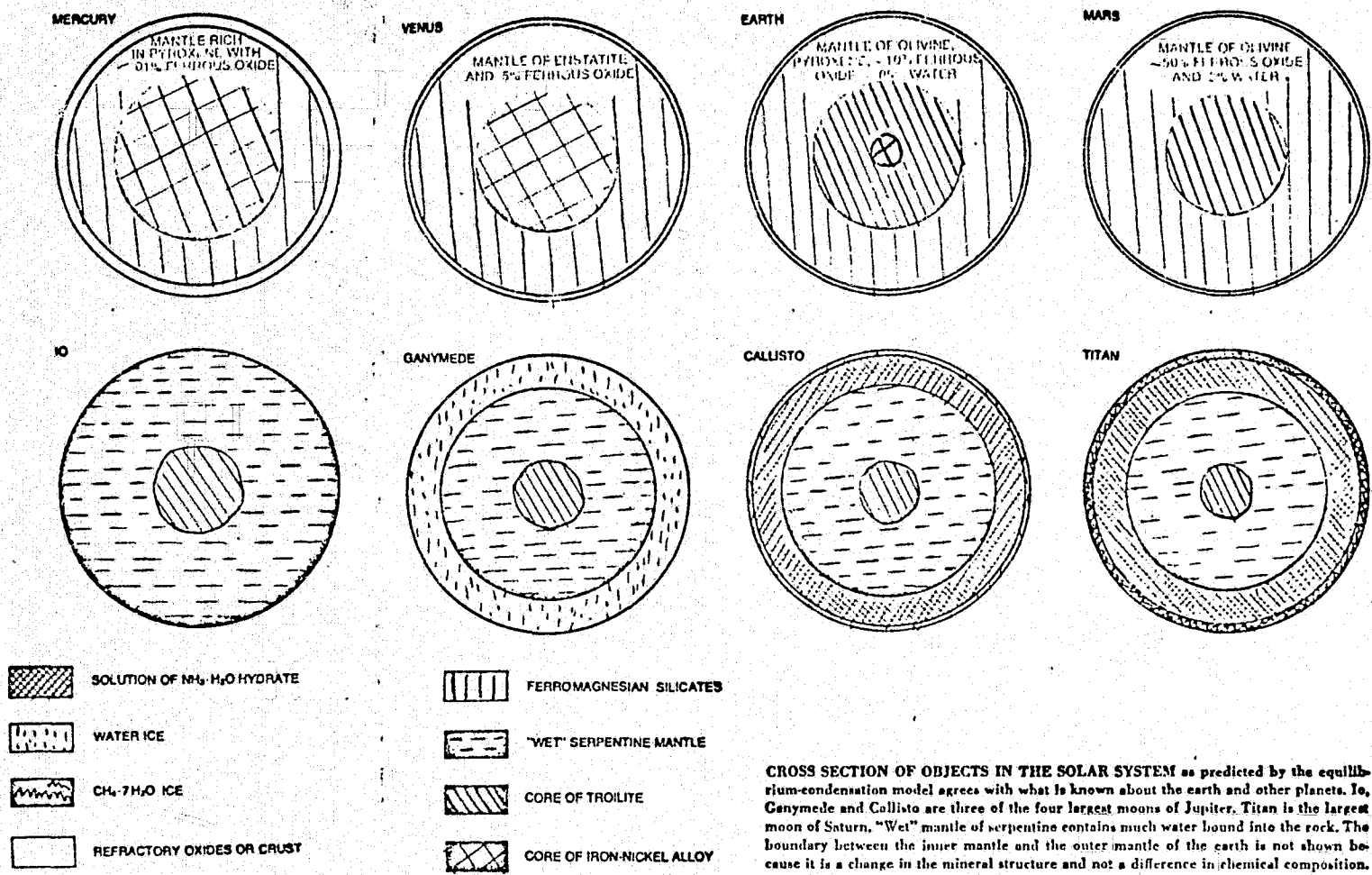
Since basalts are the products of crystallization from a magma, they are at least second generation material and are directly related to the original composition of the premelt material. Although some basaltic material has probably passed through more than one sequence of melting and differentiation, it is worthwhile to consider the possible initial starting material for the terrestrial planets. Models of condensation from a primitive solar nebula (PSN) describe simple condensation sequences that can account for most of the observed bulk densities of the planets and can also provide an estimate of their initial bulk chemistry (Lewis, 1974a, b; Barshay and Lewis, 1976).

The condensation models assume a PSN composition (Cameron, 1973) and a reasonable description of the pressure and temperature with radial distance along the PSN disk. In order to examine the predictive ability of a given model, the elemental composition of the PSN is assumed uniform throughout. This second assumption, although not purely accurate, is necessary; unconstrained deviations from simplicity can result in an infinite variety of predictions. The principles of chemical thermodynamics are applied to the assumed PSN and the primordial composition of material is predicted in the region where each planet formed. Both

equilibrium and disequilibrium models have been examined. The results of such models indicate that, with respect to major element composition and bulk density of the resulting planet, the terrestrial planets and asteroids are likely to have condensed under equilibrium or near equilibrium conditions and that the assumptions of the PSN model were thus not likely to be grossly different from reality.

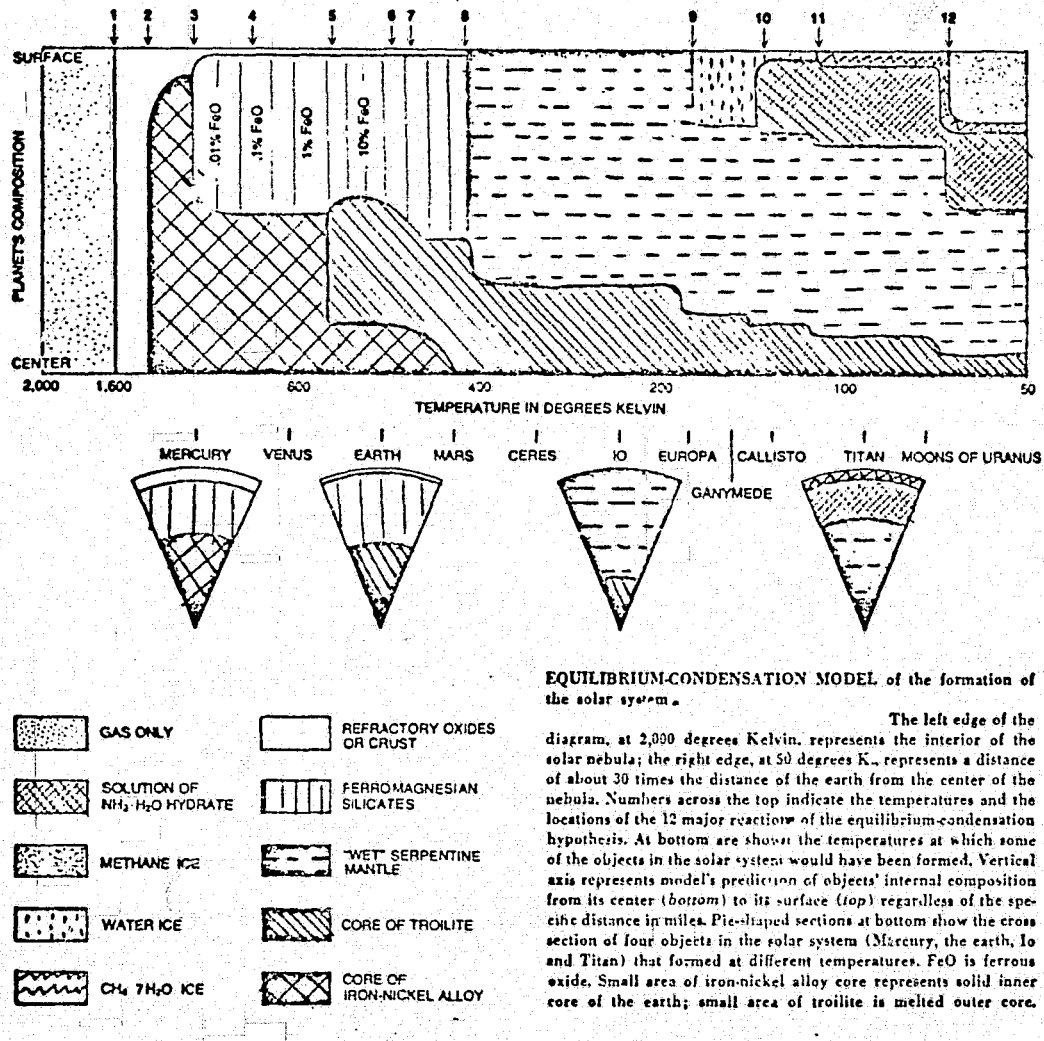
Compositions of various solar system objects predicted by the equilibrium condensation model are summarized in Figure IA1,2 (from Lewis, 1974b). Mercury is predicted to contain a major amount of refractory metals and minerals with an iron-nickel alloy core and a limited amount of  $\text{MgSiO}_3$  enstatite. Venus contains the same components as Mercury but with major amounts of enstatite and perhaps some alkali aluminosilicates. Venus is expected to have little if any sulfur or water. The Earth, by this model, contains a small amount of water and iron occurs as Fe metal, FeO, and FeS. Mars, on the other hand, contains little or no unoxidized iron and more water than the earth.

A few serious discrepancies exist, however, between predictions and observations. For example, the models do not allow the earth (density =  $5.5 \text{ g cm}^{-3}$ ) and the moon (density =  $3.3 \text{ g cm}^{-3}$ ) to form in the same region of the



CROSS SECTION OF OBJECTS IN THE SOLAR SYSTEM as predicted by the equilibrium-condensation model agrees with what is known about the earth and other planets. Io, Ganymede and Callisto are three of the four largest moons of Jupiter. Titan is the largest moon of Saturn. "Wet" mantle of serpentine contains much water bound into the rock. The boundary between the inner mantle and the outer mantle of the earth is not shown because it is a change in the mineral structure and not a difference in chemical composition.

Figure IA 1. (from Lewis, 1974b)



**EQUILIBRIUM-CONDENSATION MODEL of the formation of the solar system.**

The left edge of the diagram, at 2,000 degrees Kelvin, represents the interior of the solar nebula; the right edge, at 50 degrees K., represents a distance of about 30 times the distance of the earth from the center of the nebula. Numbers across the top indicate the temperatures and the locations of the 12 major reactions of the equilibrium-condensation hypothesis. At bottom are shown the temperatures at which some of the objects in the solar system would have been formed. Vertical axis represents model's prediction of objects' internal composition from its center (*bottom*) to its surface (*top*) regardless of the specific distance in miles. Pie-shaped sections at bottom show the cross section of four objects in the solar system (Mercury, the earth, Io and Titan) that formed at different temperatures. FeO is ferrous oxide. Small area of iron-nickel alloy core represents solid inner core of the earth; small area of troilite is melted outer core.

Figure IA 2. (from Lewis, 1974b)

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solar system without some secondary processes affecting the moon. Secondly, the nebular adiabat, along which condensation is presumed to have occurred, does not cross the graphite stability field, but carbonaceous chondrites (and probably some asteroids) are known to contain significant graphite. It is suggested by Barshay and Lewis (1976) that the carbon found in meteorites may be a remnant of interstellar dust and is an exception to the equilibrium condensation models.

Furthermore, there is an indication from the study of oxygen isotopic composition (e.g., Clayton et al., 1976) that inhomogeneities existed in the solar nebula prior to condensation. This evidence indicates that the earth, the moon, and the parent bodies of differentiated stony and stony-iron meteorites must have been derived from the same batch of solar nebular material. Five other groups of solar system material can be identified on the basis of oxygen isotopic composition, none of which can be derived from another by direct chemical fractionation processes. One of these groups, the ordinary chondrites which constitute the majority of meteorite falls on earth, cannot have been derived from the same homogeneous region of the solar nebula as the earth. The difference may imply that the earth had a higher mean condensation temperature.

### 3. Formation of Basaltic Liquids--Possible Heat Sources

Basaltic rocks by definition are igneous and crystallized from a melt. On the earth, basaltic magmas are thought to be derived by partial melting of a mafic or ultramafic mantle (See IB). The lunar mare basalts are also thought to be derived by partial melting of the interior (see IC). A number of possible heat sources for producing these melts are outlined below. A recent review of possible terrestrial melting processes is provided by Yoder, Chapter 4 (1976). The relative importance of each heat source is perhaps different for each planetary object.

#### a. Accretional Energy

The earliest history of a solar system has not been observed astronomically and is, therefore, derived from a mixture of fact and conjecture. At some time greater than 4.6 billion years ago (limiting age of meteorites; Wasson, 1974) solid particles condensed from a cooling solar nebula. These small grains began to agglomerate into larger objects. At some point the neo-planet became sufficiently large to allow gravitational accretion of enough mass to form a planet. For a given homogeneous mass  $M$  with a radius  $R$ , the accretional energy released is:

$$E = \frac{3}{5} \frac{GM^2}{R}$$

For a uniform earth, the gravitational energy per unit mass would be about  $4 \times 10^{11}$  ergs/g--enough to vaporize the earth twiceover (Verhoogen et al., 1970, p. 600-601).

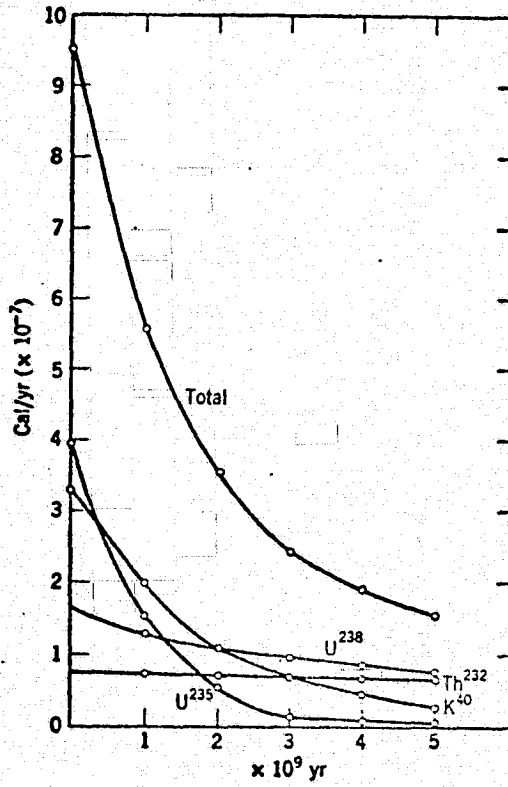
The key to how much of this accretional energy affects the original thermal state of the new planet is the degree to which the heat generated can be radiated away into space (see Mizutani et al., 1972). The amount of heat generated per unit time is a function of the rate of accretion. Accretional energy is expected to play a significant role in the initial temperature profile for a planet. Most models for the early history of the moon (e.g., Wood et al., 1970) require an original magma ocean hundreds of kilometers deep (see IC2). Accretional energy is usually cited as the probable heat source (e.g., Hubbard and Minear, 1975) with accretion time being less than 1000 years.

b. Radioactive Isotopes

Kinetic energy of the  $\alpha$ -particles and  $\gamma$  radiation emitted from radio nuclides is absorbed by surrounding atoms and converted to thermal energy. The heat produced during any given period is a function of the type and amount of radioactive isotopes. Using the estimated abundances of heat-producing elements on earth, the radiogenic heat production can be calculated as a function of

time as shown in Figure IA3 (from Mason). In the earth, the melting temperature of iron was reached after about 600 million years at a few hundred kilometers depth. Thermal evolution models of the moon (e.g., Toksöz and Solomon, 1973), Mars (Johnson et al., 1974), and Mercury (Solomon, 1976) require estimates of these long-lived radio nuclides ( $U^{238}$ ,  $U^{235}$ ,  $Th^{232}$ , and  $K^{40}$ ). Estimates are derived from solar abundances information as well as measured amounts in terrestrial, lunar, and meteoritic samples. The lunar basalts are generally believed to be results of partial melting of the interior caused by radiogenic heating about 4.0 to 3.0 AE ago (e.g., Taylor, 1975).

It is unlikely, however, for small planetary objects such as the asteroids to contain sufficient long-lived heat producing elements to allow post-accretional melting. Since the radiometric ages of most achondrites are close to 4.5 AE, an alternative heat source is required to account for the early melting of achondrite parent bodies. Short-lived radio nuclides such as  $^{26}Al$  ( $\tau_{1/2} = .72 \times 10^6$  yr.) have been suggested as a possible short term heat source in the early solar system (Reeves and Audouze, 1969). Although at first no evidence of  $^{26}Al$  could be established in meteorites, the recent discovery of a  $^{26}Mg$  excess in an Allende chondrule suggests the early existence of now extinct  $^{26}Al$  (Lee et al., 1976).



Radiogenic heat formed in the Earth by U, Th and <sup>40</sup>K separately and together plotted against time. (After Vinogradov) 1961

Figure IA 3.

c. Solar Wind Electrical Heating

Early solar wind heating of small objects assumes that young solar-type stars rapidly lose mass ( $\sim .1$  original mass) and maintain an enhanced magnetic field ( $\sim 10 \times$  present) during a T-Tauri phase of stellar evolution. As a planetoid moves through this early plasma, conditions can be favorable (e.g., no atmosphere) to allow an exchange of charge between the plasma and planet. These electric currents cause ohmic heating of the interior of the planetoid. This concept first proposed by Sonnet et al. (1969) has recently been reevaluated in light of additional information about asteroids and meteorites (Briggs, 1976). Briggs pointed out that the differences in low temperature electrical conductivity of different types of carbonaceous chondrites would cause dramatic differences in solar wind heated asteroids of such material. For objects less than 500 km in diameter, C1 and C2-type bodies would survive a T-Tauri phase of the sun intact, whereas the cores of C3 and C4-type bodies would be either metamorphosed or melted.

d. Other Heat Sources

Core formation. For most of the major terrestrial planets including the earth, a combination of the above heat sources caused (early?) extensive fractionation of the planetary material. The separation of the denser

material into a core releases additional gravitational energy (on the earth estimated to be  $10^{38}$  ergs [Verhoogen et al., 1970, p. 640]). Core formation is a major event in the thermal evolution of a planet (a summary for most terrestrial planets is given in Soloman and Chaiken, 1976).

**Impact melting.** After crustal formation (on the moon at  $\sim 4.2$  AE) the surfaces of planetary objects continue to be bombarded by various objects. The heat generated by a high velocity impact onto a surface is generally sufficient not only to vaporize the projectile, but to crush and melt the host material to some degree. Glasses found in the lunar soils and breccias have clearly had such an origin. No major bodies of impact melt, however, have been identified although a few of the returned lunar samples are hypothesized to be crystallized from an impact melt (e.g., Irving, 1975).

**Tidal friction.** When two bodies interact (without colliding) the gravitational deformation is dissipated as heat. The regular interactions of the earth and moon is estimated to account for 10% of the earth's heat flow (Verhoogen et al., 1970, p. 640). Tidal dissipation in regions of high density contrast has been proposed as a possible heat source for some lunar magmas derived along the margins of basins when the moon was closer to the earth 3.0 AE ago (Wones and Shaw, 1975).

#### 4. Formation of Basaltic Liquids--Partial Melting

If the other terrestrial planets are like the earth and moon, then basaltic (or planetary) volcanism, if it exists, is probably the result of partial melting of mantle material. Experimental data for terrestrial material will be used here to show the type of systematics that can occur in a crystal-liquid environment. A limited amount of experimental data also exists for lunar material and has been reviewed by Kesson and Lindsley (1976) (see Section IC). The formation of mafic melts from the mantle of Mars has also been considered (McGetchin and Smyth, 1977).

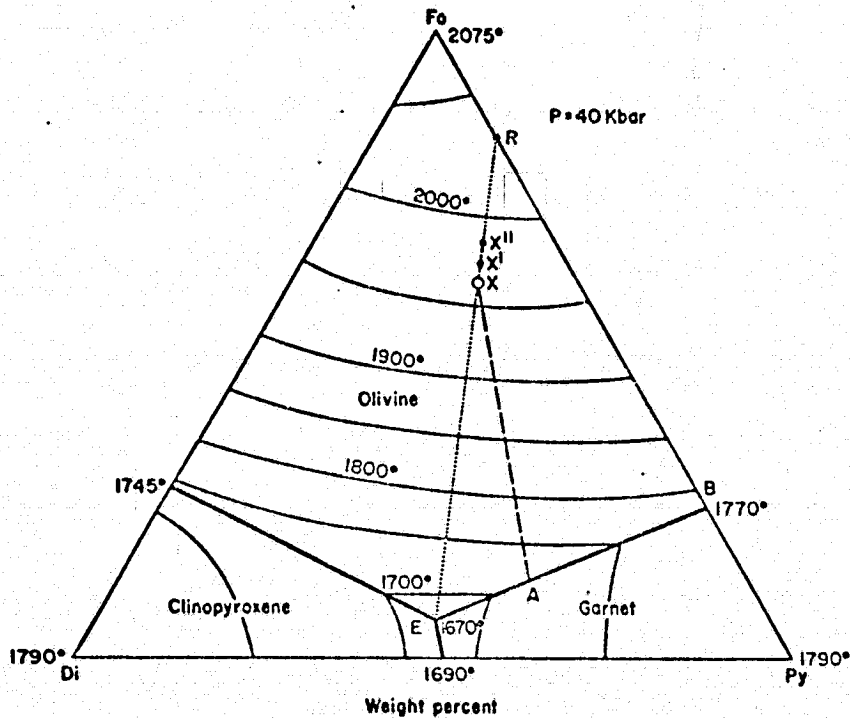
There is much confusion concerning the terms "primitive, primary and parental" as applied to igneous material (e.g., see Carmichael et al., 1974, p. 44-46). To avoid confusion in later sections, these terms will be used throughout the discussion presented here with the following meanings. Primitive refers to early solar system material, generally which has not undergone a thermal reworking. Presumably carbonaceous chondrites represent primitive material; the achondrites are less primitive being perhaps only one step removed from primitive material. Primary melts are those that have not been compositionally altered in any way since their derivation. It may be unlikely that

an ideal primary liquid ever reaches the surface of a planet. Parental melts may or may not be observed. Parental magmas are those which have generated an observed material through some fractionation or assimilation process.

During melting, the composition of a liquid produced is generally different from the composition of the remaining minerals. The behavior of major and minor elements are not the same: the major element composition of a partial melt is controlled largely by the residual mineral assemblages, whereas the trace element abundances are more sensitive to the degree of melting.

a. Major elements

The earth's upper mantle is often considered to be composed of some form of peridotite (olivine with orthopyroxene, clinopyroxene, ± spinel, ± garnet). A simplified system containing olivine (Fo), clinopyroxene (Di), and garnet (Py) at 40 k bars pressure (Figure IA-4) was discussed by Yoder (1976, chapter 6) to illustrate the principles that apply to partial melting. When the temperature of a mineral assemblage with composition X (60% Fo) reaches 1670°C, melting begins at the eutectic composition E. As more heat is added to the system, the degree of partial melting increases but the composition of the melt remains



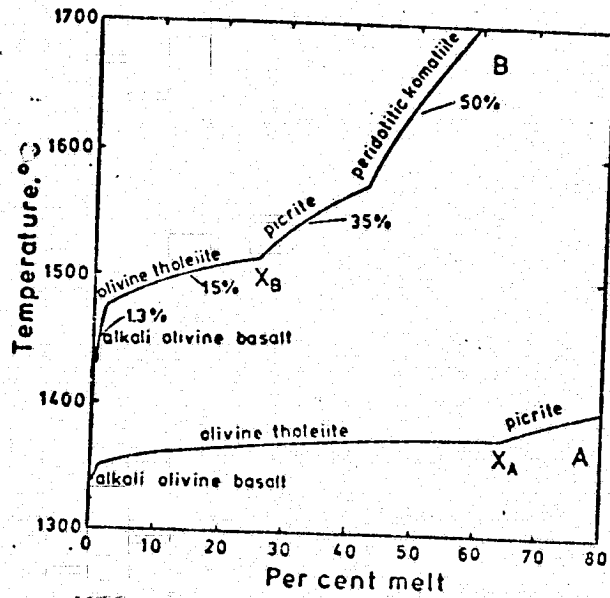
The forsterite (Fo)-diopside (Di)-pyrope (Py) system at  $P = 40$  kbar (after Davis and Schairer, 1965, p. 124, Figure 35).  $E$  is the piercing-point composition. The temperature of the beginning of melting is assumed to be sufficiently close to  $1670^{\circ}\text{C}$  that the behavior is eutecticlike.  $X$  is a bulk composition considered analogous to a peridotite, and  $X'$  and  $X''$  are successive residual compositions resulting from the fractional melting of  $X$ .  $B$  is the composition of liquid produced from the fractional melting of  $R$ . The dotted line is a construction line to illustrate the lever rule, and the dashed line is the locus of compositions of liquids produced by melting of  $X$  at temperatures indicated by the isotherms. (With permission of the Carnegie Institution of Washington.)

Figure IA 4. (from Yoder, 1976)

at E; the composition of the residual is driven toward R until the Di component is exhausted. If the melt remains in contact with the residual as more heat is added, the melt composition trends toward A and then to X (equilibrium or "batch" melting). If the melt is removed as it is produced (fractional melting), no melt will be produced as the temperature is raised from 1670 to 1770°C and a melt of composition B will occur when 1770°C is reached.

For a given source region, a large amount of liquid of a single composition can be produced; the major element composition changes as a function of how and when the liquid is removed from the residuum. Melting curves for real systems are perhaps more complicated than that represented in Figure IA-4, and depend on the starting mineral assemblage. Melting curves for two natural peridotites shown in Figure IA-5 (from Mysen and Halloway, 1977) illustrate the same principle, however: liquids of roughly the same major element composition can be produced from a given source region with distinct changes of composition as different mineral phases participate in the melting event.

The experimental data (e.g., Yoder and Tilley, 1962; Green, 1971; Mysen and Boettcher, 1975 a,b) show that the melt composition is also dependent on the temperature, pressure,  $f_{H_2}$ , and  $H_2O$  and  $CO_2$  content of the source region.



Melting curves of two natural peridotites at 20 kbar under volatile-absent conditions. A, sample 66SAL-1; B, sample 1611. Percentages on curve B denote conditions selected for the present experiments. Rock names are based on normative compositions of partial melts.  $X_A$  and  $X_B$ , see text.

Figure IA 5. (from Mysen and Hallaway, 1977)

In short, "...it appears that we can produce almost any liquid composition from andesite to olivine nephelinite by partial melting of mantle peridotite provided we select the appropriate starting material composition, temperature, pressure,  $f_{H_2}$ ,  $f_{H_2O}$ , and  $f_{CO_2}$ ." (Mysen and Boettcher, 1975b, p. 588).

b. Minor and trace elements

Only the rare earth elements (REE) and Rb and Sr concentrations in basaltic liquids will be discussed here as examples. Gast (1968) pointed out that although major element composition varies only a few percent for various basalts, the abundance of trace elements can vary as much as two orders of magnitudes. His models of REE partitioning between liquid and solids indicate that REE chemistry and major element chemistry of liquids produced by partial melting are in many cases effectively "decoupled". Some trace elements are strongly concentrated in the first melt. For example, Rb, and to a lesser extent K, are both greatly enriched in the liquid for small degrees of partial melt (Figure IA-6, after Gast, 1968). Actual partition coefficients for trace elements are difficult to predict; they are dependent on the minerals present in the source region, the size and charge of the ion, the degree of partial melting, and to some extent the temperature and pres-

Fractionation of Rb and K during ideal fractional melting of the zolite (diopside 10, enstatite 20, olivine 65, spinel 5%). Ratios for 3% equilibrium partial melting are given by circled points X (Rb), Y (K). (After Gast, 1968, p. 1073.)

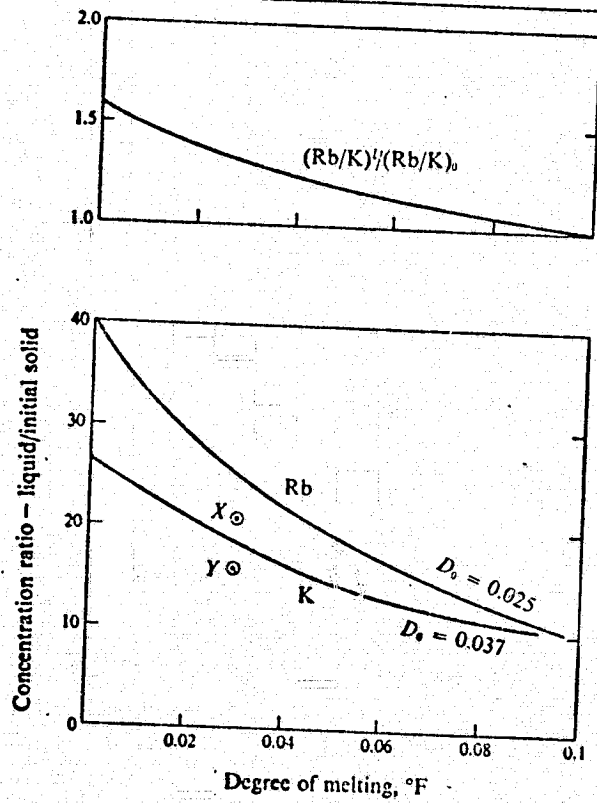
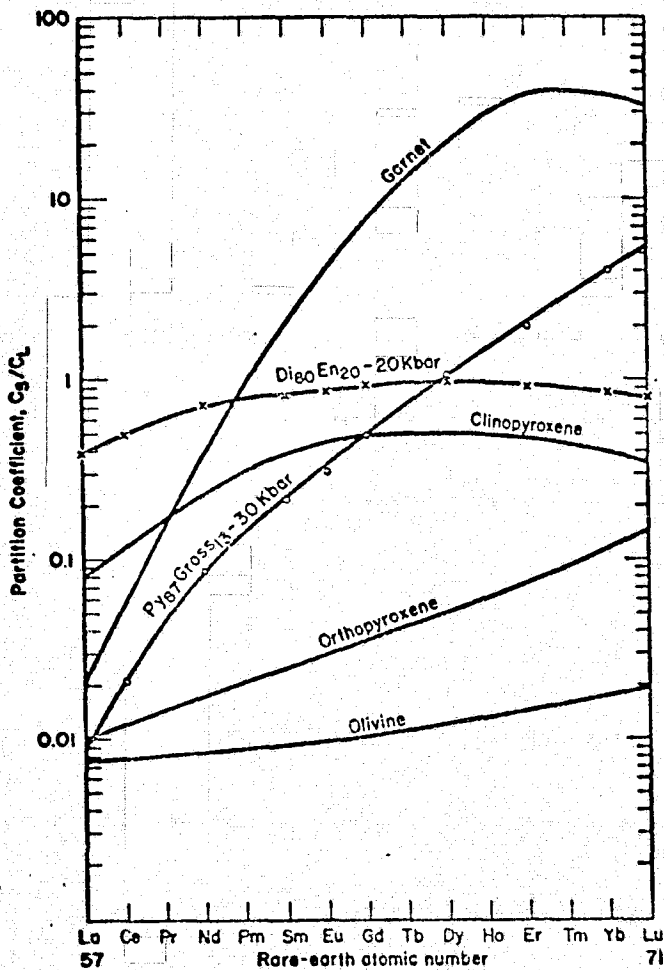


Figure IA 6. (after Gast, 1968)

sure of the melting conditions. Shown in Figure IA-7 (compiled by Yoder, 1976) are representative partition coefficients for the rare earth elements between common crystals and melt liquid. Olivine in the source region would strongly affect the degree of enrichment of REE in a liquid, whereas garnet strongly affects the relative REE pattern. The REE patterns of various partial melts from peridotite B in Figure IA-5 were examined by Mysen and Halloway (1977) and are shown in Figure IA-8. The degree of enrichment and the REE patterns are very dependent on the degree of partial melting.

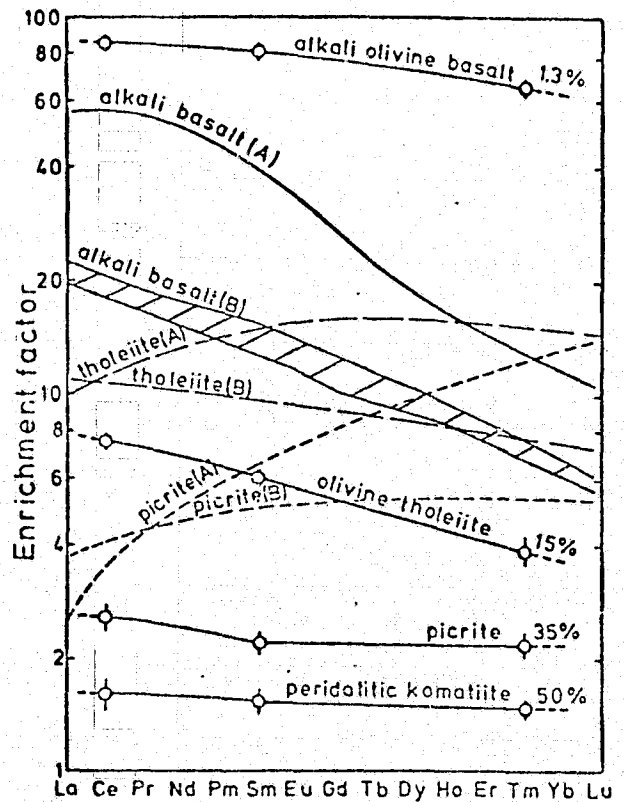
c. Isotopes

Although there is no evidence for isotopic fractionation during partial melting, special mention is due radiogenic isotope systematics such as Rb-Sr (see Faure and Powell, 1972).  $^{87}\text{Rb}$  is radioactive with a half life of about  $4.7 \times 10^{10}$  years and decays to the stable isotope  $^{87}\text{Sr}$ . During partial melting both Rb and Sr are concentrated in the liquid (Rb generally more than Sr). During crystallization Rb is incorporated preferentially into K-bearing minerals and Sr into Ca-bearing minerals. The ratio of the two stable isotopes of strontium,  $^{87}\text{Sr}/^{86}\text{Sr}$ , regularly increases in time from some initial value (I) at a rate determined by the amount of  $^{87}\text{Rb}$  present. The Rb-Sr isotope systematics can thus be used to date the crystallization



Representative partition coefficients for the rare-earth elements between natural crystals (olivine, orthopyroxene, clinopyroxene, and garnet) and deduced liquid compiled by Helmke and Haskin (1973, p. 1520, Figure 3), between synthetic clinopyroxene (Di<sub>80</sub>En<sub>20</sub>) and glass at 20 kbar (Masuda and Kushiro, 1970, p. 44, Table 2), and between synthetic garnet (Py<sub>67</sub>Gross<sub>33</sub>) and glass at 30 kbar (Shimizu and Kushiro, 1975, p. 413, Table 1). C<sub>g</sub> = concentration of rare-earth element in the residual crystal; C<sub>c</sub> = concentration of rare-earth element in coexisting liquid.

Figure IA 7. (Yoder, 1976)



Experimentally determined REE fractionation patterns of experimentally produced partial melts. Alkali basalt (A), Hawaiian alkali basalt. Alkali basalt (B), range of REE contents of alkali basalts from the Lesser Antilles. Tholeiite (A), oceanic tholeiite from the Nizca plate. Tholeiite (B), abyssal tholeiite, Baffin Bay. Picrite (A), picrite basalt, Azores. Picrite (B), picrite, Baffin Bay.

Four patterns for partial melts of sample B (in Fig. IA-5) are indicated. Figure IA 8. (Mysen and Halloway)

age for a given melt (Papanastassiou and Wasserburg, 1969). Furthermore, the initial strontium isotope ratio  $I$  describes the character of the source region at the time of separation of the melt. One of the most primitive values of  $I$  (.69898) is that measured for the basaltic achondrites (age  $\sim 4.5$  AE) (Papanastassiou and Wasserburg, 1971). Initial strontium ratios of basalts are generally compared to this value when the history of the source region is investigated.

## 5. Crystallization of Basaltic Liquids

If a basaltic liquid is allowed to cool in a closed system under equilibrium conditions, the crystallization sequence and resulting mineralogy can be well defined as a function of chemistry and pressure from observational, theoretical, and experimental data (e.g., Bowen, 1928; Yoder and Tilley, 1962; Green and Ringwood, 1967). Such an equilibrium situation rarely occurs in nature, however, and most igneous materials undergo some form of geochemical alteration where either (1) the final bulk chemistry is not the same as for the original magma and/or (2) the final mineralogy is different from the predicted for equilibrium situations.

### a. Geochemical Changes

#### i. DIFFERENTIATION/CRYSTAL FRACTIONATION.

As a basaltic body of magma cools, the first crystals to form are the high temperature minerals, such as Mg-rich olivine (fosterite,  $Mg_2SiO_4$ ) or calcium-rich plagioclase. The liquidus minerals are, of course, dependent on the composition and pressure of the melt and can be determined experimentally for a given melt composition. Once crystallization has commenced, the bulk composition of the crystals is different from the composition of the liquid (except for unique liquids of eutectic composition). Any process

that separates the crystals from the melt before the cooling sequence is complete is termed "fractional crystallization" (e.g., Carmichael et al., 1974, p. 62-65). If early-formed crystals are more dense than the residual liquid, gravitational settling will occur, thus effectively removing these minerals from the melt and changing the bulk composition of the system. For basaltic liquids, the removal of olivine or any other early phase to a cumulate layer below is a major process of chemical differentiation.

ii. ASSIMILATION. Either during transit from the source or while being emplaced, a magma may react with the surrounding host rock. The process by which components of the host rock are incorporated into the melt is called "assimilation" and produces a change in magma composition difficult to detect and evaluate. Low-temperature minerals may be melted and become part of the magma, thus altering the bulk chemistry of the magma. On the other hand, ionic exchange may occur between crystalline host phases and a saturated magma and may only affect the trace element concentrations.

b. Petrological results of cooling conditions

From the time a magma leaves the source region, it begins to cool. The texture and to some degree the mineralogy of the solidified rock is a function of the cooling

conditions.

i. CRYSTAL GRAIN SIZE. In general, only a few crystal nuclei are formed in slowly cooled liquids but the rate of crystal growth is rapid. This results in relatively few large crystals. For rapidly cooled liquids, there are many nuclei with slow growth rates resulting in abundant very small crystals. During the cooling sequence, conditions are not often constant. For example, slow cooling rates could exist during a long rise to the surface or while remaining in a magma reservoir allowing some early crystals to nucleate and grow. When the magma is extruded onto the surface, however, it is cooled more rapidly and the resulting texture is "porphyritic" with the large crystals, or phenocrysts, enclosed by a fine-grained groundmass. Similar textures can also be produced by entirely different conditions. A loss of volatiles, for example, can have a considerable effect in raising the crystallization temperatures of silicate phases and may produce a texture comparable to rapid cooling.

ii. VESICLES. If a magma is generated under pressure (at depth) and contains  $H_2O$ ,  $CO_2$ , or any gas phase, spherical or tabular cavities are formed as the gas is released at lower pressure when the magma is extruded onto

the surface. If cooling is rapid, the gas is trapped during solidification of the rock. These frozen cavities are called vesicles and are characteristic of many volcanic surfaces.

iii. MINERALOGY. Since cooling rate essentially controls whether or not equilibrium conditions are maintained, it has a major effect on the resulting mineralogy. The extreme example is a quenched liquid that is primarily glass. Intermediate are the rapidly-cooled basalts that contain substantial glass in the groundmass along with the fine-grained crystals. Non-equilibrium cooling is commonly indicated by compositional zoning of crystals in which the core is a high temperature phase and the rim is a lower temperature phase (e.g., a decrease in calcium from core to rim of a plagioclase crystal). Composition trends on crystal zoning can often be complex since the composition of the liquid can change locally in a non-uniform manner as other minerals begin to crystallize (e.g., see Hollister et al., 1971).

## 6. Surface alteration of basalts

Few basaltic surfaces are composed of unaltered material, a fact which is of interest to geochemists who study either samples or remote sensing data. As will be discussed in Sections II through V, the geochemical information that can be derived using remote sensing techniques is restricted to planetary surface material. It is thus imperative that the possible and probable effects of the environment to alter the surface mineralogy and/or geochemistry be reasonably understood. A clear description of surface alteration effects is currently perhaps the most difficult step in interpreting remotely sensed data. A rather detailed discussion of lunar soil properties is provided in Section IID. Terrestrial surfaces are perhaps an order of magnitude more complex than lunar. (Unfortunately, inadequate attention has been given to the systematics of surface alteration to be immediately useful in remote sensing exploration of terrestrial resources.) Outlined below are some of the types of environmental alteration that can and do occur on planetary surfaces. The distinction between mechanical and chemical alteration is for discussion purposes only; in reality, the two are intimately related.

### a. Mechanical alteration

A variety of processes break an original solid surface into blocks or finer particles. During the first billion years of planetary evolution, it is unlikely that any of the terrestrial objects escaped high velocity impacts by other (smaller) solid objects. The observable crust

of the moon, Mercury, Mars, and possibly Venus and the Earth have been permanently scarred by these early impacts. Crustal material was also somewhat redistributed by these events. Meteorite bombardment continues to the present day in diminishing amounts. For atmosphereless bodies, the surface has been crushed and broken to some depth which varies according to the scale of the definition. For example, the lunar "mega-regolith" (a result of the last major crustal impacts about 4 AE) probably extends to a depth of about 25 km. The mare regolith (a result of cumulative random bombardment since  $\sim$ 3.5 AE) is typically  $\approx$ 5 m. A mature lunar soil, on the other hand, has a mean particle size of about 100 $\mu$ m and is dominantly the result of cumulative micrometeoroid bombardment during the last few 100 million years.

For planets with atmospheres, a variety of erosional and depositional processes occur depending on the type of weather inherent on the planet. Conditions on earth are perhaps the most severe in that water (rain, oceans, ice) erosion has redistributed material for much of the planet's history. It is suggested from photogeologic evidence that episodic fluid erosion has also been significant on Mars. Aeolian (wind carried particles) erosion and deposition are familiar in arid regions on earth and are perhaps the dominant form of current mechanical surface

alteration on Mars.

b. Chemical alteration

Some form of geochemical alteration is associated with each of the mechanical alteration processes mentioned above. During an impact event, surface materials are mixed both vertically and laterally. A portion of the surface is melted and often mixed with the crushed but still crystalline host rock. If these impact derived materials accumulate over long periods of time, the total mineralogy of the surface can be drastically changed (see Section IID1). Contamination of the surface by the impacting object is very small, but often detectable (e.g., Anders et al., 1973).

Atmosphereless bodies have been exposed to a history of solar wind particle bombardment. An extensive literature exists concerning solar and galactic particle composition and track studies for the lunar soils and the regolith history that can be derived from such studies (e.g., Price et al., 1975). Although it has not been conclusively demonstrated, there is evidence that solar wind implanted H and C contribute significantly to the reduction of iron (to  $Fe^0$ ) during melting by micrometeorite impact of lunar soils (e.g., Housley et al., 1973).

For the earth and Mars, the dominant chemical

alteration is oxidation of surface material. A number of oxidation processes are possible but do not provide unique products; i.e., the same result can be achieved by a variety of means. Hydrothermal alteration of terrestrial basalts can occur prior to or after eruption. Oxidation effects on Mars, which may mimic terrestrial processes, are more likely due to a process by which surface material is stimulated by high energy UV radiation and oxidized by interaction with atmospheric  $O_2$  (e.g., Huguenin, 1973a, b, 1974). Huguenin (1974) estimated a 30m thick layer of clay minerals could be developed in  $10^9$  years in the Martian environment. Meteorological and biological oxidation effects are quickly (within 100 years) noticeable on earth for fresh volcanic material, except in extreme high or arid regions.

c. Current surface environment of the terrestrial planets

i. MERCURY. Atmosphereless. Meteorite bombardment extensive and continuing to present. Development of regolith and soil likely. Possibly strong interaction with solar wind although existing magnetic field should deflect a significant portion of particles. Strong thermal surface environment (-180 to +350°C).

ii. VENUS. Dense opaque atmosphere ( $\sim 10^2$  bars). Currently unaffected by small meteorites. Possible wind erosion. Possible corrosive atmospheric environment. HOT ( $480^\circ\text{C}$ ).

iii. EARTH. Atmosphere (1 bar) opaque to UV radiation. Surface unaffected by micrometeorites. Extensive fluid and wind erosion and deposition. Multitude of oxidation processes.  $3/4$  surface covered with fluid water. Extensive biological environment.

iv. MOON. Atmosphereless. Meteorite bombardment extensive and continuing to present. Well-developed regolith with accumulation of glass-rich soil. Surface material often saturated with solar wind particles (H, C, N).

v. MARS. Thin  $\approx$  transparent atmosphere ( $\sim 0.1$  bar). Surface unaffected by micrometeorites. Wind erosion and deposition significant. (Possible fluid erosion in past.) Oxidized surface (photostimulated). Developed soil.

vi. ASTEROIDS. Atmosphereless. Periodic impact events. Accumulation of mature soil improbable due to weak gravity. Relatively 'fresh' dusty broken surface probable.

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## I. BASALT TYPES: AN OVERVIEW

### B. Terrestrial basalt types

There is no universally accepted classification of terrestrial rock types. One of the problems of classification is the fact that there is a gradation from one rock type into another and boundaries are somewhat indistinct. For igneous rocks there has developed a nonstructured list of terms that are widely used to describe the interrelated mineralogy and chemistry of common rock types (see Carmichael et al., 1974, Chapter 2).

Basalts are volcanic igneous rocks; they have cooled in a low pressure environment (on the surface). Their texture is thus relatively fine grained, generally  $\lesssim 1$  mm. The mineralogy of basalts is dominated by plagioclase and pyroxene and usually includes olivine or quartz and a minor amount of opaques. The plagioclase is calcium rich ( $An_{>50}$ ); there can be one or two types of pyroxene, augite  $\pm$  Ca-poor pyroxene. Phenocrysts of olivine, pyroxene or feldspar are often present.

The chemistry of basaltic rocks is basic: 45-53%  $SiO_2$ . Iron and magnesium content is generally high ( $MgO + FeO + Fe_2O_3 \geq .15\%$ ). Calcium and aluminum are major elements with  $CaO/Al_2O_3$  generally greater than .60. Potassium, sodium and titanium are present in various proportions; these

three elements are often used to distinguish different basalt types.

Because the earth is tectonically active and many (if not all) of the surface materials are recycled a number of times, it has been difficult to develop a classification scheme for igneous rocks that relates directly to their origin. All igneous rocks are solidified melts that were derived from some pre-existing material. The chemistry of basalts suggests they are a class of igneous rocks which are probably related directly to melts from the earth's mantle. Although other types of igneous rocks are related to or associated with basalts, they will only be discussed briefly where appropriate and not in detail here.

#### 1. Geochemical Variations

There are three chemical and mineralogical descriptions of a rock that represent its geochemistry and allow basalts to be classified. (1) Bulk chemical analyses are derived from whole rock analyses and are expressed in weight percent of the major oxides. All rock geochemistry is thus expressed in the same units. (2) Modal mineralogical analyses, on the other hand, is concerned with the mineralogy of a rock; each mineral is identified petrographically, counted, and expressed as a statistical percentage. Fortunately, the number of minerals that comprise the bulk

of igneous rocks is relatively small. (3) A third representation of geochemistry is the normative composition, a recalculation of bulk chemistry into about 10 "ideal mineral" components rather than the chemical oxide analyses. Since a wide range of mineralogy can result as a magma cools (see Section A-5) most comparative discussions of basalt geochemistry are concerned with chemical oxide compositions or norms (the "mineralogical" equivalent). For ease of comparison, the chemistry rather than mineralogy of basaltic materials will be emphasized in this section. Detailed petrological studies of the mineralogy are used to understand the history (fractionation, cooling conditions, mode of emplacement, etc.) of a particular sample.

a. Relation of basalts to other igneous rocks

Chemical analyses and norms for representative igneous rocks (from Verhoogen et al., 1970) are listed in Table IB1. Approximate mineralogical compositions for common igneous rock types is illustrated in Figure IB-1 (from Mason, 1966). Frequency distribution for  $\text{SiO}_2$  in cenozoic volcanic rocks compiled by Chayes (1975) is shown in Figure IB-2. The variation of  $\text{SiO}_2$  is clearly not regular: a maximum occurs near 49% including average basalts but the distribution is strongly skewed towards higher silica content. Figure IB-3 is a variation diagram for the

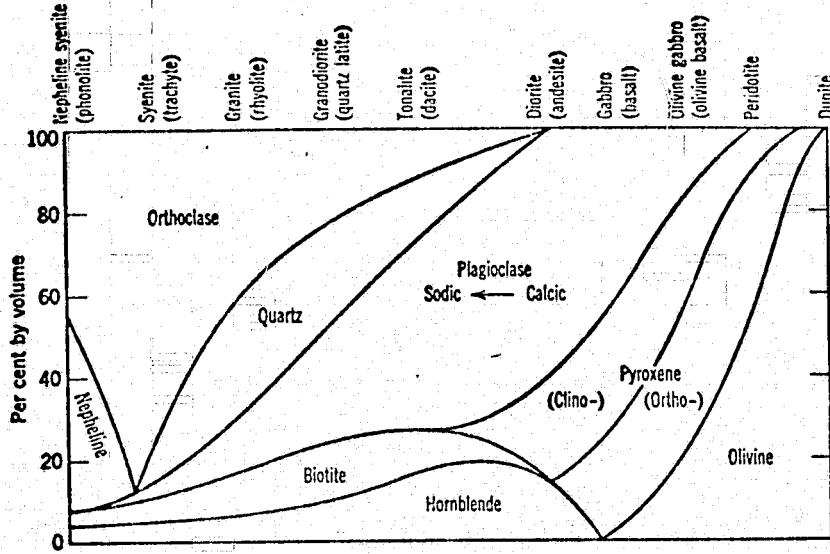
Chemical analyses and norms of some representative igneous rocks (expressed as weight percentages)

No.:	1	2	3	4	5	6
SiO <sub>2</sub>	42.86	48.04	50.02	60.31	66.57	72.60
TiO <sub>2</sub>	2.94	1.83	2.23	1.02	0.60	0.46
Al <sub>2</sub> O <sub>3</sub>	11.46	12.04	15.05	17.53	15.14	13.12
Fe <sub>2</sub> O <sub>3</sub>	3.34	2.35	3.77	3.30	1.15	1.32
FeO	9.03	8.80	7.37	3.85	1.90	1.62
MnO	0.13	0.17	0.17	0.16	0.06	0.04
MgO	13.61	14.41	7.01	2.59	0.56	0.60
CaO	11.24	8.76	10.17	5.97	1.50	2.20
Na <sub>2</sub> O	3.02	1.60	2.05	3.20	4.18	3.63
K <sub>2</sub> O	0.93	0.30	0.33	1.20	5.02	3.71
H <sub>2</sub> O	0.56	1.63	1.65	0.90	3.01	0.24
P <sub>2</sub> O <sub>5</sub>	0.52	0.12	0.27	0.14	0.19	0.04
Others	CO <sub>2</sub> 0.30				BaO 0.25	BaO 0.06
Total	99.94	100.05	100.09	100.17	100.13	99.84
Q			6.30	19.50	16.9	32.34
Or	5.56	1.67	1.67	7.12	29.5	21.68
Ab	5.76	13.62	17.29	27.04	35.6	30.39
An	15.29	24.74	31.14	28.72	7.0	8.62
Ne	10.51					
Di	29.43	14.20	13.84	0.86		1.83
Hy		27.47	17.72	9.01	3.0	1.79
Ol	20.79	9.51				
Mt	4.87	3.25	5.57	4.78	1.6	1.86
B	5.62	3.50	4.26	1.93	1.2	0.91
Ap	1.34	0.34	0.67	0.34	0.3	
Total	99.17	98.30	98.46	99.30	95.1	99.42

## Explanation of column headings:

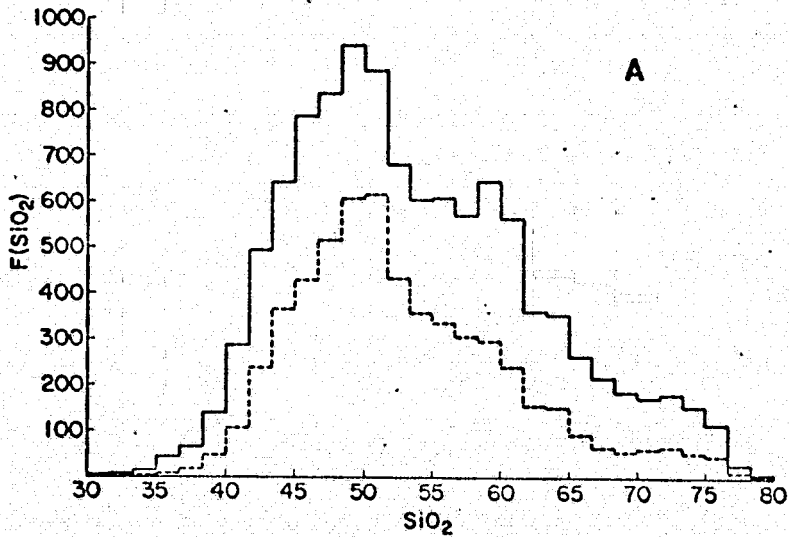
1. Highly undersaturated lava of nephelinite family, Honolulu Series, Oahu, Hawaii (H. Winchell, *Geol. Soc. Am. Bull.* vol. 58, p. 30, no. 13, 1947).
2. Somewhat undersaturated olivine basalt, Haleakala volcano, Maui, Hawaii (G. A. MacDonald and T. Katsura, *J. Petrol.* vol. 5, p. 122, no. C-122, 1964).
3. Oversaturated basalt (tholeiite) Waijanae volcano, Oahu, Hawaii (MacDonald and Katsura, *ibid.*, no. C-27, 1964).
4. Pyroxene andesite, northeastern Japan (Y. Kawano, K. Yagi, and K. Aoki, *Sci. Rept. Tohoku University*, ser. 3, vol. no. 1, p. 32, no. 70, 1961).
5. Trachyte, Main Range, Queensland, Australia (N. C. Stevens, *Proc. Royal Soc. Queensland*, vol. 87, no. 4, p. 46, no. 1965).
6. Hornblende-biotite granodiorite (granite) Mt. Hale, Southern California, Batholith (E. S. Larsen, *Geol. Soc. Amer. Mem.* 23, p. 91, 1945).

Table IB 1. (CTV)



Approximate mineralogical composition of the commoner types of igneous rocks (effusive types in brackets).

Figure IB 1.  
(Mason, 1966)



Distribution of  $\text{SiO}_2$  in 10,870 analyses of Cenozoic volcanic rocks. (A) Solid line, all data; dashed line, 5697 analyses with  $\text{H}_2\text{O} < 2$  and  $\text{Fe}^{3+}/\text{Fe}^{2+} < 1$ .

Figure IB 2.  
(Chayes, 1975)

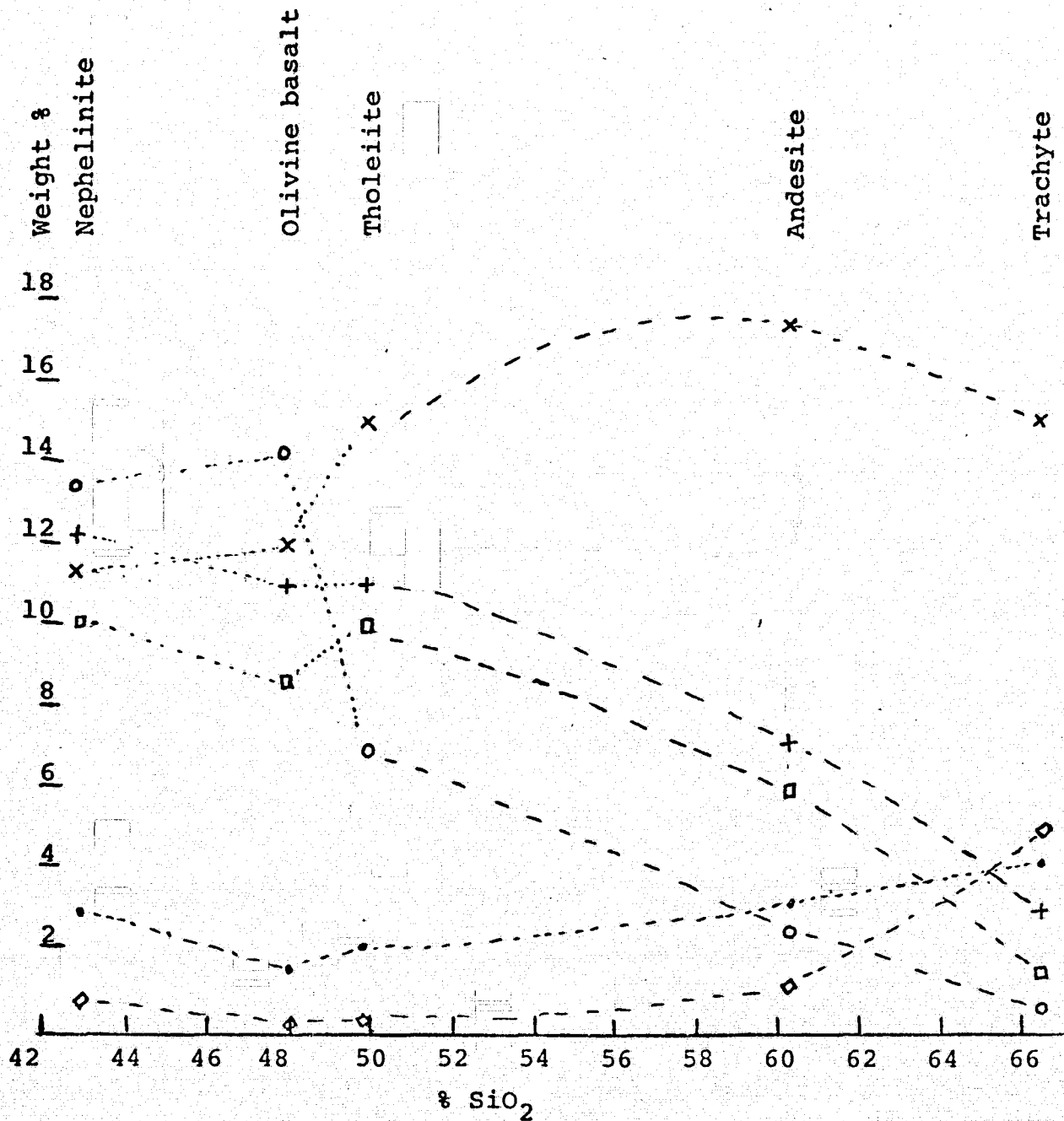


Figure IB-3 Variation diagram for representative igneous rocks

rocks of Table IB-1 illustrating the general geochemical relationship between the different rock types. As silica increases, iron, magnesium, and calcium decrease. Although  $\text{SiO}_2$  content generally identifies basalts from other igneous rocks, it is not a parameter that readily distinguishes among the variety of basalt compositions. (Most terrestrial basalts, nevertheless, can be grouped into only a few magma types (see below), although the distinguishing boundaries become very diffuse upon closer inspection.)

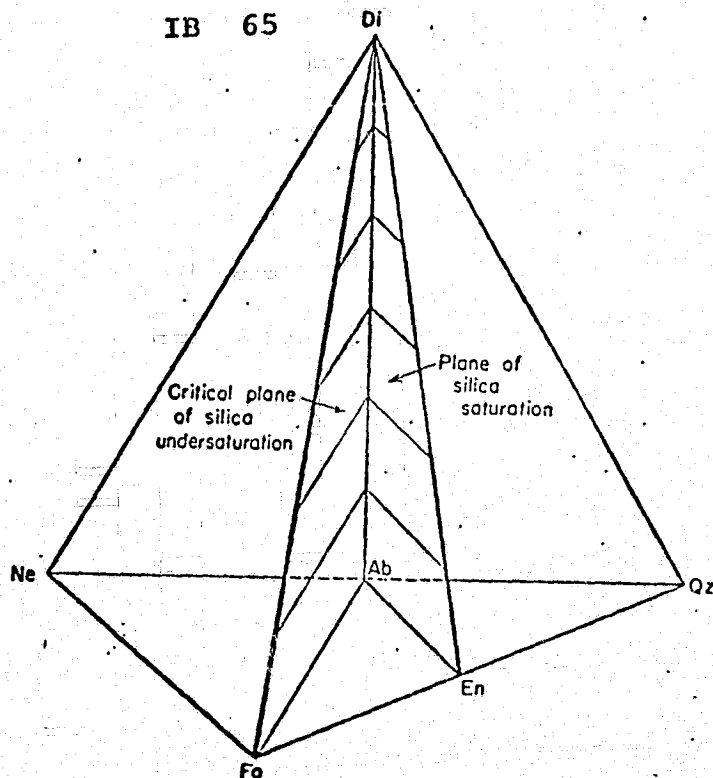
The derivation of more silicic igneous rocks from basaltic magmas through fractional crystallization (e.g., removal of Mg, Fe-rich olivine, Fe-rich opaques, or Ca-rich plagioclase) is a fairly well accepted concept, although perhaps not universally applicable. Such a trend is generally compatible with the variation diagram of Figure IB-3. The sequence of mineral-liquid reactions that occur during crystallization was described by Bowen (1928) and forms the basis for many suggested series of differentiated rocks. Such a sequence suggests the relation of basalts to other more acidic igneous rocks is one of primary to secondary, or more extensively processed, material.

b. Magma types

In the early part of the century, Bowen (1928)

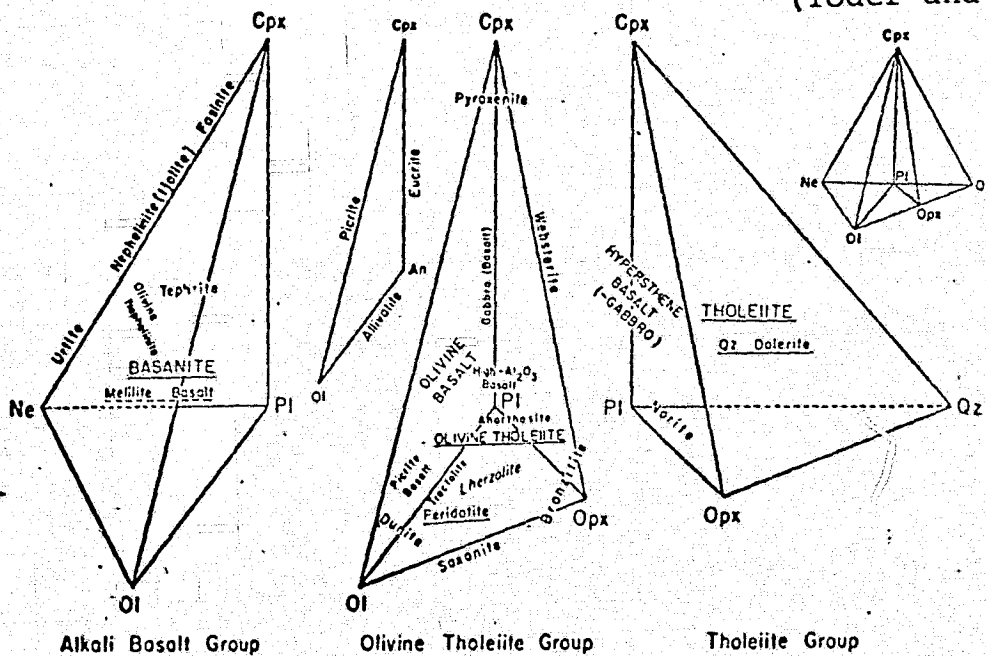
recognized and convincingly established the parental role of basaltic magma. The Scottish island of Mull became a case area where two main basaltic magma types were recognized. Kennedy (1933) then described what appeared to be two world-wide basaltic magma types: (1) olivine-basalt containing (in addition to plagioclase) olivine, augite and alkali feldspar, nepheline or zeolites; and (2) tholeiite basalt containing pigeonite, augite and siliceous material. Olivine basalt would lead to trachytes with differentiation and tholeiites would lead to rhyolites.

Experimental studies of natural and synthetic rock systems led Yoder and Tilley (1962) to describe a generalized simple basalt system that can be expressed in terms of normative components (Figures IB-4a,b from Yoder and Tilley). They further defined the chemistry of tholeiites (normative hypersthene, with modal augite or subcalcic augite, plagioclase, iron oxides,  $\pm$  olivine,  $\pm$  quartz) and alkali basalt (normative nepheline and olivine and modal high-Ca augite, plagioclase, olivine, and generally  $\text{Na}_2\text{O} + \text{K}_2\text{O} > 3\%$ ). An important criterion in their classification is the degree of silica saturation, indicated by whether or not quartz is present in the norms. Widely used terminology for three commonly observed basalt compositions has



Schematic representation of the system Di-Fo-Ne-Qz showing the plane of silica saturation, Di-En-Ab, and the critical plane of silica undersaturation, Di- $\bar{I}$ -Ab. This iron-free simple basalt system includes the principal components of the major phases of basalt.

Figure IB 4.  
(Yoder and Tilley 1962)



Exploded view of generalized simple basalt system illustrated in upper right inset. Entered are names of rocks whose major normative phases are contained in the tetrahedra. Names underlined are within tetrahedra. Names in faces are written parallel to base. Lherzolite is in OI-Cpx-Opx face. An additional inset, OI-Cpx-An, gives alternative nomenclature when plagioclase is rich in An.

emerged: (1) tholeiite--oversaturated with normative quartz and hypersthene; (2) olivine tholeiite--somewhat undersaturated, with normative hypersthene and olivine; and (3) alkali basalt--undersaturated with normative olivine and nepheline. Yoder and Tilley emphasized that there is a continuum between all basalt types even though at low pressures there is a thermal divide which roughly corresponds to the 'critical plane of silica undersaturation' (Fo-Ab-Di in Figure IB-4b). This equilibrium thermal divide, however, changes with pressure and the two basalt types may be related at depths with the alkali basalt type magma having a deeper source.

As pointed out by Carmichael et al. (1974) such simplified terminology is a geochemical convenience and does not necessarily imply the same origin for two rocks with similar gross chemistry. Not only do some basaltic materials seem out of place in such a universal classification, but the unique distinguishing characteristics of each basaltic province in reality defines an infinite number of basalt types.

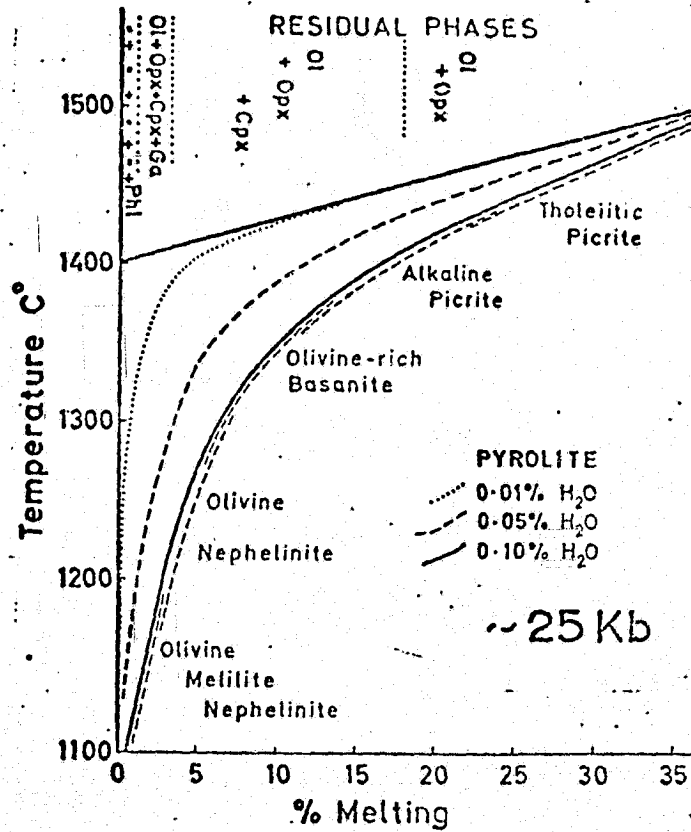
If and how one variety of basaltic material is related to another is the subject of many professional papers, each with a (slightly) different emphasis. Yoder

(1976) has written a book to expand on the concept of how common pressure-temperature-dependent phase relations control the restriction of magma types. MacDonald (1968) summarized a variety of data linking the Hawaiian alkalic rocks with the main olivine tholeiites and concludes that crystal fractionation plays the major role. Engel et al. (1965) come to a similar conclusion. A grand scheme for the derivation of all basalt types was ambitiously described by Green (1971) involving different degrees of melting at different depths with and without water and with later fractionation (Figure 1B-5a,b from Green, 1971). Allowing variations through contamination and hybrid theories (e.g., Eichelberger, 1974; Moorbath and Welke, 1968) it is truly amazing that some real similarity does indeed exist allowing a gross classification of basalt magma types.

## 2. Basalt Types Associated with Specific Environments

Great volumes of magma are being extruded along the ocean rift zones as two lithosphere plates are separating. Such extrusion of volcanic material results from perhaps the most simple terrestrial plumbing system between source (presumably the mantle) and vent. The avenues that all other terrestrial basalt magmas take to reach the surface and the processes that occur along the way are less clearly understood but are certainly dependent on the geo-

IB 68



Partial melting of a "pyrolite" model of the upper mantle.

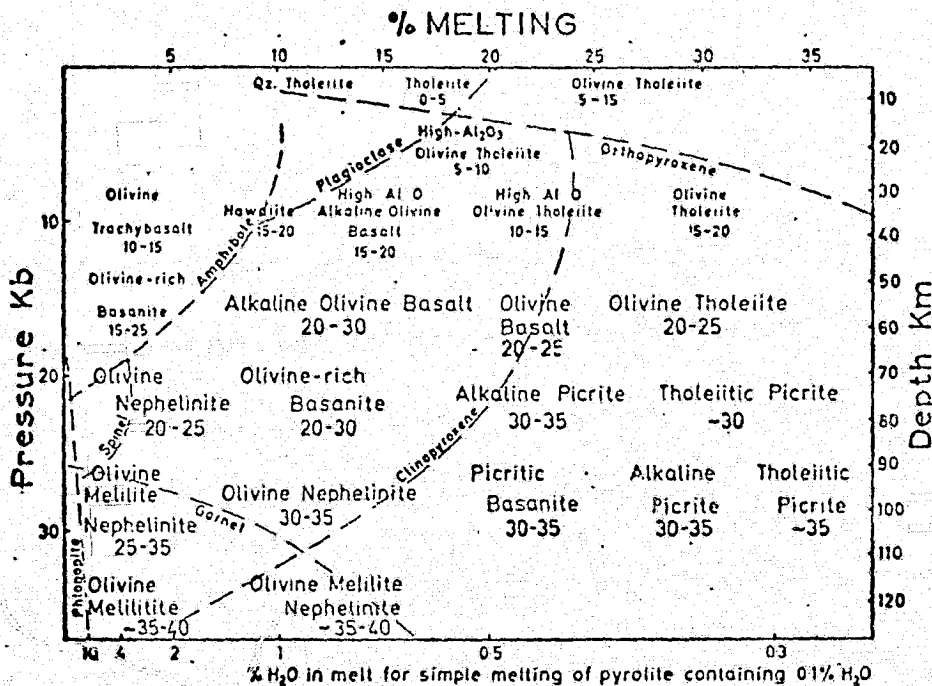


Figure IB 5. (from Green, 1971)

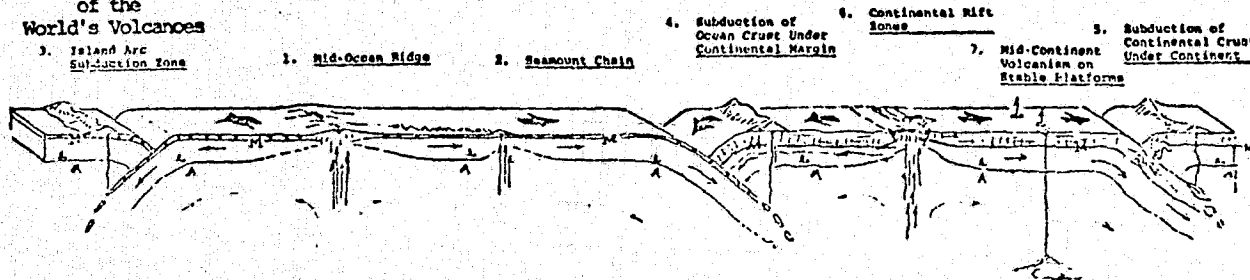
logical setting. The major basalt types and associations that occur in particular terrestrial environments (summarized in Figure IB-6, from McGetchin, 1975) are discussed briefly below. Except as otherwise noted, the tables of chemical composition and regional summaries are from Carmichael et al. (1974), referred to as CTV.

a. Ocean ridge and floor basalts

Geochemical data accumulating over the last few decades indicates that the rock type covering most of the earth's surface (the ocean floor) is a young basalt of distinctive and roughly universal composition: olivine tholeiite (Engel et al., 1965; Kay et al., 1970). This basaltic crust is formed at the spreading centers (ridges).

The geochemistry of representative oceanic tholeiites is presented in Tables IB-2a,b. In general, these basalts are characterized by normative olivine, low  $\text{Fe}_2\text{O}_3/\text{FeO}$ , low potassium, high K/Rb (700-1000), and very low  $\text{Sr}^{87}/\text{Sr}^{86}$  (.7023-.70245, Tatsumoto et al., 1965; Hart, 1971). Oceanic tholeiites show a relative depletion of large ion lithophile elements. The rare earth element pattern is roughly chondritic but with the light rare earths depleted--a pattern perhaps unique to the ocean ridge basalts (Figure IB-7, from Gast, 1972).  $\text{Al}_2\text{O}_3$  is variable

**Tectonic Setting  
of the  
World's Volcanoes**

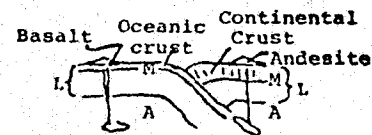


MID-OCEAN RIDGE

CONTINENTAL MARGIN

STABLE PLATFORM

Setting	Dominant Processes	Igneous Processes and Melting Mechanisms	Rock Types Produced	Structure and Style of Volcanism	Location and Example Volcano
1. Mid-Ocean Ridge	Creation of oceanic crust; lithosphere; separation of oceanic lithosphere	Partial melting of the upper mantle upon adiabatic decompression in rising convection cell	Principally primitive oceanic basalts; locally andesitic types	Effusive submarine flood basalts; rare supracrustal volcanic shields with andesitic edifices	Iceland - Hecla Galapagos - Fernandino
2. Seamount Chain	Local melting in mantle; redeposition of melt on surface	Partial melting of upper mantle due to transport over hot spot (plum) or due to decompression accompanying tectonic rifts	Primarily primitive oceanic basalts; minor alkalic basalts	Large shield volcanoes produced from effusive tholeiitic basalt flows capped with alkali basalt, largely pyroclastic	Hawaii - Kilauea Mauna Loa
3. Island Arc	Destruction of oceanic crust by partial melting	Partial melting of descending slab; locally differentiation and contamination occur	Dominantly andesite; subordinate high Al-basalt	Andesite stratovolcanoes; long repose times and episodic violent explosive activity	Japan - Asahi W. Aleutians
4. Continental Margin - Oceanic crust under continental crust	Destruction of oceanic crust by partial melting	Partial melting of descending slab; locally differentiation and contamination occur	Dominantly andesite; subordinate high Al-basalt (less basalt than #3, above)	Andesite stratovolcanoes; long repose times and episodic violent explosive activity	California - Shasta Guatemala - Pacaya Chile - Puyehue Ecuador - Cotopaxi E. Aleutians - St. Augustine
5. Continental rift zones	Deformation and partial melting of continental crust	Partial melting of descending slab; locally differentiation and contamination occur	Dominantly andesite; subordinate high Al-basalt (less basalt than #3, above)	Andesite stratovolcanoes; long repose times and episodic violent explosive activity	Italy - Etna Stronboli
6. Continental rift zones	Separation of crust-lithosphere; and volcanism	Same as #5 but with partial melting of lower crust to produce acidic volcanics in addition	Bimodal suite of volcanic rocks. Basaltic suite like #1; rhyolitic suite produced by partial melting of crust	Effusive flood basalts; explosive rhyolitic ash and caldera complexes	New Mex. - James E. Africa - Erta Ale
7. Stable platform volcanism	Penetration of stable platform by exotic volcanic types	Volatile-rich volcanics from within or below lithosphere; varying degree of partial melting	Generally exotic types - kalsherite, carbonatitic, alkalic basalts, lamprophyres. Deep seated source within the mantle; xenoliths common	Diatremes common; explosive gas-charged eruptions. Breccia dikes and pipes common	Utah - Moses Rock So. Africa - Kimberley Mine



M = Mohorovicic discontinuity (Crust-mantle boundary)  
L = Lithosphere (rigid plate)  
A = Asthenosphere (plastic interior)

Tectonic setting of the world's volcanoes.

Figure IB 6. (from McGetchin, 1975)

Chemical compositions (oxides, wt %) and CIPW norms of rocks from mid-Atlantic ridge

	1	2	3	4	5	6	7	8	9	10
SiO <sub>2</sub>	49.20	49.02	49.27	47.94	49.00	47.50	49.70	48.65	43.15	48.56
TiO <sub>2</sub>	2.03	1.46	1.26	0.75	1.46	1.83	1.49	1.44	2.70	0.24
Al <sub>2</sub> O <sub>3</sub>	16.09	18.04	15.91	17.45	15.50	16.00	14.85	15.99	13.46	18.69
Fe <sub>2</sub> O <sub>3</sub>	2.72	1.58	2.76	1.21			2.16	2.18	4.52	2.27
FeO	7.77	6.22	7.60	8.47	9.77	12.20	8.27	6.19	8.22	4.30
MnO	0.18	0.13	0.13	0.13	—	—	0.18	0.15	0.11	0.11
MgO	6.44	7.85	8.49	10.19	8.00	5.37	8.56	9.66	10.80	9.26
CaO	10.46	11.51	11.26	11.26	10.80	11.40	11.17	11.52	9.80	12.67
Na <sub>2</sub> O	3.01	2.92	2.58	2.37	2.90	2.57	2.69	2.71	3.47	1.88
K <sub>2</sub> O	0.14	0.08	0.19	0.09	0.21	0.49	0.15	0.57	1.63	0.07
P <sub>2</sub> O <sub>5</sub>	0.23	0.12	0.13	0.08	—	—	0.13	0.21	0.75	0.02
H <sub>2</sub> O <sup>+</sup>	0.70	0.64	0.35	0.23			0.61	0.75	1.21	1.72
H <sub>2</sub> O <sup>-</sup>	0.95	0.57	0.51	0.15	1.19	3.28	0.16	0.30	0.15	0.17
Total	99.92	100.14	100.44	100.32	98.83	100.64	100.12	100.32	99.97	99.96
<i>Q</i>	0.3									
<i>or</i>	0.8	0.5	1.1	0.6	1.27	2.96	1.11	3.34	9.63	0.56
<i>ab</i>	25.7	24.4	21.8	20.0	25.03	22.25	23.06	23.06	9.67	15.72
<i>an</i>	29.8	36.3	31.2	36.7	29.23	31.38	27.52	29.75	16.34	42.46
<i>ne</i>									10.67	
<i>di</i>	17.4	16.6	19.2	15.2	16.19	15.42	21.81	21.10	22.23	15.96
<i>hy</i>	16.2	7.7	13.6	4.5	9.14	6.63	13.19	1.26		14.07
<i>ol</i>		9.0	5.9	19.7	10.08	9.48	6.33	14.45	16.76	4.95
<i>mt</i>	4.0	2.3	4.0	1.8	2.22	2.22	3.25	3.25	6.55	3.25
<i>il</i>	3.8	2.7	2.4	1.4	2.83	3.56	2.89	2.74	5.13	0.46
<i>ap</i>	0.5	0.3	0.3	0.2	0.33	0.34	0.34	0.48	1.64	0.05
Total	98.5	99.8	99.5	100.1	96.32	94.24	99.50	99.43	98.62	97.48

*Explanation of column headings*

- Oceanic tholeiite, depth 2910 m; 20°40'S, 13°16'W (Engel and Engel, 1964a, D2-1)
- Oceanic tholeiite (diabase), depth 2388 m; 9°39'N, 40°27'W (Engel and Engel, 1964a, D5-5)
- Oceanic tholeiite, depth 3566 m, rift floor; 28°53'N, 43°20'W (G. D. Nicholls, 1965, table 1, analysis 1)
- High-alumina basalt, same locality as 3 (G. D. Nicholls, 1965, table 2, analysis 2)
- Oceanic tholeiite, depth 4200 m, rift floor; 30°08'N, 43°37'W (Kay et al., 1970, analysis A150-21-1C)
- Basalt, depth 3700 m; 31°49'N, 42°25'W (Kay et al., 1970, analysis GE160)
- Basalt, depth 3700 m; 31°49'N, 42°25'W (Muir and Tilley, 1966, p. 195, analysis 3)
- Basalt, depth 3600 m, rift floor; 45°44'N, 27°44'W (Muir and Tilley, 1964b, table 1, analysis 5)
- Alkali olivine basalt, depth between 2000 and 3000 m; a few kilometers northeast of St. Paul's Rocks; 1°1'N, 29°21'W (Melson et al., 1967)
- Laminated gabbro, depth 4000 to 5000 m, Romanche trench; 0°14'N, 17°7'W (Melson and Thompson, 1970)

Table IB 2a. (CTV)

Chemical compositions (oxides, wt %), CIPW norms,\* and atomic abundances and abundance ratios of trace elements and isotopes of ocean-floor lavas, rises of East Pacific Ocean

	1	2	3	4	5	6		1a	3a	4a
SiO <sub>2</sub>	49.80	49.13	48.30	59.00	49.90	50.10	Ba	25	19.4	54.8
TiO <sub>2</sub>	2.02	1.23	2.19	1.75	1.08	2.18	Ce		16.5	75
Al <sub>2</sub> O <sub>3</sub>	14.88	14.97	14.30	12.60	17.30	13.80	Cs		0.074	0.082
Fe <sub>2</sub> O <sub>3</sub>	1.55	3.28					Co	35		
FeO	10.24	5.72	11.70	12.00	7.60	12.30	Cr	160		
MnO	0.21	0.16					Ni	58	58	10
MgO	6.74	7.68	6.70	1.70	7.08	6.11	Pb	0.49		
CaO	10.72	12.68	10.10	5.60	12.78	10.90	Rb	1.06	5	7.25
Na <sub>2</sub> O	2.91	2.37	2.75	4.25	2.45	2.83	Sr	110 (86)	107	105
K <sub>2</sub> O	0.24	0.16	0.18	0.65	0.18	0.16	Th	0.21		
P <sub>2</sub> O <sub>5</sub>	0.28	0.15					U	0.09		
H <sub>2</sub> O <sup>+</sup>	0.54	1.06	1.29	1.78	0.80		V	400		
H <sub>2</sub> O <sup>-</sup>	0.06	1.25					Zr	150		
Total	100.19	99.84	97.51	99.33	99.17	98.38	K/Rb	1890	310	770
<i>Q</i>		0.79		13.66		0.04	Cs/Rb		0.014	0.011
<i>or</i>	1.1	0.89	1.10	3.92	1.08	0.96	Th/U	2.3		
<i>ab</i>	24.6	20.01	24.08	36.70	20.99	24.25	Sr <sup>87</sup> /Sr <sup>86</sup>	0.7025		
<i>an</i>	26.7	29.75	27.06	13.66	36.12	24.79	Pb <sup>206</sup> /Pb <sup>204</sup>	18.24		
<i>di</i>	22.0	25.73	19.59	11.84	21.90	23.96	Pb <sup>207</sup> /Pb <sup>204</sup>	15.53		
<i>hy</i>	13.7	12.79	17.73	14.27	13.16	19.27	Pb <sup>208</sup> /Pb <sup>204</sup>	38.03		
<i>ol</i>	5.7		3.54		2.12		U <sup>238</sup> /Pb <sup>204</sup>	6.4		
<i>mt</i>	2.3	4.76		2.25	2.22	2.20				
<i>il</i>	3.7	2.36	4.30	3.39	2.08	4.19				
<i>ap</i>	0.6	0.35	0.34	0.33	0.33	0.33				
Total	100.4	97.43	99.99	99.99	99.98	99.99				

\* For analyses 3-6, norms are calculated (Kay et al., 1970) assuming a low state of oxidation of Fe (Fe<sub>2</sub>O<sub>3</sub> = 1.50 percent) and reasonable values of MnO (0.18 percent) and P<sub>2</sub>O<sub>5</sub> (0.15 percent).

*Explanation of column headings*

- 1 Glassy basalt, East Pacific rise, depth 2300 m; 12°52'S, 110°57'W (Engel et al., 1965, table 1, analysis PVD-3)
- 1a Trace-element and isotopic data for analysis 1 (Engel et al., 1965; Tatsumoto et al., 1965; Tatsumoto, 1966)
- 2 Basalt, Mohole drill core, depth 3746 m; off Guadalupe Island (East Pacific rise), 28°59'N, 117°30'W (Engel and Engel, 1961, p. 1799, analysis 1)
- 3, 3a Basalt, East Pacific rise, depth 3120 m; 7°08'N, 103°15'W (Kay et al., 1970, p. 1593, analysis V2023)
- 4, 4a "Andesite" glass, East Pacific rise, depth 3182 m; 5°31'S, 106°46'W (Kay et al., 1970, p. 1593, analysis V21-0)
- 5 Basalt, Gordo rise, depth 2500 m; 41°15'N, 127°28'W (Kay et al., 1970, p. 1592, analysis 13E)
- 6 Basalt, Juan de Fuca rise, depth 2502 m; 44°36'N, 130°19'W (Kay et al., 1970, p. 1591, analysis 2C)

Figure IB 2b. (CTV)

and sometimes high.

In spite of their general primary nature, most dredged basalts nevertheless show small compositional variations that are ascribed to shallow differentiation or fractionation in the ascending magma (Kay et al., 1970; Frey et al., 1974; Mazzullo and Bence, 1976). Removal of plagioclase and olivine at shallow depth from a mantle derived liquid would account for most of the observed major and minor element trends. For example,  $TiO_2$  is concentrated in the more fractionated magmas (Figure IB-8, from Mazzullo and Bence).

Samples obtained from the fracture zones and aseismic ridges associated with the mid-Atlantic rift valley are more fractionated types of basaltic rocks than those from the rift itself (Hekinian and Thompson, 1976). The transform fault rocks are depleted in olivine and have a higher range of  $TiO_2$  content. Variation diagrams for basalts from the mid-ocean ridge, fracture zones, and aseismic ridges are shown in Figure IB-9 (from Hekinian and Thompson). Some recovered rocks are gabbroic and may be part of an intrusive layer (Miyashiro et al., 1970).

The basalt cores obtained from the Atlantic floor away from the rift axis show no obvious systematic compositional differences as a function of distance from

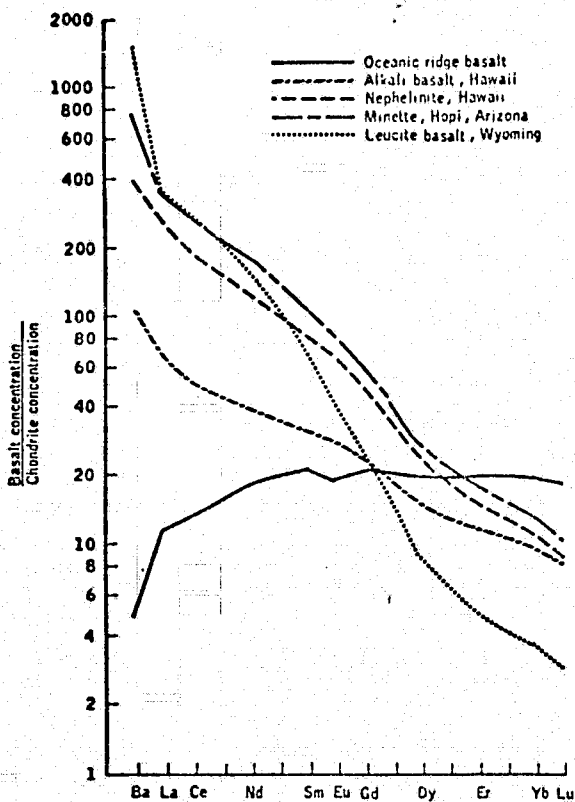


Figure IB 7.  
(Gast, 1972)

Chondrite normalized terrestrial LIL abundance patterns. Data are from Kay (1970), Kay *et al.* (1970), and unpublished data. Uranium data on oceanic basalts were furnished by Tatsumoto (personal communication).

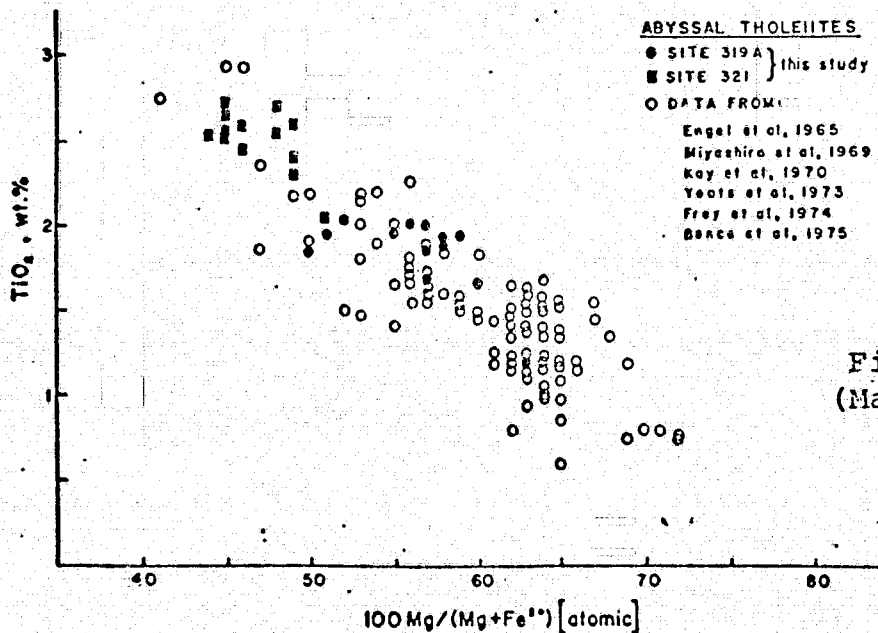
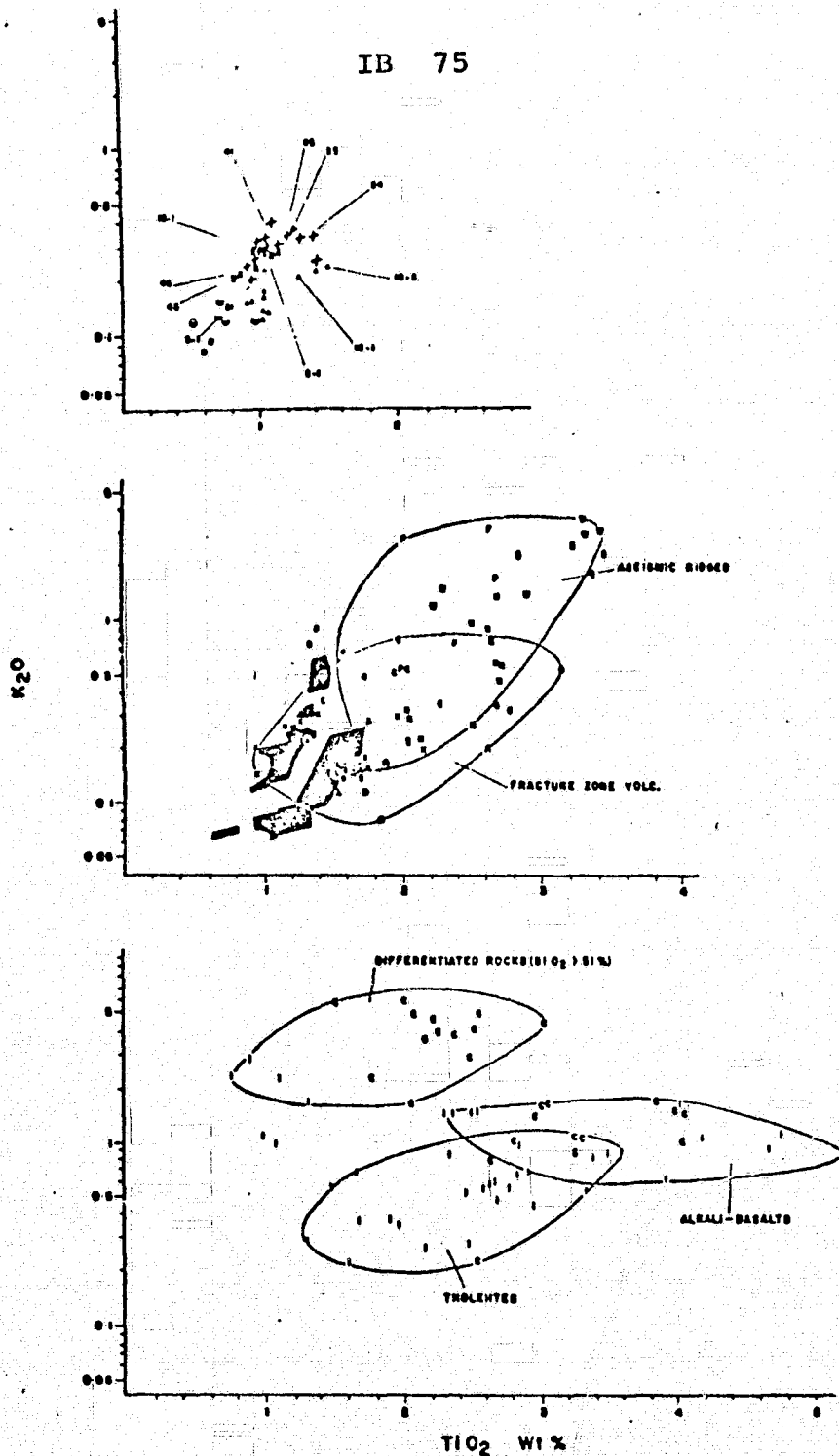


Figure IB 8.  
(Mazzullo and Bence)

TiO<sub>2</sub> versus Mg/(Mg + Fe<sup>2+</sup>) diagram, showing whole rock chemical data from leg 34 basalts and additional data from other abyssal tholeiites. All data are plotted assuming Fe<sup>2+</sup>/(Fe<sup>2+</sup> + Fe<sup>3+</sup>) = 0.1 (atomic).

IB 75



K<sub>2</sub>O-TiO<sub>2</sub> Variation diagram of basaltic and other differentiated rocks from various oceanic provinces. The top diagram represents the various types of basaltic rocks encountered near 36° 50' N.

The middle diagram shows the field of M.A.R. Rift Valley rocks (black area) compared to transform faults and aseismic ridge rocks. S, E, W and N indicate the volcanics from the Cocos, the Iceland-Faeroes, the Walvis and the Ninetyeast Ridge respectively. The bottom diagram shows the volcanics from Iceland (I) and those from the Azores (C). The alkali-basalts data are from Esenwein (1929), Sigvaldason (1969) and Thorarinson et al. (1973). The andesitic rocks are from Carmichael (1964) and from Girod (1972). The tholeiites are from Carmichael (1964) and from Sigvaldason (1969)

Figure IB 9. (Hekinian and Thompson, 1976)

the ridge (Frey et al., 1974). No oceanic rock has been recovered older than Triassic and there is no indication that the source region for North Atlantic ridge basalts has evolved since the late Mesozoic. Because the Atlantic Ocean basement appears to be roughly spatially and temporally homogeneous and composed of large ion lithophile element depleted tholeiites, Frey et al. support the suggestion by Kay et al. that the source rock for this material (the upper mantle) is a (extensive) residue from an earlier melting event.

Although an increasing number of exceptions will be found (e.g., Bryan et al., 1976), it is clear from the data in hand that the major rock type that forms the oceanic crust is tholeiitic basalt depleted in large ion lithophile elements. The source of these basalts (the upper mantle) and the geological processes of formation, fractionation, and extrusion have produced on a global scale immense volumes of magma with roughly the same composition. Our understanding of these young oceanic basalts is closely associated with plate tectonics. It is unclear, however, whether any comparable terrestrial basaltic volcanism occurred prior to the breakup of the continents during the Triassic period.

b. Ocean Islands

The volcanic islands and seamounts that occur in the ocean environment show a much more complicated variation of chemistry than the great volumes of ocean floor basalt. Almost all islands show a variety of igneous rock types, the sequence of which can generally be interpreted as differentiation series. Although a few trends are similar in widely separated islands, it is clear that there are significant differences in either the source composition, the melting conditions, or the sequence of processes (including fractionation) that occur prior to extrusion. For the discussion below, the volcanic island provinces have been grouped according to whether the series of observed rocks seem to follow a sequence with (i) tholeiite--rhyolite, (ii) alkali basalt--trachyte--phonolite, or (iii) both.

i. THOLEIITE SERIES. Islands that occur on or near a current oceanic ridge invariably consist to some degree of tholeiitic basalt although the chemistry of such basalts is not identical to the ocean ridge basalts. Potassium is generally slightly higher for non-ridge tholeiites. The rare earth element patterns of island basalts usually show no depletion in LIL elements.

Iceland, centered on the Atlantic rift, consists predominantly of oversaturated (olivine-free)

tholeiites (Carmichael, 1964). About 10% of the total accumulation of volcanics is rhyolite and andesite. A low pressure fractionation scheme is proposed from a parental olivine tholeiite.

Olivine tholeiites also predominate on Ascension Island (on the mid-Atlantic ridge) as well as on a number of Pacific islands such as Guadalupe, Clarion, and a few of the Samoan islands. Minor amounts of silicic trachyte and andesite occur as associated volcanics for these islands.

ii. ALKALI BASALT SERIES. Most of the non-ridge oceanic islands and seamounts contain basalts enriched in alkalis and undersaturated in silica. These alkali basalts show a rare earth pattern regularly enriched in the light rare earths (Figure IB-7). The associated rocks (differentiates) are highly alkaline trachytes, phonolites, or peralkaline rhyolites. A potassium variation diagram (Figure IB-10 from Engel et al., 1965) shows a fairly sharp distinction between the chemistry for these oceanic island basalts and those from the ocean ridges.

Various rock series can be found on a number of ocean islands with alkali basalt magma generally the hypothesized parent. One extreme, represented by the islands of Trinidad and Fernando de Noronha, far off the

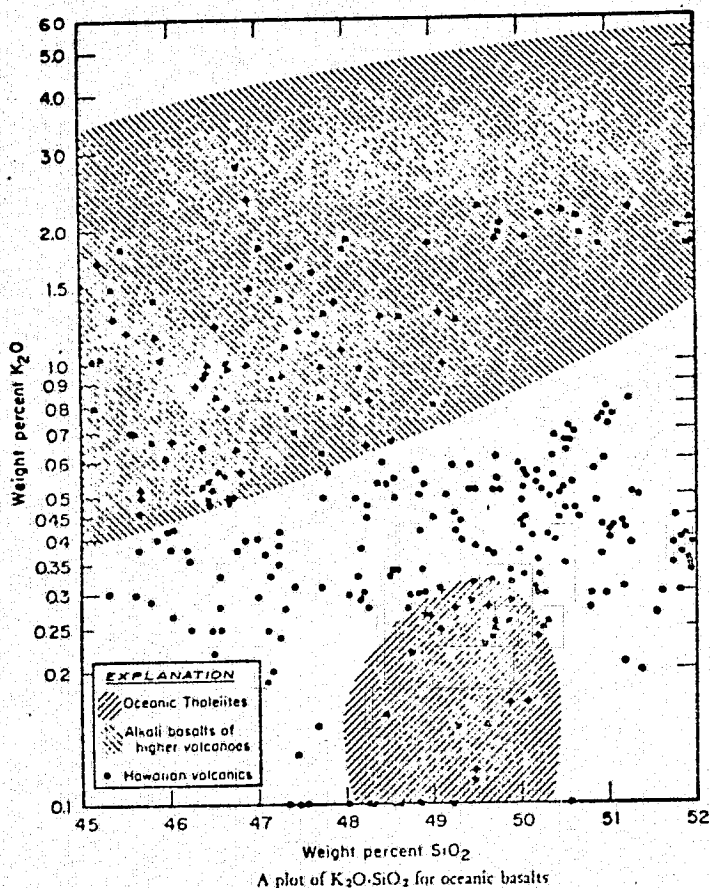


Figure IB 10. (Engel et al, 1965)

Chemical compositions (oxides, wt %) of volcanic rocks from Trinidad and Fernando de Noronha

	1	2	3	4	5	6	7	8	9
SiO <sub>2</sub>	39.00	40.08	44.80	51.16	38.42	42.68	44.23	54.82	60.81
TiO <sub>2</sub>	3.60	2.30	1.60	0.49	4.01	2.00	4.33	0.50	0.65
Al <sub>2</sub> O <sub>3</sub>	11.86	15.67	17.76	21.53	13.55	16.65	10.12	22.46	18.88
Fe <sub>2</sub> O <sub>3</sub>	6.20	6.75	5.55	2.64	3.32	5.08	3.50	1.84	2.57
FeO	9.55	5.17	3.81	1.86	9.40	8.11	6.58	0.72	0.00
MnO	0.19	0.15	0.20	0.07	0.21	0.20	0.18	0.12	—
MgO	12.31	4.49	3.47	0.68	12.54	5.57	11.70	0.07	0.61
CaO	10.40	10.60	7.80	1.92	11.75	11.00	11.45	1.42	1.70
Na <sub>2</sub> O	3.68	6.49	6.87	10.53	3.72	5.04	3.20	10.22	6.20
K <sub>2</sub> O	1.80	1.35	3.87	5.69	0.86	1.69	1.12	5.93	5.80
P <sub>2</sub> O <sub>5</sub>	0.55	1.40	0.88	0.05	1.01	0.67	0.78	0.12	
H <sub>2</sub> O <sup>+</sup>	0.50	3.40	1.10	1.72	1.15	0.52	2.04	0.82	
H <sub>2</sub> O <sup>-</sup>	0.30	2.20	2.30	0.45	n.d.	0.78	0.50	0.02	2.22
Cl				0.34				0.28	
SO <sub>3</sub>				1.14			0.31*	0.98	
Total	99.94	100.05	100.01	100.27	99.94	99.99	100.04	100.32	99.44

\* CO<sub>2</sub>.

*Explanation of column headings*

- 1 Nephelinite (ankaratrite), Trinidad (Almeida, 1961, p. 168, table 19, no. 6)
- 2 Nephelinite, Trinidad (Almeida, 1961, p. 122, table 9, no. 1; p. 168, table 19, no. 10)
- 3 Sanidine nephelinite (grazinite), Trinidad (Almeida, 1961, p. 137, table 13, no. 1; p. 168, table 19, no. 17)
- 4 Phonolite (tinguaite), Trinidad (Almeida, 1961, p. 108, table 6, no. 3; p. 168, table 19, no. 24)
- 5 Nephelinite (ankaratrite), Fernando de Noronha (Almeida, 1955, p. 154, no. 30)
- 6 Alkali basalt, Fernando de Noronha (Almeida, 1955, p. 150, no. 18)
- 7 Nepheline basanite, Fernando de Noronha (W. C. Smith and Burri, 1933, p. 430; Almeida, 1955, p. 150, no. 19)
- 8 Sodalite phonolite, Fernando de Noronha (W. C. Smith and Burri, 1933, p. 412; Almeida, 1955, p. 145, no. 3)
- 9 Trachyte, Fernando de Noronha (Almeida, 1955, p. 147, no. 10)

Table IB 3. (CTV)

coast of Brazil, is a nepheline--phonolite series. All of the rocks are exceptionally high in alkali and low in silica and contain high normative nepheline (Table IB-3). Another extreme is represented in the rocks of Tristan de Cunha (Table IB-4) where alkali basalts exist in only minor amounts relative to the (differentiated) trachybasalts and trachytes that are greatly enriched in potassium (and thus, Rb, Ba, Sr, Th and U).

More typical (?) of the oceanic alkaline basalt series are the rocks of St. Helena (Baker, 1969) an island about 800 km east of the mid-Atlantic ridge. Alkali basalts make up ~70% of exposed rock generally followed by mugearites, trachytes, and phonolites (Table IB-5). Chemical variation diagrams (Figure IB-11a,b,c,) are consistent with a fractional crystallization model (early → late: olivine, chrome spinel, pyroxene, plagioclase, alkali feldspar) of parental alkali olivine basalt liquid (or + ab + ne = 35).

### iii. THOLEIITE--ALKALI BASALT ASSOCIATIONS.

Although tholeiites and alkali basalts appear to be distinct and separate parental material in a wide variety of oceanic environments, there are also significant oceanic provinces where they are not only both present, but perhaps related.

Galapagos. The currently active Galapagos

Chemical compositions (oxides, wt %), CIPW norms, and atomic abundances (ppm) of trace elements in volcanic rocks of Tristan da Cunha (P. E. Baker et al., 1964) and Gough Island (Le Maitre, 1962, table 10)

	1	2	3	4	5	6	7		1a	2a	3a	4a	5a	6a	7a
SiO <sub>2</sub>	42.43	46.01	54.95	58.0	47.73	48.79	54.41	Ba	750	950	1200	1000	800	700	1400
TiO <sub>2</sub>	4.11	2.19	1.58	1.2	3.30	3.18	1.67	Co	40	20			22	100	5
Al <sub>2</sub> O <sub>3</sub>	14.15	16.84	19.63	19.5	15.53	17.39	17.37	Cr	65	18			220	100	
Fe <sub>2</sub> O <sub>3</sub>	5.84	7.61	1.62	1.7	2.02	2.48	4.02	Li	4	4	10	15	16	5	7
FeO	8.48	5.37	3.31	2.2	8.95	7.39	3.29	Ni	50				100	30	2
MnO	0.17	0.18	0.18	0.1	0.14	0.10	0.12	Pb	10	18	14	28			
MgO	6.71	4.75	1.42	1.0	8.37	4.00	2.27	Rb	110	110	230	350	100	40	100
CaO	11.91	9.36	5.73	3.3	8.71	8.97	4.36	Sr	1000	1100	1300	650	650	1000	850
Na <sub>2</sub> O	2.77	3.74	5.89	6.5	2.89	3.28	4.94	V	400	200	95	50	140	250	80
K <sub>2</sub> O	2.04	2.72	4.95	5.3	1.70	2.28	4.69	Zr	200	360	350	350	125	220	450
P <sub>2</sub> O <sub>5</sub>	0.58	1.18	0.43	0.2	0.29	0.26	0.46								
H <sub>2</sub> O <sup>+</sup>	0.34	0.01	0.00	0.2	0.18	0.98	0.86								
H <sub>2</sub> O <sup>-</sup>	0.44	0.08	0.01	0.1	0.06	0.76	1.50								
Total	99.97	100.04	100.05*	99.3	99.87	99.86	100.09†								
<i>or</i>	12.06	16.08	29.27	31.33	10.01	13.34	27.50								
<i>ab</i>	6.95	21.63	29.99	40.12	24.10	27.77	39.30								
<i>an</i>	20.16	21.13	13.57	8.38	24.46	26.13	11.40								
<i>ac</i>	8.93	5.43	9.66	8.06			1.42								
<i>di</i>	28.01	13.73	9.60	5.36	13.69	13.57	5.19								
<i>ky</i>							0.86								
<i>ol</i>	5.48	4.03	0.94	0.55	16.57	6.99	2.40								
<i>nt</i>	8.47	11.03	2.34	2.46	3.02	3.71	5.80								
<i>il</i>	7.81	4.16	3.00	2.23	6.23	6.08	3.19								
<i>ap</i>	1.37	2.78	1.01	0.47	0.67	0.67	1.34								
Total	99.24	100.00	99.38	99.01	99.61	98.26	97.84								

\* Including Cl = 0.27, F = 0.08.

† Including F = 0.13.

*Explanation of column headings*

Numbers in parentheses refer to specimen numbers in sources cited.

- 1, 1a Alkali basalt (6), Tristan da Cunha
- 2, 2a Trachybasalt (369), Tristan da Cunha
- 3, 3a Trachyandesite (657), Tristan da Cunha
- 4, 4a Trachyte (560), Tristan da Cunha
- 5, 5a Olivine basalt (G111), Gough Island
- 6, 6a Trachybasalt (G22), Gough Island
- 7, 7a Trachyandesite (G86), Gough Island

Table IB 4. (CTV)

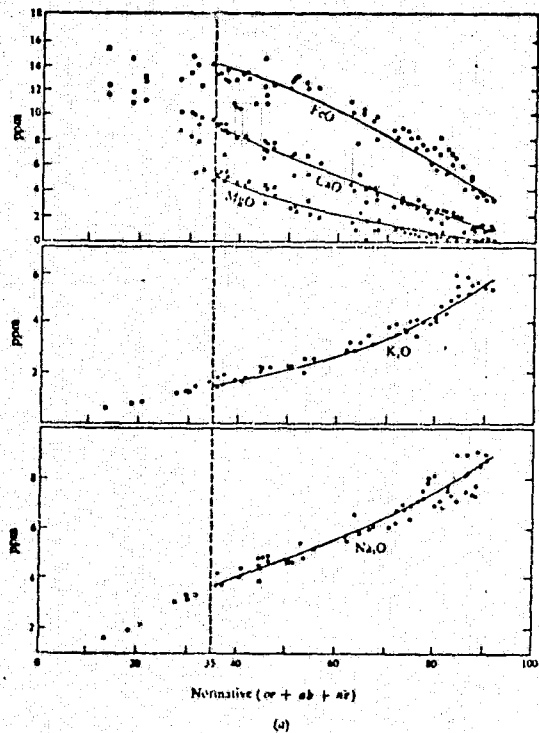
Chemical composition (oxides, wt %), CIPW norms, and atomic abundances (ppm) of trace elements of volcanic rocks of St. Helena (I. Baker, 1969, table 2) and Mauritius (Shand, 1933)

	1	2	3	4	5	6		1a	2a	3a	4a
SiO <sub>2</sub>	45.50	54.88	59.64	59.92	46.90	60.69	Ba	290	790	1100	1100
TiO <sub>2</sub>	3.44	1.11	0.42	0.06	3.31	0.15	Cr	36			
Al <sub>2</sub> O <sub>3</sub>	15.71	17.41	17.67	19.86	15.05	19.74	Rb	40	50	90	220
Fe <sub>2</sub> O <sub>3</sub>	3.61	2.44	2.59	1.69	1.11	1.92	Sr	510	545	280	85
FeO	8.64	6.79	3.92	1.70	10.46	2.33	Zr	195	490	645	1085
MnO	0.22	0.23	0.22	0.18	0.16	0.17					
MgO	5.37	1.88	0.50	0.05	8.41	0.01	Rb/Sr	0.08			
CaO	9.43	3.48	2.09	1.07	10.92	1.02	K/Rb	300	400	350	180
Na <sub>2</sub> O	3.47	6.04	7.62	8.94	3.13	7.95					
K <sub>2</sub> O	1.38	2.89	4.02	4.93	0.27	5.50					
P <sub>2</sub> O <sub>5</sub>	0.29	0.41	0.29	0.17	Tr.	Tr.					
H <sub>2</sub> O <sup>+</sup>	0.60	0.77	0.08	0.07	0.27	0.58					
H <sub>2</sub> O <sup>-</sup>	2.49	1.53	1.04	1.29	0.23	0.28					
Total	100.15	99.86	100.10	99.93	100.22	100.34					
<i>or</i>	8.16	17.08	23.76	29.14	1.67	32.80					
<i>ab</i>	24.79	51.07	53.68	46.68	23.35	47.16					
<i>an</i>	23.22	11.86	2.14		26.13	1.67					
<i>ne</i>	2.48	0.02	5.85	15.18	2.27	10.79					
<i>ac</i>				0.83							
<i>di</i>	15.97	0.32	6.65	4.29	21.84	2.97					
<i>ol</i>	9.60	10.15	1.96	0.03	16.74	1.81					
<i>mt</i>	5.23	3.54	3.76	2.04	1.62	2.55					
<i>il</i>	6.53	2.11	0.80	0.11	6.23	0.30					
<i>ap</i>	1.42	1.82	0.19	0.17	—	—					
Total	97.40	97.97	98.79	98.47	99.85	100.05					

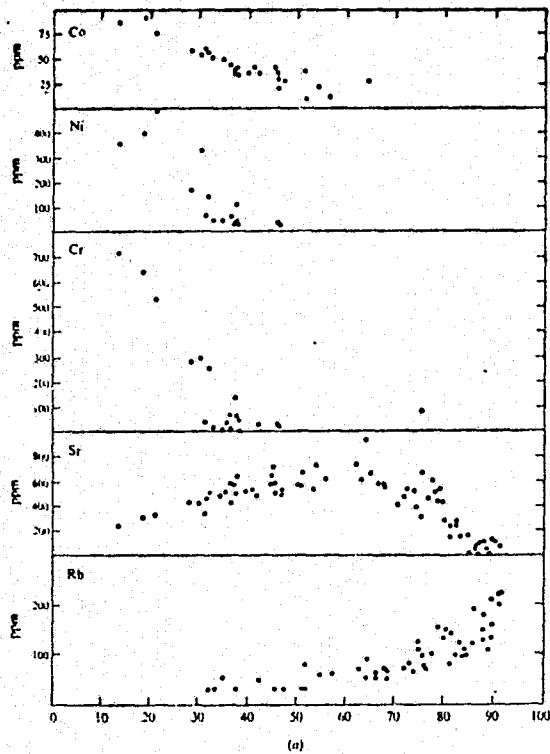
*Explanation of column headings*

- 1, 1a Basalt, St. Helena (803)  
 2, 2a Mugearite (trachyandesite), St. Helena (822)  
 3, 3a Trachyte, St. Helena (763)  
 4, 4a Phonolite, St. Helena (11)  
 5 Olivine basalt, Mauritius (A, p. 5)  
 6 Phonolitic trachyte, Mauritius (A, p. 8)

Table IB 5. (CTV)



Chemical variation in lavas of St. Helena (data from I. Baker, 1969). Total iron plotted as FeO.



Trace- and minor element variation in lavas of St. Helena. (Data from I. Baker, 1969.) Abundances (ppm) plotted against normative (or + ab + ar). (a) Co, Ni, Cr, Sr, Rb.

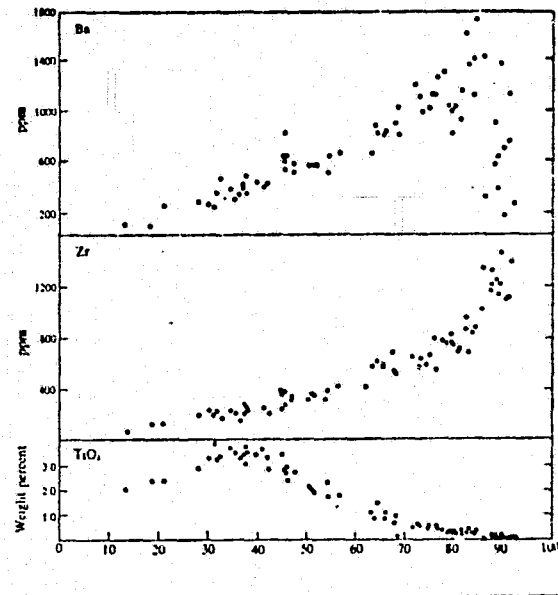


Figure IB 11. (CTV)

islands are located on a platform (1.5 km below sea level) east of the East Pacific Rise. The islands are composed predominantly of basalts (Table IB-6) that range from slightly saturated tholeiites to alkali basalts (McBirney and Williams, 1969) but that are spatially separated. Alkali basalts make up the central and eastern islands whereas tholeiitic lavas and their derivative trachytes made up the western islands.

Hawaii. The Hawaiian Islands are a series of volcanoes extending into the Emperor Seamounts that form a line along the central floor of the Pacific plate, increasing in age towards the west (Figure 1B-12; from Dalrymple et al., 1973). Due to the great diversity of rock types found on the islands and the constant level of recent activity, Hawaii has become a testing ground for many studies of terrestrial basalts. The age of the ocean floor in this part of the Pacific is about 100 million years (Larson and Chase, 1972). Most investigators agree that the origin of the Hawaiian chain is due to the relative motion of the Pacific plate and some melting zone within the mantle, which has been active for at least 50 m.y. (for a review of proposed hypothesis concerning this melting zone, see Dalrymple et al., 1973). From estimates of extrusion rates, Mauna Loa could have grown to its present

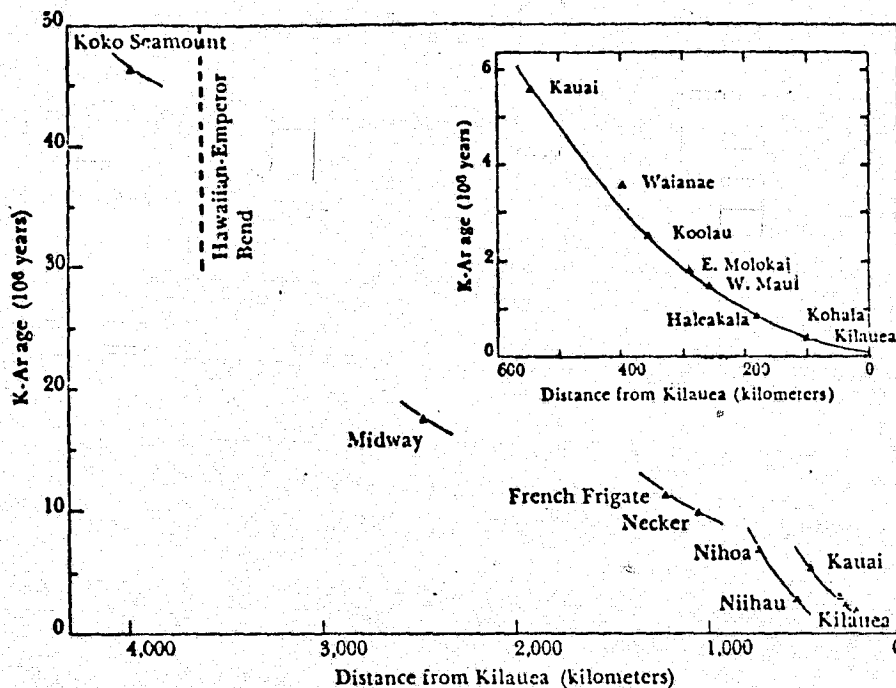
Chemical compositions (oxides, wt %), CIPW norms, and atomic abundances of trace elements; volcanic rocks of western Galápagos

	1	2	3	4	5		2a	4a
SiO <sub>2</sub>	47.01	48.45	48.54	55.34	66.87	Ba	76	285
TiO <sub>2</sub>	3.20	3.39	4.20	1.93	0.66		(65)	(280)
Al <sub>2</sub> O <sub>3</sub>	15.57	13.75	14.49	14.02	12.55	Co	55	54
Fe <sub>2</sub> O <sub>3</sub>	2.32	4.72	4.67	3.31	1.84	Cr	47	17
FeO	11.57	8.60	9.16	8.73	2.53		(142)	
MnO	0.20	0.20	0.16	0.17	0.09	Mn	1440	1600
MgO	5.25	6.05	4.58	2.71	0.60	Ni	70	14
CaO	9.77	10.71	8.29	6.54	1.10	Sr	235	250
Na <sub>2</sub> O	3.00	2.79	3.58	4.54	5.32		(323)	(266)
K <sub>2</sub> O	0.31	0.50	0.84	1.33	3.08	V	310	98
P <sub>2</sub> O <sub>5</sub>	0.32	0.36	0.47	0.65	0.05	Zr	(255)	(538)
H <sub>2</sub> O <sup>+</sup>	1.40	0.36	0.68	0.64	4.66			
H <sub>2</sub> O <sup>-</sup>	0.24	0.02	0.16	0.18	0.33			
<b>Total</b>	<b>100.16</b>	<b>99.90</b>	<b>99.82</b>	<b>100.09</b>	<b>99.68</b>			
<i>Q</i>	—	2.21	2.04	5.95	20.82			
<i>or</i>	1.83	2.95	4.96	7.86	18.20			
<i>ab</i>	25.39	23.61	30.29	38.42	45.02			
<i>an</i>	28.10	23.52	20.99	13.95	1.27			
<i>di</i>	15.17	22.00	13.86	11.97	3.24			
<i>hy</i>	11.92	11.12	11.00	11.16	2.15			
<i>ol</i>	5.93	—	—	—	—			
<i>mt</i>	3.36	6.84	6.77	4.80	2.67			
<i>il</i>	6.08	6.44	7.98	3.67	1.23			
<i>ap</i>	0.76	0.75	1.11	1.54	0.12			
<b>Total</b>	<b>98.54</b>	<b>99.44</b>	<b>99.00</b>	<b>99.32</b>	<b>94.72</b>			

*Explanation of column headings*

- 1 Basalt, Albemarle Island (McBirney and Williams, 1969, p. 124, table 3, no. 8)
  - 2 Tholeiitic basalt, Albemarle Island (McBirney and Williams, 1969, p. 121, table 2, no. 63)
  - 3 Ferrobasalt, Jervis Island (McBirney and Williams, 1969, p. 147, table 10, no. 65)
  - 4 Icelandite, Duncan Island (McBirney and Williams, 1969, p. 146, table 10, no. 71)
  - 5 Siliceous trachyte pumice, Alcado volcano, Albemarle Island (McBirney and Williams, 1969, p. 146, table 10, no. 130)
- 2a, 4a Trace-element abundances, same rocks as analyses 2 and 4. Determinations by emission spectrography. Values in parentheses determined by neutron activation (Cr) or x-ray fluorescence (Ba, Sr, Zr).

Table IB 6. (CTV)



Potassium-argon age of tholeiitic volcanism on the Hawaiian-Emperor chain as a function of distance from Kilauea. Inset shows the detailed relations for the principal Hawaiian islands. Lines connect ages of shield volcanoes that lie on common or adjacent loci. Although the ages of the volcanoes increase more or less sys-

tematically to the northwest along the chain, the age-distance relations are not linear. (Data from McDougall 1964; Dalrymple 1971; Doell and Dalrymple 1973; McDougall and Swanson 1972; Clague and Dalrymple 1973; and unpublished data of the U.S. Geological Survey.)

Figure IB 12. (Dalrymple et al., 1973)

size (about 9 km above the ocean floor) in 1.5 million years (MacDonald and Abbot, 1970) and the smaller Kilauea could have required less than .4 my (Swanson, 1972).

The lavas forming the islands follow a well-defined sequence: the major mass of each volcano is made up of tholeiites, followed by and interbedded with small amounts (~1%) of alkalic lavas (Table IB-7 from MacDonald, 1968). After a period of volcanic quiescence, nephelinitic magmas are erupted. No sharp chemical boundary exists between these three major rock suites as is demonstrated by the alkali variation diagram of Figure IB-13 (from MacDonald, 1968) and the potassium variation diagram in Engel et al. (Figure IB-10). When fewer data were available, the vague distinction between alkali and tholeiitic basalts was originally believed to be a sharp chemical boundary (Tilley, 1950).

Short-term fractionation trends within a given suite of rocks is rarely disputed for Hawaiian lavas. Although the dominant material that makes up the volcanoes is a slightly saturated tholeiite, it is largely agreed (e.g., O'Hara, 1965; MacDonald, 1968; Carmichael et al., 1974) that the primary magma type was olivine tholeiite with much of the olivine lost either during the ascent from the source (~60km) or during fractionation in a near surface magma

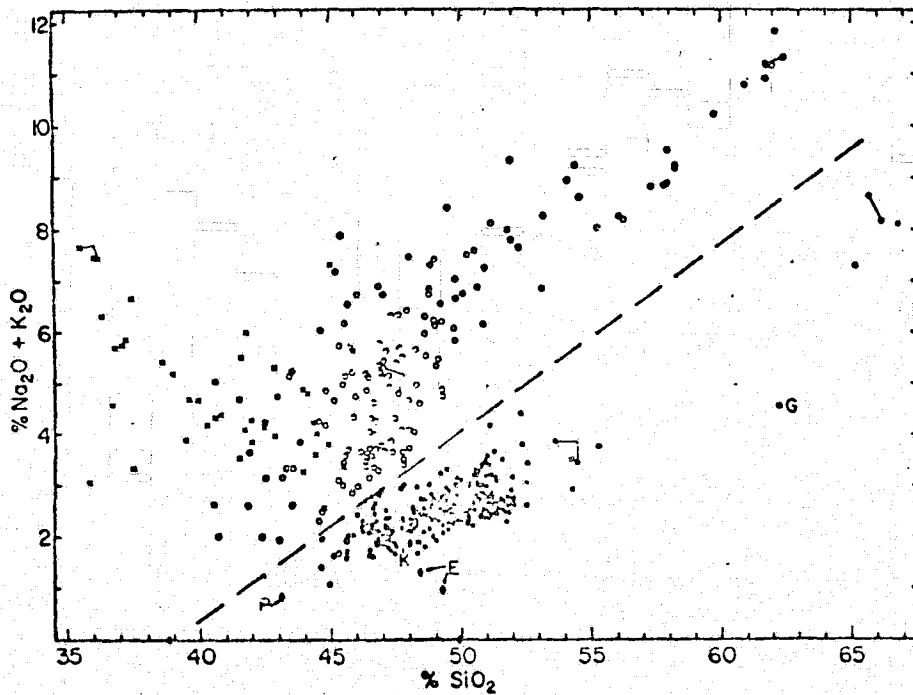
AVERAGE COMPOSITIONS OF HAWAIIAN LAVAS

Type of rock	Tholeiitic suite		Alkalic suite							Nephelinitic suite					
	Oceanite	Tholeiite and olivine tholeiite	Ankaramite	Alkalic olivine basalt	Feldspar-phyric basalt	Hawaiite	Mugearite	Benmoreite	Soda trachyte	Alkalic olivine basalt (basanitoid)	Basanite	Mimosite	Nephelinite	Ankaratrite	Melilitic nephelinite
Number of analyses	14	200	9	35	6	62	23	5	5	11	11	4	10	1	7
SiO <sub>2</sub>	46.4	49.4	44.1	45.4	46.6	47.9	51.6	57.1	61.7	44.8	44.1	41.3	39.7	39.4	36.6
Al <sub>2</sub> O <sub>3</sub>	8.5	13.9	12.1	14.7	16.8	15.9	16.9	17.6	18.0	12.7	12.7	10.4	11.4	10.2	10.8
Fe <sub>2</sub> O <sub>3</sub>	2.5	3.0	3.2	4.1	4.5	4.9	4.2	4.8	3.3	3.2	3.6	5.6	5.3	6.5	5.7
FeO	9.8	8.5	9.6	9.2	8.1	7.6	6.1	3.0	1.5	9.4	9.1	8.3	8.2	7.0	8.9
MgO	20.8	8.4	13.0	7.8	5.8	4.8	3.3	1.6	1.4	11.4	11.2	13.8	12.1	14.1	12.6
CaO	7.4	10.3	11.5	10.5	9.3	8.0	6.1	3.5	1.2	11.4	10.6	12.1	12.8	12.3	13.6
Na <sub>2</sub> O	1.6	2.2	1.9	3.0	3.2	4.2	5.4	5.9	7.4	2.7	3.6	2.8	3.8	2.7	4.1
K <sub>2</sub> O	0.3	0.4	0.7	1.0	0.8	1.5	2.1	2.8	4.2	0.9	1.0	0.9	1.2	1.2	1.0
TiO <sub>2</sub>	2.0	2.5	2.7	3.0	3.3	3.4	2.4	1.2	0.5	2.3	2.6	2.7	2.8	3.3	2.8
P <sub>2</sub> O <sub>5</sub>	0.2	0.3	0.3	0.4	0.4	0.7	1.1	0.7	0.2	0.5	0.5	0.7	0.9	0.8	1.1
MnO	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.1	0.1
FeO:Fe <sub>2</sub> O <sub>3</sub>	4.0	2.8	3.0	2.2	1.8	1.6	1.4	0.6	0.4	2.9	2.5	1.5	1.5	1.1	1.6

TABLE 8. (CONTINUED)

NORMS (CIPW)																
Q	.....	2.2	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	
or	1.7	2.2	3.9	6.1	5.0	8.9	12.2	16.7	25.0	5.6	6.1	5.6	.....	.....	.....	
ab	13.1	18.3	11.5	20.4	27.2	35.1	45.6	49.8	62.4	12.1	11.0	2.6	.....	.....	.....	
an	15.6	27.0	22.5	23.6	28.9	20.0	15.8	12.8	3.3	19.7	15.6	13.1	10.6	12.2	8.1	
ne	.....	.....	2.6	2.6	.....	0.3	.....	.....	.....	6.0	10.5	11.4	17.3	12.2	18.7	
lc	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	5.7	5.7	4.8	
di	wo	8.5	9.3	13.6	10.8	6.0	6.3	2.9	.....	0.7	13.9	13.9	17.6	17.5	18.1	11.4
	en	6.1	6.0	9.4	6.9	4.0	4.2	2.5	.....	0.6	9.2	9.5	13.4	13.0	15.0	8.3
hy	fs	1.6	2.6	3.0	3.2	1.6	1.6	.....	.....	.....	3.7	3.3	2.4	2.8	0.8	2.0
	en	12.7	14.9	.....	.....	4.4	.....	2.7	4.0	0.4	.....	.....	.....	.....	.....	.....
cl	fo	23.2	.....	16.2	8.8	4.3	5.5	2.1	.....	.....	13.5	13.0	14.8	12.0	14.3	16.2
	fa	6.3	.....	6.0	4.4	2.1	1.2	.....	.....	.....	5.8	4.9	3.2	2.9	1.0	4.0
g	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	1.7	.....	7.6	
mt	3.5	4.4	4.6	6.0	6.5	7.2	1.9	7.0	4.2	4.6	5.3	8.1	7.7	9.5	5.5	
ll	3.8	4.7	5.2	5.8	6.2	6.5	4.6	2.3	0.9	4.4	5.0	5.6	5.3	6.2	5.3	
sp	0.3	0.7	0.7	1.0	1.0	1.7	2.7	1.7	0.3	1.3	1.3	1.7	2.0	2.0	2.7	
c	.....	.....	.....	.....	.....	.....	.....	0.2	.....	.....	.....	.....	.....	.....	.....	

Table IB 7. (Macdonald, 1968)

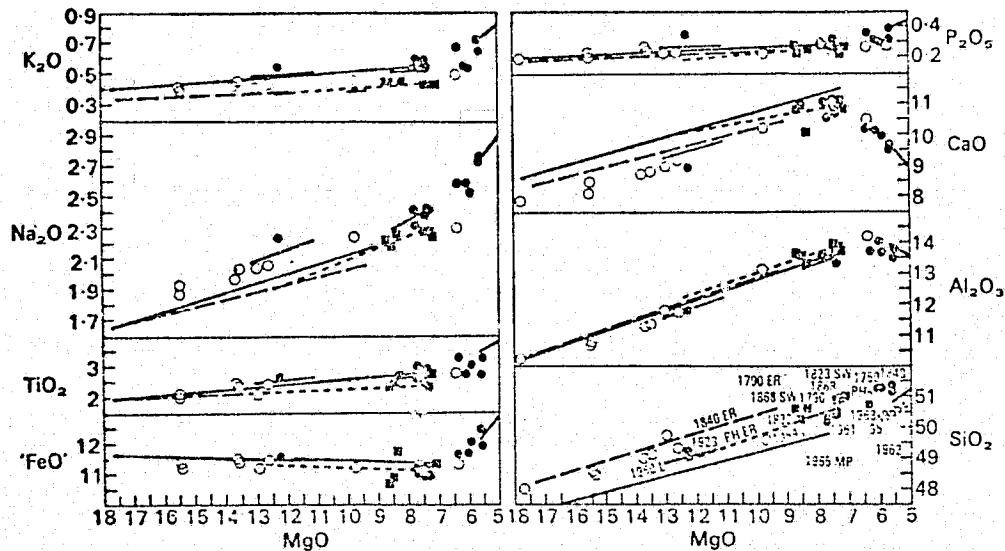


Alkali:silica diagram of Hawaiian rocks. Solid circles, tholeiitic rocks; open circles, rocks of the alkalic suite; crosses, posterosional rocks of the nephelinic suite; G, granophyre associated with the Palolo quartz diabase; open triangles, low-K tholeiites of the deep Pacific basin (Engel and others, 1965); K, weighted average of first phase of 1959 eruption of Kilauea (Murata and Richter, 1966); P, pyrolite (Green and Ringwood, 1963); E, "eclogite" inclusions from Salt Lake tuff (Yoder and Tilley, 1962, p. 482; Macdonald and Katsura, 1964, p. 123). The pairs of points connected by lines represent analyses of the same rock powder by two different laboratories.

Figure IB 13. (Macdonald, 1968)

chamber. Magma chambers are known to exist for the two most recently active volcanoes: for Kilauea it is estimated to be 3-5 km below the current summit, whereas for Mauna Loa the chamber is deeper--10-15 km (e.g., Wright and Fiske, 1971; Lipman, 1977). There is no obvious link in the plumbing of these two neighboring volcanoes. The chemistry of a series of summit and flank eruptions for Kilauea was studied by Wright and Fiske (1971) and showed clear olivine fractionation control which presumably occurred largely in the magma chamber (Figure IB-14 from Wright and Fiske). Rift lavas show distinct fractionation from parental summit material.

There is less agreement as to the relation between the various rock suites. The regularity of the tholeiite--alkali basalt--nepheline basalt sequence, the gradational geochemistry of these basalt types and their similarity in  $Sr^{87}/Sr^{86}$  (Table IB-8) indicate that they are related in some way. The processes involved in their petrogenesis must be repeatable. Green and Ringwood (1967) and Green (1971) have stressed results from experimental studies that the composition of a partial melt is greatly controlled by conditions of pressure, temperature, and water content in the source region. Their 'pyrolite' models of the upper mantle (peridotite composition) can produce magmas of tholeiitic or alkaline composition by changing the



MgO variation diagrams: differentiated lavas of the east rift zone and selected undifferentiated lavas of historic age. Average control lines (essentially straight olivine control) have been computed for the several eruptions labeled on the SiO<sub>2</sub>-MgO plot: Ancient Makaopuhi lava lake = PH ER (analyses from Moore & Evans, 1967, and additional unpublished analyses); 1840 undifferentiated lavas = 1840 ER; later lavas of the 1960 eruption = 1960 L (analyses from Murata & Richter, 1966a); March, 1965, eruption in Makaopuhi crater = 1965 MP (analyses from Wright *et al.*, 1968, and additional unpublished analyses of Makaopuhi lava lake). A control line for the early 1955 lavas = 1955 E, obviously not olivine control, is also plotted. A control line for undifferentiated lava from Kilauea summit, the latter part of the 1959 eruption in Kilauea Iki crater = 1959 L is given for reference.

All other eruptions are identified as follows on the SiO<sub>2</sub>-MgO plot: Filled squares = undifferentiated historic lavas, Kilauea summit, south-west and east rift zones. Filled circles = differentiated lavas of historic age, east rift zone. Open circles = differentiated lavas of prehistoric age, east rift zone. All lavas represented by open circles are submarine except that labeled PH.

Figure IB 14. (Wright and Fiske, 1971)

## Isotopic compositions of Sr\* and Pb in oceanic volcanic series

PROVINCE	ROCKS	Sr <sup>87</sup> /Sr <sup>86</sup>	Pb <sup>206</sup> /Pb <sup>204</sup>	
Réunion (McDougall and Compston, 1965)	Basalts (5)	0.7040-0.7046		
	Mugearites (3)	0.7042-0.7044		
	Syenite (1)	0.7046		
Easter Island (Hedge and Peterman, 1970; Tatsumoto, 1966a)	Basalts (2)	0.7030, 0.7036	19.280, 19.301	
	Icelandite (1)	0.7030	19.253	
	Rhyolite obsidian (1)		19.308	
Hawaiian Islands (Hedge, 1966; Tatsumoto, 1966b Hedge and Peterman, 1970)	Hawaii	Kilauea tholeiites (3)	0.7039-0.7041	
		Hualalai, alkali basalt		17.92
		Hualalai, trachyte	0.7035	18.08
		Mauna Kea, picrite basalt	0.7034	18.48
	Oahu	Mauna Kea, hawaiiite	0.7033	18.47
		Koolau basalts	0.7039	18.09
		Waianae basalts	0.7032	
		Honolulu series (nephelinites)	0.7031	18.17, 18.24
Iceland (Moorbath and Walker, 1965)	Pleistocene-Recent basalts (3)	0.7028-0.7033		
	Recent obsidians (1)	0.7017		
	Tertiary basalts (6)	0.7021-0.7032		
	Tertiary rhyolites (3)	0.7014-0.7015		
	Late Tertiary granophyres (6)	0.7010-0.7033		
Guadalupe Island (Tatsumoto, 1966a; Peterman and Hedge, 1971)	Alkali basalts (4)	0.7033-0.7036	20.172-20.436	
St. Helena (Hedge, 1966)	Alkali basalts (2)	0.7031, 0.7032		
	Phonolites (2)	0.7047, 0.7054		
Tutuila, Samoa (Hedge, 1966)	Basalt	0.7057		
	Trachytes (2)	0.7062, 0.7066		
Upolu and Savaii, Samoa (Hedge et al., 1972)	Potassic alkali basalts (10)	0.7051-0.7066		
Gough Island (Gast et al., 1964)	Basalt	0.7045	18.36	
	Trachyandesite, trachybasalt	0.7050, 0.7043	18.37, 18.43	
	Trachytes (2)	0.7094, 0.7050	18.63, 18.73	
Ascension Island (Gast et al., 1964)	Basalts (2)	0.7025	18.43, 19.55	
	Trachyandesite	0.7025		
	Trachytes (2)	0.7045, 0.7073	19.72	
	Obsidian		19.50	

\* All values normalized to Sr<sup>86</sup>/Sr<sup>88</sup> = 0.1194, Sr<sup>87</sup>/Sr<sup>86</sup> = 0.7080, for standard E. and A. SrCO<sub>3</sub>.

Table IB 8. (CTV)

the degree and/or depth of partial melting (see Figures IB-5). These systematics, along with Yoder and Tilley's (1962) low pressure thermal barrier, have led most proponents (e.g., MacDonald, 1968) of primary liquid differentiation producing alkali basalts to hypothesize deep-seated processes, perhaps involving remelting at depth. Others (e.g., Verhoogen et al., 1970; Carmichael et al., 1972) would prefer an independent origin (change of melting parameters) for each magma type. Currently neither alternative can be entirely ruled out.

Canary Islands. The seven Canary Islands lie off the coast of Africa with the eastern two bordering on the continental coast. The islands are shield volcanoes formed largely of alkali olivine basalt (Table IB-9) with the youngest to the west. The basalts seem to differentiate with time towards phonolites by removal of olivine, titanomagnetite, then plagioclase (Ridley, 1970). The two eastern (older) islands present a more complicated picture. They not only contain the typical alkali basalt differentiation sequence, but also an old intrusion of tholeiitic composition and a historic flow of olivine tholeiite with normative hypersthene (analysis #6 in Table IB-9). Not only does the tholeiite on these Canary islands

## Chemical analyses (oxides, wt %) and CIPW norms of rocks from the Canary

Islands	1	2	3	4	5	6	7	8	9
SiO <sub>2</sub>	42.50	43.20	46.10	45.86	47.80	49.30	52.55	59.20	64.20
TiO <sub>2</sub>	2.20	3.93	2.38	2.95	2.22	2.06	1.23	0.73	0.35
Al <sub>2</sub> O <sub>3</sub>	13.50	16.30	14.24	15.81	13.19	13.43	19.32	20.10	16.29
Fe <sub>2</sub> O <sub>3</sub>	5.16	8.09	3.95	4.65	2.21	1.02	2.76	1.26	2.77
FeO	7.14	4.69	7.90	5.24	8.88	9.23	2.33	1.12	0.62
MnO	0.17	0.16	0.15	0.18	0.14	0.15	0.18	0.15	0.05
MgO	10.44	5.16	9.25	6.31	10.50	11.39	1.74	0.91	1.05
CaO	10.99	10.74	10.45	9.77	9.90	9.39	4.97	2.40	2.36
Na <sub>2</sub> O	3.28	3.12	3.10	5.04	3.19	2.48	8.12	7.32	7.00
K <sub>2</sub> O	1.20	1.64	0.63	3.11	0.93	0.74	4.88	5.00	4.40
P <sub>2</sub> O <sub>5</sub>	1.04	0.88	0.53	0.70	0.58	0.36	0.34	0.16	0.05
H <sub>2</sub> O <sup>+</sup>	2.67	2.03	1.12		0.24	0.17			
H <sub>2</sub> O <sup>-</sup>							}0.40	1.36	0.66
Rest			0.10*	0.37			1.18†	0.20*	
Total	100.29	99.94	99.90	99.99	99.78	99.72	100.00	99.91	99.80

\* CO<sub>2</sub> † Cl = 0.24; SO<sub>3</sub> = 0.94.

*Explanation of column headings*

Analyses 1-6 are cited from Ibarrola (1969), analysis 7 from Hernández-Pacheco (1969), and analyses 8 and 9 from Muñoz (1969a). Tables and numbers given in parentheses refer to these works.

- 1 Olivine basalt (first episode), Lanzarote (table 1, no. 7)
- 2 Olivine basalt (first episode), Gomera (table 4, no. 8)
- 3 Olivine basalt (first episode), Fuerteventura (table 2, no. 12)
- 4 Average Quaternary basalt (tephrite), Gran Canaria (table M4, no. 2)
- 5 Olivine basalt, last Quaternary episode, Lanzarote (table 5, no. 27)
- 6 Olivine basalt, historic flow of 1730-1736 eruptions, Lanzarote (table 5, no. 30)
- 7 Tahitite, Gran Canaria (table 1, no. 4)
- 8 Nepheline syenite, central stock, Pájara, Fuerteventura (table 1, no. 3)
- 9 Syenite, La Peñitas ring dike, Pájara, Fuerteventura (table 1, no. 16)

Table IB 9. (CTV)

not conform to the compositional trend of the majority of neighboring islands, but the relation of tholeiite to alkali basalt is sequentially opposite that observed in the well-documented Hawaiian sequence.

In summary, for the ocean environment, basalts and their derivatives are young ( $\sim$  200 my) and uncontaminated by continental material. A variety of basalt compositions can be observed, perhaps the most primary occurring at the spreading ocean ridges. All are presumed to be derived from the upper mantle by one or more sequences of partial melting. The oceanic upper mantle itself is roughly homogeneous in major element geochemistry, but may be heterogeneous either vertically or laterally in incompatible elements.

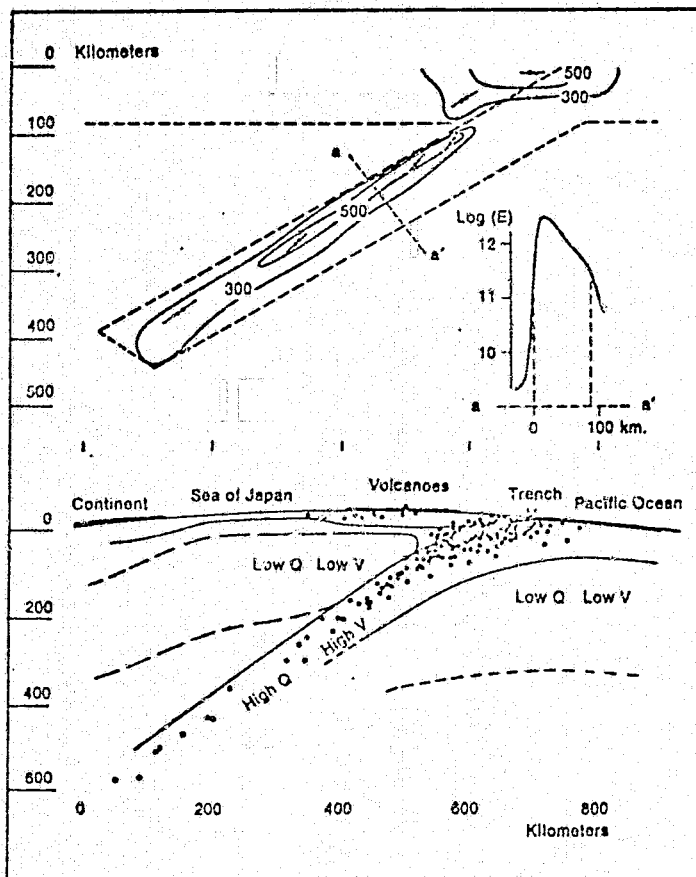
c. Subduction Zone

Most recent, active volcanoes are located along margins of lithospheric plates which are in motion relative to each other a few centimeters a year. Three quarters of the world's active surface volcanoes occur along the Pacific margin (CTV, p.528). The ocean floor basalts are formed as two oceanic plates separate. When two plates collide, one is generally forced to descend into the mantle creating a subduction zone of intense earth-

quake activity (Figure IB-15 from Toksoz, 1972) and initiating a variety of volcanic materials. The rock type generally associated with subduction zone volcanics is andesite (the rock name being derived from the Andes of South America). Since it is the nature of this tectonic setting that the subducted (oceanic) plate encounters changing temperatures and pressure (and perhaps  $H_2O$  content) as it descends, the chemistry of igneous melts associated with this environment would be expected to be different from those associated with simple partial melting of the upper mantle that occurs in the ocean environment.

Although the subduction zone continental igneous rocks are predominantly andesite and related differentiates, basalts are also abundant in the island arc environment, especially in younger arcs (Baker, 1968). Again there is no uniform sequence for all environments but a few patterns can be recognized:

(1) Chemically, the circumoceanic basalts are different from ocean floor or ocean island basalts. No material associated with island arcs or continental margins contains the very low strontium ratio ( $<.703$ ) of abyssal basalts (Table IB-10). For some provinces, the initial strontium isotopic ratios for a range of rock types are essentially uniform implying a genetic relationship. The moderately



Cross-section of calculated stress model (upper) and the distribution of earthquakes (lower) in descending lithosphere under Japan. The units of stress are "bars" (very closely, atmospheric pressures). Inset diagram indicates the Young's modulus variation inside the slab due to temperature effects. Arrows point the direction of maximum principal stress. In the lower diagram earthquake

foci are indicated by black dots. The shallow quakes are located along the boundary between the descending lithosphere and the stable lithosphere. Deep quakes are located along the interior of the slab where the slab is coolest and where calculated stresses are maximum. Low Q (high attenuation of seismic waves) and Low V (low velocity) indicate the asthenosphere.

Figure IB 15. (Toksoz, 1972)

Average initial strontium isotopic values ( $Sr^{87}/Sr^{86}$ ) for selected rock types of active volcanic arcs and continental margins (all data adjusted to a value of 0.7080 for MIT  $SrCO_3$ )\*

	BASALT ( $< 52\% SiO_2$ )	BASALTIC ANDESITE ( $52-55\% SiO_2$ )	ANDESITE ( $55-63\% SiO_2$ )	DACITE ( $63-68\% SiO_2$ )	RHYOLITE ( $> 68\% SiO_2$ )
New Britain	0.7035	0.7036	0.7036	0.7036	0.7035
Tonga	—	0.7037	0.7042	0.7043	—
Marianas	0.7042	—	0.7042	0.7038	—
Izu Islands	0.7036	—	0.7040	—	0.7034
Caribbean					
St. Kitts	0.7036	0.7040	0.7038	—	—
St. Vincent	0.7042	0.7040	0.7039	—	—
Carricou	0.7052	—	0.7054	—	—
North Japan	0.7043	—	0.7041	—	—
California					
Mt. Shasta	—	0.7035	0.7030	0.7032	—
Mt. Lassen	0.7039	0.7032	0.7040	—	—
Medicine Lake	0.7034	0.7037	—	—	0.7040
Central America	0.7035	0.7042	0.7036	0.7036	0.7042
New Zealand					
Taupo	0.7042	—	0.7055	0.7051	0.7053
Average†	0.7040	0.7038	0.7040	0.7037	0.7038

\* Data taken from Pushkar (1968), Peterman et al. (1970a, 1970b), Hedge (1966), Hedge and Knight (1969), Hedge and Lewis (1971), Ewart and Stipp (1968), Oversby and Ewart (1972), Ewart et al. (1973).

† Excludes New Zealand average andesite, dacite, and rhyolite.

Table IB 10. (CTV)

low ratios for the entire series of rock types imply minor, if any, crustal contamination.

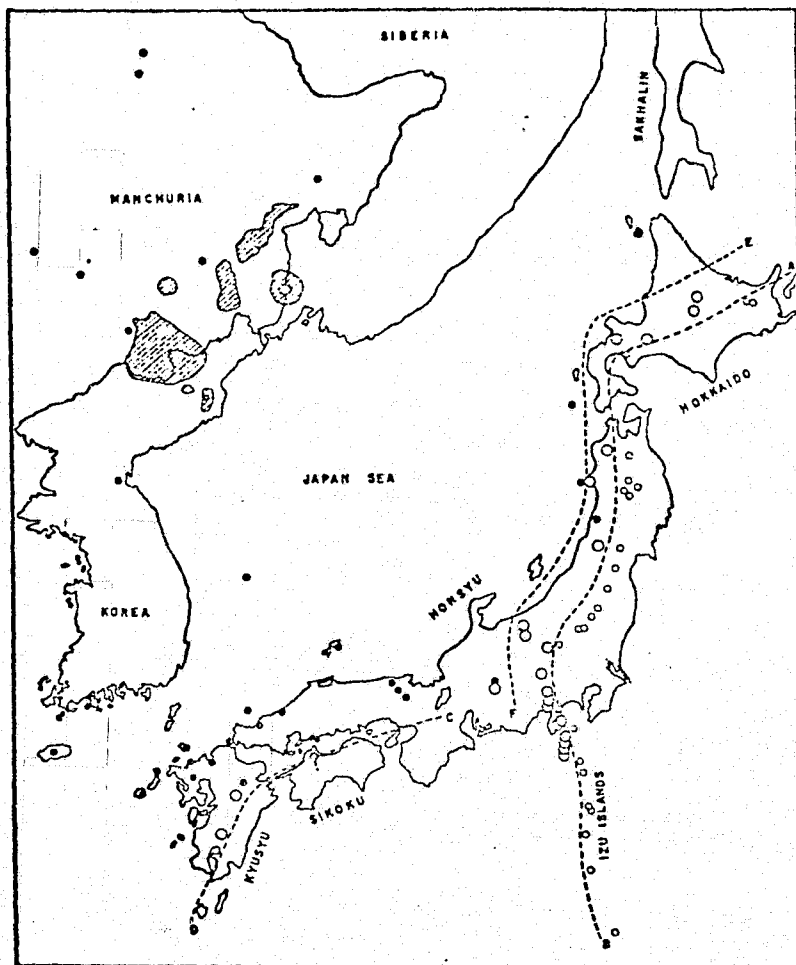
(2) Kuno (1967) noted that volcanoes extruding tholeiitic basalts occurred closest to oceanic trenches (start of the subduction zone) and that volcanoes with high-aluminum basalts and alkali-olivine basalts occurred respectively further inland (Figure IB-16 from Kuno, 1967). The correlation of this pattern with deep foci earthquakes suggested the generation of different basalts at successively greater depths.

(3) The association of magma generation with descending lithospheric plates is further strengthened by the tendency of  $K_2O$  content of lavas (andesite) to be correlated with depth of the subduction zone as inferred from earthquake foci (Figure IB17 from CTV, p. 562).

d. Continental basalts

A thorough discussion of continental volcanics must include extreme compositions (e.g., carbonatites) and modes of emplacement (e.g., kimberlites) and is well beyond the scope of this summary. Two common types of continental basaltic provinces will be mentioned here: the voluminous tholeiitic flood basalts and the complex alkali basalt regions. In both cases, plate tectonics have been implicated in the ultimate origin of the basalts. However, only in

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a Distribution of tholeiite (●), high-alumina basalt (○) and alkali olivine basalt (⊙) among Quaternary volcanoes of Japan, Korea and Manchuria. The shaded areas in east Manchuria are Tertiary lava plateaus mostly of alkali olivine basalt.

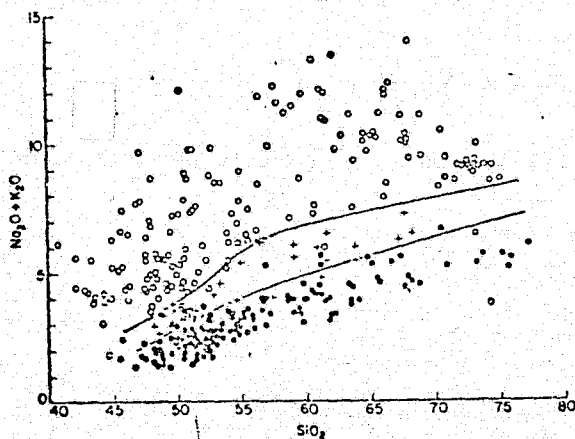
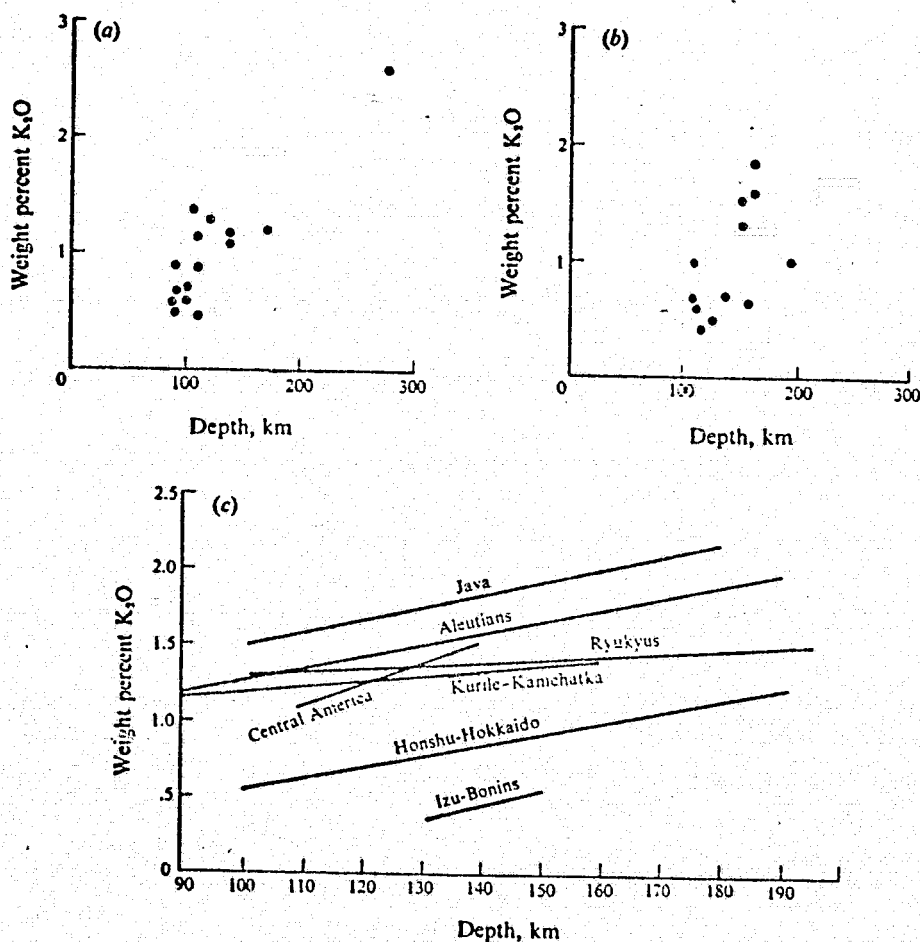


Figure IB 16.  
(Kuno, 1967)

b Alkali-silica relations of tholeiite series (●) (basalt, andesite, dacite and rhyolite; high-alumina basalt series (+) (basalt, andesite and dacite); alkali rock series (○) (basalt, mugearite, trachyandesite, trachyte and alkali rhyolite), occurring along a traverse zone from the Izu Islands, passing through central Honshū, to the Japan Sea region including Korea and Manchuria.



Plot of  $K_2O$  content (at 55 percent  $SiO_2$ ) of lava series against depth in kilometers from the volcano to the inclined seismic zone beneath: (a) Kuriles-Kamchatka; (b) Honshu, Japan; (c) Circum-Pacific volcanic arcs. Length of lines represent range of data for each region, and their position is obtained by statistical regression of published analyses (after Dickinson, 1970 and Nielson and Stoiber, 1973).

Figure IB 17. (CTV)

the case of alkali basalt association with continental rift zones has it been substantiated. An example of each province will be discussed.

i. THOLEIITIC FLOOD BASALTS. Although granitic rocks are the dominant continental plutonic rock type, the most voluminous continental lavas are basalts (CTV, p. 427). Immense outpouring over short time periods of essentially tholeiitic lava has occurred within the last 150 m.y. on a number of continental regions. These occur as a series of relatively thin flows that cover an area up to  $10^6$  km<sup>2</sup>. The total volume of such regions of flood basalt can approach  $10^6$  km<sup>3</sup>, many times the volume of basalts that built the shield volcano Mauna Loa (see Table IIB-11, compiled from CTV, and Head, 1975). These tholeiitic lavas are generally similar to the oceanic tholeiites but are more potassium-rich and show enrichment of light rare earth elements.

A substantial amount of geochemical data has become available for the Columbia River basalt group (Figure IIB-18, from Verhoogen et al., 1970, p. 299). The geology and geochemistry of this area was extensively described by Waters (1961) and further major and minor element abundances were measured by McDougall (1976). The Columbia River basalts are oversaturated tholeiites; the independent

Table IB2-11 Estimated volume of tholeiitic flood basalts.

<u>Location</u>	<u>Estimated volume (<math>\times 10^6 \text{ km}^3</math>)</u>
Decca, India	.7
Columbia and Snake River	.3
Siberia	.25
Parana, Brazil	.20
(Mauna Loa	.05)
Bushveld (plutonic complex)	.10

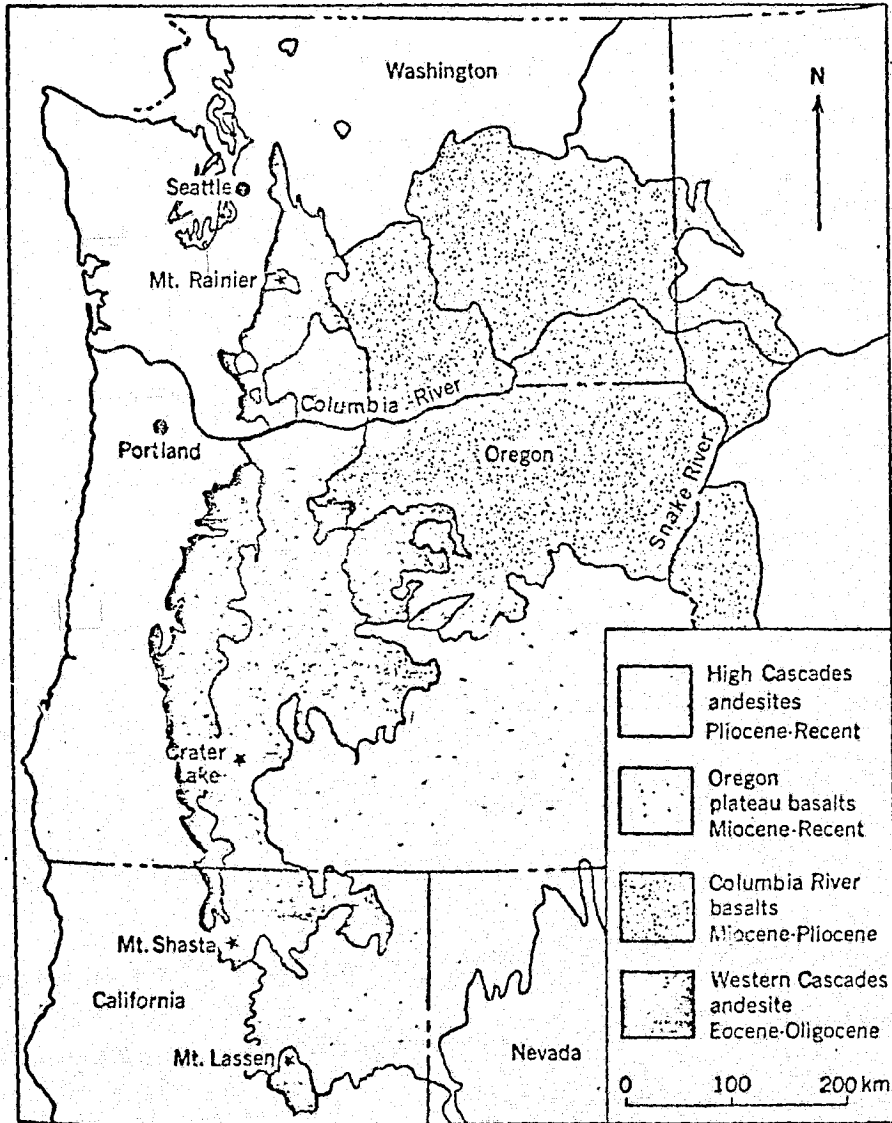


Figure IB 18. Columbia and Snake River Basalts (Verhoogen et al., 1970)

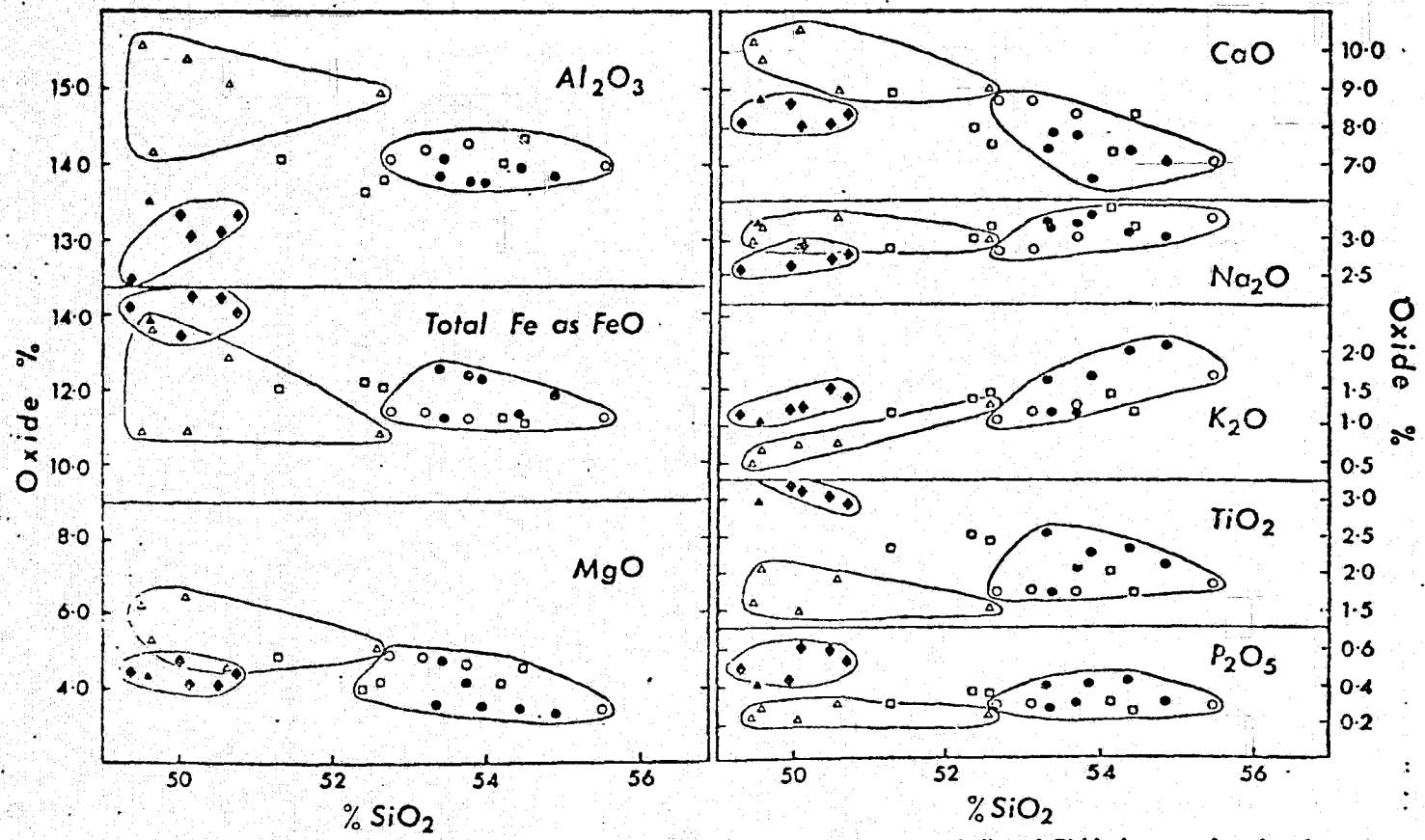
Oregon plateau basalts to the south are much higher in  $\text{Al}_2\text{O}_3$  content. Four distinct basalt sequences of the Columbia River group were described from these studies, all tholeiites: Picture Gorge basalts and Lower, Middle, and Upper Yakima basalts. The Lower and Middle Yakima basalts can be easily distinguished by  $\text{TiO}_2$  content (Siems et al., 1974). Variation diagrams for major element chemistry for these sequences are shown in Figure IB-19 (from McDougall, 1976). Waters showed that these basalts were erupted from fissure systems into a broad basin.

The Sr isotopic studies of McDougall (Figure IV-20) distinguished the Picture Gorge basalts from the Yakima and indicated they were likely to have been derived from upper mantle sources with little modification. The Yakima basalts with higher initial Sr show a correlation of increasing  $^{87}\text{Sr}/^{86}\text{Sr}$  with decreasing age. This pattern could either be the result of progressive crustal contamination or derivation from inhomogeneous upper mantle source materials. McDougall suggests the ultimate source of the Yakima lavas was the upper mantle with partial melting caused by tectonic events related to plate motions in the Pacific NW during that time.

ii. ALKALI BASALT REGIONS (RIFT ZONES?).

Alkali basalts frequently dominate a small continental province, but almost always with a very complex association

▲ Picture Gorge Basalt ▲ Imnaha basalt ▽ Lower Yakima basalt, Imnaha Valley ● Lower Yakima basalt, Grande Ronde sequence ○ Lower Yakima basalt, Crescent Bar ◆ Middle Yakima basalt, Vantage area

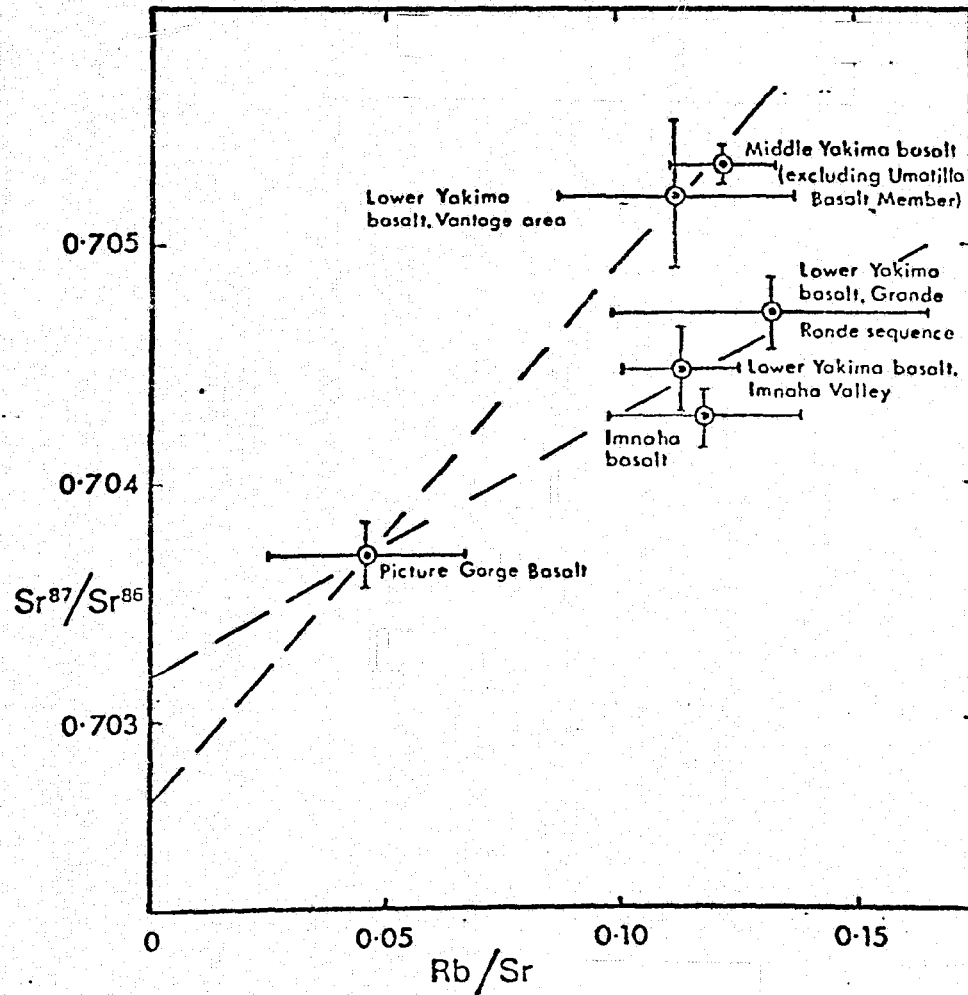


Oxide percentages of major elements plotted relative to SiO<sub>2</sub> for samples from sequences indicated. Fields drawn enclose data from Picture Gorge Basalt, lower Yakima basalt from Crescent Bar and Grande Ronde, and middle Yakima basalt.

Figure IB 19. (McDougall, 1976)

Plot of average initial  $Sr^{87}/Sr^{86}$  ratio relative to  $Rb/Sr$  ratio for each of main sequences of basalt sampled in Columbia River Group. Bars represent one standard deviation for each parameter in each case (not standard deviation of mean).

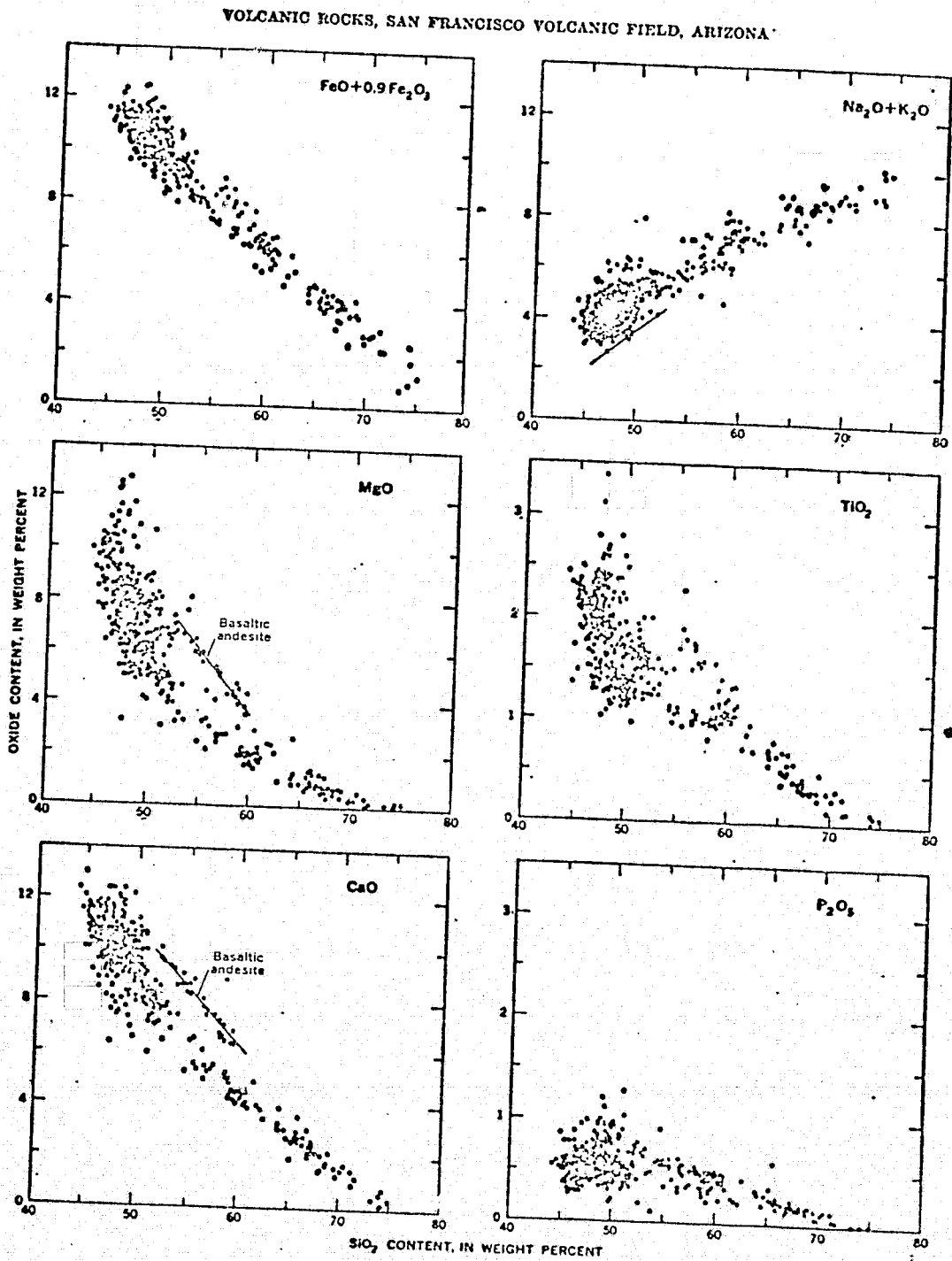
Figure IB 20.  
(McDougal, 1976)



of other magma types. The range of rock types associated with continental alkali basalts is much greater than those associated with oceanic alkali basalts (CTV, p. 505-506). From high pressure experiments, alkali basalts are assumed to have a deep source (between 40-100 km) which is consistent with derivation from continental sub-crustal material. (Only rarely are tholeiitic basalts associated with continental alkali basalts and the question of fractionation relationships between the two basalt types, as is often suggested for oceanic basalts, does not generally occur.)

The tectonic setting for continental alkali basalts is perhaps simply a zone of crustal weakness. The classic example, in complexity, occurs in the East African rift zones. In this region (e.g., CTV, Chapter 10) are found alkali olivine-trachyte-phonolite series, nephelinites, and carbonatites in somewhat random temporal and spatial relationship.

Another continental alkali basalt region, the northern and eastern San Francisco volcanic field in Arizona (Moore et al., 1976), consists of flows of predominantly alkali-olivine basalt, most of which occurred during the last 3 m.y. Associated with these basalts are high alumina basalt, basaltic andesite, andesite, dacite, rhyodacite, and rhyolite. Chemical variation diagrams



—Silica variation diagrams. San Francisco volcanic field. Line on plot of Na<sub>2</sub>O + K<sub>2</sub>O relative to SiO<sub>2</sub> separates alkali olivine basalts from tholeiitic basalts of Hawaii (MacDonald and Katsura, 1964).

Figure IB 21. (Moore et al, 1976)

(Figure IB-21) from Moore et al.) suggest these basalts are consanguineous although no simple evolution pattern is recognized. Strontium isotope values are low for all members of this series (.7026 - .7050) implying derivation from the mantle with little contamination of crustal material. (It is perhaps noteworthy that contemporaneous with some of the San Francisco volcanics 3 m.y. ago were the basalts of the Taos plateau a few hundred kilometers to the east. The Taos basalts are predominantly olivine tholeiites with less extensive alkali basalts and associated rock types [Lipman, 1977].)

In summary, a variety of basalt types exist on the current surface of the earth, most of which are ultimately related to mantle sources. Although chemical fractionation processes and host material contamination often clearly affect compositional trends, it also seems apparent that the general geophysical environment (plate tectonics, type and thickness of crustal material) plays a major role in determining the basalt types erupted on the surface. The source region of basalts for each of the provinces discussed above differ in probable temperature and pressure (depth) environment.

A variety of questions, often with no obvious answers, exist for oceanic basalts as well as those found well within a thick continental plate. Are the volcanics derived from partial melting of crustal or mantle material? What degree of melting is involved? If the magma originated in the mantle, to what degree has it been contaminated by crustal material? How does the sub-continental mantle compare with the oceanic mantle? To what degree is it heterogeneous vertically or laterally? These questions are difficult to answer not only because of the likely complexity of each volcanic event, but also because often not all the relevant information is available (e.g., major element chemistry, minor element chemistry, REE, absolute age, isotopic chemistry, areal and volume estimates, etc.).

Increasingly sophisticated thermal models, experimental petrology results, and isotopic analyses are becoming available and may eventually lead to answers for the above questions. For example, a recent discussion of Nd and Sr isotopic abundances for continental and oceanic rocks (DePaolo and Wasserberg, 1976) provided evidence to indicate that all types of continental igneous rocks (flood basalts to carbonatites) are derived from a mantle reservoir with chondritic REE patterns. Mid-ocean ridge basalts, on the other hand, are derived from a different (shallower?)

ancient reservoir and the source area for ocean island basalts is intermediate between the two.

Probing the lateral and vertical compositional variations of the earth's mantle seems within reach of planetary scientists using a synthesis of current geophysics (thermal models, plate tectonics, seismic internal structure patterns), experimental petrology (melting conditions as a function of P, T, and composition), and basalt geochemistry (major and minor elements, and isotopic abundances). Currently, the terrestrial data are incomplete and scattered in a variety of forms. Such a synthesis is equally applicable for probing the compositional structure of each of the other terrestrial planets, but fewer of the variables are defined.

### 3. Terrestrial Volcanism in Time

A characteristic feature of terrestrial basalts in the preceding discussion is their youth; most basaltic material that has been seriously studied is less than 200 million years old, or very recent in terms of geologic time. Basaltic materials from other solar system objects are much older: the basaltic achondrites are generally much older than ~3000 m.y.; the lunar basalts are between ~3000 and 3900 m.y. old; and the surface age of the largest volcano on Mars is estimated to be at least 1000 m.y. (Soderbloom,

1974). What, then, can be said about the early history of terrestrial volcanism?

Only recently has data become available concerning ancient terrestrial material. A variety of models concerning the early history of the earth have been proposed, many of which have been included in a recent anthology of papers edited by B.F. Windley (1976a). The earth appears to be tectonically unique in the solar system and has recycled much of its surface material perhaps more than once. It has not been defined to what extent the current patterns of terrestrial volcanism, which are closely linked to plate tectonics, were also active in earlier periods of earth history. The Archean rock associations and geochemistry, for example, seem to be similar to modern continental margins and island arcs (Windley, 1976b) implying possibly similar tectonic processes. Burke et al. (1976) describe the early Archean deformation of greenstone terrains as "permobile" deformation, or without rigid behavior during collision. The model preferred by Green (1975) to describe the character of Archean greenstone belts includes a steeper Archean geotherm, a thin lithosphere, and an athenosphere with about 5% melting. His model does not allow subduction of basaltic oceanic crust, but instead postulates such a crust is "scraped off" as the lithosphere is subducted and,

eventually, forms the greenstone belts. A thin Archean lithosphere was also proposed previously by Hart et al. (1970) on the basis of trace element abundances of Archean greenstones. These trace element studies indicated that the Archean upper mantle was not depleted in K, Rb, Cs, Sr, and Ba, unlike the present upper mantle (source region for ocean floor basalts). Glikson (1976), on the other hand, points out that the coherent pattern in the orientation of Archean greenstone belts in Gondwanaland argues against large scale plate dispersions prior to the breakup of Gondwanaland 300 million years ago. He suggests some of the earliest earth volcanics (up to 2.6 AE) may have been "relic terrestrial maria" similar to those common to the moon. An earlier model by Green (1972) suggested a similar link between the Archean greenstone belts and lunar maria.

Although data concerning the terrestrial Archean rocks are incomplete and there seems to be no general consensus on how to interpret these data, they do provide a few extremely significant facts and implications:

(a) Ancient crustal material does exist on earth.

Table IIB-12 (from Jahn and Nyquist, 1976) summarizes the Rb-Sr isochron ages for a variety of Archean rocks. Granitic as well as mafic surface material was formed during the

Age and initial  $Sr^{87}/Sr^{86}$  data for terrestrial Archean rocks

Rock unit	$t = \text{age in b.y.} \pm 2$	$I = (Sr^{87}/Sr^{86})_0 \pm 2$	References	Comments
<b>Minnesota, U.S.A.</b>				
Ely Greenstone	2.69 ± 0.08	0.70056 ± 26	Jahn and Murthy (1975)	$I$ obtained from WR isochron
Newton Lake Fm	2.65 ± 0.11	0.70086 ± 24	Jahn and Murthy (1975)	$I$ obtained from WR isochron
Vermilion Granite	2.70 ± 0.05	0.70041 ± 29	Jahn and Murthy (1975)	$I$ obtained from WR isochron Basic data from Peterman et al. (1972) and Jahn and Murthy (1975)
	2.68 ± 0.10	0.7006 ± 12	Peterman et al. (1972)	$I$ obtained from WR isochron
<b>Northern Light</b>				
Gneiss	2.74 ± 0.10	0.7007 ± 4	Hanson et al. (1971)	$I$ obtained from WR isochron
Saganaga Tonalite	2.71 ± 0.56	0.7010 ± 6	Hanson et al. (1971)	$I$ obtained from WR isochron
Icarus Pluton	2.69 ± 0.48	0.7100 ± 14	Hanson et al. (1971)	$I$ obtained from WR isochron
Giants Range Granite	2.67 ± 0.07	0.7003 ± 19	Prince and Hanson (1972)	$I$ obtained from WR isochron
Morton-Montevideo Gneisses	3.80	0.700	Goldich and Hedge (1974)	Both $t$ and $I$ are best values from an eye-ball fit through six data points
<b>Canada</b>				
<b>Pyroxenes, Superior Prov.</b>	2.70	0.7010 to 0.7014	Hart and Brooks (1974)	Seven age-corrected $I$ values for primary pyroxenes from two extrusive and three intrusive mafic units
<b>Keewatin +</b>				
Coutchiching	2.69 ± 0.08	0.7003 ± 13	Hart and Davis (1969)	$I$ obtained from WR isochron
<b>Chibougamau</b>				
Greenstone belt:				
Dore Lake Complex	2.72 ± 0.50	0.7011 ± 10	Jones et al. (1974)	$I$ obtained from WR isochron
Anorthosite-1		0.70068 ± 6	Jahn (unpublished)	Age-corrected $I$ value; Sr composition measured at JSC; $t = 2.70$ b.y.
Anorthosite-2		0.70150 ± 7	Jahn (unpublished)	Age-corrected $I$ value; Sr composition measured at JSC; $t = 2.70$ b.y.
Chibougamau Pluton	2.56 ± 0.16	0.7007 ± 4	Jones et al. (1974)	$I$ obtained from WR isochron
Meta-meladiorite		0.70133 ± 6	Jahn (unpublished)	Age-corrected $I$ value; Sr composition measured at JSC; $t = 2.6$ b.y.
<b>Yellowknife:</b>				
SE Granodiorite	2.64 ± 0.08	0.7011 ± 16	Green and Baadsgaard (1971)	$I$ obtained from WR isochron
<b>Metavolcanics</b>	2.63 ± 0.16	0.7022 ± 23	Green et al. (1968)	$I$ obtained from WR isochron
<b>South Africa</b>				
Onverwacht:				
Komati	3.50 ± 0.20	0.70048 ± 5	Jahn and Shih (1974)	$I$ obtained from mineral isochron
Anorthositic norite		0.70061 ± 6	Jahn and Shih (1974)	Age-corrected $I$ value ( $t = 3.5$ b.y.)
Anorthosites and komatiites		0.7000 to 0.7006	Allsopp et al. (1973)	Age-corrected $I$ values assuming $t = 3.2$ b.y.
Middle Marker Horizon	3.36 ± 0.07	0.7015 ± 13	Hurley et al. (1972)	$I$ obtained from WR isochron
<b>Rhodesia:</b>				
Bulawayan	3.0-3.1	0.70110 ± 5	Jahn (unpublished)	Age-corrected $I$ value assuming $t = 3.0$ b.y. Measured = 0.70122
Bulawayan limestones		0.7014 to 0.7019	Bell and Blenkinsop (1974)	Measured values; uncorrected
Great Dyke	2.53 ± 0.09	0.7024 ± 8	Davies et al. (1970)	$I$ obtained from WR isochron
<b>West Greenland:</b>				
Amitsoq Gneisses:				
Narsaq Area	3.75 ± 0.09	0.7015 ± 8	Moorbath et al. (1972)	All $I$ 's obtained from WR isochrons;
Qilangarsuit	3.74 ± 0.10	0.7009 ± 11	Moorbath et al. (1972)	Best age = 3.70-3.75 b.y.
Praestefjord	3.69 ± 0.23	0.7001 ± 17	Moorbath et al. (1972)	Best $I = 0.7010-0.7015$
Isua Area	3.70 ± 0.14	0.7011 ± 20	Moorbath et al. (1972)	

Note: All  $I$  values have been adjusted based on NBS-SRM987 Sr Standard = 0.71025 or E & A Sr Standard = 0.7081.

Table IB 12. (Jahn and Nyquist, 1976)

first billion years of earth history and still constitutes a minor part of the earth's crust. Although the earth is a dynamically active planet, it has failed to destroy a few pieces of its earlier history--it only deformed them badly.

(b) The moon, with its ancient surface, and perhaps also Mars and Mercury, provide excellent laboratories for examining the early history of the solar system. All the terrestrial planets may very well have been affected by the same processes in their early history. Further examination of the Moon, Mars, and Mercury in conjunction with investigations concerning the terrestrial Archean environment should eventually allow a unified theory of early planetary evolution to be developed.

## I. BASALT TYPES: AN OVERVIEW

## C. Lunar Basalt Types

Lunar samples have been returned from nine areas on the front side of the moon by the Apollo and Luna landing missions. The locations of these areas are shown in Figure IC-1 and the coordinates are given in Table IC-1 (updated from Taylor, 1975). Six of these landing areas were in mare regions.

Although it has only been eight years since the first samples were returned, it is easy to forget some of the major facts the samples provided immediately. Lunar rocks have their own distinct character--different from terrestrial samples and different from meteorites. The lunar samples contain essentially no water (and no organic matter). The lunar surface is differentiated. The lunar maria are basaltic in composition and igneous in origin. The lunar rocks are old ( $\approx 3.0$  AE).

The returned lunar samples fall into two distinct families: the highland feldspathic rocks and the mare basalts. The light-colored highland rocks comprise over 80% of the lunar surface and have been intensely brecciated by repeated impacts. The highlands are the oldest lunar surface and consist of a variety of recognizable, although often mixed, rock types (summarized by Taylor, 1975, Chap. 5). The high-

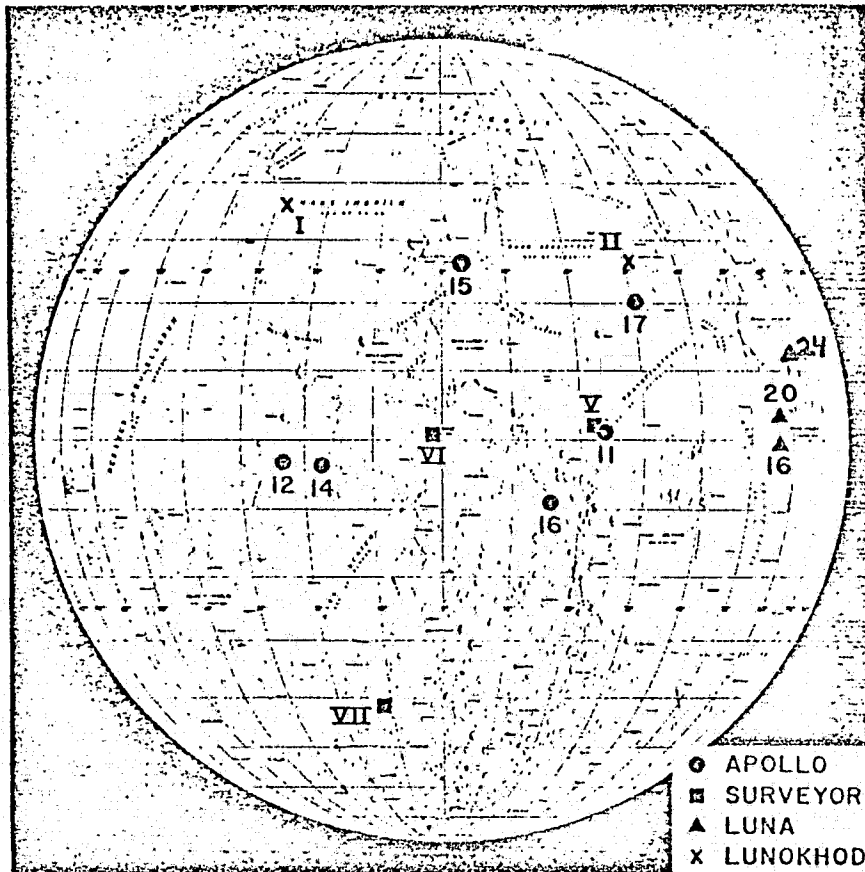


Figure IC1 Apollo, Luna, and Surveyor landing sites.

The Apollo Lunar Landings.

Mission	Landing site	Latitude	Longitude	LVA duration (hours)	Traverse distance (km)	Date
11	Mare Tranquillitatis	0°67'N	23°49'E	2.24	—	July 20, 1969
12	Oceanus Procellarum	3°12'S	23°23'W	7.59	1.35	Nov. 19, 1969
14	Fra Mauro	3°40'S	17°28'E	9.23	3.45	Jan. 31, 1971
15	Hadley-Apennines	26°06'N	3°39'E	18.33	27.9	July 30, 1971
16	Descartes	8°60'S	15°31'E	20.12	27	April 21, 1972
17	Taurus-Littrow	20°10'N	30°46'E	22	30	Dec. 11, 1972

Mission	Landing site	Latitude	Longitude	Date
Luna 16	Mare Fecunditatis	0°41'S	56°18'E	Sept., 1970
Luna 20	Apollonius highlands	3°32'N	56°33'E	Feb., 1972
Luna 24	Mare Crisium	12° 45'N	62° 12'E	Aug. 1976

(B) Russian Lunar Traverse Vehicles.

Vehicle	Landing site	Date	Traverse length
Lunokhod 1	Western Mare Imbrium	Nov., 1970	20 km
Lunokhod 2	Le Monnier Crater, Eastern Mare Serenitatis, 180 km north of Apollo 17 site	Jan., 1973	30 km

Table IC 1. (Taylor 1975)

land crust contains abundant feldspar. Although some of the highland rock types have basaltic compositions (e.g., Fra Mauro basalts), the average composition of the highlands is too low in mafic elements ( $MgO + FeO < 16\%$ ) and too high in  $Al_2O_3$  ( $>20\%$ ) to be considered basaltic. The Fra Mauro basalts, which may have been the products of early near surface partial melting (e.g., Walker et al., 1973), may make up 20% of the crust (Taylor, 1975, p. 252). The impact history of the lunar highland crust, however, has obscured any early magmatic history and the possibility of lunar highland basalts will not be discussed here.

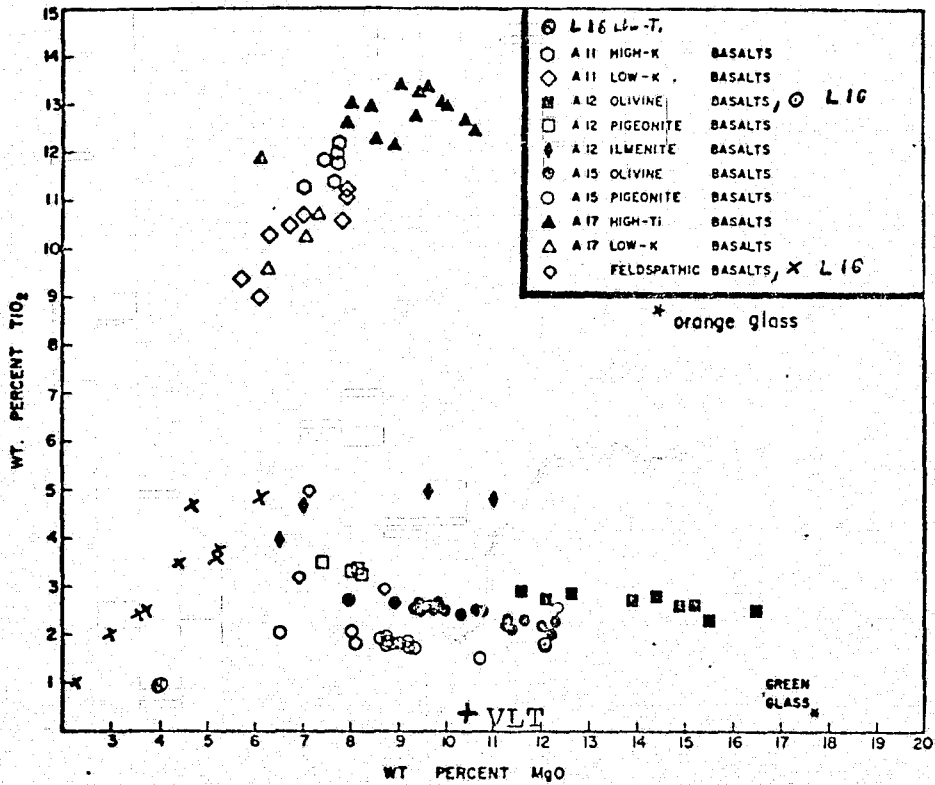
The dark mare basalts are perhaps less complicated. They are igneous rocks that fill the crustal lowlands, but are themselves old by terrestrial standards (3.8 - 3.0 AE). The lunar mare basalt types identified in the returned samples are discussed in the following sections along with the implications these basalts have for lunar evolution. Remote sensing techniques (Sections II-V) are then used to extend this information to unsampled areas in the attempt to understand the global surface geochemistry of the moon.

1. Major Element Geochemistry of Lunar Mare Basalt Types

An immense literature on lunar samples studies exists, largely in the 7 x 3 Proceedings Volumes, and cannot be

reviewed here in detail. Much of the discussion of lunar basalt types that follows is derived from two excellent recent reviews which include extensive references. Taylor (1975) has summarized and synthesized in a readable book much of the geological, geochemical and geophysical data for the moon that has resulted from the Apollo missions. The chemistry, mineralogy and petrology of lunar mare basalts is extensively reviewed by Papike et al. (1976). A variety of lunar basalt type classifications exist in the literature. The classification by Papike et al., which incorporates many of these previous classifications, is based primarily on major element chemistry and is used as the basis for the discussion presented here. A recent pamphlet by McGee et al. (1977) provides a useful introduction to the petrology and geochemistry of representative lunar igneous rocks.

One of the most useful chemical factors that distinguishes various mare basalt types is  $TiO_2$  content. Most of the sampled lunar mare basalts can be placed into one of two distinct chemical groups, high- and low-titanium basalts, as shown in the variation diagram of Figure IC-2 (from Papike et al., 1976; updated). To date no samples have been obtained for intermediate titanium basalts. There is also an apparent correlation between measured crystallization



TiO<sub>2</sub> versus MgO (wt %) variation diagram for mare basalt major-element chemistry.

Figure IC 2. (update of Papike et al., 1976)

ages and  $\text{TiO}_2$  content--the high-Ti basalts (Apollo 11, 17) are older than the low-Ti basalts (Apollo 12, 15, Luna 16). The bulk chemistry of these two major groups does not allow a petrogenetic relation through fractional crystallization, a distinction confirmed by melting experiments (Kesson, 1975; Walker et al., 1975).

The mare basalt samples can be further subdivided into 10 distinct groups within which the major element variations have likely been caused by near surface fractionation. Major element chemistry for these basalt groups are provided in Table IC-2 (from Papike et al., 1976). The average modal mineralogy for nine of these groups is presented in Table IC-3 and summarized in Figure IC-3 (from Papike et al., 1976).

a. High-Ti

The high titanium basalts of Apollo 11 fall into two groups which can be distinguished largely by  $\text{K}_2\text{O}$  content. The low-K basalts ( $\text{K}_2\text{O} \sim .06-.09\%$ ) are somewhat older than the high-K ( $\text{K}_2\text{O} \sim .30\%$ ). The petrological variations observed in a variety of Apollo 11 low-K basalts are consistent with these basalts representing the upper portion of a single cooling unit. Initial  $^{87}\text{Sr}/^{86}\text{Sr}$  data for these two high-Ti groups indicate they cannot be derived from the same source region (Papanastassiou and Wasserberg, 1971).

Apollo 11											
	Low K			High K			Apollo 17				
	10003	10050	Avg (8)	10017	10057	Avg (6)	75055	70215	71055	74275	Avg (30)
SiO <sub>2</sub>	39.76	40.62	40.67	40.64	39.79	40.37	41.27	37.79	38.14	38.43	38.84
TiO <sub>2</sub>	10.50	9.61	10.18	11.78	11.44	11.77	10.17	12.97	13.41	12.66	12.35
Al <sub>2</sub> O <sub>3</sub>	10.43	10.87	10.40	7.98	10.84	8.84	9.75	8.85	8.62	8.51	8.84
FeO	19.80	16.51	18.68	19.65	19.35	19.28	18.24	19.66	19.20	18.25	18.94
MnO	0.30	0.26	0.27	0.24	0.20	0.24	0.29	0.27	0.26	0.25	0.27
MgO	6.69	7.82	6.92	7.68	7.65	7.56	6.84	8.44	9.04	10.26	8.52
CaO	11.13	12.65	11.70	10.65	10.08	10.59	12.30	10.74	10.77	10.38	10.80
Na <sub>2</sub> O	0.40	0.35	0.41	0.51	0.54	0.52	0.44	0.36	0.31	0.37	0.32
K <sub>2</sub> O	0.06	0.06	0.07	0.29	0.32	0.31	0.09	0.05	0.06	0.07	0.05
P <sub>2</sub> O <sub>5</sub>	0.12	0.07	0.09	0.16	0.17	0.17	0.07	0.09	0.08	0.07	0.06
S	0.18	0.11	0.16	0.22	0.22	0.22	0.19	0.18	0.17	0.14	0.17
Cr <sub>2</sub> O <sub>3</sub>	0.26	0.35	0.29	0.36	0.36	0.36	0.27	0.41	0.41	0.64	0.49
Mg (Mg + Fe)	0.48	0.48		0.41	0.41		0.40	0.43	0.46	0.50	

Avg (8) indicates an average of eight samples. References and analyses used for averages are available on request

Apollo 12											
	Olivine			Pigeonite		Ilmenite			Aluminous Basalts		
	12009	12040	Avg (9)	12021	Avg (4)	12022	12064	Avg (4)	12038	14053	Luna 16
45.03	43.88	44.32	46.68	46.46	42.77	46.30	44.47	46.83	46.22	43.80	
2.90	2.45	2.65	3.53	3.35	4.85	3.99	4.63	3.24	2.93	4.90	
8.59	7.27	8.03	10.78	10.38	9.08	10.73	9.76	12.48	12.99	13.65	
21.03	21.09	21.11	19.31	19.72	21.75	19.89	20.78	17.76	16.95	19.35	
0.28	0.27	0.28	0.26	0.27	0.25	0.27	0.27	0.25	0.26	0.20	
11.55	16.45	14.07	7.39	7.94	11.01	6.42	8.52	6.86	8.68	7.05	
9.42	8.01	8.61	11.38	11.03	9.47	11.78	10.78	11.49	11.15	10.40	
0.23	0.17	0.22	0.31	0.28	0.38	0.29	0.32	0.65	0.44	0.33	
0.06	0.05	0.06	0.07	0.07	0.07	0.07	0.07	0.07	0.10	0.15	
0.07	0.06	0.08	0.09	0.11	0.13	0.04	0.10	0.14	0.11	0.12	
0.06	0.04	0.06	0.07	0.07	0.08	0.07	0.08	0.07	0.13	0.17	
0.55	0.63	0.63	0.40	0.47	0.56	0.37	0.42	0.31	0.40	0.28	
0.49	0.58		0.41		0.47	0.38		0.41	0.48	0.39	

Apollo 15						
	Olivine			Pigeonite		
	15119	15555	15650	Avg (19)	15499	15076
45.23	44.57	45.33	44.98	47.75	48.44	47.98
2.64	2.10	2.25	2.41	1.81	1.92	1.82
9.24	8.69	8.17	8.80	9.23	8.97	9.46
22.25	22.53	22.17	22.37	20.37	20.33	20.13
0.31	0.29	0.26	0.30	0.29	0.29	0.28
8.93	11.36	12.27	10.42	9.03	8.61	8.74
10.55	9.40	8.98	9.79	10.40	10.52	10.54
0.30	0.27	0.27	0.28	0.34	0.34	0.31
0.05	0.04	0.06	0.05	0.06	0.07	0.06
0.09	0.06	0.12	0.08	0.08	0.07	0.07
0.05	0.06	0.07	0.07	0.07	0.08	0.06
0.47	0.61	0.50	0.57	0.52	0.31	0.47
0.42	0.47	0.50		0.44	0.43	

Table IC 2. Major element chemistry of lunar basalt types. (Papike et al., 1976)

Lunar basalt type Textural Sequences

Group	Modal Mineralogy*				No. Averaged	Textural Sequence†
	cpx	ol	plag	opaq		
Apollo 11 low K	44-55 (49)	0-7 (2.2)	24-38 (31)	10-18 (14)	14	10020, 10045, 10062, 10003, 10050, (10044, 10058, 10047)
Apollo 17 low K	44-52 (48)		29-33 (31)	13-15 (14)	2	(75035, 75055)
Apollo 11 high K	47-59 (53)	0-1 (0.1)	15-27 (20)	15-26 (19)	9	10049, 10022, 10057, 10017, 10024, 10072
Apollo 17 very high Ti	42-51 (47)	1-10 (4.5)	14-29 (23)	18-30 (24)	5‡	74235, 70215, 74275, 71055, 75075, 70017, 70035
Apollo 12 ilmenite	55-61 (59)	0-20 (3.5)	12-33 (25)	7-11 (9)	6	12022, 12063, 12051, 12064
Apollo 12 olivine	32-64 (53)	36-11 (20)	5-27 (19)	3-14 (7)	14‡	12009, 12004, 12002, 12075, 12018, 12020, 12040, 12035
Apollo 12 pigeonite	62-71 (68)	0-4 (1.4)	17-31 (21)	5-12 (9)	7	12052, 12053, 12065, 12021
Apollo 15 olivine	61-64 (63)	5-9 (7)	22-27 (24)	4-6 (5.5)	3	15545, 15556, 15016, 15555
Apollo 15 pigeonite	46-70 (61)		24-48 (33)	3-4 (3.6)	3‡	15597, 15595, 15499, 15476, 15495, 15475, 15076, 15085, 15058, 15065

Values in parentheses are averages.

\*Clinopyroxene is abbreviated cpx; olivine, ol; plagioclase, plag; and opaques, opa.

†From fine at left to coarse at right.

‡Does not include vitrophyres.

MODAL MINERALOGY AND BULK COMPOSITION

Table IC 3. (Papike et al., 1976)

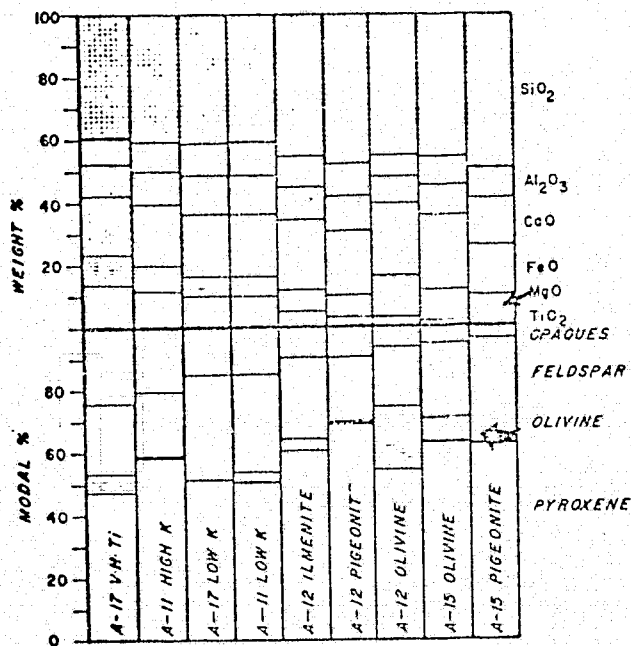


Diagram showing correlation between major-element chemistry and modal mineralogy for major mare basalt groups. Histogram is ordered left to right in terms of decreasing TiO<sub>2</sub> content. An obvious correlation between TiO<sub>2</sub> (wt %) and modal abundance of opaques is present.

Figure IC 3. (Papike et al., 1976)

There are also two major basalt types at Apollo 17. The minority low-K basalt samples are very similar in composition and in age to the Apollo 11 low-K basalts. The Apollo 17 very high-Ti samples are distinguished from other high titanium basalts by their higher  $\text{TiO}_2$ , FeO, and MgO content. The major element variations within the very high-Ti suite of rocks can be interpreted in terms of multiple crystal fractionation. Both Apollo 11 and 17 low-K basalt compositions could possibly have been produced by near surface fractionation from a magma similar to that which formed the very high-Ti basalts.

b. Low-Ti

Three major groups of low titanium basalts can be distinguished at the Apollo 12 site: olivine basalts, pigeonite basalts, and ilmenite basalts. A few samples of feldspathic basalts are also found. These three basalt types can be distinguished in the chemical variation diagrams of Figure IC-4 (from Papike et al., 1976; Rhodes et al., 1977). The major element variations for the Apollo 12 olivine basalts fall along olivine control lines indicating the variations are due largely to the loss or gain of olivine. The coarser-grained samples are interpreted as olivine cumulates. The Apollo 12 pigeonite basalts contain large zoned phenocrysts or pyroxene. They show lit-

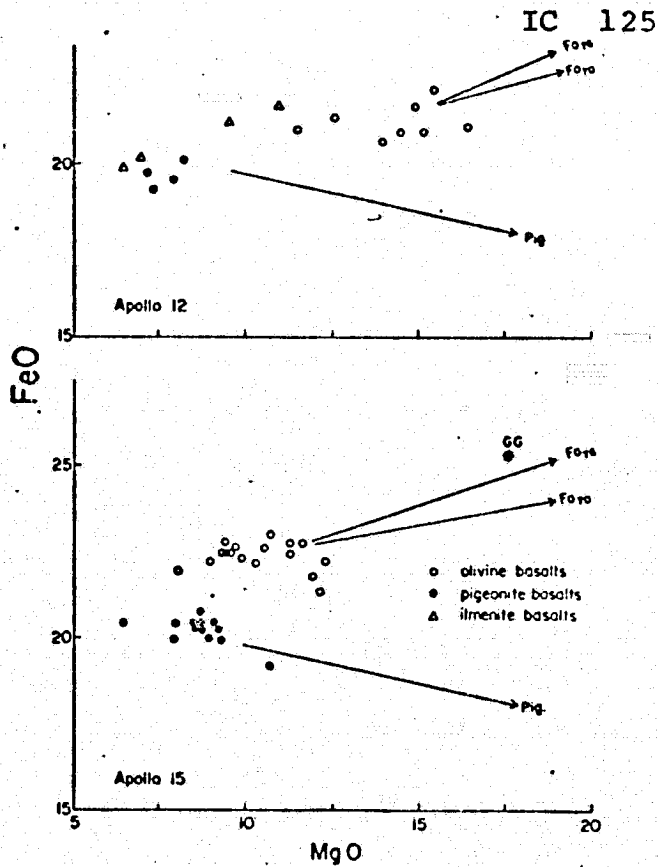


Fig. 4a. FeO versus MgO (wt %) variation diagrams for low-Ti mare basalts. Olivine and pigeonite control lines are indicated. G.G. refers to Apollo 15 emerald green glass.

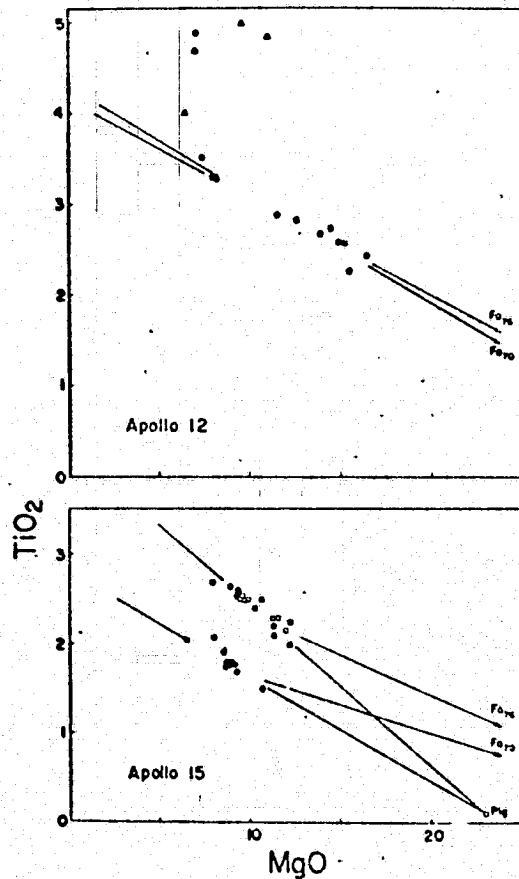


Fig. 4b. TiO<sub>2</sub> versus MgO (wt %) variation diagrams for low-Ti mare basalts. Olivine and pigeonite control lines are indicated.

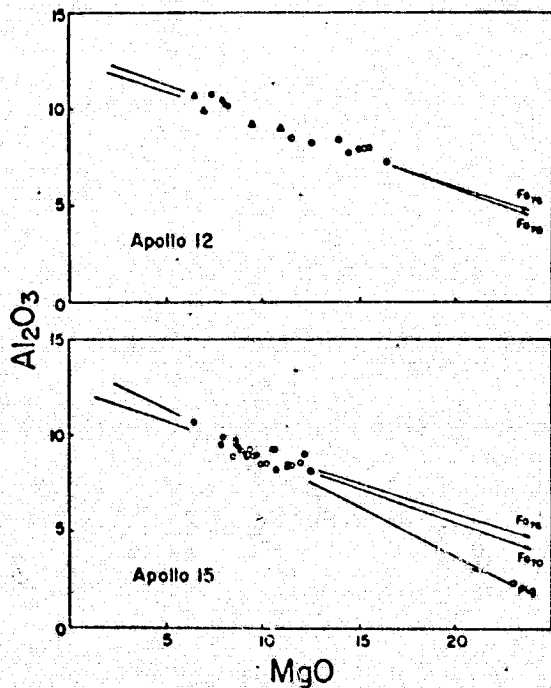


Fig. 4c. Al<sub>2</sub>O<sub>3</sub> versus MgO (wt %) variation diagrams for low-Ti mare basalts. Olivine and pigeonite control lines are indicated.

Figure IC 4. (Papike et al., 1976)

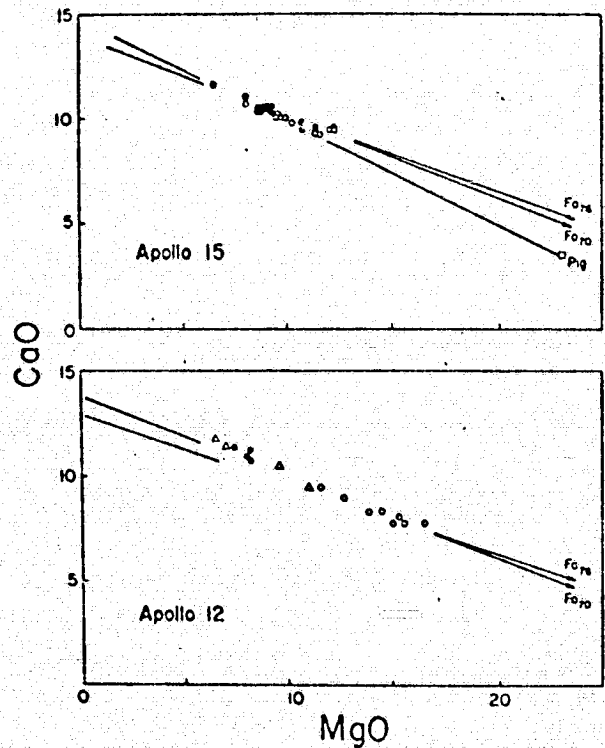
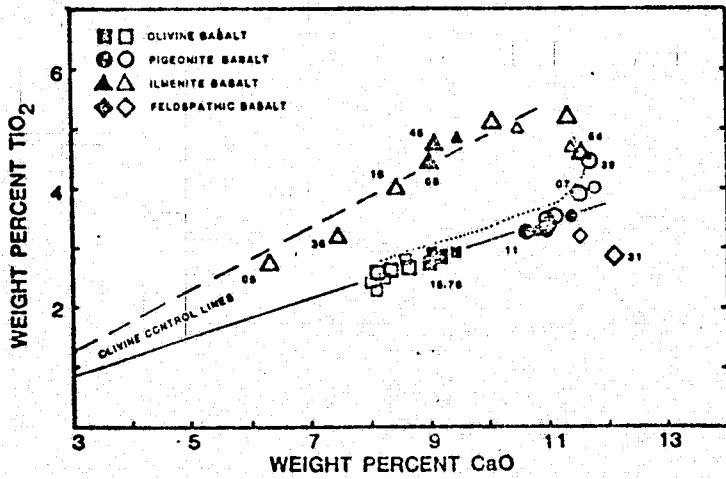


Fig. 4d. CaO versus MgO (wt %) variation diagrams for low-Ti mare basalts. Olivine and pigeonite control lines are indicated.



- TiO<sub>2</sub> vs CaO for Apollo 12  
mare basalts.

The liquid line of descent for  
12002(12) is shown as a dotted  
line.

Figure IC 4e. (Rhodes et  
al., 1977)

the evidence of near surface fractionation although the relation of their major element chemistry to that of the olivine basalts (Figure IC-4) indicates they both could have been derived from similar parental magma. The Apollo 12 ilmenite basalts have recently been extensively studied (Rhodes et al., 1977, Nyquist et al., 1977). These results indicate that the major element compositional variations within the ilmenite basalt group, can be attributable to near surface fractionation involving olivine + ilmenite (Figure IC-4e). The rare earth and isotopic data indicate the ilmenite basalt parental magma is probably derived from a distinctly different source region than that for the olivine-pigeonite basalts. Remote sensing data (see Section III) show two surface units in the Apollo 12 region with the landing site being within a few kilometers of a unit of slightly higher  $TiO_2$  content.

The mare basalts of Apollo 15 include two basalt types similar to those at Apollo 12 but lower in  $TiO_2$ : olivine basalts and pigeonite basalts. The Apollo 15 olivine basalts are high in FeO (Figure 4a). The major element variations are consistent with a limited amount of olivine fractionation, and, together with textural evidence, indicate the samples are not olivine cumulates but are probably samples from the upper portion of cooling units.

The composition of the Apollo 15 pigeonite basalts do not fall on the same olivine control lines as the olivine basalts, although the two basalt types could be products of fractionation of a single primary magma not sampled. The major element variations of the pigeonite basalts are consistent with a limited amount of pigeonite (and possibly olivine and chrome spinel) fractionation.

c. Other (Feldspathic, VLT, Luna 24) basalts

It is possible, if not probable, that some of the minor rock types found in the lunar samples are, in fact, regionally important. Three general groups in particular may be significant: feldspathic basalts, very low titanium (VLT) basalts, and the recently returned Luna 24 basalts.

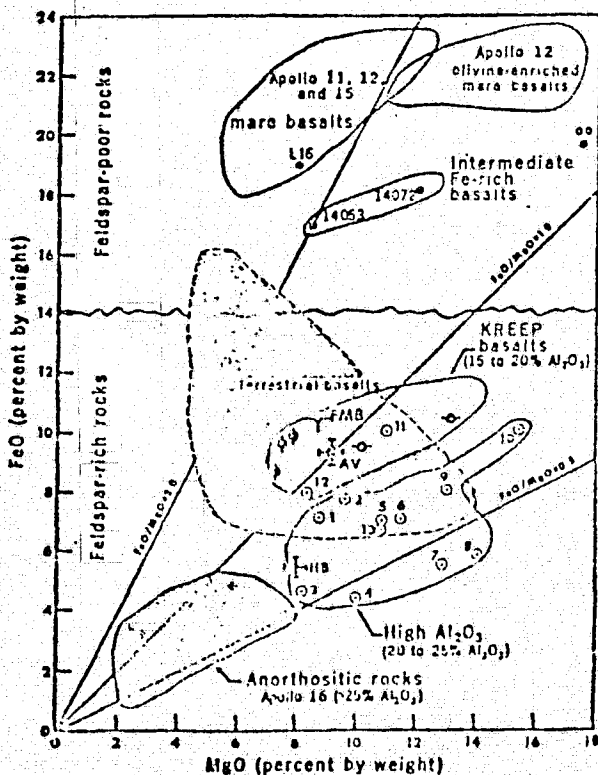
The feldspathic basalts (high aluminum) are exemplified by a heterogeneous and perhaps unrelated number of samples found as minor components at several landing sites (Hubbard et al., 1973; Ridley, 1975, Irving, 1975). Their major element chemistry (Table IC-4, from Ridley, 1975) and petrographic relationships indicate they are indeed basaltic. Their high MgO + FeO content (Figure IC-5, from Hubbard et al., 1973) and medium TiO<sub>2</sub> (Table IC-4) distinguishes them from possible highland melt rocks. They are distinguished from most other mare basalts by their

Major-element chemistry of aluminous mare basalts.

	1	2	3	4	5	6	7	8
SiO <sub>2</sub>	45.9	45.2	46.3	46.82	45.50	43.80	40.2	43.7
TiO <sub>2</sub>	7.3	2.57	2.79	2.33	4.04	4.90	7.2	3.3
Cr <sub>2</sub> O <sub>3</sub>	0.2	0.51	0.37	0.32		0.28		
Al <sub>2</sub> O <sub>3</sub>	14.4	11.1	13.7	13.81	13.95	13.65	13.2	13.2
MnO				0.26	0.26	0.20		
FeO	10.7	17.8	17.0	15.92	17.77	19.35	16.5	16.5
MgO	6.7	12.2	8.54	6.30	5.95	7.05	8.5	9.9
CaO	11.3	9.84	11.2	11.57	11.96	10.40	12.0	11.1
Na <sub>2</sub> O	1.0	0.32	0.44	0.80	0.63	0.38	0.3	0.2
K <sub>2</sub> O	0.1	0.08	0.11	0.25	0.21	0.15	0.10	0.16

- (1) Average composition of basaltic clasts in 14063, broad beam analysis.
- (2) XRF analysis of 14072. Hubbard *et al.* (1972).
- (3) XRF analysis of 14053. Hubbard *et al.* (1972).
- (4) Average 7 vitrophyric and variolitic clasts. 14321. Grieve *et al.* (1975).
- (5) Basalt fragment in Luna 16 soil. Albee *et al.* (1972).
- (6) Luna 16 basalt. Vinogradov *et al.* (1971).
- (7), (8). Preferred glass compositions in Apollo 11 and 12 soils, respectively. Reid *et al.* (1972).

Table IC 4. (Ridley, 1975)



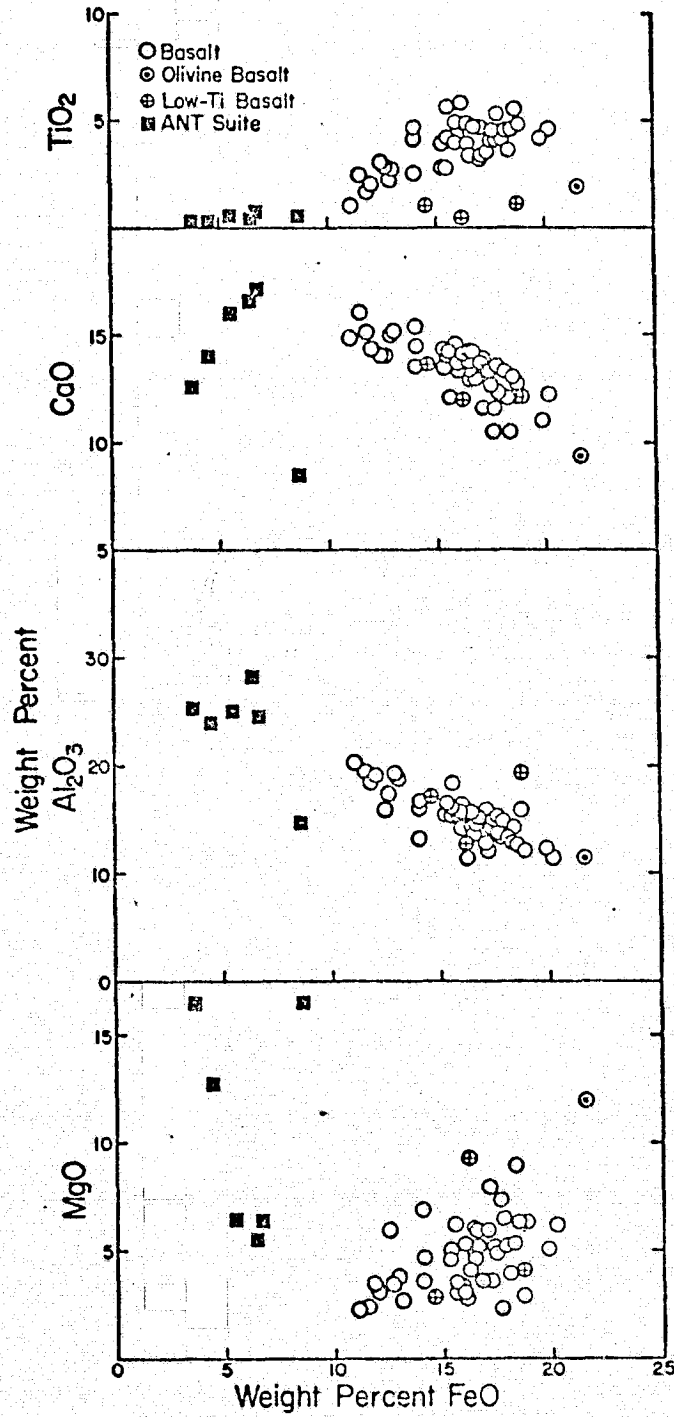
The FeO, MgO, and Al<sub>2</sub>O<sub>3</sub> contents of different lunar rock types: +, rocks with more than 25 percent Al<sub>2</sub>O<sub>3</sub>; ⊙, rocks with 20 to 25 percent Al<sub>2</sub>O<sub>3</sub>; ⊚, Apollo 15 KREEP basalts; ⊙, Apollo 16 KREEP basalts; AV, mean KREEP composition from broad-beam microprobe analyses of individual fragments; HB, "highland basalt"; FMB, mean composition of Fra Mauro basalt glasses; GG, Apollo 15 green glass; L16, Luna 16 basalt. Points labeled 1, 2, 3, . . . correspond to samples 66095, 61156, 60335, 61016 (dark), 14073, 14083 (light), 15273.4.2, 67295, 22007, 15445.17, 22006, 14310, and 15273.4.3, respectively.

Figure IC 5. (Hubbard *et al.*, 1973)

higher  $\text{Al}_2\text{O}_3$  content, although they are frequently simply grouped with the low-Ti basalts. The few that have been dated demonstrate the heterogeneity of this 'group': 14053 is about 3.95 AE (Papanastassiou and Wasserberg, 1971; Turner et al., 1973) and predates most mare basalts, whereas a sample from Luna 16 (B-1) is about 3.45 AE (Papanastassiou and Wasserberg, 1972; Huenke et al., 1972).

Recently a comprehensive study of the mineralogy and petrology of 58 lithic fragments from the Luna 16 core has been made (Kurat et al., 1976). Three separate basalt types were identified in the samples. Fragments of the main type of Lunar 16 basalt show compositional variations (Figure IC-6, from Kurat et al.) indicative of near surface fractionation involving plagioclase, olivine, and ilmenite. These Luna 16 basalts are distinct among the returned samples and seem to form a class by themselves with low FeO and MgO, and high  $\text{Al}_2\text{O}_3$ , FeO/MgO, CaO and alkali content. Remote sensing data (Pieters and McCord, 1975) suggest the Luna 16 basalts are part of a family of basalts in the eastern maria.

A few fragments of a very low titanium basalt have been recently discovered in the Apollo 17 drill core (Vaniman and Papike, 1977). Preliminary major element chemistry of this basalt shown in Table IC-5 (Papike, per-



Bulk compositions of 58 igneous lithic fragments from Luna 16.

Figure IC 6. (Hurat et al., 1976)

March 12, 1977

PRELIMINARY Data on the  
composition of Very Low Titanium

## Mare Basalt (Apollo 17)

	phaneritic fragment <sup>1</sup> PTS 70008370 (1.75 mm fragment)	phaneritic fragment <sup>1</sup> PTS 70008356 (2.2 mm fragment)	vitrophyre <sup>2</sup> PTS 70007328
SiO <sub>2</sub>	48.74	48.03	48.75
Al <sub>2</sub> O <sub>3</sub>	11.40	11.25	10.04
TiO <sub>2</sub>	0.50	0.36	0.69
FeO	18.98	18.23	17.54
MnO	0.26	0.26	0.30
MgO	9.45	11.00	11.79
CaO	10.15	10.16	9.35
Na <sub>2</sub> O	0.15	0.15	0.06
K <sub>2</sub> O	0.04	0.01	0.02
Cr <sub>2</sub> O <sub>3</sub>	0.60	0.60	0.69
sum	100.27	100.10	99.63
Fe/ (Fe+Mg)	.53	.48	.46
CaO/Al <sub>2</sub> O <sub>3</sub>	.89	.90	.93

<sup>1</sup>recombination of microprobe mineral data using modal analyses obtained by counting 1000 points in each fragment. Averages of many probe analyses (e.g., 200 pyroxene spots in the fragment from PTS 70008356) were used to account for mineral zonation.

<sup>2</sup>recombination of microprobe mineral and glass analyses using modal data obtained by counting 1000 points in the vitrophyre.

<sup>3</sup>average of eight overlapping 50 micron spots within a single vitrophyre (about 85% glass, 15% small crystals up to 10 microns in diameter)

Table IC 5. (Papike, 1977, preliminary communication)

sonal communication) indicates they clearly are mare basalts and not highland melt rocks. Their very low  $\text{TiO}_2$  content (which is comparable to the Apollo 15 green glass) distinguishes them from other mare basalts. The regional source area of these basalts has not yet been identified.

On August 28, 1976, the Soviet Luna 24 automatic probe obtained a core sample from SE Mare Crisium and returned it to earth. Preliminary analyses of some of the basalt fragments indicate a low titanium basalt ( $\sim 1\%$   $\text{TiO}_2$ ) with high aluminum content and low KREEP (Barsukov et al., 1977). Although it is still uncertain which unit in Mare Crisium was sampled (see Section III E-1), this low-Ti, high-Al basalt may be unique among the samples.

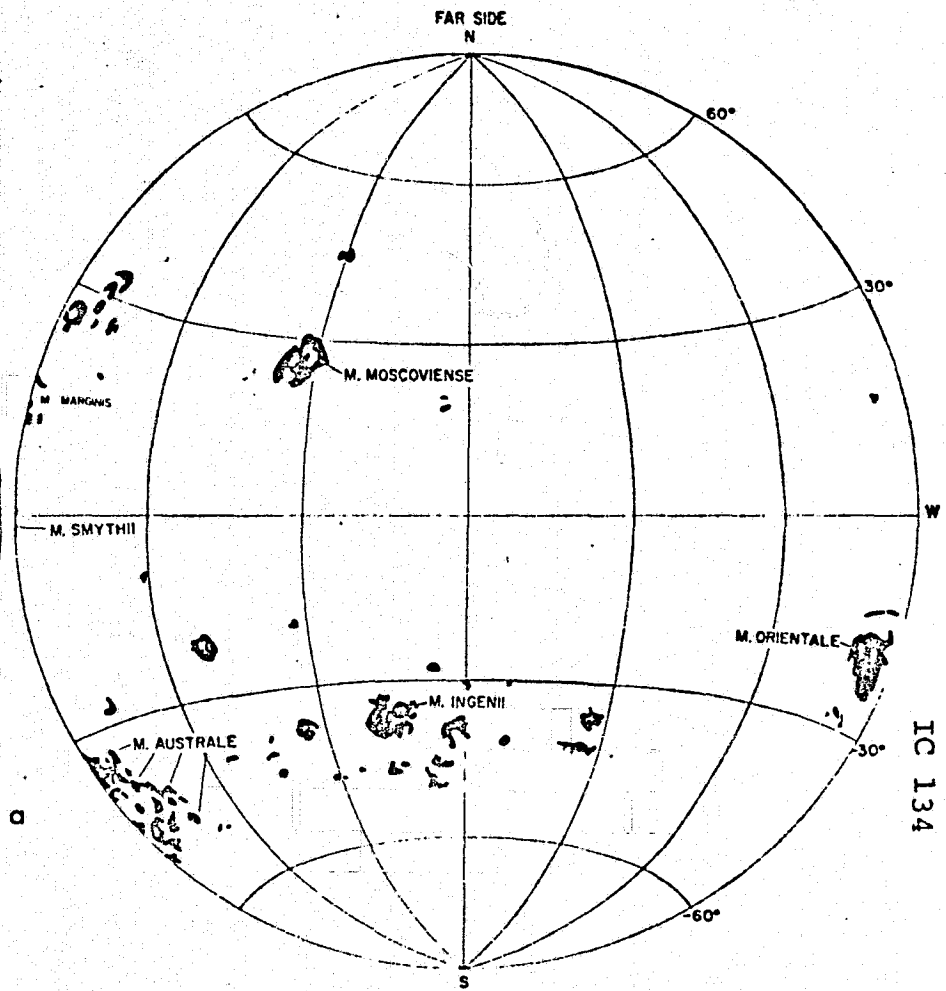
## 2. Role of Basalts in Current Understanding of the Moon

### a. Comparison with terrestrial basalts

About 17% of the lunar surface is covered with mare basalts (Head, 1975, 1976) compared to perhaps >75% for the present terrestrial (e.g., oceanic) surface. The distribution of mare basalts is shown in Figure IC-7 (from Head, 1976). The style of volcanic eruption is dramatically different for the two planets. Most of terrestrial volcanic deposits are associated with recent plate tectonics, whereas the moon has maintained a rigid lithosphere for at least



Distribution of mare deposits  
denotes mare; O, oceanus; S, sinus; P, palus; L, lacus.



on the lunar near side and far side (Lambert equal area projection). (a) M

IC 134

Figure IC 7. (Head, 1976)

3.5 AE. The closest recent terrestrial analogue to lunar volcanic deposits are the flood basalts and Hawaiian lavas although the origin of even these terrestrial basalts is likely to be tied to terrestrial tectonics. A summary of estimated volumes for lunar, terrestrial and Martian volcanics is provided in Table IC-6 (compiled from Head, 1975; CTV).

In major element chemistry lunar basalts are strongly enriched in FeO and in some cases  $TiO_2$  and depleted in  $Al_2O_3$  with respect to most terrestrial basalts (Figure 8, from Gast, 1972).  $SiO_2$  content for lunar basalts is low with respect to terrestrial values (Table IC-2). The Hi-Ti basalts are not strictly basalts unless their  $SiO_2$  content is recalculated for a bulk chemistry with lower amounts of  $TiO_2$ . The ranges of CaO and MgO content are roughly comparable for lunar and terrestrial basalts. Both lunar basalts and basaltic achondrites are depleted in  $K_2O$  and  $Na_2O$  with respect to terrestrial basalts (Figure IC-9, from Gast). The composition of lunar feldspars reflect this low K and Na content (Figure IC-10, from Papike et al., 1976).

The lunar basalts are in extremely reduced state and contain no water of lunar origin. Almost all iron occurs as  $Fe^{2+}$  with up to a few tenths of a percent being reduced

Table IC-6 Estimated volumes of basalt.  
(Head, 1975, CTV)

<u>Location</u>	<u>Estimated volume (x10<sup>6</sup>km<sup>3</sup>)</u>
Earth: Decca, India	.7
Columbia and Snake River	.3
Siberia	.25
Parana, Brazil	.20
(Mauna Loa	.05)
Bushveld (plutonic complex)	.10
Mars: Olympus Mons	2.6
Moon: Tranquillitatis	.26
Fecunditatis	.13
Nectaris	.08
H <sub>2</sub> , h <sub>2</sub> Imbrium	.04
Circular maria	5.2
Whole moon basalts	10.

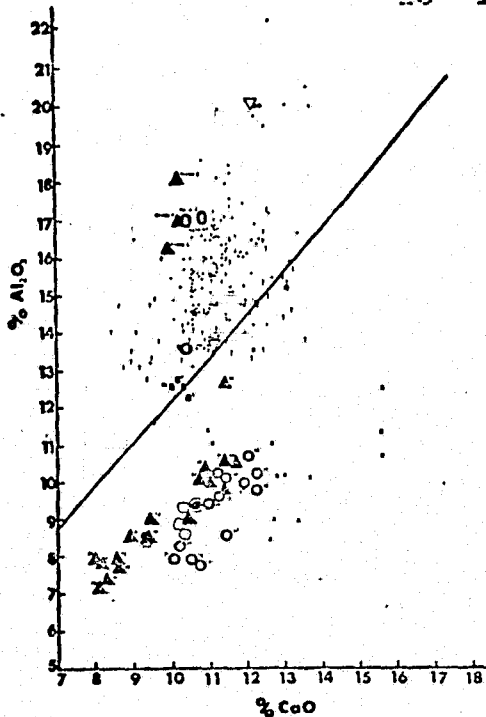
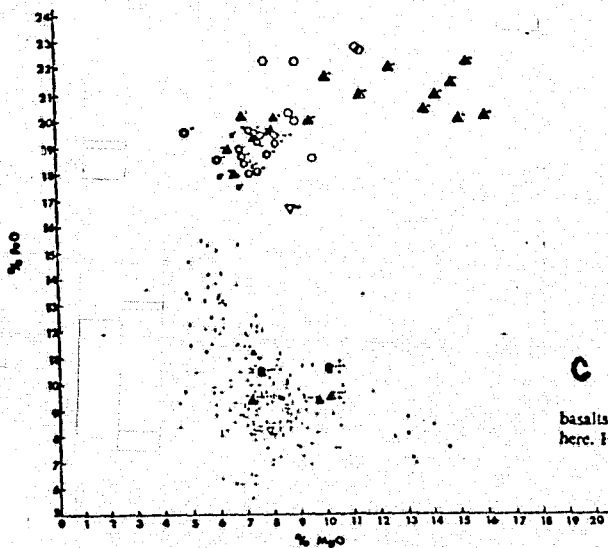


Figure IC 8. (Gast, 1972)

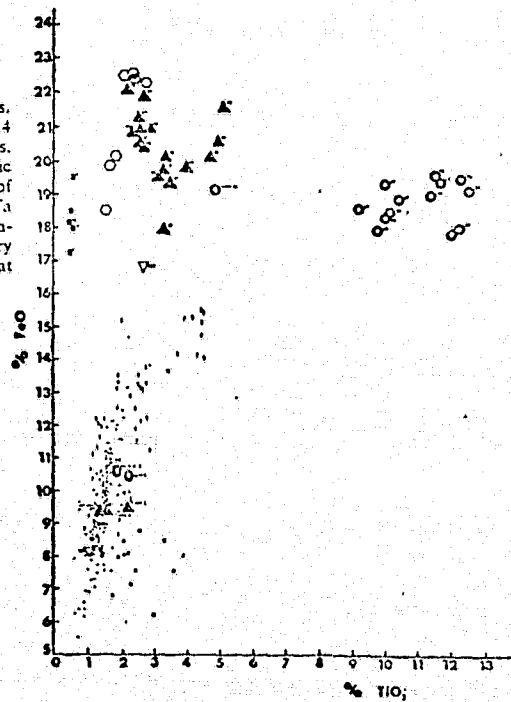
a

CaO and Al<sub>2</sub>O<sub>3</sub> contents of lunar basalts, basaltic achondrites, and terrestrial basalts. Symbols are as follows: ○ Apollo 11 basalts, ▲ Apollo 12 basalts, □ KREEP, ◇ Apollo 14 samples, ▽ Apollo 15 basalts, ◻ basaltic achondrites, • oceanic ridge basalts, † Icelandic basalts, and × post-erosional Hawaiian basalts. The solid line shows the locus of chondrites and basaltic achondrites. Both terrestrial and lunar basalts appear to divide into groups that fall on either side of this line; i.e. the Al-rich (ORB on Earth and non-mare basalt on the Moon) group with an Al/Ca ratio greater than the meteorite ratio and the Al-poor rocks with low Al/Ca ratios (nepheline normative basalts on Earth and mare basalts on the Moon) lower than the meteorite ratio. Given the very plausible assumption that the Al/Ca ratio of both planets is meteoritic, this observation suggests that two distinct petrogenetic fractionation trends are involved in producing liquids with these compositions.



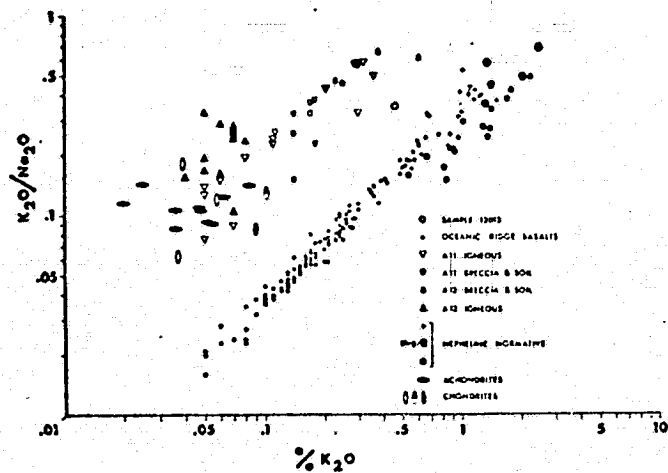
b

FeO and MgO contents of lunar basalts, basaltic achondrites, and terrestrial basalts. Symbols same as Figure a. The three groups of terrestrial basalts shown here were chosen to illustrate the extremes in total Fe (as FeO), FeO, Al<sub>2</sub>O<sub>3</sub>, and TiO<sub>2</sub> contents found in terrestrial basalts. A much more extensive compilation - e.g. Manson (1967) - of terrestrial basalts would still show a very marked gap between the total Fe content of lunar and terrestrial basalts. The high MgO Apollo 12 basalts are probably the result of olivine (Fa 68-72) accumulation.



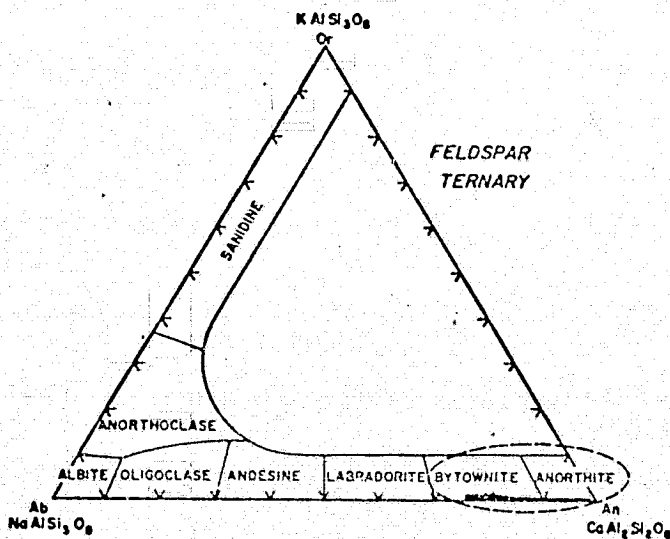
c

Total Fe (as FeO) and TiO<sub>2</sub> contents of lunar basalts, basaltic achondrites, and terrestrial basalts. Symbols same as Figure a. The extreme TiO<sub>2</sub> contents of Apollo 11 basalts are clearly shown here. However, other mare basalts (Luna 16) also tend to have relatively high TiO<sub>2</sub> contents relative to terrestrial basalts.



$K_2O$  and  $Na_2O$  content of terrestrial and lunar basalts.

Figure IC 9. (Gast, 1972)

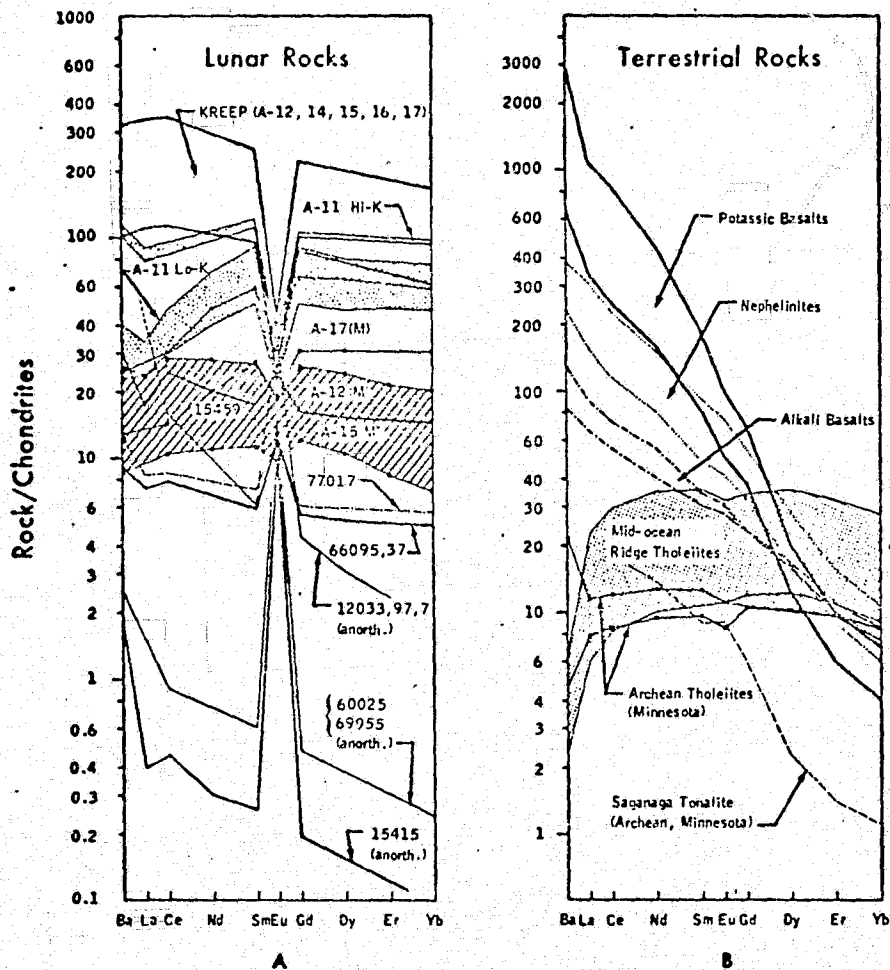


Feldspar ternary plot showing limited compositional range of feldspars in mare basalts.

Figure IC 10. (Papike et al., 1976)

to metallic iron. Unlike terrestrial basalts, no detectible  $\text{Fe}^{3+}$  is found in lunar basalts. The reduced state of titanium and chromium ( $\text{Ti}^{3+}$ ,  $\text{Cr}^{2+}$ ) are rare in terrestrial basalts but occur in significant amounts in lunar basalts (Burns et al., 1973). The oxygen fugacity value of lunar samples for a given temperature is several orders of magnitude below those for terrestrial basaltic lavas (Sato et al., 1973).

The rare earth element patterns for most terrestrial basalts are strongly enriched for the light rare earths (Figure IC-11, from Jahn and Nyquist, 1976). This pattern is found in none of the lunar basalts although the high titanium basalt REE pattern is somewhat similar to that for terrestrial oceanic tholeiites. A notable distinction of lunar mare basalts is the "europium anomaly" exhibited in their REE pattern. This relative depletion of Eu is linked to the lunar reducing conditions which allow  $\text{Eu}^{2+}$  to be stable. The geochemical behavior of  $\text{Eu}^{2+}$  is distinct from  $\text{Eu}^{3+}$ ;  $\text{Eu}^{2+}$  can readily enter the Ca site in feldspar lattices. The europium anomaly is generally interpreted as indicating the removal of feldspar from the source region of mare basalts prior to their generation (see summary in Taylor, 1975, p. 154-159).



Ba and REE abundance patterns for various lunar and terrestrial rocks. The difference in the shapes of the patterns is very striking. In the case of lunar rocks, KREEP and mare basalts (A-11, Hi-K, Lo-K, A-17(M), A-12(M), and A-15(M)) have distinct negative Eu anomalies. In contrast, all anorthositic rocks (15459, 77017, 66095, 12033, 60025, 69955 and 15415) have positive Eu anomalies but much lower total REE abundances. VHA basalts or troctolitic rocks (not shown) have REE abundances intermediate of those of KREEP and anorthositic rocks. In the terrestrial case, two Archean tholeiites from Minnesota have REE patterns comparable with the oceanic ridge or island arc tholeiites; and an Archean tonalite shows a strong depletion in heavy REE. Sources: Gast et al. (1970), Hubbard and Gast (1971), Hubbard et al. (1972, 1974), Schnetzler et al. (1972), Lafl and Schmitt (1973), Philpotts et al. (1974), Rhodes and Hubbard (1973), Kay et al. (1970), Kay and Gast (1973), Arth and Hanson (1972) and Jahn et al. (1974)

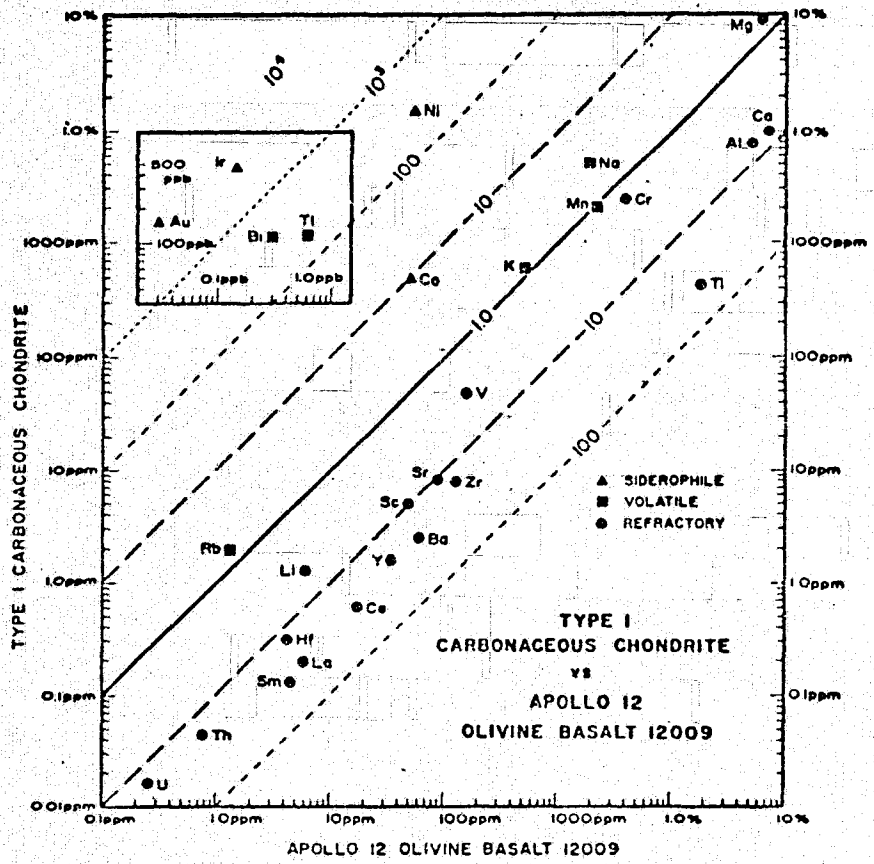
Figure IC 11. (Jahn and Nyquist, 1976)

Comparisons of element abundances between Apollo 12 basalt 12009 (a possible primary partial melt from the interior) and type 1 carbonaceous chondrites (primitive solar system material) is shown in Figure IC-12, from Taylor, 1975. Taylor (1975) suggests three and perhaps four stages of fractionation for lunar material to account for these differences: (1) pre-accretional fractionation (loss of volatiles and probably also siderophiles), (2) post-accretional wide-spread early melting (see below) with fractionation creating a zoned lunar mantle, (3) later partial melting of this mantle leading to further element fractionation, and (4) possible fractionation of this melt en route to the surface and/or during surface crystallization.

Similar elemental comparisons can be made between terrestrial mantle derived basalts (e.g., oceanic tholeiites) and Apollo 15 olivine basalt 15555. Figure IC-13 (from Taylor, 1975) illustrates the general depletion of K and Na in lunar basalts and the enrichment of Fe and Cr.

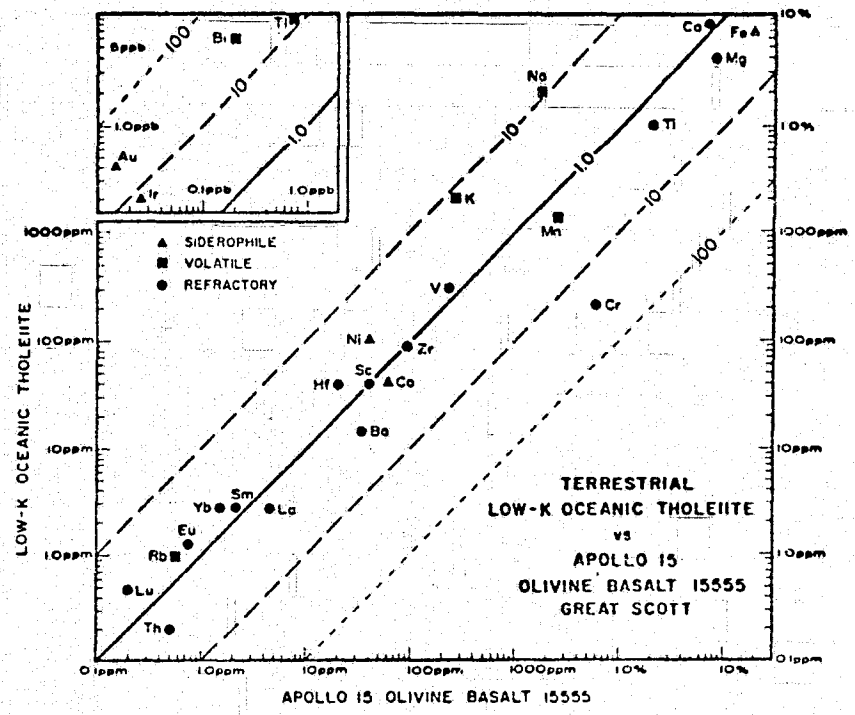
b. Chronology

One of the most notable characteristics of lunar rocks is their age. Lunar mare basalts crystallization ages range from about 3.2 to 3.85 AE. A summary of many of the isotopic age dating results and appropriate references is provided by Taylor (1975) and shown in Table IC-7.



Comparison of the average composition of Type I carbonaceous chondrites (data from Mason) taken as representative of the abundances of the nonvolatile elements in the solar system nebula, with those in a primitive Apollo 12 olivine basalt 12009. This sample is one of the best examples of an unfractionated primary melt from the lunar interior. Points lying along the 45° diagonal full line indicate equality in composition. Increasing distance from the diagonal line indicates increasing disparity in composition. The dashed lines indicate limits for differences by factors of 10, 100, 10<sup>3</sup>, and 10<sup>4</sup>. Elements lying to the right of the equal abundance line are enriched in the lunar sample. Those lying to the left are depleted.

The volatile, siderophile and refractory elements are indicated by distinctive symbols. Note the very wide dispersion of abundances, with some volatile and siderophile elements depleted in the lunar sample by factors of 10<sup>3</sup>-10<sup>4</sup>. Note the rather uniform enrichment of the refractory elements by factors of 10-20. Three stages of fractionation separate 12009 from the primitive abundances. These are (a) precreation separation of volatile and siderophile elements, (b) differentiation following initial lunar melting at about 4.5-4.6 aeons to form source region of maria basalts, (c) partial melting in source region at 3.2 aeons to produce 12009. This olivine basalt reached the surface as a liquid, so that its composition is not affected by a fourth stage of subsequent fractional crystallization.



Comparison of the element abundances in low-K oceanic tholeiites (taken as representative of an important terrestrial primary magma) with that of Apollo 15 olivine basalt 15555, a typical lunar primary magma. Construction of the diagram as for Fig. 4.31.

Note that there are many similarities in composition but that a few decisive differences remain. These include the diagnostic lunar element Cr, greatly enriched in the lunar basalt, and the alkali elements Na, and K, which show their characteristic lunar depletion. The rare volatile (Bi, Tl) and siderophile (Au, Ir) elements also show a characteristic lunar depletion.

Figure IC 13. (Taylor, 1975)

Figure IC 12. (Taylor, 1975)

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Ages of Maria Basalts.

	Age (aeons)	Rock type	Sample no.	Method	Reference	T BABI (p. 65) (aeons)
Apollo 14	3.96	Al-basalt	14053	Rb-Sr	101	4.60
	3.95	Al-basalt	14053	<sup>40</sup> Ar- <sup>39</sup> Ar	102	—
	3.95	Al-basalt	14321	Rb-Sr	101	4.24
Apollo 17	3.83	High Ti	75055	Rb-Sr	103	—
	3.82	High Ti	70035	Rb-Sr	104	—
	3.76	High Ti	75055	<sup>40</sup> Ar- <sup>39</sup> Ar	102	—
	3.74	High Ti	75083	<sup>40</sup> Ar- <sup>39</sup> Ar	105	—
Apollo 11	3.82	Low K	10062	<sup>40</sup> Ar- <sup>39</sup> Ar	106	—
	3.71	Low K	10044	Rb-Sr	107	4.52
	3.63	Low K	10058	Rb-Sr	107	—
Apollo 11	3.68	High K	10071	Rb-Sr	107	3.86
	3.63	High K	10057	Rb-Sr	107	3.90
	3.61	High K	10024	Rb-Sr	107	3.84
	3.59	High K	10017	Rb-Sr	107	3.80
Luna 16	3.56	High K	10022	<sup>40</sup> Ar- <sup>39</sup> Ar	108	—
	3.45	Al-basalt	B-1	<sup>40</sup> Ar- <sup>39</sup> Ar	109	—
	3.42	Al-basalt	B-1	Rb-Sr	110	—
Apollo 15	3.44	Quartz basalt	15682	Rb-Sr	111	4.06
	3.40	Quartz basalt	15085	Rb-Sr	111	—
	3.35	Quartz basalt	15117	Rb-Sr	111	—
	3.33	Quartz basalt	15076	Rb-Sr	111	—
	3.32	Olivine basalt	15555	Rb-Sr	111	—
	3.31	Olivine basalt	15555	<sup>40</sup> Ar- <sup>39</sup> Ar	112	—
	3.26	Quartz basalt	15065	Rb-Sr	111	—
	3.36	Olivine basalt	12002	Rb-Sr	107	4.56
Apollo 12	3.30	Olivine basalt	12063	Rb-Sr	107	4.33
	3.30	Olivine basalt	12040	Rb-Sr	107	4.64
	3.27	Quartz basalt	12051	<sup>40</sup> Ar- <sup>39</sup> Ar	108	—
	3.26	Quartz basalt	12051	Rb-Sr	107	4.45
	3.24	Olivine basalt	12002	<sup>40</sup> Ar- <sup>39</sup> Ar	108	—
	3.24	Quartz basalt	12065	<sup>40</sup> Ar- <sup>39</sup> Ar	108	—
	3.18	Quartz basalt	12064	Rb-Sr	107	4.98
	3.16	Quartz basalt	12065	Rb-Sr	107	4.38

Table IC 7. (Taylor, 1975)

The oldest major group is the high-titanium basalts although some of the Apollo 14 feldspathic basaltic clasts are older. Figure IC-14 (from Papike et al., 1976) illustrates the bi-modal distribution of the basalt ages of returned samples into old high-Ti and 'young' low-Ti groups although there is also a hint of an old low-Ti group.

Internal mineral Rb-Sr isochrons used for obtaining many of the above crystallization ages (e.g., Figure IC-15, from Papanastassiou and Wasserberg, 1971) also provide an initial  $^{87}\text{Sr}/^{86}\text{Sr}$  (I) value. Measured values for initial  $^{87}\text{Sr}/^{86}\text{Sr}$  for lunar samples are shown in Figure IC-16 (from Jahn and Nyquist, 1976). The range of these I values indicate that the source regions of the mare basalts were heterogeneous at the time of partial melting. The lunar mantle must have developed a heterogeneous nature prior to ~4.0 AE.

c. Two stage melting hypothesis

An increasing amount of data indicates the outer 100-500 km of the moon was partially or totally molten very shortly after accretion of the planet. Most theories of mare basalt petrogenesis call for a later partial remelting of various portions of the solidified magma ocean. A few of the most compelling arguments for this popular hypothesis of early lunar evolution are outlined below.

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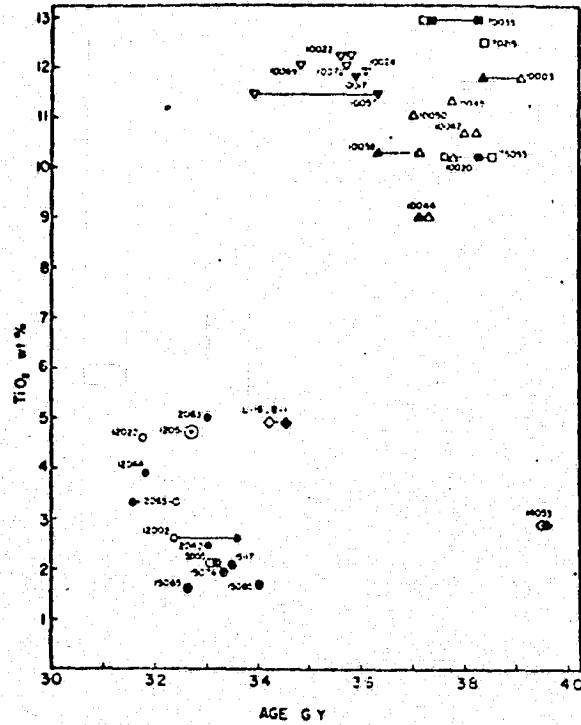
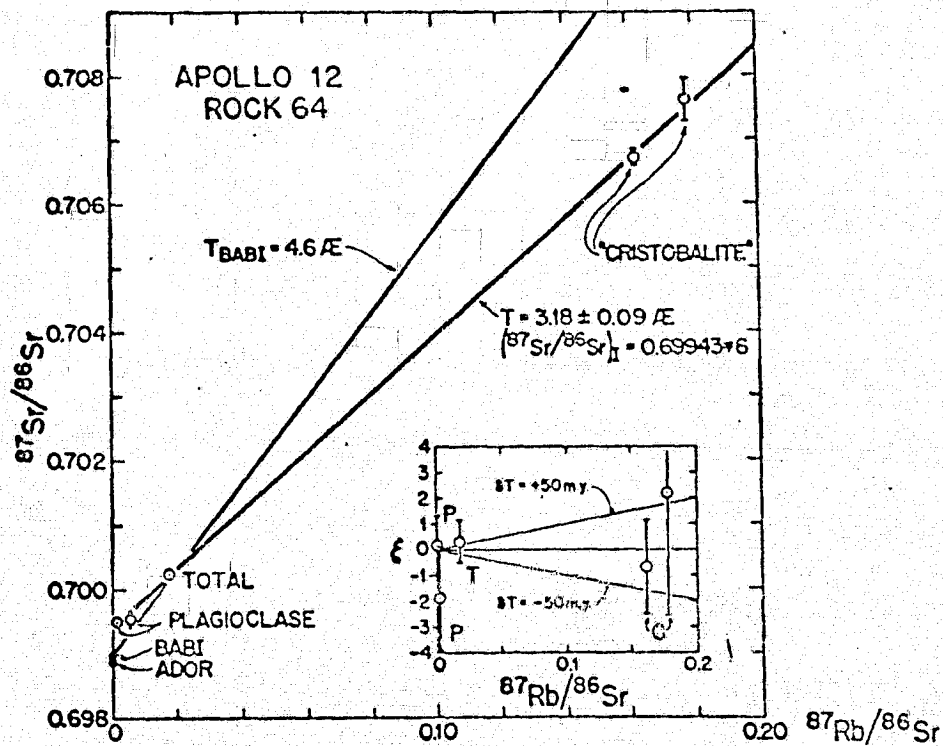


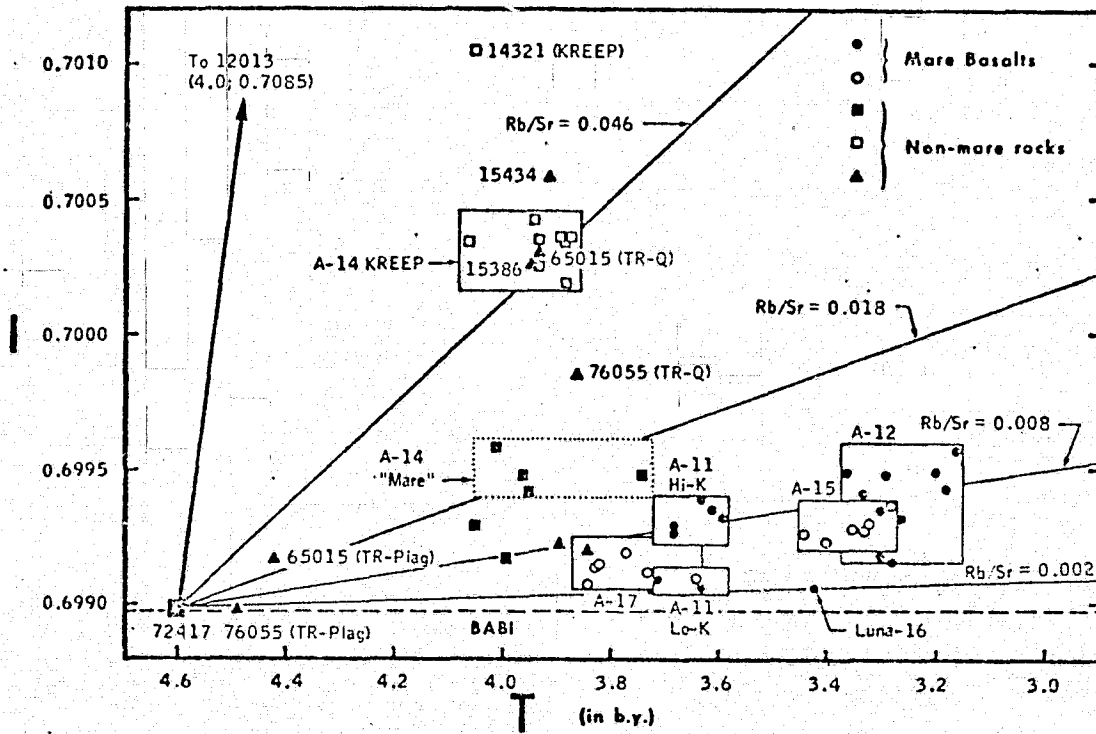
Figure IC 14. (Papike et al., 1976)

Ages of mare basalts. Open symbols indicate <sup>40</sup>Ar <sup>39</sup>Ar, and solid symbols Rb-Sr. Triangles indicate Apollo 11 low K, and inverted triangles Apollo 11 high K.



Rb-Sr evolution diagram for rock 12064. The best fit parameters as well as the reference 4.6 AE isochron and BABI composition are shown. The total range of enrichment in (<sup>87</sup>Sr/<sup>86</sup>Sr)<sub>m</sub> is 1.2%.

Figure IC 15. (Papanastassiou and Wasserburg, 1971)



Sr isotope evolution diagram (or *I, T* plot) for lunar samples. The majority of the data presented here are from the works of the C.I.T. group. Other data are as reported by workers at the Australian National University, the University of Minnesota, the Johnson Space Center, the University of California at Los Angeles, and U.S. Geological Survey, Denver

Figure IC 16. (Jahn and Nyquist, 1976)

The composition of the lunar crust was first inferred from anorthosite fragments found in Apollo 11 soil (Wood et al., 1970). If the major element composition of lunar material was roughly chondritic, the original surface would have to be melted to about 200 km depth to account for the presence of such an observed anorthositic crust (a floating cumulate formed by crystal fractionation). The europium anomaly observed in all lunar basalts (Figure IC-11) further suggests the prior removal of large amounts of plagioclase from the source region of mare basalts (e.g., Philpotts and Schnetzler, 1970; Taylor and Jakes, 1974).

Rb-Sr isotopic data provide further evidence that Rb last separated from Sr in a global fractionation event around 4.4 - 4.5 AE (Papanastassiou and Wasserburg, 1971). A model age ( $T_{\text{BABI}}$ ) of when Rb and Sr were last fractionated can be estimated from whole rock data assuming an initial  $^{87}\text{Sr}/^{86}\text{Sr}$  value equal to that found in basaltic achondrites (Figure IC-15). Most  $T_{\text{BABI}}$  ages cluster around 4.5 AE (see Table IC-7) indicating the moon, or specifically the source regions of most basalts, behaved approximately as a closed system with no major later fractionation in Rb/Sr. The discovery of a dunite differentiate with a crystallization age of 4.55 AE (Papanastassiou and Wasserburg, 1975) tightens the Rb-Sr case for early lunar

differentiation.

Sm-Nd isotopic systematics for an Apollo 17 basalt (Lugmair et al., 1975) indicate a fractionation event of the parent material at 4.35 AE. Independent U-Pb chronologies for Apollo 17 basalts (Tera and Wasserburg (1974) also indicate similar ages (4.43 AE) for an early lunar differentiation, with later derivation of the basalts from a closed U/Pb system.

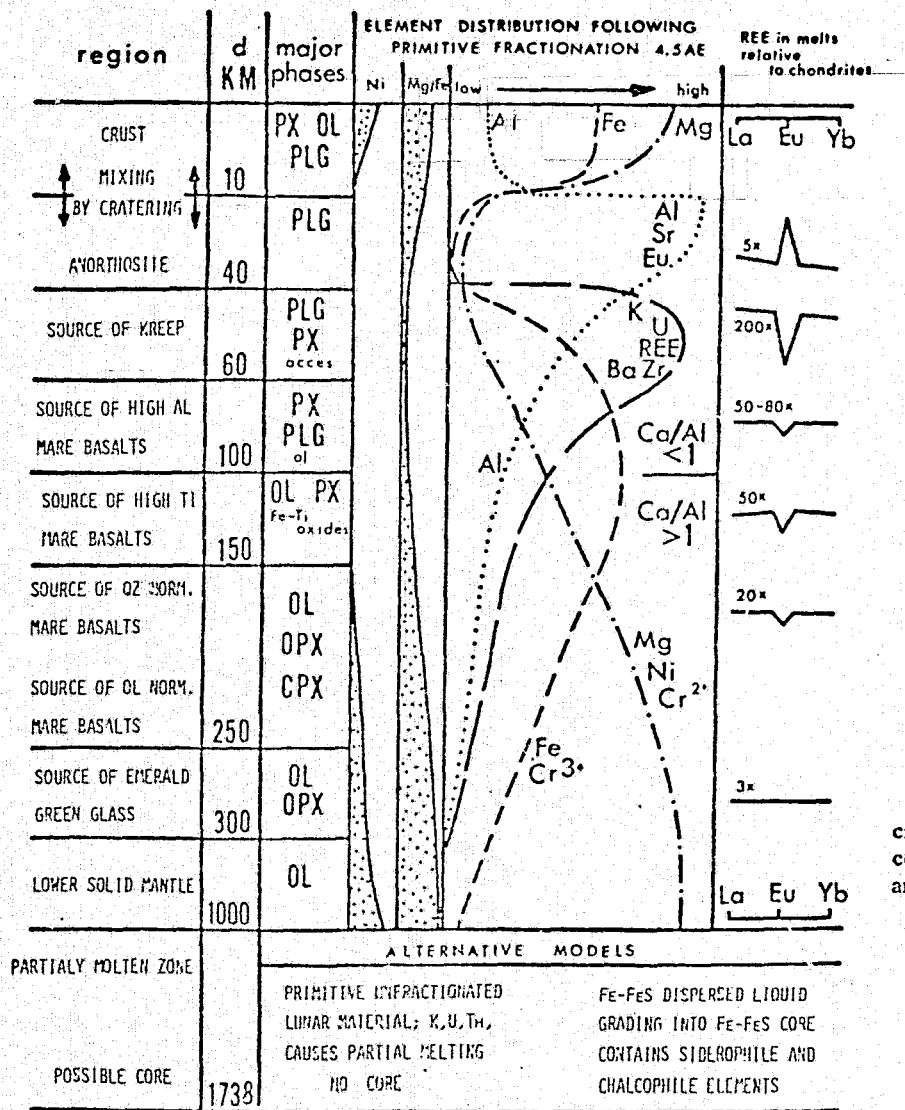
Solomon and Chaiken (1976) recently examined a variety of plausible thermal evolution models to account for the age of observable structural features (or lack of) on the lunar surface. There are no features indicative of large compressional or tensional stresses since the formation of the major basins and emplacement of the lunar maria. The only thermal models that can account for this lack of whole planet volume change through 4 AE requires an initial hot surface and cold interior. They argue the original "magma ocean" could not have been deeper than 300 km nor shallower than 100 km.

d. Mare basalt source regions

The mare basalt magmas are believed to have been derived as partial melts generally from the lunar mantle which, from the previous discussion, include a 100-300 km deep fractionated zone originating from an early magma

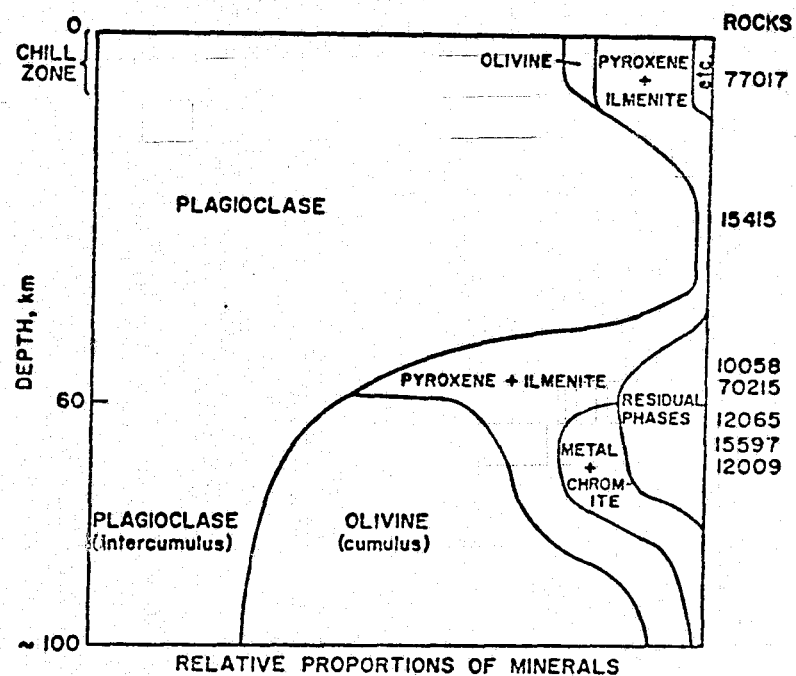
ocean. The details of the geochemical structure of the mantle are not entirely agreed upon although common features include basal zones of olivine and pyroxene cumulates, crustal plagioclase (late floating cumulate) and some sub-crustal residual material that is generally enriched in incompatible elements (Taylor and Jakes, 1974; Hollister, 1975; Hubbard and Minear, 1975; Wood, 1975; Walker et al., 1975; Taylor, 1975; Longhi, 1977; Solomon and Longhi, 1977). Two such models of the lunar mantle composition are shown in Figure IC-17 (from Taylor, 1975) and Figure IC-18 (from Hollister, 1975).

Hypotheses for the genesis of mare basalts are circularly dependent on models of the lunar interior. Most current models involve simple remelting of various cumulate layers with or without assimilation of other heterogeneous pockets or more primitive (deeper, unfractionated) material. An unbiased critical review of competing models of mare basalt petrogenesis has not been presented and is beyond the scope of this description. Discussion of a variety of these models include Walker et al. (1975), Ringwood and Green (1975), Hubbard and Minear (1975), Drake and Consolmagno (1976), O'Hara et al. (1975), Kesson and Ringwood (1976), Ma et al. (1976), Ringwood and Kesson (1976), Shih and Schonfeld (1976), and Taylor and Jakes (1974, 1977).



Geochemical model of the lunar interior at about 4.5 aeons.

Figure IC 17. (Taylor, 1975)



Relative proportions of minerals in shell A as a function of depth, after crystallization of shell A. The mineral proportions do not take into account the change in composition by addition of melt from below 100 km. The thickness of the chill zone is arbitrary. The source region, based on the model, for some well-characterized samples is indicated.

Figure IC 18. (Hollister, 1975)

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High pressure equilibrium experiments on lunar basalts (e.g., Hays and Walker, 1974) can be used to estimate depth of partial melting and the mineralogy of the residuum (e.g., Ringwood and Essene, 1970) assuming the sample represents the equivalent of a primary (unfractionated) liquid. There is still much discussion concerning the degree to which particular samples can be considered quenched primary melts or the products of significant fractionization prior to extrusion (O'Hara et al., 1975). The extensive literature concerning experimental studies has been recently evaluated and reviewed (Kesson and Lindsley, 1976). There is a widespread opinion that the low-Ti parental magma formed by partial melting of an olivine-pyroxene mantle at depths of 200-500 km. There is less agreement concerning the high-Ti basalts with the two prevalent views being (a) partial melting of an olivine-pyroxene-ilmenite source at ~120 km, or (b) partial melting at or below 240 km of an olivine-pyroxene source that is either intrinsically enriched in Ti or contaminated by Ti enriched phases formed at different depths. Both the orange and green glass present anomalies and may represent partial melts of an olivine-pyroxene source at depths of 400 km or more. (Alternatively, these unusual glasses may be products of large degrees of partial melt.)

Although no experimental data exist for the Luna 16 high alumina basalts, their bulk chemistry suggests they could be derived by partial melts of the lower crust or upper mantle ( $\sim 40$ -100 km) (Kurat et al., 1976). The low initial strontium ratio for Luna 16 basalts (see Figure IC-16) suggests a distinct source region comparable in Rb/Sr to that for the low-K high-Ti basalts.

### 3. Unanswered Questions

Further detailed and extensive study of existing lunar samples will certainly clarify some of the models for mare basalt petrogenesis and lunar evolution in general. Before any of these models can be generally accepted, the following questions must be answered concerning the relation of the lunar samples to the whole moon:

(a) How representative are the samples of the major geochemical units on the moon? What is the relative global proportions of rock types identified in the samples? What is the geochemistry of unsampled regions?

(b) What is the temporal sequence of mare basalt types? Are the current patterns seen in the lunar samples representative?

Answers to these questions are essential in determining the composition and degree of heterogeneity (vertical and lateral) of the lunar mantle. The temporal and

volumetric relations between global basalt types put constraints on thermal models related to basalt petrogenesis. Until further samples are obtained, remote sensing techniques are the only means by which to derive answers to the above questions. Information concerning regional and global geochemistry that can be derived from spectral reflectance measurements is discussed in the remaining sections.

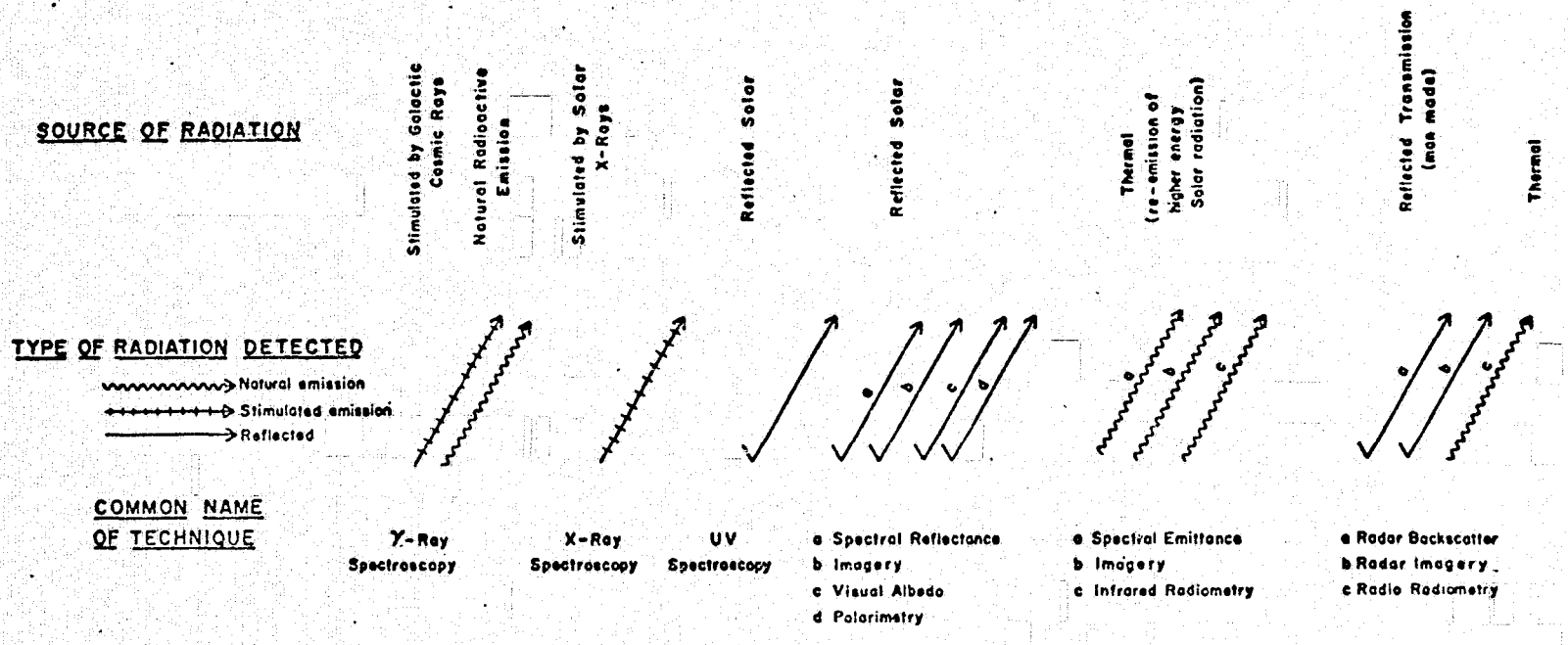
II. REMOTE SENSING OF THE MOON

## II. REMOTE SENSING OF THE MOON

### A. General: Current Techniques of Remote Sensing

Definition and characterization of planetary surface units provide the basic framework for understanding the planet as it currently exists as well as the placing constraints on its past history. A variety of techniques currently exist which provide information about a surface from remote observations (Figure IIA-1 and Table IIA-1, from Head et al., 1976). All but one of these remote sensing techniques involve the detection of electromagnetic radiation; only segments of the electromagnetic spectrum, however, have been utilized in remote observations (see Figure IIA-1). Many observations can only be made from orbit due to the absorption characteristics of the earth's atmosphere.

Existing remote sensing techniques were examined in detail by Head et al. (1976, 1977) and subdivided into three types according to how the information they provide is used for mapping the geology and geochemistry of unvisited planetary surfaces: (1) Characterizing (C in Table IIA-1) techniques allow the classification and characterization of chemical composition, lithology, or physical features. (2) Defining techniques (D) help to define the boundaries and extent of units. (3) Supporting techniques (S) pro-



**COMMON UNITS**

Regions of Transparent Atmosphere (Earth)

**GENERAL UNITS**

Wavelength (meters)

Energy (eV)

Frequency (Hz)

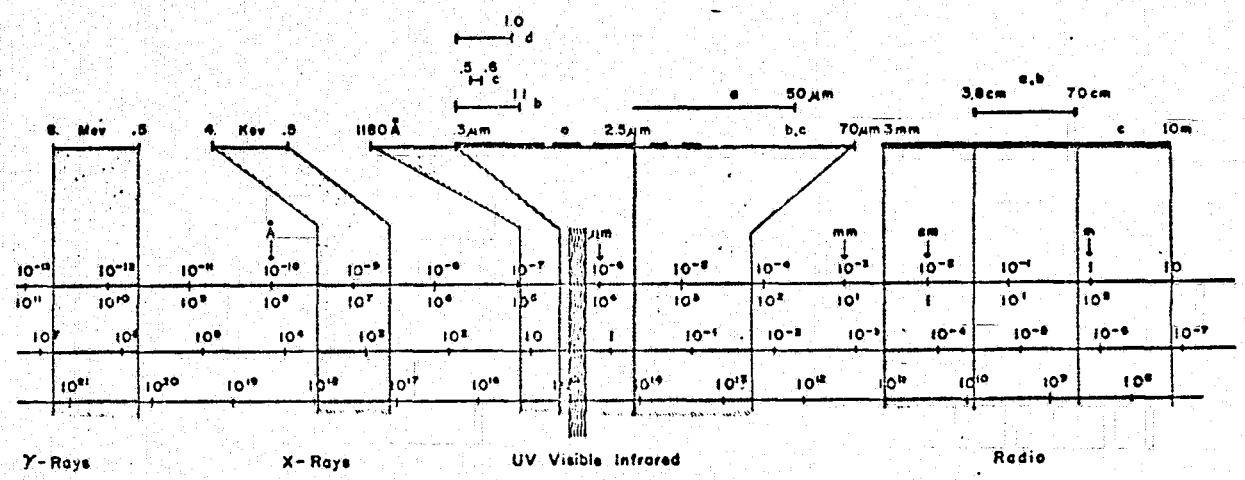


Figure IIA 1. Radiation utilized by remote sensing techniques. (Head et al., 1976, 1977)

<u>Technique</u>	<u>Type</u>	<u>Use or potential use in mare areas</u>
γ-ray spectroscopy	C	Provides regional variations in Th, K, Ti, Fe, Mg, concentration; low spatial resolution but general correlation with mapped units.
X-ray spectroscopy	C	Provides regional variations in Al/Si ratios. Regional variations seen between and within maria.
UV spectroscopy	S	Potential supporting technique.
Spectral Reflectance	C	Provides quantified information on composition, mineralogy, and soil maturity; e. g., TiO <sub>2</sub> content of mature mare soil, major mafic mineral content and composition.
Spectral Imagery	D	Reveals boundaries of units characterized by specific spectral reflectance properties. When images are quantitative, some characterizing properties can be mapped if the correct spectral bands are used.
Albedo	D, C	Used qualitatively in basic distinction of mare units from highlands. Quantitative albedo measurements are used to subdivide mare units. Within an area of similar maturity, can be used as an indication of compositional variation.
Polarimetry	S	Provides information on the surface regolith (as a combined function of complex refractive index and particle size).
Visual Observations, Photography	D, C	Used in basic definition of unit boundaries and extent (regional topography, and local topography such as flow fronts). Other observations such as surface texture and qualitative estimates of crater density are useful in unit characterization.
Infrared Radiometry	S	Measures thermal inertia $\sqrt{K\rho C}$ ; Surface density (blockiness) major influence but other parameters may have an effect.
Spectral Emittance	C(?)	May provide information on Si/O ratio and also mineralogy, if soil particle size is large. Little data available.
Radar (Backscatter and Imagery)	D, C, S	Radar imagery helps define extent of some units. Backscatter may also be helpful as a characterizing and supporting technique by providing information on crater density, block distributions, regolith thickness, and composition.
α-particle	S	Provides regional variations in <sup>222</sup> Ra and <sup>210</sup> Po.

Table IIA 1. Uses of remote sensing techniques.  
(Head et al., 1976, 1977)

vide additional useful information but are not necessarily fundamental to the unit characterization or definition.

The remote sensing techniques that will be discussed in detail in the following sections and used for geochemical exploration are spectral reflectance and multi-spectral imaging. Both techniques utilize the spectral character of reflected solar radiation in the visible and near infrared spectral region. The basic principles of the techniques (described in Section IIC and IID) rely on well-defined absorption characteristics of transition metal ions (Fe, Ti, Cr, etc.) in minerals. Spectral reflectance measurements refer to continuous spectra, generally measured for a single small area with 50-300Å resolution through the spectral range of .3 to 2.5µm. Spectral reflectance measurements provide information about the mineralogy of the surface material and are thus a characterizing technique. Multi-spectral imaging is used to map a particular spectral parameter (intensity ratio) in two dimensions. If the spectral characteristics of a unit can be uniquely defined with a few parameters, multi-spectral imagery obtained at the characteristic wavelengths is used to map the regional extent of the unit. Multi-spectral imagery is thus a defining technique and is currently used through the spectral range .35 to 1.0µm (being limited by the sensitivities of two-dimensional detectors available).

## II. REMOTE SENSING OF THE MOON

## B. Outline of Previous Work in Spectral Reflectance

Prior to the late 1960's, most of the astronomical observations concerning the color of the moon was either photographic imagery or broad band photometry. The photometry was obtained primarily with the classical astronomical filters: U ( $.35\mu\text{m}$ ), B ( $.45\mu\text{m}$ ), V ( $.55\mu\text{m}$ ). This early work showed that small color differences exist from place to place on the moon, but no geological or geochemical significance was discerned or attempted. A thorough summary of this data can be found in McCord (1968a).

In the late 1960's, two types of research were begun which have since both merged and expanded into the general field of spectral reflectance of the lunar surface. In 1968 high precision photoelectric measurements were obtained of the reflectance from  $.40$  to  $.80\mu\text{m}$  of many (83) small lunar areas relative to a standard lunar area (McCord, 1969a). Such measurements were later expanded to include the spectral range from  $.40$  to  $1.10\mu\text{m}$  for about 20 areas (McCord and Johnson, 1969). Concurrently, spectral reflectance measurements in the visible and near infrared were being made in the laboratory for samples of powdered rocks and minerals (White and Keester, 1966; Adams and Filice, 1967). It was recognized that the spectral differences observed remotely using a telescope were intimately related

to and dependent on the geochemistry of the material observed (Adams, 1968; Adams and McCord, 1970). Strong supporting independent evidence for this hypothesis was provided by the interpretation of transmission spectra of oriented crystals using crystal field theory (e.g., Burns and Fyfe, 1967; Burns, 1970a). The interpretations of transmission spectra of minerals provided sound theoretical basis for linking the spectral characteristics of rocks and minerals to their geochemistry (see Section II-C-1).

An abundant amount of research concerning spectral reflectance ensued, largely as an outcome of this initial pioneering work by McCord and Adams. For this review, the relatively recent telescopic and laboratory research results reported in the scientific literature have been divided for convenience into six categories and are outlined below. Some papers have significant amounts of material that fall into more than one category. Not included are a scattering of additional broad-band lunar photometry that either confirm earlier work or are not relevant to a discussion of spectral features of lunar material.

1. Laboratory Reflectance Measurements: General

During the last decade, a large number of reflectance spectra have been obtained in the laboratory for powdered samples of rocks, minerals, and soils. The spectral region of interest here is from .30 to 2.5 $\mu$ m. The laboratory

measurements have been obtained generally with a high resolution (50Å) spectrometer that measures the diffuse reflectance of a sample relative to that of a "white" standard such as smoked MgO.

About ten years of research on the spectral reflectance of lunar, meteoritic, and terrestrial samples is presented by Adams (1974, 1975). The direct relationship between laboratory reflectance measurements of returned lunar soil and telescopic measurements of the surface at the landing area was described by Adams and McCord (1970). A sequence of research reports followed as more lunar samples became available (Adams and McCord, 1971a, b, 1972, 1973). The problems of how the optical properties of lunar soils are related to those of the parent rock was discussed by Adams and McCord (1971a) and Nash and Conel (1973). A review of the spectral reflectance properties of lunar soil is in preparation (Adams and Charette, 1977).

An extensive catalogue of spectral reflectance curves for a variety of terrestrial rocks and minerals with roughly-defined geochemistry is provided by Hunt and Salisbury (1970, 1971, 1976a, b), Hunt et al., (1971a, b, 1972, 1973a, b, c, 1974), and Salisbury et al. (1975). Reflectance spectra of carefully chosen meteoritic samples have been extensively analyzed and catalogued by Gaffey (1973, 1976). The problems of deconvolving a rock spectrum to

obtain information about the mineral components is discussed by Adams (1976) and in more detail by Gaffey (1977). Variations in the spectral properties of minerals in a prepared mixture is presented by Pieters (1972, 1974) and Nash and Conel (1974).

## 2. Telescopic Reflectance Measurements: General

Most of the telescopic spectral reflectance measurements have been made with some form of a photoelectric filter photometer. The wavelength range and resolution has varied, but to date most measurements have been limited in the near-infrared to the spectral sensitivity of an S-1 photomultiplier detector ( $\lambda < 1.1 \mu\text{m}$ ) (e.g., see McCord et al., 1972a), although a few low resolution lunar spectra have been obtained to  $2.5 \mu\text{m}$  (McCord and Johnson, 1970).

A variety of reports have been presented containing a large number of telescopic lunar reflectance spectra (McCord et al., 1972a, b; Pieters and McCord, 1976; Pieters, 1977). The time variation of relative lunar color was examined by McCord (1969b) and found to be small with no evidence for significant luminescence (McCord, 1967). The general phase function of the moon was described by Hapke (1971) and was measured by Lane and Irving (1973). The wavelength dependence of the phase coefficient was measured also by Lane and Irving for nine filters from  $.35$  to  $1.0 \mu\text{m}$

and later by Nygard (1975) with 25 filters from .3 to 1.1 $\mu$ m.

The spectral types evident in spectra of about 200 small (10 to 20 km diameter) lunar areas were presented by McCord et al. (1972a). An analysis of the relationship between the spectra of these areas and the geologic units defined by the USGS mapping program was summarized by Charette et al. (1974). Spectral data for the then current and proposed Apollo landing sites were presented by McCord et al. (1972b).

### 3. Telescopic Spectral Images

Three distinctly different techniques have been used to obtain spectral images of the moon: (1) photography, (2) line scan, and (3) vidicon imagery. A thorough review and discussion of multispectral mapping is given by McCord (1976).

The most widely used photographic color-difference images were published by Whitaker in 1972. Higher quality pictures have since been produced, but are unpublished (Whitaker, personal communication). Barbarshov (1973) used the same photographic technique to produce a color difference image of the full moon. He calibrated and digitized the image and published a low spatial resolution map (in color) of the color difference data.

A low spatial resolution spectral map of Mare Serenitatis and Mare Tranquillitatis was produced by Soderblom (1970) from photoelectric line scans with high photoelectric precision

using three visible filters. Later a similar high-precision low-resolution spectral map of the full moon was produced (Soderblom et al., 1976).

Digital vidicon spectral images have been obtained by two independent groups, one at MIT and the other at JPL. The latter has concentrated on intensive study of specific regions (e.g., Johnson et al., 1975b), and have often integrated other forms of remote sensing with the spectral data (see Section 6 below). The MIT group (McCord et al., 1976, 1977) has undertaken a program to map the entire lunar frontside at high spatial resolution ( $\sim 2$  km) with three carefully chosen spectral band-passes. A number of spectral images from this larger program have also been used in specific studies of lunar geology (see Section 6 below).

#### 4. Laboratory Reflectance Measurements: Applied

As the number of available laboratory lunar spectra increased, the information that could be derived from spectral reflectance data of the moon became increasingly more defined. With the return of the first samples (Adams and Jones, 1970; Adams and McCord, 1970; Conel and Nash, 1970), the direct link between laboratory and telescopic measurements was made. The darkening of lunar soil with surface age was first discussed by Adams and McCord in 1971b. Effects of dark basalt glasses (artificial) in lunar soils was further discussed by Adams

and McCord (1971a) and by Nash and Conel (1973). The spectral dominance of pyroxenes in reflectance spectra of lunar rocks was described by Adams and McCord (1971a, 1972) along with a discussion of the geochemical interpretations that can be derived from the energy of characteristic absorption bands. Diffuse reflectance measurements of pyroxenes were interpreted in more detail by Adams (1974).

Using reflectance spectra of magnetic separates for numerous Apollo 16 soils, Adams and McCord (1973) linked the spectral maturation of lunar soils with the accumulation of the dark glass-rich agglutinates. The effects of maturation on a reflectance spectrum of lunar soil was further characterized by Adams and Charette (1975). The related geochemical processes involved with agglutinate formation were discussed by Charette and Adams (1975a) and Rhodes et al. (1975).

Laboratory reflectance measurements of mature mare soil samples were used to define the relationship between  $TiO_2$  content and continuum slope in the blue and ultraviolet (Charette et al., 1974). An understanding of this relationship led to the prediction of high-titanium material in the Apollo 17 region (Pieters et al., 1973). A clarification of this technique for determining  $TiO_2$  content from remote observations is presented by Adams (1977). The identification of the dark mantling material at Apollo 17 resulted from laboratory spectra of the orange and black glasses obtained at Shorty crater

(Pieters et al., 1974; Adams et al., 1975). A method for determining the FeO content for lunar soils from reflectance measurements is being examined by Charette and Adams (1977).

##### 5. Telescopic Reflectance Measurements: Applied

After the general spectral types of lunar maria and high-land regions had been defined (McCord et al., 1972a), it became possible to use telescopic spectral reflectance measurements to solve specific problems of geology and geochemistry. Many regions of dark mantling material were shown to have unique spectral characteristics (Pieters et al., 1973) that were later linked to the orange and black glass beads of Apollo 17 (Pieters et al., 1974; Adams et al., 1975). From spectra for over 30 areas in Mare Humorum (Johnson et al., 1973) it was determined that two basaltic units filled the central basin, one a medium-high titanium basalt, the other a low titanium basalt. The geochemical units of this region were later more explicitly defined using more remote sensing information (see Section 6 below). The sampled and unsampled mare basalt types were defined and partially characterized from spectra of over 120 mature mare regions (Section III; Pieters and McCord, 1976). This characterization of basalt types allowed a regional description of the basalt types at the recent Luna 24 landing site in Mare Crisium

using vidicon spectral imagery (Pieters et al., 1976). A characterization of the spectral types for spectra of highland and mare craters was presented by Pieters (1977).

6. Spectral Reflectance: Application with other forms of remote sensing

A current trend of applied research is to integrate data from numerous remote sensing sources for one specific region or problem. It is recognized that the whole is often greater than the sum of its parts. Spectral reflectance lends itself particularly well to synthesis studies that aim at in-depth description of regional geology or geochemistry. In some cases, however, the results of synthesis studies have not been significantly useful. The approach of the synthesis team as well as the inherent value of the data are not uniform for all studies reported. An evaluation of the various remote sensing techniques currently in use can be found in Head et al. (1976a, b) as they apply to unit definition and characterization.

Most remote sensing synthesis studies have been concerned with particular regions. These include Mare Imbrium (Schaber et al., 1975), Mare Serenitatis (Thompson et al., 1974), Mare Humorum (Pieters et al., 1975), Lamont (Johnson et al., 1975a), Aristarchus (Zisk et al., 1976), and

Flamsteed (Pieters et al., 1977). Saunders et al., (1977) have converted a variety of remote sensing data to a database that can be conveniently manipulated using computer processing techniques. A few studies have been directed towards understanding more completely "special" types of lunar areas such as areas of dark mantling material (Pieters et al., 1973; Adams et al., 1975) or areas of suspected highland volcanism or sources of KREEP material (Malin, 1973; Head and McCord, 1977).

## II. REMOTE SENSING OF THE MOON

### C. Principles of Spectral Reflectance

#### 1. Optical Properties of Minerals

In order for reflection spectroscopy to be accepted and used as a technique for remotely determining surface mineralogy, the principles of physics and chemistry that govern optical absorption in minerals must be sufficiently understood. Absorption (transmission) spectra have been obtained for a variety of terrestrial, meteoritic, and lunar rocks and minerals generally through the spectral range from 0.30 to 2.50  $\mu\text{m}$  (e.g., Burns et al., 1975, Bell et al., 1975). Over this spectral range there are primarily three general types of absorption features observed: (a) bands due to crystal field electronic transitions within a transition metal ion, (b) bands due to intervalence charge transfer transitions between two neighboring ions, and (c) molecular vibrational bands. Since the few vibrational bands (attributable to  $\text{OH}^-$ ,  $\text{CO}_3^{2-}$  ligands as well as  $\text{H}_2\text{O}$  and  $\text{CO}_2$  molecules) are not observed in lunar material, only the electronic absorptions will be discussed here. The strength and sharpness of absorption features in a spectrum varies somewhat as a function of temperature (Sung et al., 1977) and pressure (Mao and Bell, 1972).

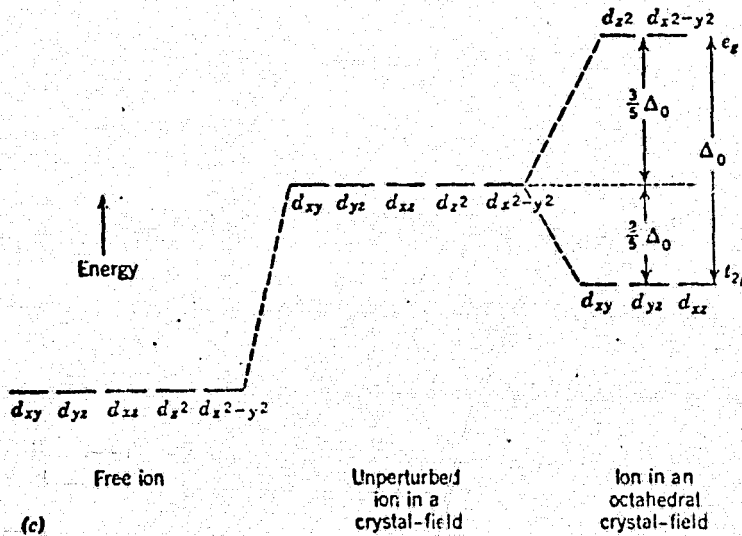
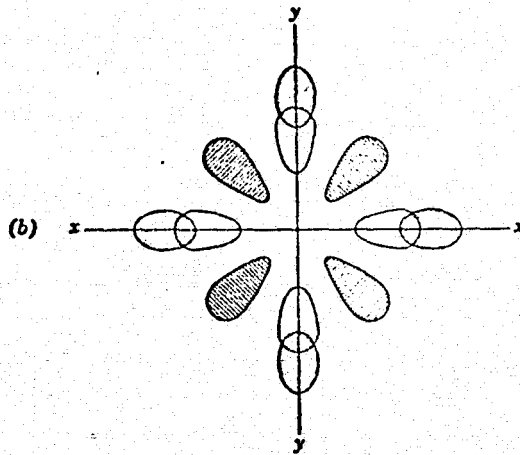
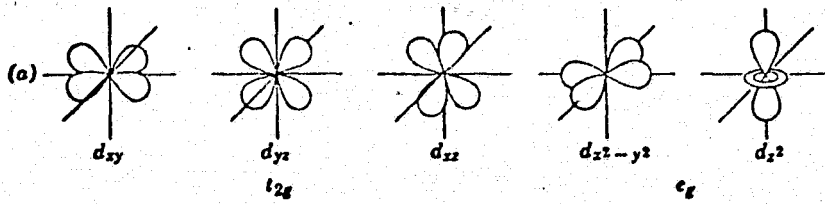
a. Crystal Field Transitions of Transition Metal Ions

The first series transition metals (Table TIIC-1; from Burns, 1970a) constitute about 40% of the earth (primarily iron) and play at least a comparable role in extraterrestrial material (Mason, 1966). In rock forming minerals, ions of these elements are characterized by incompletely filled inner  $3d$  atomic orbitals (see Table 1). In an isolated environment, the five  $d$  orbitals of transition metal ions are degenerate although the spatial distribution of electron density in each orbital is distinct (Figure II-C-1a from Burns, 1970a). When a transition metal ion is incorporated into a crystal structure or is surrounded by negatively charged ligands, the energy level of the  $d$  orbitals lose their degeneracy. Electrons in the orbitals are repelled by the surrounding ligands with the resulting energy level of each orbital being dependent on the orientation of the electron orbital with respect to the ligands. In crystal field theory, the anions are considered as point charges and any covalent interaction between anion and cation is ignored. For an ion in a regular octahedral crystal field (six-fold coordination) the  $d_{z^2}$  and  $d_{x^2-y^2}$  orbitals, which are oriented along

*Electronic configurations and crystal field stabilization energies of transition metal ions  
in octahedral co-ordination*

Number of 3d electrons	Ion	High-spin state				Low-spin state			
		Electronic configuration		Unpaired electrons	CFSE	Electronic configuration		Unpaired electrons	CFSE
$t_{2g}$	$e_g$	$t_{2g}$	$e_g$						
0	Ca <sup>2+</sup> , Sc <sup>3+</sup> , Ti <sup>4+</sup>			0	0			0	0
1	Ti <sup>3+</sup>	↑		1	$\frac{2}{5}\Delta_0$	↑		1	$\frac{2}{5}\Delta_0$
2	Ti <sup>2+</sup> , V <sup>3+</sup>	↑ ↑		2	$\frac{3}{5}\Delta_0$	↑ ↑		2	$\frac{3}{5}\Delta_0$
3	V <sup>2+</sup> , Cr <sup>3+</sup> , Mn <sup>4+</sup>	↑ ↑ ↑		3	$\frac{4}{5}\Delta_0$	↑ ↑ ↑		3	$\frac{4}{5}\Delta_0$
4	Cr <sup>2+</sup> , Mn <sup>3+</sup>	↑ ↑ ↑ ↑		4	$\frac{4}{5}\Delta_0$	↑↓ ↑ ↑		2	$\frac{4}{5}\Delta_0$
5	Mn <sup>2+</sup> , Fe <sup>3+</sup>	↑ ↑ ↑ ↑ ↑		5	0	↑↓ ↑↓ ↑		1	$\frac{1}{5}\Delta_0$
6	Fe <sup>2+</sup> , Co <sup>3+</sup> , Ni <sup>4+</sup>	↑↓ ↑ ↑ ↑ ↑		4	$\frac{2}{5}\Delta_0$	↑↓ ↑↓ ↑↓		0	$\frac{1}{5}\Delta_0$
7	Co <sup>2+</sup> , Ni <sup>3+</sup>	↑↓ ↑↓ ↑ ↑ ↑		3	$\frac{3}{5}\Delta_0$	↑↓ ↑↓ ↑↓ ↑		1	$\frac{3}{5}\Delta_0$
8	Ni <sup>2+</sup>	↑↓ ↑↓ ↑↓ ↑ ↑		2	$\frac{3}{5}\Delta_0$	↑↓ ↑↓ ↑↓ ↑ ↑		2	$\frac{3}{5}\Delta_0$
9	Cu <sup>2+</sup>	↑↓ ↑↓ ↑↓ ↑↓ ↑		1	$\frac{1}{5}\Delta_0$	↑↓ ↑↓ ↑↓ ↑↓ ↑		1	$\frac{1}{5}\Delta_0$
10	Zn <sup>2+</sup> , Ga <sup>3+</sup> , Ge <sup>4+</sup>	↑↓ ↑↓ ↑↓ ↑↓ ↑↓		0	0	↑↓ ↑↓ ↑↓ ↑↓ ↑↓		0	0

Table IIC 1. (Burns, 1970a)



Orientations and relative energy levels of  $d$  orbitals in an octahedral crystal-field. (a) Electron distribution in  $d$  orbitals; (b) the  $x$ - $y$  plane of a transition-metal ion in an octahedral crystal-field. The  $d_{xy}$  orbital is cross-hatched, the  $d_{x^2-y^2}$  orbital is open, and the ligands are stippled. (c)  $d$  orbital energy level diagram.

Figure IIC 1. (Burns, 1970a)

the axes of the octahedron, are of higher energy than the  $d_{xy}$ ,  $d_{yz}$ , and  $d_{xz}$  (Figure II-C 1c). Distortions of the octahedral site will cause further splitting of the  $d$  orbital energy levels. This energy separation of the  $d$  orbitals is largely dependent on the type and valence of transition metal ion as well as the crystal field environment it is subjected to (Burns and Fyfe, 1967).

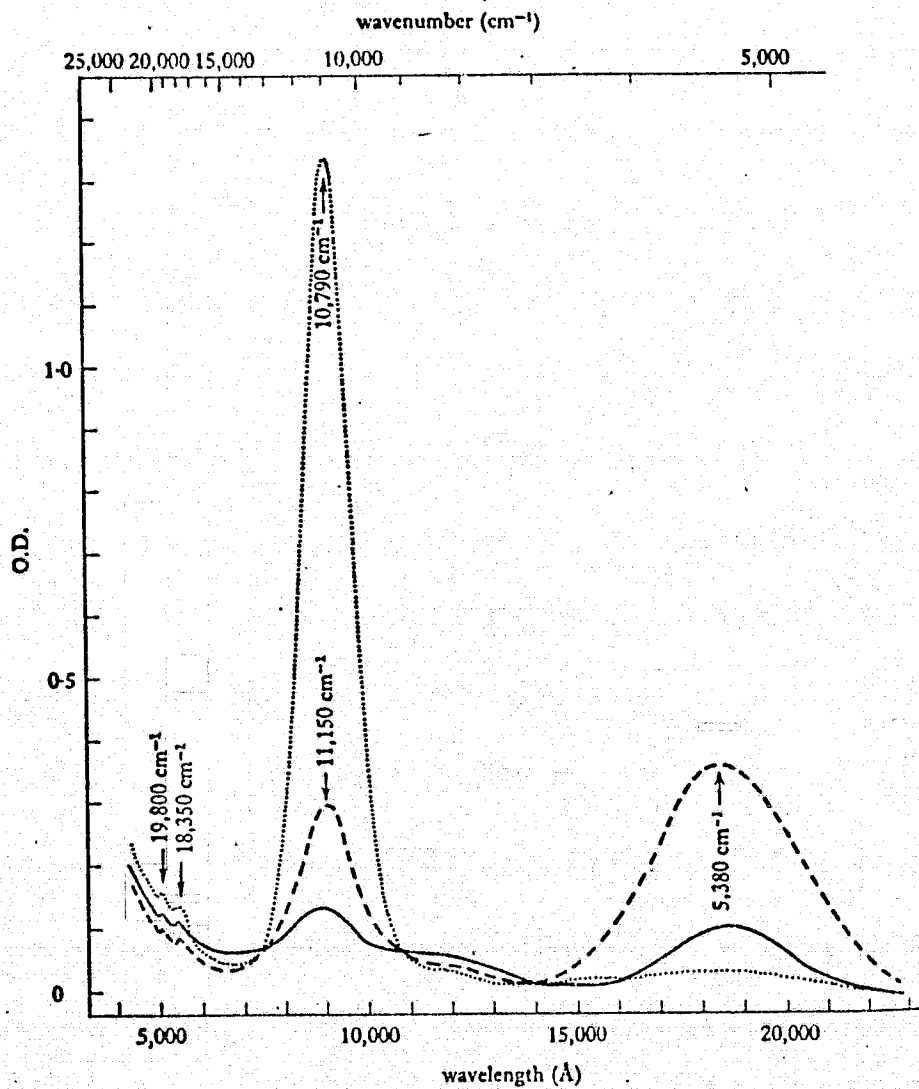
Absorptions occur in the spectra of minerals containing transition metal ions as a  $d$  orbital electron is excited to a higher energy state by an incoming photon. The energy of the absorption is equivalent to the difference in energy between the ground state and excited state of the ion. Absorption features characteristic of particular ions, in particular crystal structures, can be interpreted and predicted by the application of crystal field theory to describe the relative energies of the  $d$  orbitals (Burns, 1970a). Absorption characteristics of a material are thus directly linked to its mineralogy.

Two types of crystal field  $d$  orbital transitions can be considered: (1) spin allowed and (2) spin forbidden, with the latter being about two orders of magnitude less intense than the former. Both

types occur in the transmission spectrum of an orthopyroxene shown in Figure IIC -2 (from Burns 1970a). The two broad absorptions centered near 1 and 2 micrometers are due primarily to  $\text{Fe}^{2+}$  (with six d orbital electrons) in the distorted M2 octahedral site of the pyroxene. These absorptions for which both the ground state and the excited state have the same number of unpaired electrons are spin allowed. The sharp weak absorptions that occur at higher energies here are attributable to spin forbidden transitions of  $\text{Fe}^{2+}$  (Burns and Vaughan, 1975) for which the excited state has a spin multiplicity different from the ground state.

b. Charge Transfer Transitions

Many of the intense mineral absorptions that occur in the visible and ultraviolet arise from electronic transitions between neighboring ions. Since there is possible overlap between electron orbitals of neighboring ions, the anions should not be treated as point charges in this interaction and crystal field theory is inappropriate to interpret the anion-cation and cation-cation charge transfer transitions. Recently, molecular orbital theory, which provides a more complete description of the electric structure



Polarized absorption spectra of orthopyroxenes. ...  $\alpha$  spectra; ---  $\beta$  spectra; —  $\gamma$  spectra.  
 (a) bronzite,  $\text{Fs}_{14-5}$ , from Bamle, Norway.

Figure IIC 2. (Burns, 1970a)

of a compound, has been used successfully for the calculations of the energy of many such transitions (Tossel et al., 1974; Loeffler et al., 1974, 1975) and has achieved good agreement with experimental values for band centers (Burns et al., 1976).

Oxygen-to-metal charge transfers bands generally occur well into the ultraviolet and are thousands of times more intense than the crystal field bands discussed above (Bell et al., 1975). The side wings of these bands extend into the visible and often into the near infrared causing the continuum of most silicates to be curved with increasing absorption towards the ultraviolet.

Cation-cation intervalence charge transfer absorptions occur between metal ions in adjacent coordination sites sharing a common edge or face and are directional or strongly polarization dependent. Since both the concentration of metal ions and the degree of orbital overlap affect the intervalence charge transfer, such absorptions are observed with a wide range of intensities (Burns and Vaughan, 1975). Some of the most common homonuclear intervalence transfer transitions are  $\text{Fe}^{2+} \rightarrow \text{Fe}^{3+}$  and  $\text{Ti}^{3+} \rightarrow \text{Ti}^{4+}$ . A variety of heteronuclear charge transfer processes can also

occur between the different oxidation states of iron and titanium, the most common of which is  $\text{Fe}^{2+} \rightarrow \text{Ti}^{4+}$  (Burns and Vaughan, 1975).

Since a variety of charge transfer and crystal field absorptions overlap in the visible and ultraviolet, it has been difficult to empirically interpret absorption features in this part of the spectrum. Bell et al. (1975) and Huguenin (1975) have made considerable progress in band assignment for  $\text{O}^{2-} \rightarrow \text{Fe}^{3+}$  in ferric oxides. Burns et al. (1976) have sorted out a number of the numerous transitions involving titanium and iron in terrestrial and lunar materials.

## 2. Reflectance of Minerals and Rocks

### a. Physical components of reflection.

The optical properties of minerals discussed above have been examined primarily using transmission spectroscopy of oriented individual grains where the percent transmission or absorbance is measured through the material as a function of wavelength. Reflection spectroscopy, on the other hand, is defined as the wavelength dependence of the ratio of the light reflected from a surface to that incident upon it. Most natural surfaces are rough or particulate and the reflected radiation contains both (1) a specular

and (2) a diffuse component (Figure IIC -3). The strength of each component in natural reflection is directional, i.e., dependent on the geometry of illumination and observation.

The specular component of directional reflectance from a silicate powder or soil is a first surface reflection and is described by Fresnell's equations for reflection from a dielectric. Figure IIC -4 shows the percent reflectance for a single surface reflection as a function of phase angle for a silicate with a real refraction index of 1.5. The phase angle is the angle between the source-object-detector, or the sum of the angle of incidence and the angle of reflection. The Brewster angle for such a material is near 57 degrees and the total reflected radiation is therefore 100% polarized at a phase angle of about 114 degrees. This specular component of reflection contains no measurable spectral information in the spectral range between .3 and 2.5 $\mu$ m. The variations of the absorption coefficient in the visible and near infrared are too small to significantly effect the refractive index.

The diffuse component of surface reflection contains radiation that has been transmitted through

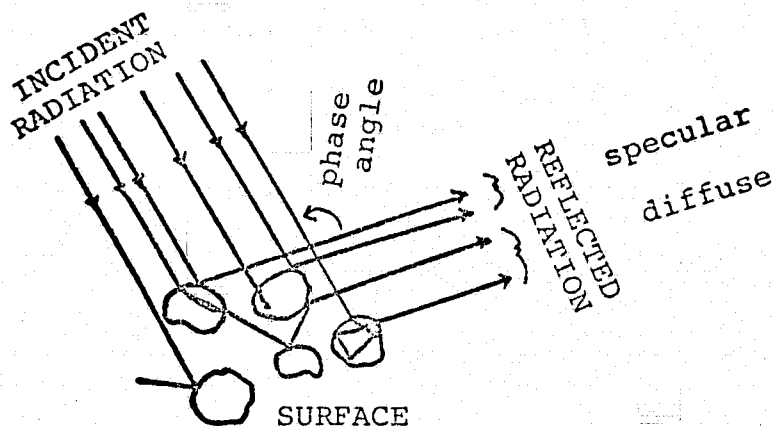


Figure IIC-3  
Components of reflected radiation

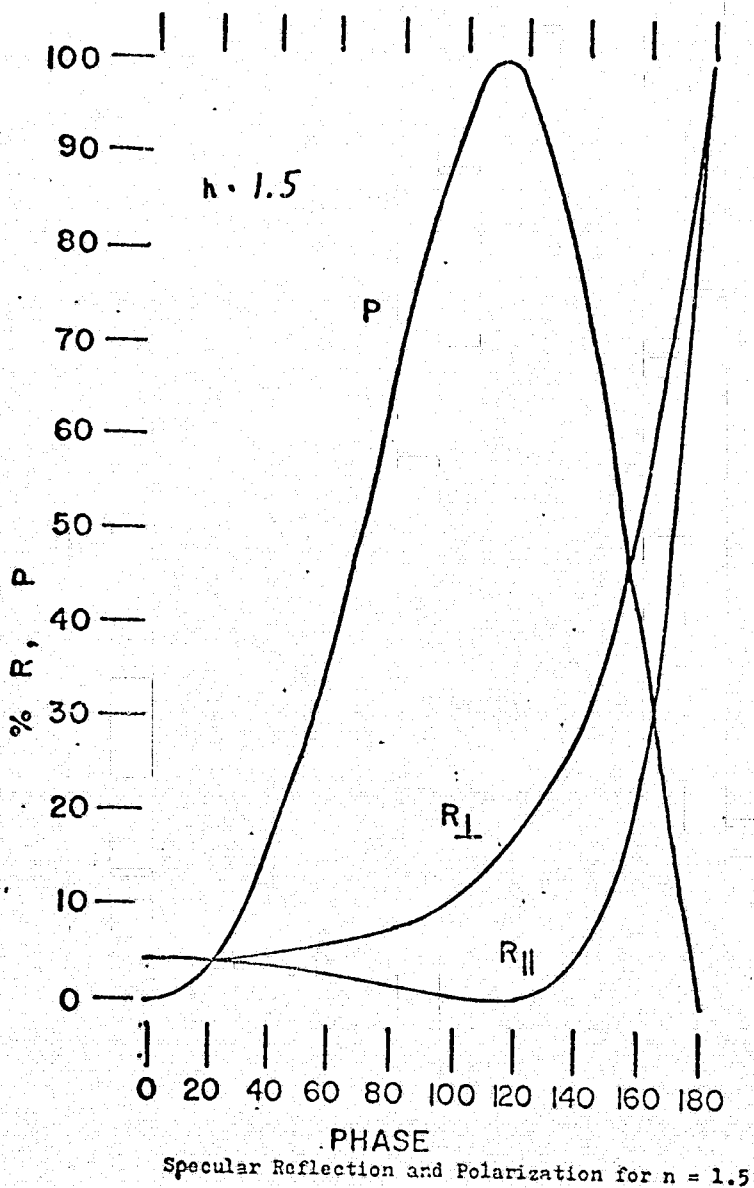


Figure IIC 4.

one or more grains and scattered randomly back into space. This body component contains all the spectral information of transmitted light; it is the sum of transmission through randomly oriented crystals.

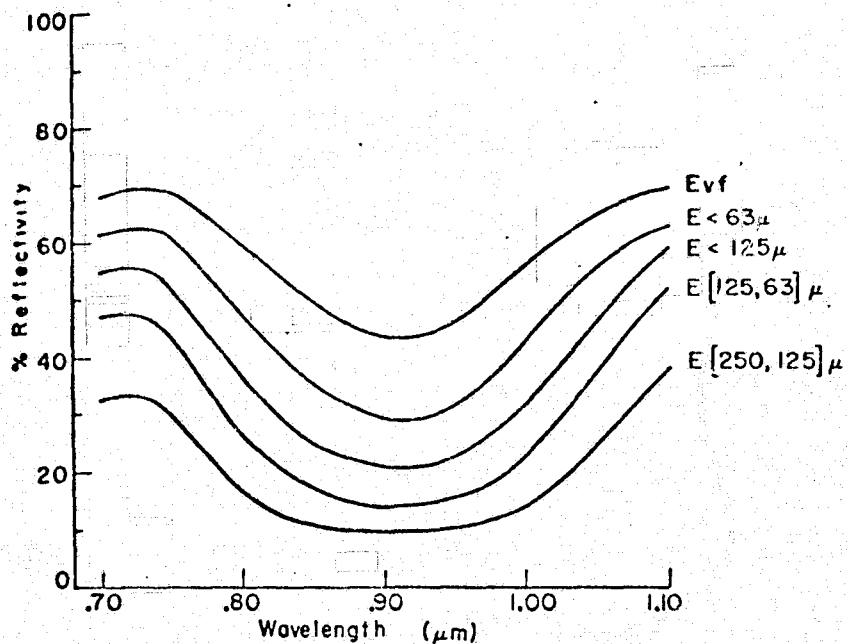
This random orientation of surface crystals, however, causes any phase or polarization information from birefringent materials to be scrambled and lost.

Nearly "pure" diffuse reflectance can be measured in the laboratory by placing the sample in a white diffusing sphere. Such an arrangement either (1) causes diffuse light to be incident upon the sample, or (2) allows the detector to view only the integrated reflected radiation

Most of the specular component of reflection is thus eliminated.

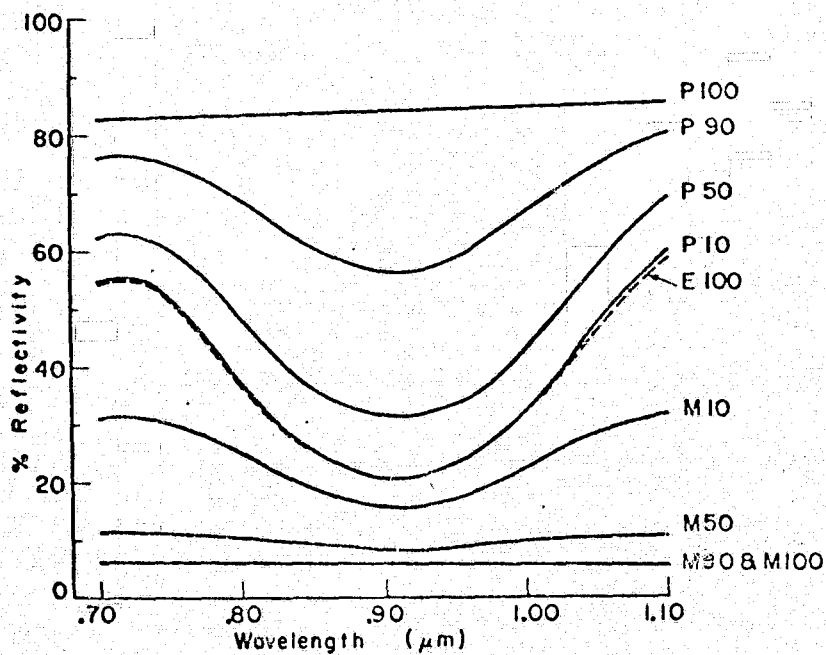
b. Mean optical path length (MOPL)

For any given material the strength of absorption features in a reflectance spectrum varies according to both the absorption coefficient and the average distance light is transmitted through the particles, or the mean optical path length (MOPL). For a mixture of materials, this MOPL is a complicated function of both the particle size and the general opacity of the components. Figure IIC -5 illustrates the effects of particle size on the strength of an



Diffuse reflection spectra of the absorption band of  $\text{En}_{59}$  for various particle sizes.

Figure IIC 5.



Diffuse reflection spectra of the absorption band of  $\text{En}_{59}$  for mixtures with plagioclase and magnetite. The particle size for all mixtures is  $< 125 \mu\text{m}$ .

Figure IIC 6.

absorption band centered at  $.91\mu\text{m}$  in the spectrum of an orthopyroxene,  $\text{En}_{89}$ . A measure of the band depth is listed in Table IIIC.-2. As the particle size decreases, the sample becomes brighter and the band depth decreases due to a reduction of the MOPL with increased scattering at particle interfaces. For a near transparent substance, such as  $\text{SiO}_2$  glass, the MOPL was estimated to be between 2-4mm for particles between  $500\text{-}250\mu\text{m}$  in diameter,  $<2\text{mm}$  for particles between  $250\text{-}125\mu\text{m}$ , and only hundreds of micrometers for particles  $<125\mu\text{m}$  (Pieters, 1972, Appendix A). These estimates of MOPL are upper limits; the MOPL for semitransparent materials would be smaller.

The general opacity of a material strongly affects the MOPL of reflected radiation. Figure IIC -6 contains spectra for mineral mixtures with particle size held constant. These mixtures contain various amounts of (1) a material with a well-defined absorption band (the orthopyroxene of Figure IIC -5), (2) a very absorbing but featureless material (magnetite), and (3) a featureless transparent material (iron-free plagioclase). The addition of transparent material brightens the mixture and the addition of absorbing material darkens the mixture, but in a nonlinear manner.

Table IIC 2. Band depth of samples measured

1	2	3	4
		$R_p(.91)/R_p(.73)$	
<u>SAMPLE</u>	<u>PARTICLE SIZE</u> (microns)	<u>DIFFUSE</u>	<u>90° phase</u>
E [250, 125]	250 > ps > 125	.29 <sup>1</sup>	.25
E [125, 63]	125 > ps > 63	.29	.31
E < 125	less than 125	.37	.38
E < 63	less than 63	.47 <sup>2</sup>	.44
Evf	very fine	.62 <sup>2</sup>	.56
P100	less than 125	1.01	1.04
P90	less than 125	.74	.77
P50	less than 125	.50	.53
P10	less than 125	.37	.40
E100	less than 125	.37	.38
M10	less than 125	.49	.52
M50	less than 125	.74	.75
M90	less than 125	.87	.91
M100	less than 125	.87	.92

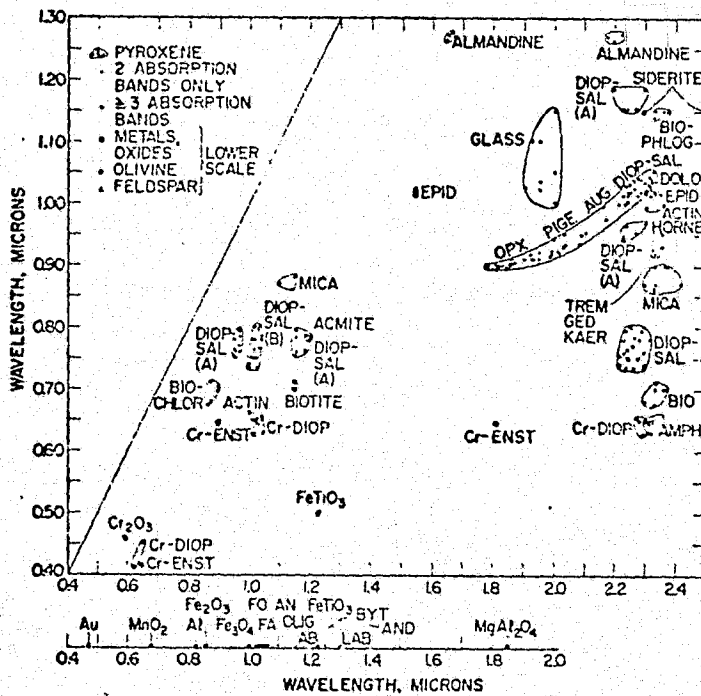
1 saturated band

2 partially backed

The reduction of band depth (Table IIC -2), by an addition of 10% magnetite, is equal to the reduction effects of a 50% addition of plagioclase. The absorbing material severely reduces the MOPL whereas the transparent material simply dilutes the spectral features.

c. Spectral features of minerals

The absorption features in reflectance spectra for a variety of mineral powders have been measured in the laboratory (see Section IIB for a review). Most of these features can be well understood from crystal field theory if the geochemistry of the material is sufficiently well defined. The uniqueness of absorption features in well characterized minerals has been examined by Adams (1975) and is best summarized in the " $\lambda$ - $\lambda$ " figure of IIC - 7 (from Adams, 1975). To form this omni-mineral figure, the energy of each absorption in a reflectance spectrum was discerned, and the short wavelength (high energy) band position was plotted against the long wavelength (low energy) band position. Those minerals that have more than two absorption features occur on the plot more than once; those minerals that have a single absorption band are plotted separately. Many of the sub-



Plot of centers of absorption bands in the diffuse reflectance spectra of several common minerals. Minerals having only one main band are plotted on the lower scale. Minerals with two or more bands in their spectra are designated by points in the main upper field. Hydroxyl and water bands at 1.4 and 1.9  $\mu\text{m}$  are not used. Bands for mineral species and for glasses plot in separate fields. See text for discussion. Diop-sal = diopside-salite; bio-phlog = biotite-phlogopite; aug = augite; pige = pigeonite; opx = orthopyroxene; dolo = dolomite; epid = epidote; actin-hornb = actinolite-hornblende; trem-ged-kaer = tremolite-gedrite-kaersutite; amph = amphibole; enst = enstatite; chlor = chlorite; Fo = forsterite; Fa = fayalite; Ab = albite; Olig = oligoclase; And = andesine; Lab = labradorite; Byt = bytownite; An = anorthite.

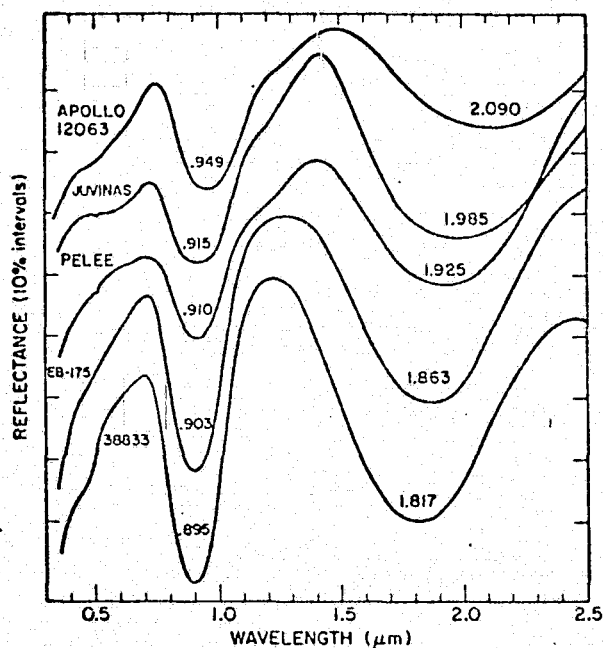
Figure IIC 7. (Adams, 1975)

tleties of a reflection spectrum are lost in such a figure, but it illustrates the fact that most minerals can be characterized by their absorption features. The major non-opaque constituents of lunar basalts (pyroxene, olivine, plagioclase, and glass) are discussed below.

i. PYROXENES. Two major spin-allowed  $\text{Fe}^{2+}$  absorption bands occur in the diffuse reflectance spectrum of pyroxene (Figure IIC -8). These bands centered near  $1\mu\text{m}$  and  $2\mu\text{m}$  define a pyroxene trend in Figure IIC2-5; they vary as a function of pyroxene composition in a well-ordered manner. Figure IIC -9 is a similar  $\lambda$ - $\lambda$  plot with only pyroxenes included. The distinction between orthopyroxenes and clinopyroxenes is clear in this figure. The trend of band centers proceeds to the right with increasing Fe and Ca content of the pyroxenes. The quantification of this compositional trend is shown in Figure IIC 10

(from Adams, 1975). Several pyroxenes (e.g., clinopyroxenes with  $\text{Cr}^{3+}$  or  $\text{Fe}^{3+}$ ) have a more complicated spectrum with more than two bands present. These pyroxenes define different but distinct regions in Figure IIC - 7.

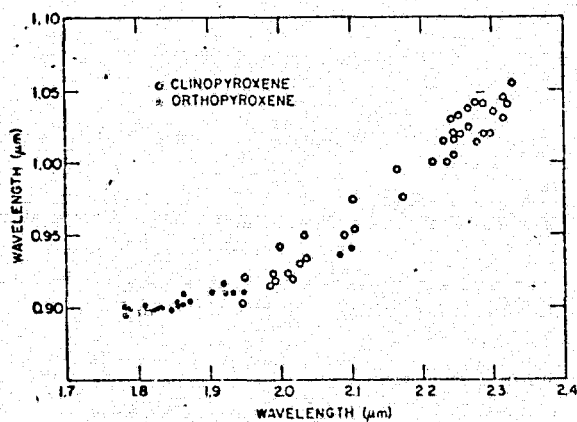
ii. OLIVINE. Several transitions occur for  $\text{Fe}^{2+}$



Diffuse reflectance spectra (relative to MgO) of pyroxenes. Samples 38833, EB-175, and Pelee are orthopyroxenes and increase in Fe in that order. Juvinas is a pigeonite, and Apollo 12063 is a subcalcic augite.

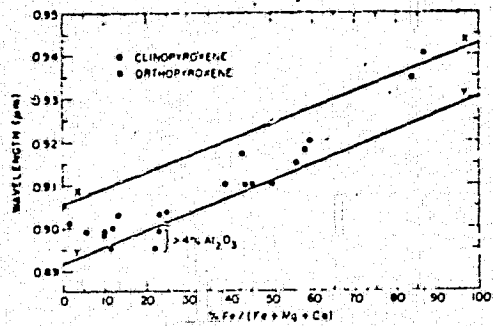
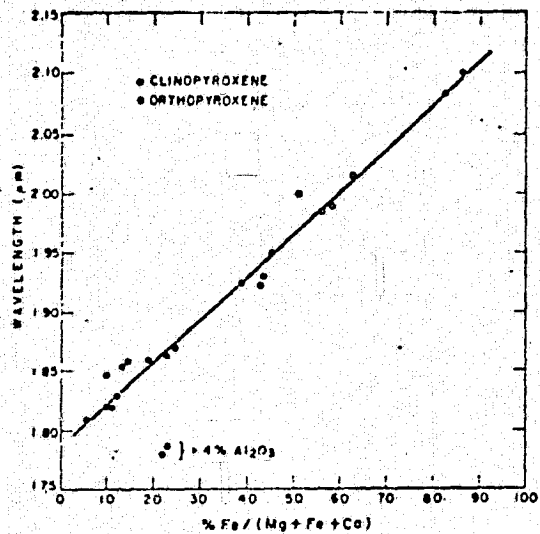
Note the two prominent  $\text{Fe}^{2+}$  absorption bands for all samples, and the shift of both bands to longer wave-lengths from bottom to top of the figure.

Figure IIC 8. (Adams, 1974)

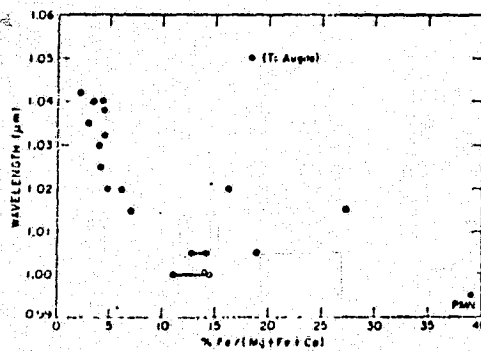
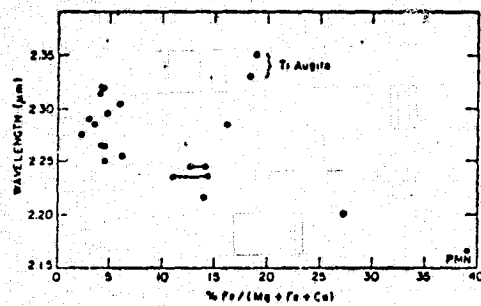


Wavelengths of centers of absorption bands near  $1 \mu\text{m}$  and  $2 \mu\text{m}$  in the diffuse reflectance spectra of pyroxenes. Orthopyroxene bands shift to longer wavelengths with increasing amounts of iron, whereas clinopyroxene bands shift to longer wavelengths with increasing calcium content.

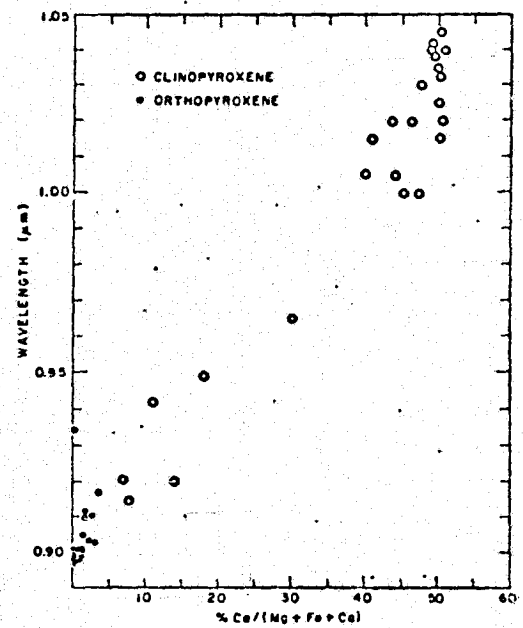
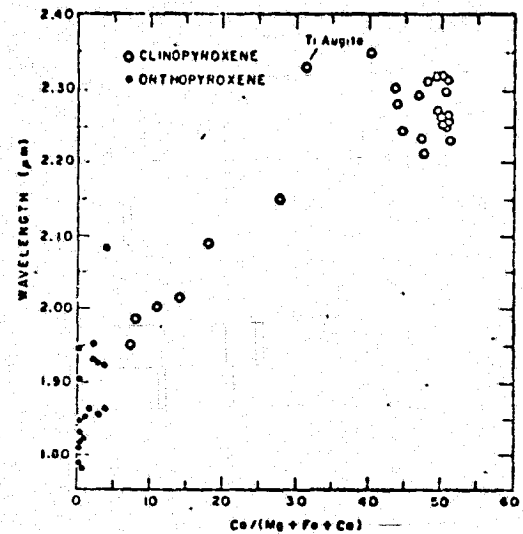
Figure IIC 9. (Adams, 1974)



Absorption band centers of orthopyroxenes and low-calcium clinopyroxenes as a function of  $\text{Fe}/(\text{Mg} + \text{Fe} + \text{Ca})$ . Lines labeled X and Y (bottom) indicate polarized absorption bands (Burns, 1965).



Centers of main absorption bands of calcic clinopyroxenes as a function of  $\text{Fe}/(\text{Mg} + \text{Fe} + \text{Ca})$ . Tie lines connect end members of zoned pyroxenes. PMN represents pyroxmangite.



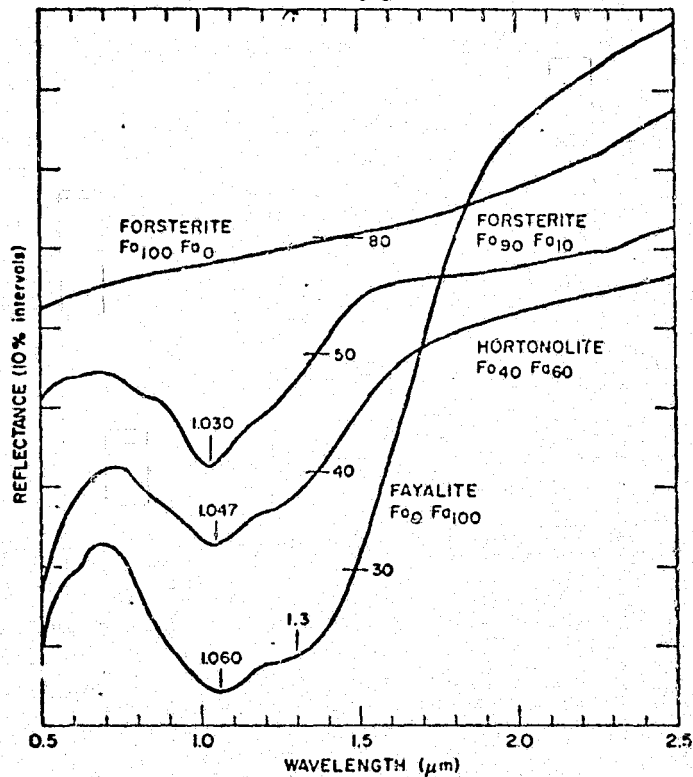
Centers of the main  $\text{Fe}^{2+}$  absorption bands of pyroxenes versus  $\text{Ca}/(\text{Mg} + \text{Fe} + \text{Ca})$ . Bands shift to longer wavelengths with increasing Ca.

Figure IIC 10. (Adams, 1974)

in the centrosymmetric  $M_1$  and the non-centrosymmetric  $M_2$  sites of olivines (Burns, 1970b). In a diffuse reflectance spectrum, the combination of these transitions is a broad non-symmetric band with a minimum between 1.0 and 1.1 $\mu$ m (Figure IIC-11). The depth of this composite band increases with increasing iron and the center shifts to longer wavelengths. The overall extinction coefficient for olivine is much less than that for pyroxenes; pure olivine ( $\sim$ Fo<sub>50</sub>) is relatively bright.

iii. FELDSPARS. Pure plagioclase has a featureless spectrum (see Figure IIC-6). However, when even small amounts of Fe<sup>2+</sup> and Fe<sup>3+</sup> are accommodated in the Ca or Al sites, distinct absorption features are discernable. When only Fe<sup>2+</sup> is present (as in the lunar case), the plagioclase feldspars show a broad crystal field absorption band with a minimum between 1.1 and 1.3 $\mu$ m as seen in Figure IIC-12 (from Adams and McCord, 1971a). It appears that the position of this band is controlled by the % anorthite (Adams, 1975).

iv. GLASS. Two broad absorptions generally occur near 1 and 2 $\mu$ m if Fe<sup>2+</sup> is present in a pure (not devitrified) glass. These two bands are not only much broader than those that occur in pyroxenes, but they are also closer together (Figure IIC-13). The 1 $\mu$ m band is due to Fe<sup>2+</sup>



Diffuse reflectance spectra (relative to MgO) of olivine minerals. The  $\text{Fe}^{2+}$  absorption band near  $1 \mu\text{m}$  increases in depth and shifts to longer wavelengths, and the  $1.3 \mu\text{m}$  inflection enlarges with increasing iron content of the minerals. Horizontal lines crossing the curves that are labeled with a two-digit number indicate percent reflectance.

Figure IIC 11. (Adams, 1975)

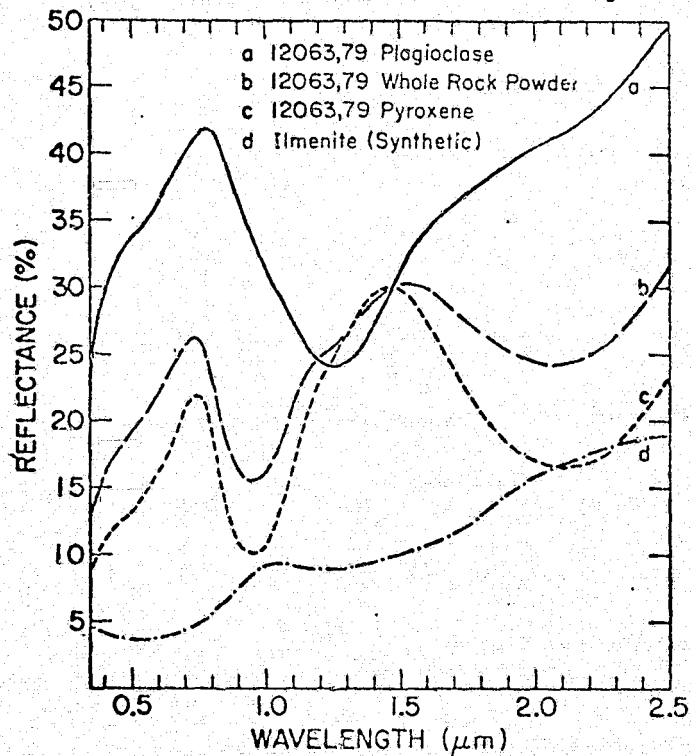


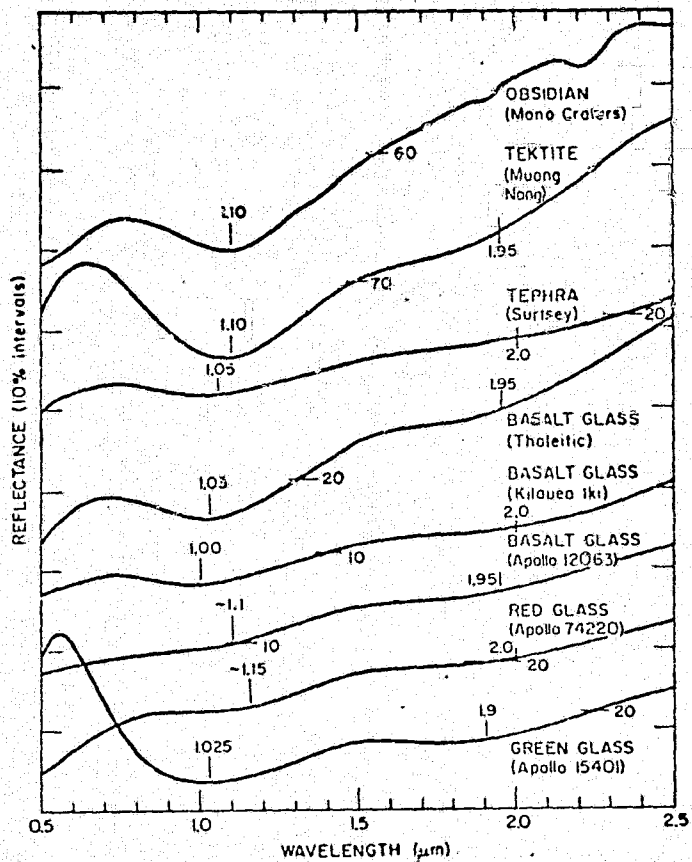
Figure IIC 12.  
(Adams and McCord, 1971a)

Spectral reflectivity of Apollo 12 basalt powder 12063 and plagioclase and pyroxene separates from the same rock. Ilmenite is a synthetic sample.

in octahedral sites whereas the  $2\mu\text{m}$  band, when present, is possibly due to  $\text{Fe}^{2+}$  in a tetrahedral site (Boon and Fyfe, 1972). Transmission spectra of synthetic glass indicate that the strength of the  $1\mu\text{m}$  band is directly related to the iron concentration in the glass (Bell et al., 1976). For lunar materials, the term "glass" has often been used to refer to the dark glass-welded aggregates abundant in soils. These agglutinates are an inhomogeneous mess of glass, mineral fragments, submicroscopic particles and sometimes native iron. A broad weak band exists near  $1\mu\text{m}$  in agglutinate spectra (Adams and Charette, 1977). The strength of this band appears to be related to FeO content of the soil but in a complicated manner dependent on the other ill-defined components of the agglutinates (see Section IIC-1a).

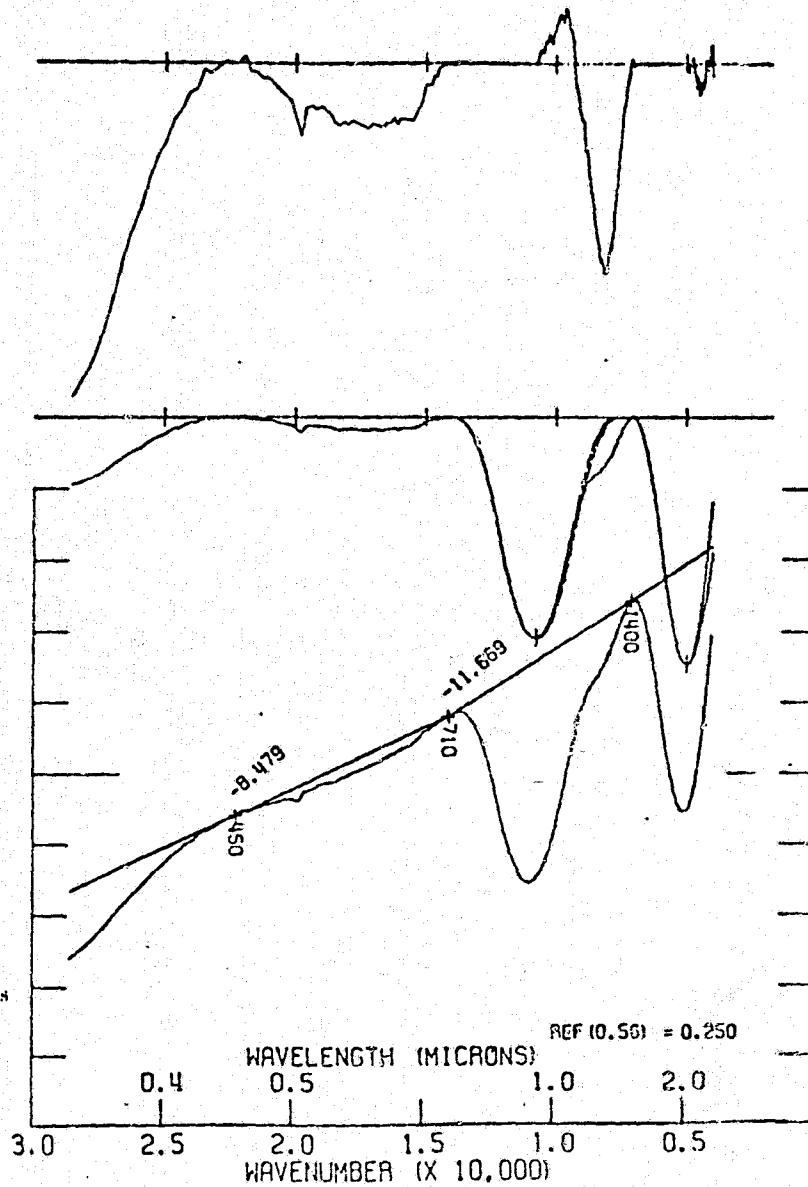
d. Spectral components of a whole rock spectrum

A rock is usually composed of a mixture of minerals and the strength of an absorption band is a function of the composition of the absorbing mineral, the average particle size, and also the opacity of the other components, with the latter often being a dominant effect. The spectral component from each constituent mineral of lunar rock 12063 can be seen in Figure IIC-12 (from Adams and McCord, 1971a). The whole rock spectrum is clearly dominated by



Diffuse reflectance spectra of iron-bearing glasses. The broad  $Fe^{2+}$  bands above  $1 \mu m$  and below  $2 \mu m$  are characteristic of glasses.

Figure IIC 13. (Adams, 1975)



2/28/73#7 HARAIYA ASU

(26)

Figure IIC 14. (Gaffey, 1975)  
Deconvolution of the Haraiya meteorite spectrum.

the pyroxene component (51% modal abundance). The prominent  $1.25\mu\text{m Fe}^{2+}$  band of the plagioclase (27%) is seen as a subtle but distinct inflection in the whole rock spectrum. Ilmenite darkens the spectrum.

Subtle features can nevertheless often be deconvolved from a spectrum using computer processing techniques (Gaffey, 1975). There are two basic difficulties which must be overcome in order for a rock or soil spectrum to be deconvolved into the spectral components for each mineral present. First, the general slope of the continuum must be removed. (This means the wings of poorly-defined high energy (UV) absorptions that always effect the slope of the visible and near infrared spectrum must be eliminated.) Secondly, the stronger absorption bands need to be sufficiently well defined and removed so that remnant weaker bands can be examined. Figure IIC-14 illustrates such a deconvolution of the spectrum for the Haraiya meteorite (from Gaffey, 1975). As described in the previous section, the wavelength and symmetry of each absorption band is directly a function of the mineralogy of the sample; the relative strength of bands from different minerals in a whole rock spectrum can be used to estimate the relative modal abundances of constituent minerals (Gaffey, 1975).

## II. REMOTE SENSING OF THE MOON

### D. Spectral Reflectance of Lunar Material

The application of spectral reflectance measurements to understanding the lunar surface incorporates both the laboratory measurements of returned lunar samples and the telescopic measurements of small lunar regions (10 to 20 km in diameter). An additional source of spectral information comes from multi-spectral imagery which allows general spectral differences to be mapped in two dimensions. (Most of the spectral data relevant to the following discussion is derived from coordinated research of two groups--one organized by McCord at MIT/Hawaii and the other by Adams at University of Washington, Seattle.)

#### 1. Soils

As mentioned in the review of Section IIB, the laboratory measurement of a representative soil sample from Apollo 11 was in close agreement with the telescopic measurement (.3 to 1.1 $\mu$ m) of the landing site (Adams and McCord, 1970). Figure IID-1 (from McCord and Adams, 1973) illustrates the agreement between the telescopic spectra and the laboratory spectra of soils returned from subsequent missions. The early studies by Adams and McCord demonstrated (1) that the telescopic spectra of small regions were com-

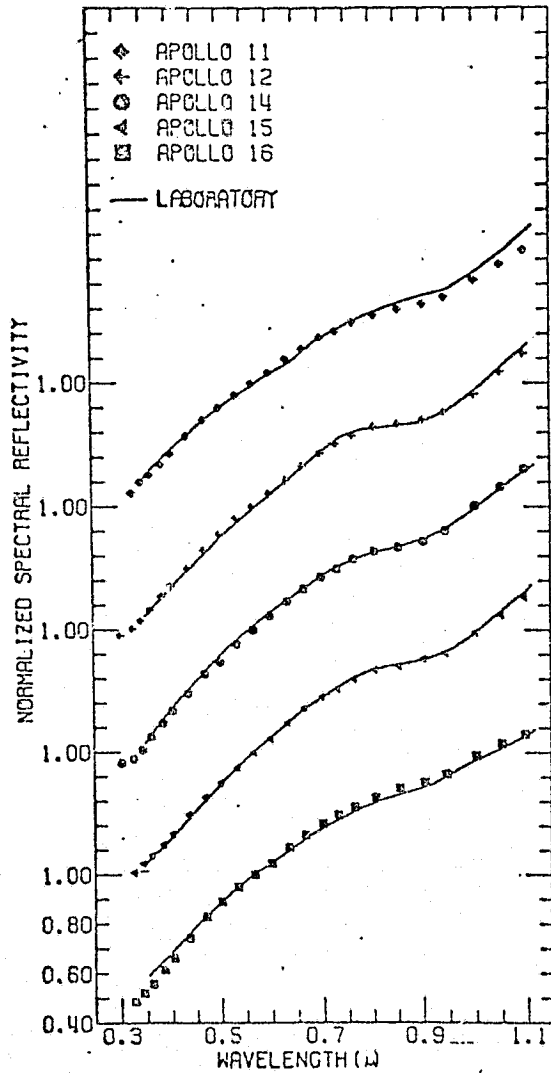


Figure IID 1.  
Spectral reflectance of soil from the Apollo landing sites. Points are telescopic spectra; solid lines are laboratory measurements.

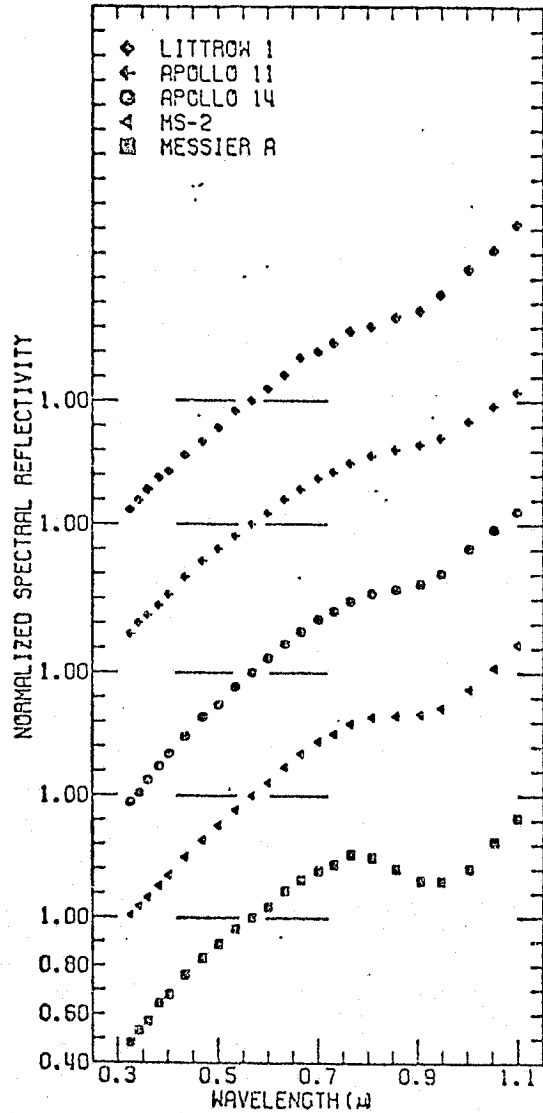


Figure IID 2.  
Spectral reflectance measurements for representative lunar areas (telescopic).

23

parable to spectra of soils, rather than rock fragments or chips, and (2) that well-chosen soils can be considered representative of the surrounding 10 - 20 km regions measured telescopically.

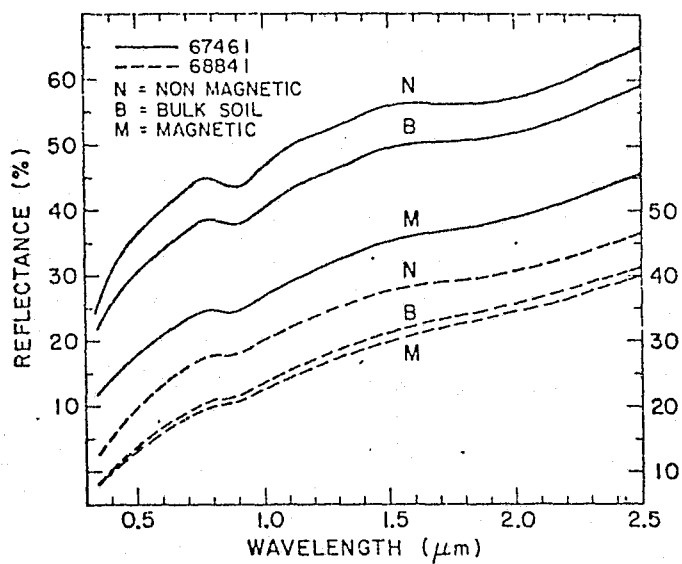
When compared to other solar system objects and terrestrial samples, mature lunar soil has a distinct spectrum: a steeply sloped red continuum with a very weak absorption band near  $1\mu\text{m}$  and sometimes also  $2\mu\text{m}$  (Figure IID-2). Mercury is the only object with a comparable red spectrum, but without distinct absorption bands (McCord and Adams, 1972; Vilas and McCord, 1976; McCord and Adams, 1976; Adams and McCord, 1977). Only through extensive study of the lunar samples has the significance of this peculiar character of lunar soil spectra been illuminated.

Lunar soils are largely derived from the prevalent rock type of the region (with a small amount of contamination due to lateral transport of impact material). The main weathering mechanism is largely micrometeoroid bombardment, although the processes involved in soil development are not all defined. As the soil matures, particles are finely crushed and broken and also welded together by impact generated melts into friable aggregates called agglutinates (e.g., McKay et al., 1974). The subsequent accumulation of these dark agglutinitic particles dramatically

changes the mineralogy of the soils from that of the parent rocks, although the bulk composition only undergoes a minor (contamination) alteration. As a soil matures, the spectral changes that occur are mostly due to the replacement of mineral fragments with the absorbing glass-welded agglutinates (Adams and McCord, 1973). The spectrum of a mature soil is almost wholly dominated by the spectral characteristics of the multi-component agglutinates. A discussion of the special character of agglutinates is thus presented here before a general discussion of soil spectra.

a. Agglutinates

Adams has used a magnetic separation technique to separate the agglutinitic and non-agglutinitic components of a soil (Adams and McCord, 1973). Shown in Figure IID-3 (from Adams and McCord, 1973) are spectra for bulk soil (B) samples 67461 and 68841, the agglutinitic (magnetic--M) separate, and the non-agglutinitic (non-magnetic--N) component. The albedo of the bulk soil is intermediate between that of the agglutinitic and non-agglutinitic concentrates. Mature soils contain  $\approx 60\%$  agglutinates as measured by the magnetic separation technique. The chemistry of lunar soil may be somewhat partitioned between agglutinitic and non-agglutinitic components with the ferromagnesian elements concentrated in the agglutinates (Rhodes et al.,



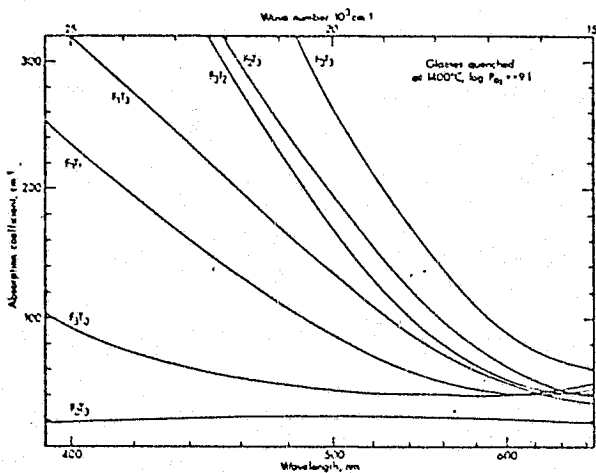
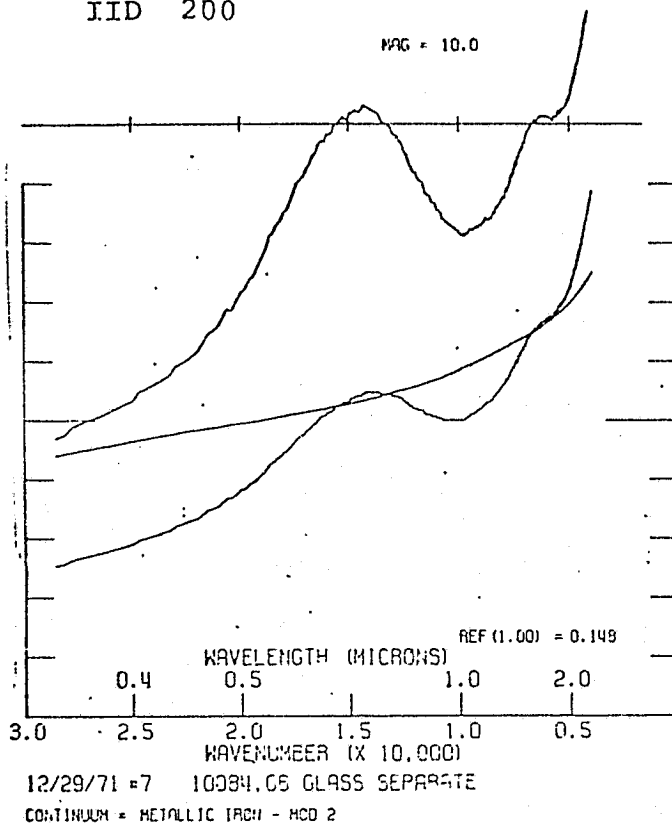
Spectral reflectance curves (relative to MgO) of Apollo 16 soil samples 67461 and 68841, and of their magnetic and non-magnetic fractions. The magnetic fraction consists of glass-welded aggregates (agglutinates) and always is darker than the parent soil.

Figure IID 3. (Adams and McCord, 1973)

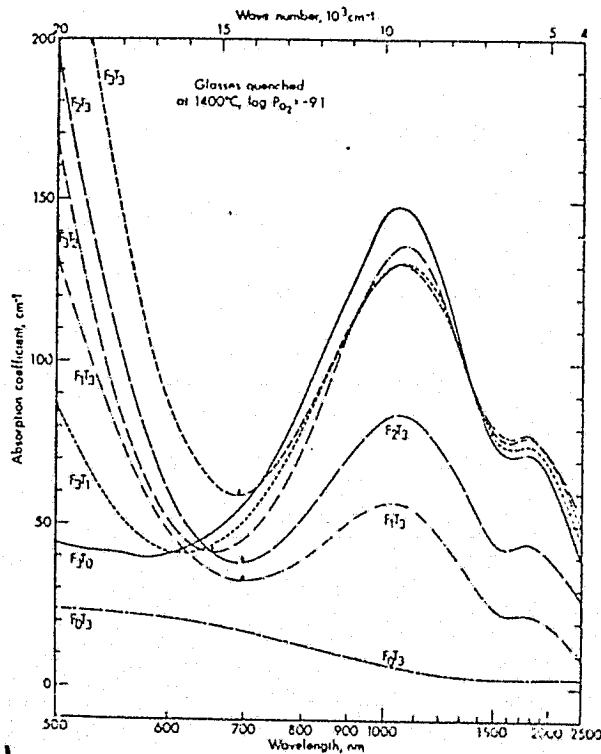
1975). Since it is these agglutinates that dominate the spectrum of such soils, it is useful to understand the spectral nature of a number of key components that have a special character in lunar agglutinates: Fe-Ti bearing glasses, Fe<sup>0</sup> (microscopic), and ilmenite and other opaque phases. The spectral character of the mineral fragments, which are significant components in agglutinates, but often with subtle spectral effects, were discussed in greater detail in Section IIC-2c.

i. Fe-Ti RICH GLASS. A relatively pure Fe-Ti rich glass sample was separated from soil sample 10084. The spectral reflectance properties of this glass are shown in Figure IID-4. The color of lunar glasses is caused primarily by various electronic transitions between and within Fe<sup>2+</sup>, Ti<sup>3+</sup> and Ti<sup>4+</sup> ions. Using transmission spectroscopy and controlled experimental techniques, Bell et al. (1976) have carefully documented the spectral effects of iron and titanium concentrations in synthetic glasses as a function of oxygen fugacity. These concentration systematics were shown to be equally valid for reflectance spectra of glasses (Wells and Hapke, 1977). Figures IID-5a,b (from Bell et al., 1976) illustrate the following spectral characteristics for glasses produced at oxygen fugacities comparable to the lunar condition: (1) An iron-rich, titanium

Figure IID 4.  
Spectral reflectance  
of a glass separate  
from Apollo 11 soil.



a. Spectral region 400-600 nm for the glasses whose slope at 20,000 wave numbers is plotted in Fig. 5.



b. Continuation of the observations plotted in Fig. a to 2500 nm (four wave numbers).

Figure IID 5. (Bell et al., 1976)  
Absorption of glasses containing  
various amounts of iron and ti-  
tanium. F<sub>0</sub>--F<sub>3</sub> indicates increasing  
iron; T<sub>0</sub>--T<sub>3</sub> indicates increasing  
titanium.

free glass is weakly absorbing throughout the visible region (minimum absorption between .5 and .6 $\mu\text{m}$ ; i.e., green glass) and contains a well-defined  $\text{Fe}^{2+}$  broad absorption band (crystal field) centered around 1.0 $\mu\text{m}$ . (2) If titanium is included in an iron-rich glass, a strong absorption occurs toward the blue and ultraviolet (i.e., red glass), the strength of which is related to the amount of titanium. The strength of this absorption is also very sensitive to the effective oxygen fugacity. The  $\text{Fe}^{2+}$  1.0 $\mu\text{m}$  feature is only slightly affected by the addition of titanium. (3) Only the concentration of iron affects the strength of the broad 1.0 $\mu\text{m}$  feature. (4) The concentration of both iron and titanium affect the strength of the UV absorption (with titanium perhaps playing a stronger role.) (5) In a Ti-rich glass without iron, a relatively weak absorption is detected near .5 $\mu\text{m}$  due primarily to  $\text{Ti}^{3+}$  (crystal field absorption) although  $\text{Ti}^{3+} \rightarrow \text{Ti}^{4+}$  may also occur (charge transfer absorption). This feature is associated with a purple color of such glasses and increases in strength with lower oxygen fugacity. These transitions were identified by Burns et al. (1976).

Due to the strong absorption towards the ultraviolet, iron- and titanium-bearing glasses are certain to have a significant effect on the continuum slope of reflectance spectra for mature lunar soils. The very broad

1 $\mu$ m feature common to reflectance spectra of all mature lunar soils is typically a result of FeO in the agglutinitic glass (Charette and Adams, 1977b).

ii. Fe<sup>0</sup>. Small particles of iron metal were observed in the first returned lunar soils (Agrell, 1970). From magnetic and Mossbauer techniques, it was further shown (e.g., Housley et al., 1973) that submicroscopic metal grains are enriched in mature soils. Evidence has been accumulating that these fine metal grains have a significant effect on the spectrum of a lunar soil even though they account for <1% of the bulk composition. Hapke et al. (1975) propose that the principle darkening effect for lunar soils is submicroscopic Fe<sup>0</sup> formed in impact events and solar wind sputtering. Gaffey (1976) showed that although metal iron is opaque to visible and near infrared radiation, its spectral character in reflected radiation is a distinctive smooth but red continuum. For meteorite specimens, a red continuum is clearly associated with Fe<sup>0</sup> content in a mineral mixture. This is an observed and well-documented effect (Gaffey, 1974, 1976) although theoretical descriptions of how iron particles interact with light in a mineral mixture are complicated by the fact that iron is a conducting material rather than a semi-transparent dielectric. A key difference between lunar agglutinitic

metal and that found in meteorites is particle size--the lunar particles in agglutinates are submicroscopic ( $\sim 200\text{\AA}$ ) whereas the meteoritic iron is easily identified with normal petrographic techniques. Mie theory calculations by Conel (1970) show that the extinction due to scattering and absorption of radiation by submicroscopic particles of iron or ilmenite is much stronger in the blue and ultraviolet than in the near infrared. The overall spectral effects of submicroscopic metal iron inclusions, therefore, is to preferentially absorb blue and ultraviolet radiation (whatever the cause). With further theoretical and laboratory analyses, a distinction between scattering and reflection effects should be possible.

Recently Bell and Mao (1977) have been able to obtain a transmission spectrum for a portion of a lunar agglutinate that is rich in submicroscopic  $\text{Fe}^0$  particles. For comparison they produced thin ( $10 - 100\text{\AA}$ ) coatings of  $\text{Fe}^0$  on transparent powders (silica glass). The results are shown in Figure IID-6. Their data show an increased absorption towards the ultraviolet associated with the metal iron. Although they could not guarantee that no oxidation had occurred during the laboratory analysis for both the lunar sample and the coated grains, the similarity between the two iron-bearing spectra is striking.

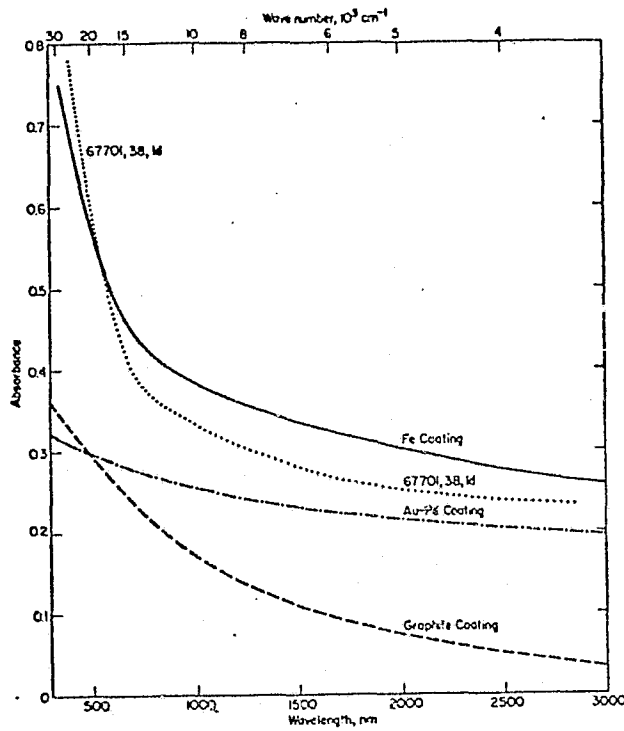


Figure IID 6.  
(Bell and Mao,  
1977)

Optical spectra of Fe, Au-Pd alloy, and graphite coatings on silica glass, and the optical spectra of agglutinate specimen 67701,38,ld.

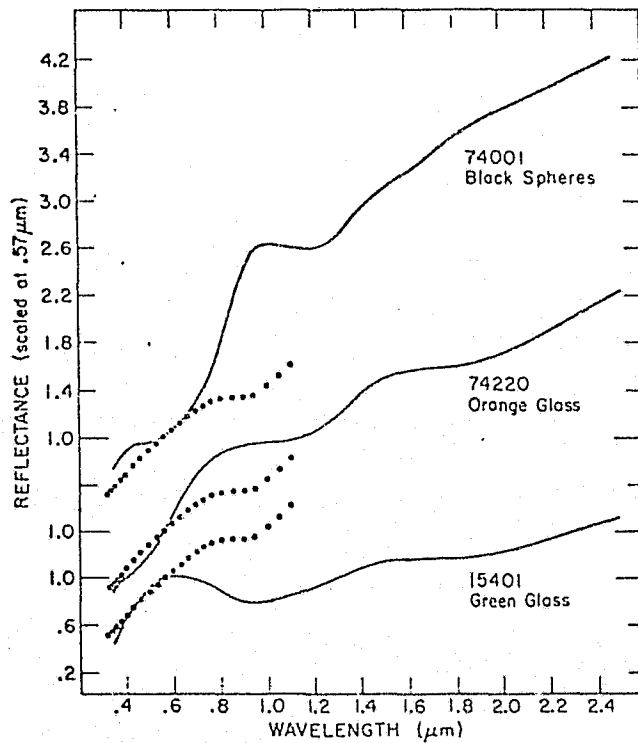


Figure IID 7.  
(Adams et al.,  
1975)

Spectral reflectance for the orange glass and related soils. Superimposed on each spectrum is the spectrum for the standard lunar area MS-2 (closed circles).

A purely empirical piece of evidence for the effects of  $\text{Fe}^0$  on mature soil spectra results from attempts to characterize the lunar continuum sufficiently well to examine superimposed absorption characteristics. It was found that a fit of Gaffey's (1977) iron metal continuum to a lunar soil spectrum accounted remarkably well for the red slope in the near infrared. Such a continuum was removed from the spectra in Figure IID-9-16 to examine the  $1\mu\text{m}$  feature in better detail. For very dark soil, it appears the continuum approximation of  $\text{Fe}^0$  is not as good as for bright soil.

Although the effects of submicroscopic iron particles on a reflectance spectrum of lunar soils are currently poorly understood, it is likely that they are significant. Small amounts of metallic iron probably contribute to the darkening of the soil and help account for the distinctive red continuum of lunar soils.

iii. ILMENITE AND OPAQUES. Ilmenite is the most abundant opaque mineral in lunar samples and accounts for up to 25% of modal abundances for the igneous basalts. Ilmenite is present in lunar agglutinates as remnant mineral fragments, some of which are very finely dispersed throughout the glassy matrix. Although the dominant effect of opaque ilmenite particles in a soil is to darken the

soil and lower the spectral contrast of other mineral features (see Section IIC-2d), ilmenite is not without its own spectral character. Two prominent absorption features near .65 and 1.25 $\mu$ m were observed for the unique thin ilmenite plates in the glassy spheres of 74001 (Figure IID-1-7, from Adams et al., 1975). The finest grain size in basaltic rocks do not generally allow the spectral character of ilmenite to contribute to a rock spectrum. However, if the particles are sufficiently small, they could interact significantly with the reflected radiation. Small ilmenite particles primarily contribute to the darkness of the agglutinitic material but for some samples may also allow the 1.25 $\mu$ m ilmenite band to slightly affect the symmetry of the 1 $\mu$ m absorption in a soil spectrum.

iv. MINERAL COMPONENTS. The spectral characteristics of the other major mineral components in lunar basalts, and hence in soil agglutinates, was discussed in Section IIC. Briefly, these include: two strong, symmetric absorption bands near 1 and 2 $\mu$ m from pyroxenes; a broad, nonsymmetric absorption band near 1 $\mu$ m from olivine; and a weak band near 1.3 $\mu$ m from iron-bearing plagioclase. Mineral spectral features are strongest in immature soils and barely discernable (if at all) in mature soils.

b. Soil spectra

The mineralogy of mature lunar soils is largely a combination of remnant mineral and lithic fragments and complex, multi-component agglutinates. Table IID-1 lists various soil components, their spectral absorptions and the approximate magnitude of effect from these absorptions on a soil spectrum. In order to derive useful compositional information from a bulk soil spectrum, it is necessary to isolate and identify each particular absorption.

i. STRONG ABSORPTIONS. Although albedo differences are close to a factor of two between maria and highland mature regions, it is remarkable that the spectral differences between such regions (.3 - 1.1 $\mu$ m) are generally less than  $\sim 10\%$  (McCord et al., 1972). There are at least three competing strong absorptions that dominate the general structure of a mature (agglutinate-rich) mare soil spectrum throughout the visible and near infrared spectral region: the Fe-Ti bearing glass, the finely dispersed iron, and opaques (primarily ilmenite). All of these absorptions darken the soil, preferentially through the visible toward the ultraviolet. All three of these components exist in mare soils and are thus expected to contribute spectrally to some degree. Which (if any) plays the primary role in all lunar soil spectra is the subject of much

TABLE IID-1 Effect of various absorption features on soil or rock spectra.

<u>Soil or Rock Component</u>	<u>Absorptions</u>	<u>Effects on Spectra of:</u>		
		<u>Mature Soil</u>	<u>Immature Soil</u>	<u>Rocks</u>
Pyroxene	.90-1.0 $\mu$ m	subtle	strong	dominant
	1.8-2.3 $\mu$ m	subtle	strong	dominant
Plagioclase	1.3 $\mu$ m	trace?	subtle	strong
Olivine	1.0 $\mu$ m	-	subtle	strong
Ilmenite	UV-IR	strong	subtle	subtle
	.6, 1.2 $\mu$ m	trace?	-	-
Glass (Fe-Ti bearing)	UV-IR	strong	subtle	-
Glass (Fe-rich)	1.0 $\mu$ m	subtle	subtle	-
Fe <sup>0</sup>	UV-IR	strong?	subtle	-

discussion and will not be resolved here. A few observations and conclusions, however, can be made independent of this controversy.

All the strong darkening effects depend on Fe or Fe + Ti. Thus, for soils of equal maturity (same accumulation of agglutinates) the more Fe- and Ti-rich soils are darker. This point is obvious in the gross distinction between highland and maria and also seems to hold for Apollo 14 soils relative to Apollo 16. For mare soils, however, other factors are also present.

For mare soils, the strong absorption in the blue and ultraviolet approaches saturation. For example, as the iron and titanium content increase, soils do not get redder as they would if Fe-Ti spectral absorption were the only effect (Bell et al., 1976); instead they get bluer (Charette et al., 1974). This can be understood by referring to the simple diagram in Figure IID-8. Shown are three hypothetical spectra with stronger absorption in the blue (B) relative to the red (R). Whatever is responsible for the B/R color, spectrum II is affected by twice as much absorbing material as I, and spectrum III by twice as much as II. In transmitted light, the absorption strengths would follow the ratio 1:2:4. If the increased

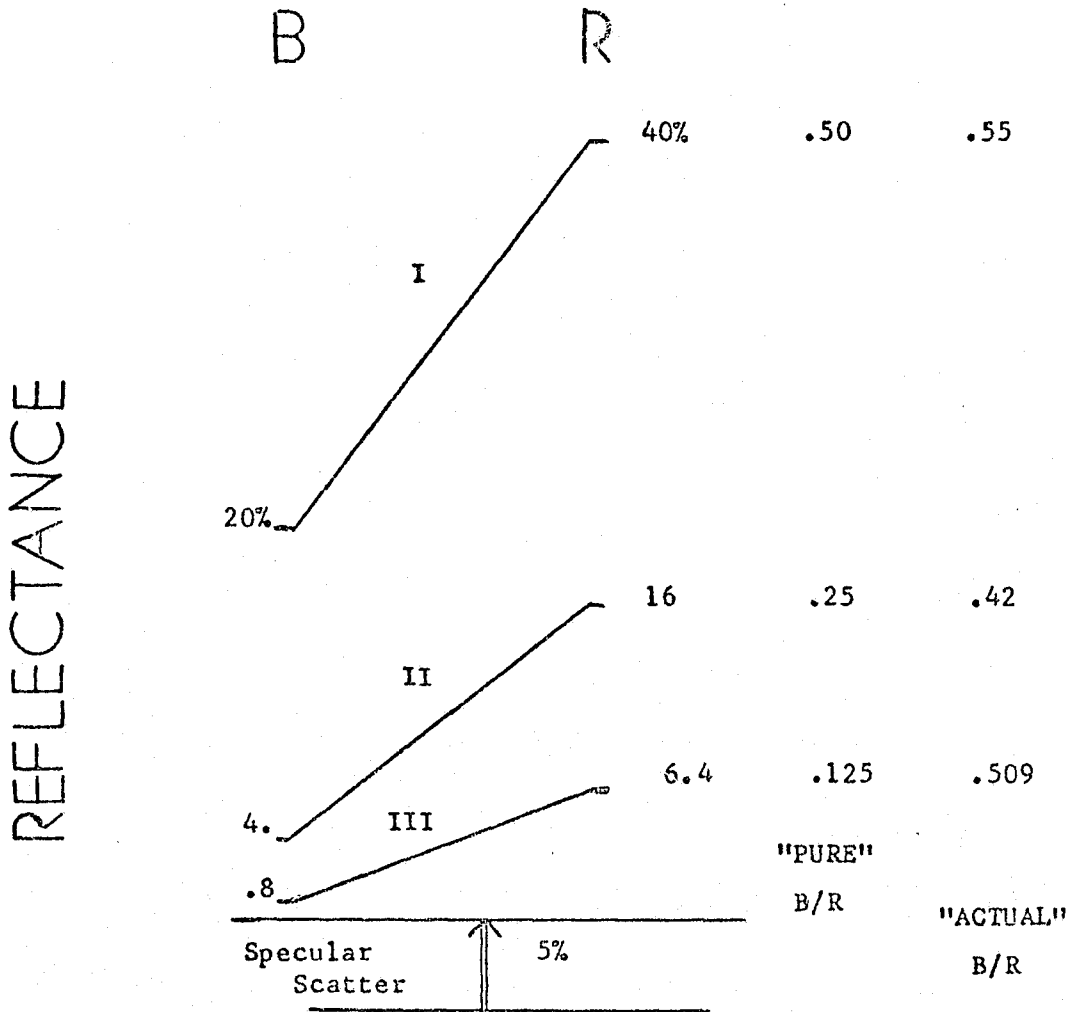


Figure IID 8. Effects of specular scatter on spectral contrast.

absorption was the only factor that influenced the reflectance, the color ratio B/R should be .5, .25 and .125 for spectra I, II, and III respectively. However, even totally opaque particles scatter a significant amount of radiation from specular reflection without imparting any spectral information. This specular scattered component is estimated to be 5% here. (A lower bound of 3-1/2% can be inferred from the far UV albedo measurements of the moon (Luke et al., 1973) since all surface material is effectively opaque at those wavelengths.) When the scattered component is included in reflected radiation, the actual color ratios measured (see Figure IID-8) would show spectrum II to indeed be redder than I (although not as much as it should be). The dark spectrum III, however, would actually be bluer than spectrum II. Thus, increasing a (red) absorption strongly darkens the soil so much that spectral contrast is gradually lost and the spectrum becomes flatter, or relatively bluer. Highland soils may be bright enough to discern real absorption differences (e.g., becoming redder with increasing iron in Fe-Ti-bearing glasses). Mare soils are nearly saturated in the blue and ultraviolet and become bluer with increasing Ti-content, presumably because of Fe-Ti glass absorption. The concept of spectral contrast for mare soils has recently been reexamined by Adams (1977).

Thus, two general spectral parameters for mature soils that can be interpreted compositionally are albedo and continuum slope. Note that the interpretation of measured continuum slope is dependent on the overall brightness of the surface.

ii. SUBTLE ABSORPTIONS. There are a number of small subtle features superimposed on the general red continuum of lunar soils, the most prominent of which is a small absorption near  $1\mu\text{m}$  in the spectra of Figure IID-1 and 2. In order to examine these in greater detail, the red continuum must be estimated and removed. A number of techniques for determining this continuum have been tried. If this step is not performed correctly, then measured band centers can be offset from actual centers and erroneous compositional interpretations can result.

One of the most promising estimates for the lunar continuum is the iron metal continuum used by Gaffey (1977). Such a continuum has been used for a variety of lunar soil samples shown in Figures IID -9-16 (plotted as a function of energy). In these figures a metal continuum has been fitted through the laboratory sample spectra at  $.73$  and  $1.5\mu\text{m}$ . The spectra have been scaled to unity at  $1.0\mu\text{m}$ . The fit seems to be a reasonably good

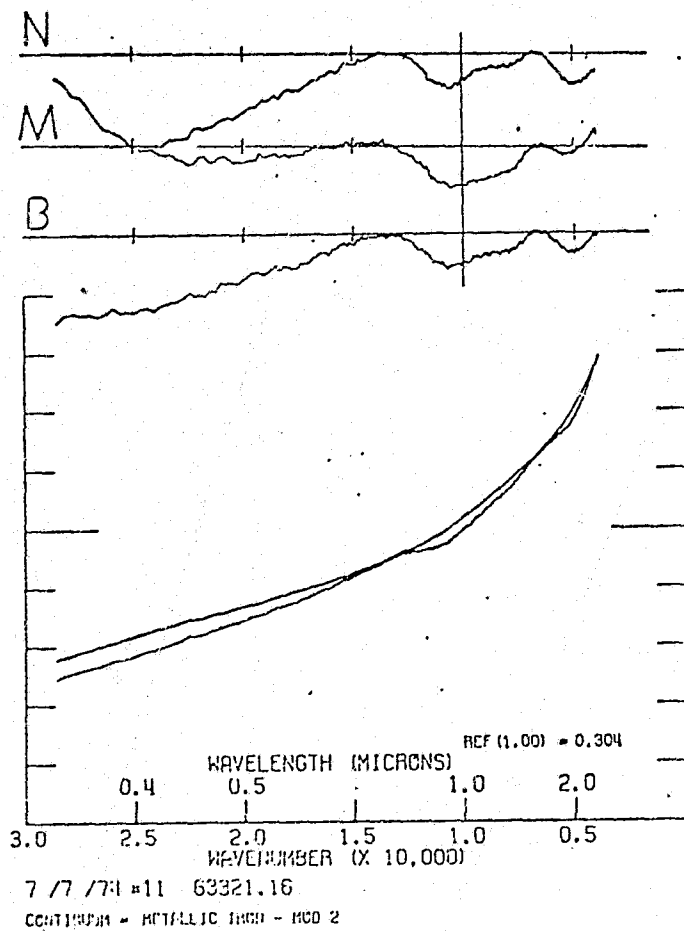


Figure IID 9. Spectral reflectance for lunar soil 63321. An iron continuum has been fit to the spectrum. Shown above the spectrum are residual absorptions after the removal of the continuum for the bulk soil (B), magnetic separate (M), and nonmagnetic separate (N).

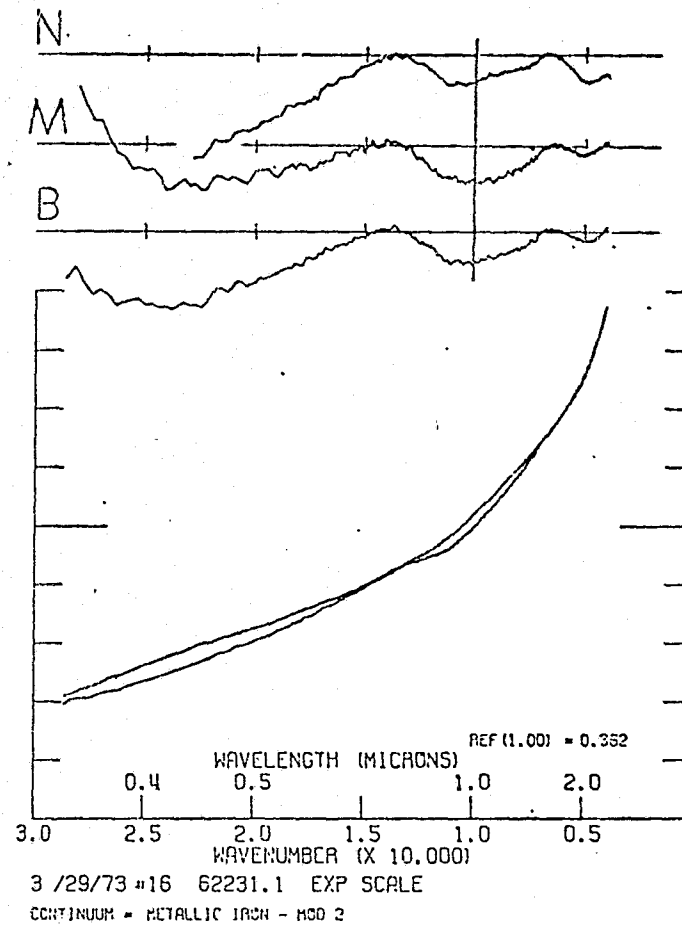


Figure IID 10. Similar to Figure IID 9, for lunar soil 62231.

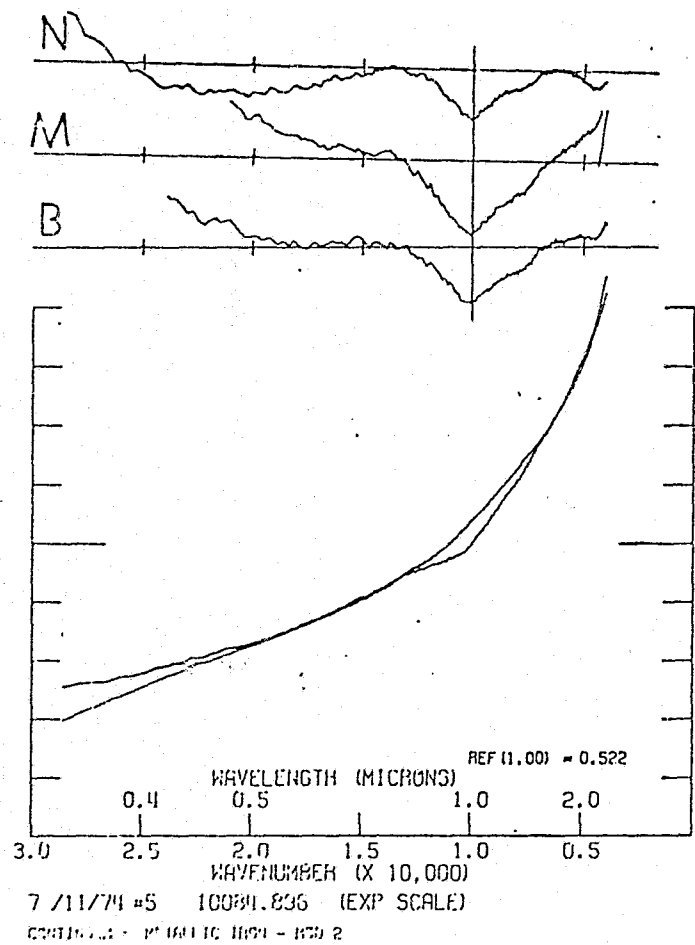


Figure IID 11.  
Similar to Figure IID 9,  
for lunar soil 10084.

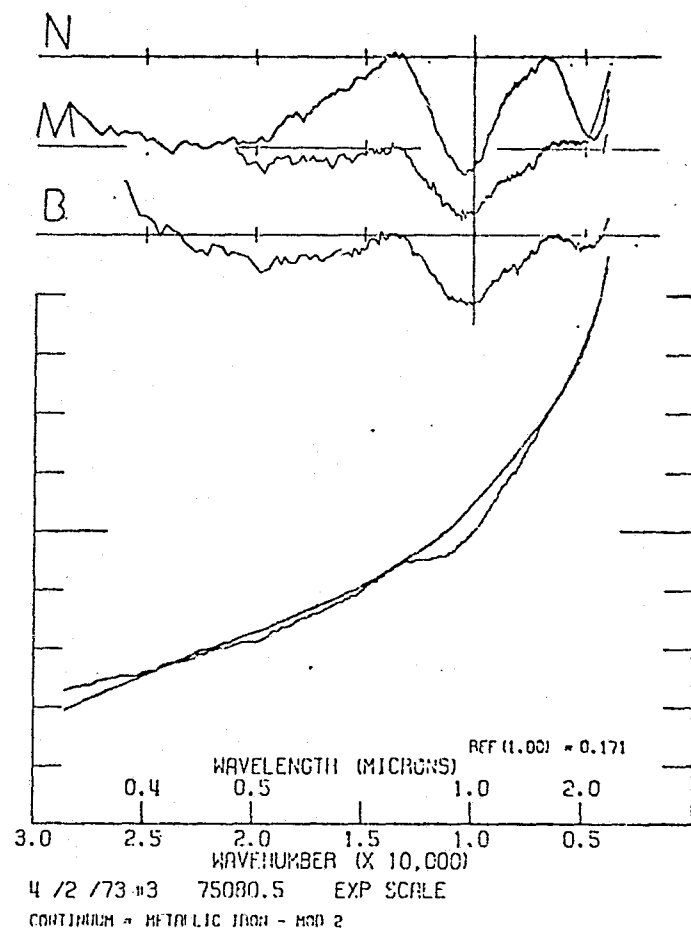


Figure IID 12.  
Similar to Figure IID 9,  
for lunar soil 75080.

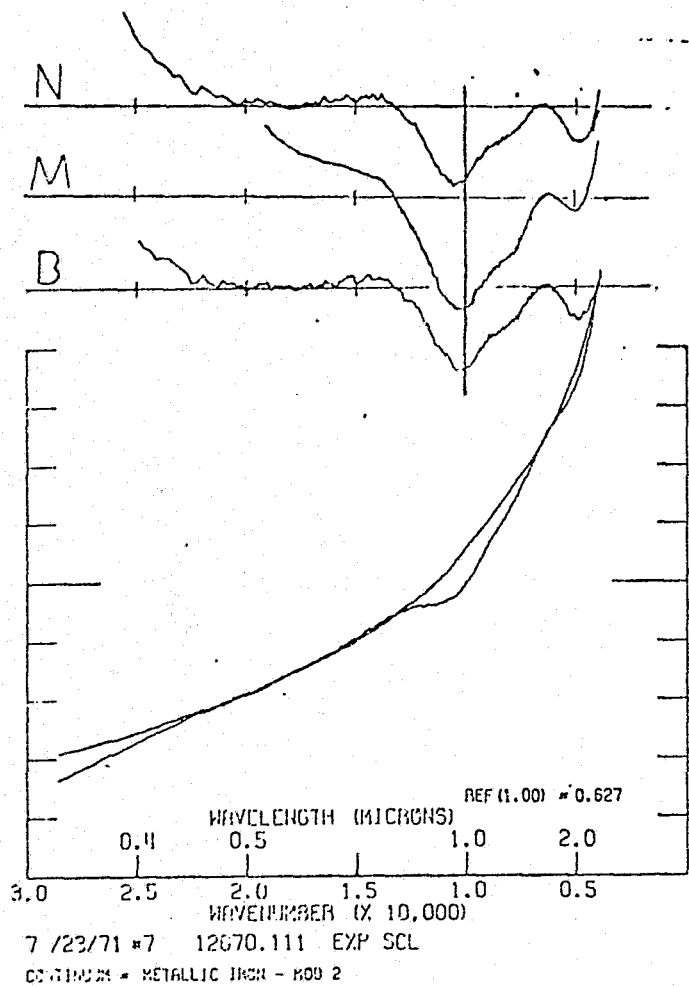


Figure IID 13.  
 Similar to Figure IID 9,  
 for lunar soil 12070.

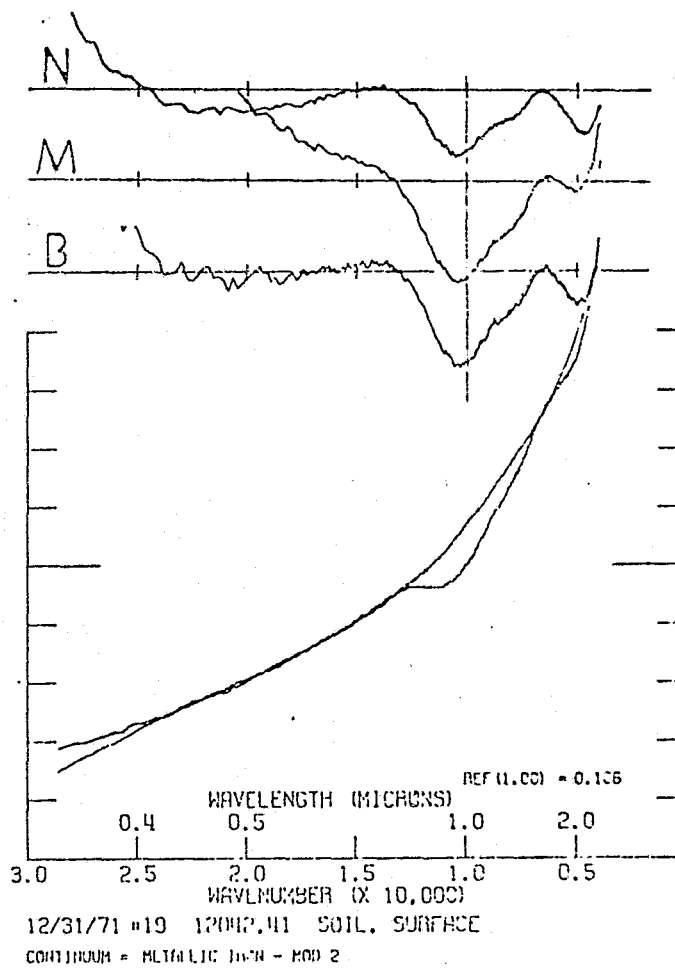
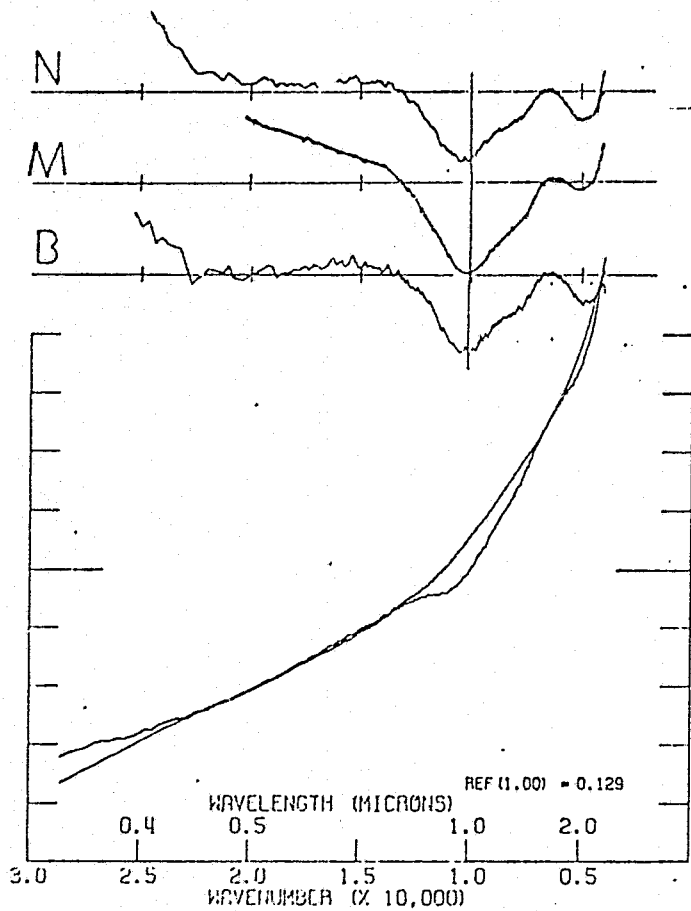
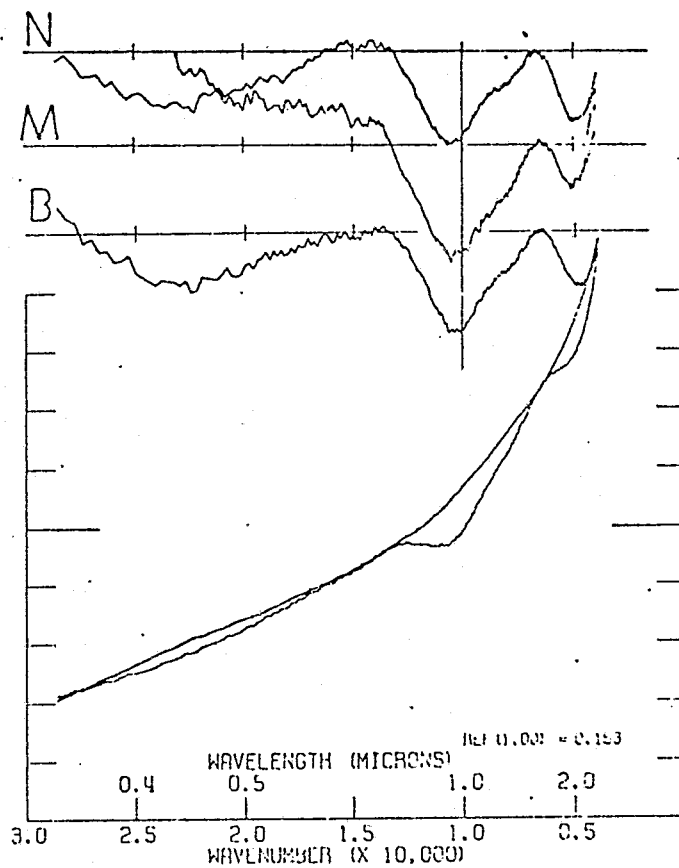


Figure IID 14.  
 Similar to Figure IID 9,  
 for lunar soil 12042.



12/27/71 #21 15021.114  
 CONTINUA = METALLIC IRON - ROD 2

Figure IID 15.  
 Similar to figure IID 9,  
 for lunar soil 15021.



1 / 5 / 72 #9 15471.37  
 CONTINUA = METALLIC IRON - ROD 2

Figure IID 16.  
 Similar to Figure IID 9,  
 for lunar soil 15471.

approximation above  $5000\text{\AA}$ . A ratio has been formed between the continuum and bulk soil spectrum (B) and is displayed in the figures above the spectrum. Similar ratio spectra were formed for the magnetic (M) and nonmagnetic (N) soil separates and are displayed above the bulk soil spectra. Most of the remnant mineral fragments are concentrated in the nonmagnetic fractions (Adams and McCord, 1973).

In most of the spectra presented in Figures IID9-16, the following subtle features can be noted: (1) There is a broad composite feature near  $1\mu\text{m}$  with the peak off-center to shorter wavelengths and a distinct shoulder towards longer wavelengths. (2) There is often a second less intense feature near  $2\mu\text{m}$ , the intensity of which varies greatly. It is unclear whether the variations shortwards of  $.7\mu\text{m}$  are real or simply an effect of the continuum removal procedure. For the reasons outlined below the  $1\mu\text{m}$  soil feature is interpreted as a composite pyroxene and glass band and the  $2\mu\text{m}$  feature is primarily the 2nd pyroxene band (Adams and Charette, 1977).

Sample 63321 (Figure IID-9) is an immature soil (magnetic component or agglutinate content  $\sim 32\%$ ) and the non-magnetic fraction (N) is dominated by mineral fragments. The pair of absorptions at  $\sim .91$  and  $1.9\mu\text{m}$  can be clearly interpreted as orthopyroxene (Adams, 1974). A

second band, almost resolvable at  $\sim 1.25\mu\text{m}$ , is likely to be due to feldspar. This interpretation is in good agreement with the mineralogy of the sample and the Apollo 16 site in general. The spectral features of these two minerals are severely degraded in the bulk soil spectrum (B) by a broad deep feature near  $1\mu\text{m}$  most noticeable in the magnetic separate (M).

This broad  $1\mu\text{m}$  feature is tentatively identified as the  $\text{Fe}^{2+}$   $1\mu\text{m}$  glass band. Examine the spectrum for mature soil 62231 (Figure IID-10) which has a bulk composition similar to 63321 ( $\text{FeO} \sim 4\%$ ,  $\text{TiO}_2 \sim .35\%$ ) but with agglutinates dominating the soil mineralogy. The bulk sample and non-magnetic separate spectra only contain a trace of the orthopyroxene features. The magnetic separate spectrum is dominated by the broad  $1\mu\text{m}$  feature. Since very few mineral fragments remain in this agglutinate sample, the  $1\mu\text{m}$  feature is likely due to the  $\text{Fe}^{2+}$  glass band described in Section IID1a. (The band observed here may be slightly offset to shorter wavelengths by a hint of remnant orthopyroxene.) Even though it is weak in this sample, the broad symmetric band at  $\sim 1.0\mu\text{m}$  contains the same spectral properties as those identified in glass separates (Figure IID-4) and with transmission spectroscopy of glass (Figure IID-5).

Turning to the more mafic mare soils (Figures IID1-11-16), all spectra appear to contain a combination of various proportions of pyroxene and glass features. The non-magnetic fraction of the least mature soil shown here (75081, agglutinates ~30%) shows a spectrum dominated by clinopyroxene with characteristic bands near .95 and 2.1 $\mu$ m (Adams, 1974). There is little evidence for any contribution of the glass band. In the magnetic separate spectrum, the .95 $\mu$ m band is degraded and the 2.1 $\mu$ m band almost eliminated as the broad glass band begins to appear.

All the other mare soils show a composite clinopyroxene and glass band at 1 $\mu$ m. The glass band is the strongest in the magnetic agglutinate fractions and weakest in the non-magnetic fractions. The .95 - 1.0 $\mu$ m clinopyroxene band seen in the non-magnetic separates shifts to slightly longer wavelengths as it is diluted by the 1 $\mu$ m glass band.

Note that for glass-rich mare soils of approximately equal FeO content, the depth of the 1 $\mu$ m band is less strong for the dark soils (Apollo 11) than for lighter soils (Apollo 12 and 15). This lower spectral contrast of the 1 $\mu$ m band is probably due to an effect similar to that illustrated in Figure IID-8: if significant absorption has already occurred throughout the spectrum (dark

soil), the effects of a second absorption (the glass band) are not as strong as they would be for brighter soils. On the other hand, since highland mature soils are generally bright, an increase in the FeO content is expected to increase the  $1\mu\text{m}$  band depth (Charette and Adams, 1977a).

In summary, subtle spectral features can be identified in spectra of mature lunar soils: a substantial iron glass band and also weaker pyroxene bands (Adams and Charette, 1977). It is unclear whether small components of plagioclase and ilmenite in mare soils can contribute a detectable feature near  $1.3\mu\text{m}$ . For less mature soils, the pyroxene bands are stronger and a plagioclase band may be identified. If these weak bands can be accurately resolved, the composition of the pyroxene and (perhaps) plagioclase can be identified. As mentioned previously, correct identification is in part dependent on how accurately the effects of the red continuum have been removed. Since the strength of the  $1\mu\text{m}$  glass band has been shown to vary as a function of FeO content (Bell et al., 1976), it should be possible to estimate the FeO content of a mature soil from the depth of the band (Charette and Adams, 1977b). In order to accomplish this, the  $1\mu\text{m}$  pyroxene band must also be removed from the composite  $1\mu\text{m}$  band, and the lower

spectral contrast effects of dark soils would have to be calibrated to lighter soils. Even without spectral contrast calibration, differences in the depth of the  $1\mu\text{m}$  feature can nevertheless be interpreted in terms of relative FeO content if one compares mature soils of similar albedo and assumes that the minor pyroxene content is similar and that the soil development processes have homogenized all soils to the same degree.

## 2. Rocks and Craters

Since most of the lunar surface is covered with a mature soil, the emphasis of many detailed spectral reflectance studies of lunar material has been directed towards understanding the properties of these soils. It is recognized that the mineralogy of lunar rocks, on the other hand, is used by petrologists to identify the major lunar basalt types. Currently, there is a lot of activity (Adams and associates) to obtain spectra for a sufficient number of lunar rock powders and chips in order to identify spectral characteristics of the various distinct rock types identified by the petrologists. Although a number of preliminary reports have become available (e.g., Charette and Adams, 1975, 1977a) this work is still in progress. Thus, this section will only briefly discuss some of the current laboratory reflectance results, the likely output, and the difficulties inherent in the undertaking. Telescopic spectra of craters (the closest lunar surface analogue to rock samples) are discussed in Section IV.

Shown in Figure IID-17 are reflectance spectra (unscaled) of a variety of lunar mare rock powders and chips. The spectra have been grouped by Adams and Charette

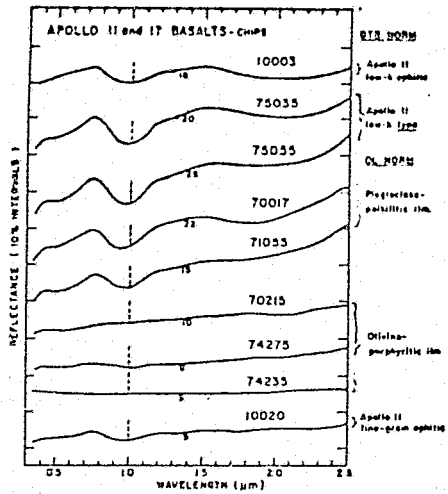


FIGURE a. Apollo 11 and 17 basalt (interior) chips. Basalt classifications by Warner (2) for Apollo 11 and Papike et al. (7) for Apollo 17. Dashed vertical lines are placed at 1.0  $\mu\text{m}$  over each spectra. The integers below each spectra are the absolute reflectance values at the marked wavelength.

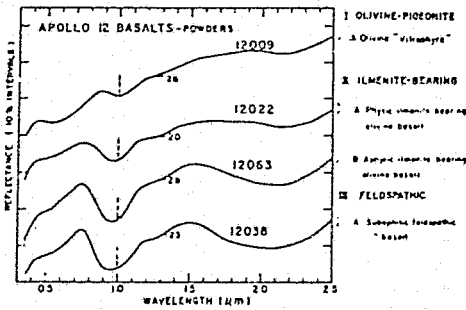


FIGURE b. Apollo 12 basalt powders. Classification by James and Wright (3).

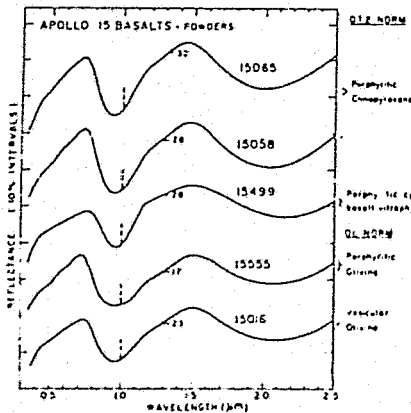
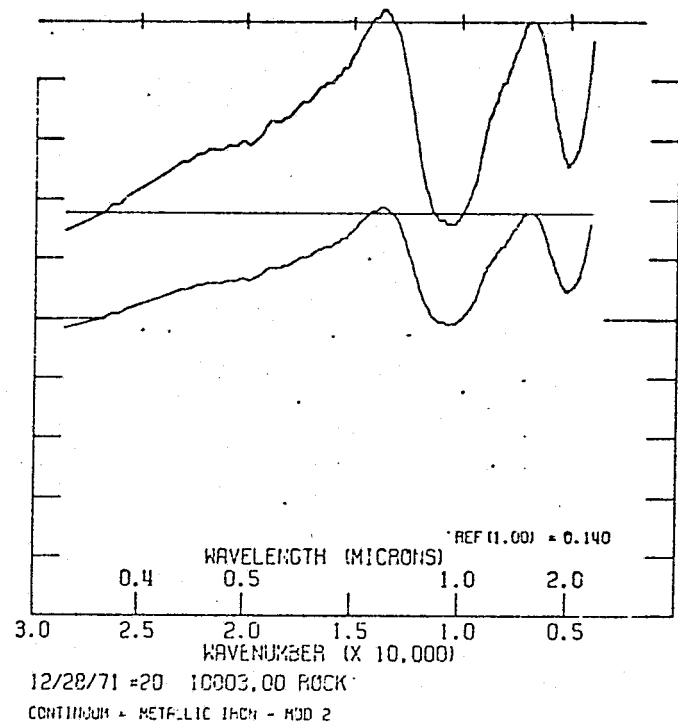
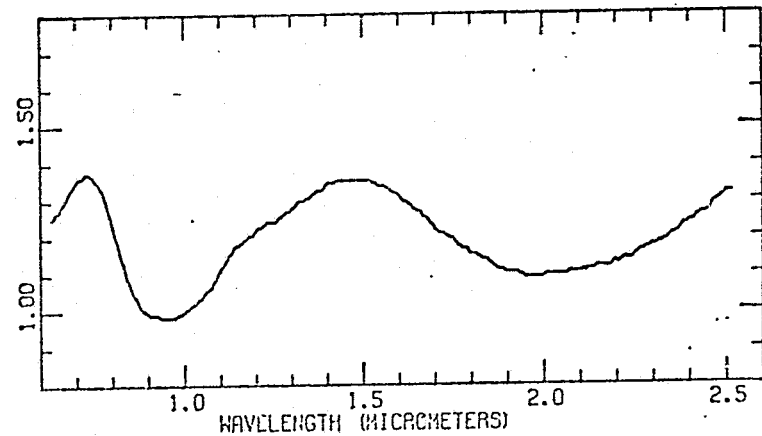


Figure IID 17. (Charette and Adams, 1975)

FIGURE c. Apollo 15 basalt powders. Classification by Rhodes and Hubbard (4).



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0 12/28/71 = 20 10003.00 ROCK Figure IID 18. Spectral reflectance of lunar sample 10003 (expanded vertical scale).

according to rock types identified by Warner (1971), James and Wright (1972), Rhodes and Hubbard (1973), and Papike et al. (1974). The albedo for each spectrum is marked for one wavelength. Some of the principle conclusions by Charette and Adams include: (1) Differences in basalt types can be recognized in reflectance spectra as the presence or absence of olivine bands, pyroxene bands at different wavelengths, plagioclase bands, and ilmenite bands. (2) Intra-site differences are often more easy to distinguish than inter-site differences since some basalt types having different origins may nevertheless have similar spectra (mineraology).

The spectral contrast of the spectra in Figure IID-17 is dependent on the effective grain size--which is smaller for powders than for chips. The grain size of the rock and its degree of brecciation also effect this spectral contrast. Fortunately, computer processing techniques can be used to enhance the details of a spectrum. Compare the spectra of 10003 in Figure IID-18 with the top spectrum in Figure IID-17a. As discussed in Section IIC2, one should be able to deconvolve such spectra into features due to specific mineral components such as pyroxene, plagioclase, olivine, ilmenite and glass. If a spectra deconvolution is successful the composition of minerals such as pyroxene can be identified. To some extent, the relative

proportions of mineral phases can also be described from the relative strength of mineral absorption bands.

Thus, the first step in laboratory spectral reflectance studies of lunar rocks and powders by Adams and associates is to identify the specific mineral spectral features that can be associated with the major lunar rock types. This project necessarily draws on an extensive background of reflectance studies of minerals and rocks. A number of related programs will likely ensue in order to have a solid background for interpreting remotely obtained spectra for fresh lunar surfaces:

(a) The individual spectral features for minerals in mixture of minerals are quantitatively identified using computer processing techniques. This would eventually allow the inverse process--the creation of likely spectra for possible mineral mixtures. This step is essential if quantitative identification of the mineralogy of unknown assemblages is desired.

(b) It is unlikely that most lunar craters have excavated a single rock type. It is thus necessary to deal with the complicated spectral character of known and possible mixtures of unrelated rock types. For example, varying the proportions in a mixture of pigeonite, clinopyroxene, olivine, and plagioclase should have small, but definable,

effects on a rock spectrum. Since the variations will be subtle as the proportion of minerals change, computer processing techniques will be required for detection and quantification of the spectral changes.

(c) The mineralogy that can be determined from reflectance spectra must be translated into terms that are meaningful to other geochemists studying lunar samples. Currently, for example, the pyroxenes in lunar mare rock types are distinguished by the compositional trends during crystallization. Spectral reflectance techniques, on the other hand, will be able to identify the composition (Ca, Fe, Mg) of the average pyroxene or perhaps the relative proportions of different pyroxenes (pigeonite, sub-calcic augite, etc.). The average pyroxene composition of a regional basalt type will be a major parameter for distinguishing mare basalt types using remote sensing techniques. The concept of "average pyroxene composition" needs to be considered as a possible meaningful geochemical parameter by lunar geochemists who are concerned with lunar basalt types on the smaller scale of the lunar samples.

### 3. Telescopic Spectral Measurements

Although the general lunar spectrum is distinct from spectra of all other solar system objects (except Mer-

cury) the spectral variations from place to place on the moon are small (<10%). Much of the effort concerning telescopic lunar spectral reflectance measurements has been to identify such spectral differences and to derive interpretations from the parallel lunar sample studies for observed spectral characteristics. Quantitative analysis requires precision better than 1% which, for telescopic observations, has only been available for about 10 years (McCord, 1968a). Current telescopic data available with adequate precision include spectra (.3 to 1.1 $\mu$ m) obtained for small lunar areas (10-20 km diameter) with a filter photometer (McCord, 1968a) and two-dimensional spectral images at wavelengths between .4 and 1.0 $\mu$ m obtained with a vidicon digital imaging system (McCord et al., 1975, 1976b).

It is anticipated that near-infrared spectra (.65 - 2.5 $\mu$ m) will soon be available for selected lunar regions using recently developed telescopic instrumentation (McCord, 1977). Continuous infrared spectra are exceptionally difficult to obtain due to atmospheric absorptions. Preliminary results (McCord, 1977) show that thermal emission begins to become detectable beyond 2.1 $\mu$ m under full moon viewing conditions. Theoretical calculations, (R. Clark, in preparation) indicate, however, that this minor component of lunar radiation between 2 - 2.5 $\mu$ m can be defined

sufficiently to be removed.

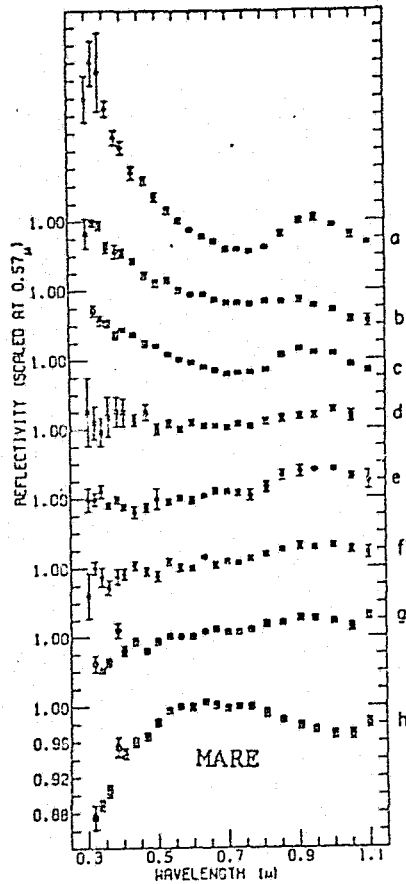
a. Relative Reflectance Spectra

Spectra for a variety of lunar regions were shown in Figures IID-1 and 2. Both the strong and subtle absorptions that affect the lunar spectrum were discussed in the previous sections. To distinguish the small differences between spectra, a technique is used by which each spectrum is divided by the spectrum of a standard area with the resulting spectrum being a relative reflectance spectrum. The standard area commonly used is MS-2 in central Mare Serenitatis ( $18^{\circ} 40' N$ ,  $21^{\circ} 21' E$ ). The spectrum for MS-2 is second from the bottom in Figure IID-2. Note that this spectrum is not perfectly smooth and contains an easily identifiable  $1\mu m$  feature. Thus, although all systematic spectral differences observed in relative reflectance spectra are significant, the position and symmetry of absorption features are not absolutely accurate but rather are relative to the features observed for MS-2. Relative reflectance spectra can be obtained from telescopic observations with minimal processing since both the atmospheric and instrumental effects usually cancel in the ratio.

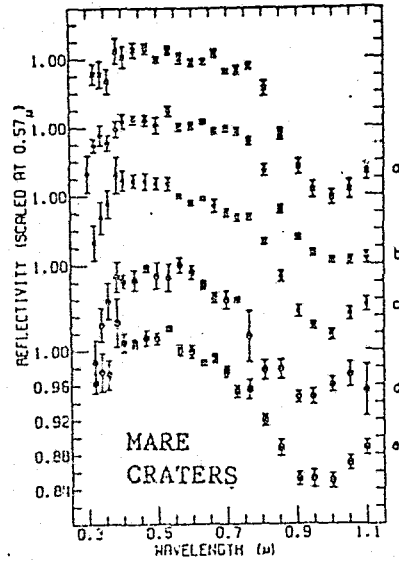
When relative reflectance spectra were obtained for over 150 regions, a classification was made according to the general morphology of the region: mare, upland, mare crater, and upland crater. The spectral types evident

in spectra for these groups were discussed by McCord et al. (1972a) and by Charette et al. (1974). A fifth group of regions (dark mantled areas) formed an additional distinct spectral type and was discussed by Pieters et al. (1973). Examples of relative reflectance spectra for the four major spectral types are shown in Figure IID-19. Spectra for the maria show a wide range of continuum slope in the blue and ultraviolet that was correlated with  $\text{TiO}_2$  content (see Section IID1). Mare crater spectra show a strong absorption feature near  $1\mu\text{m}$  characteristic of pyroxenes (see Section IV). Most highland regions appear spectrally similar (although work in progress indicates there are a variety of subtle distinctions between highland surface units). Highland craters can be easily distinguished from mare craters. See Section IIB for a review of some of the applications of these lunar spectral types.

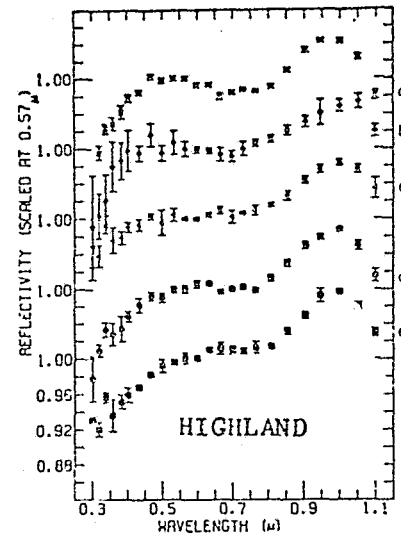
An important aspect of this classification for telescopic data was that similar spectral types could be seen in laboratory data (Adams and McCord, 1972). Recall that relative reflectance spectra show the very small and subtle spectral differences between two surfaces. The fact that the same systematic small features also appear in laboratory measurements of soils shows fairly conclusively that the laboratory measurements are indeed comparable to



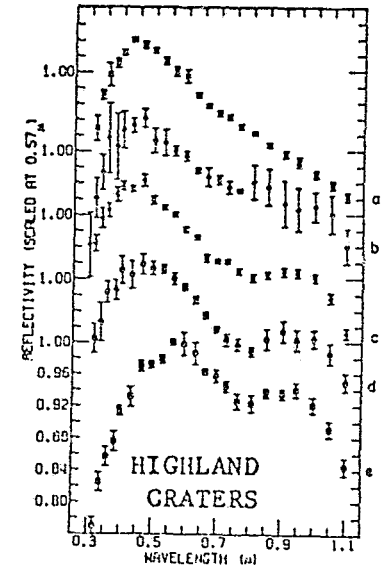
Relative spectral reflectivity for a series of background mare areas ranging from 'blue' mare (Sea of Tranquility 1) to 'red' mare (Sea of Cold 2). (a) Sea of Tranquility 1; (b) Sea of Moisture 0; (c) Luna 16 landing site; (d) Fra Mauro 15; (e) Le Monnier; (f) Sea of Moisture 51; (g) Plato C; (h) Sea of Cold 1.



Relative spectral reflectivity for (a) Linne; (b) Sea of Moisture 45; (c) Messier A; (d) Kepler; (e) Misting C.



Relative spectral reflectivity for (a) Descartes 3; (b) Fra Mauro 7; (c) uplands 7; (d) Sea of Moisture 41; (e) Alphonsus 2.



Relative spectral reflectivity for (a) Proclus; (b) Curviorinus 2; (c) Descartes 2; (d) Fra Mauro 6; (e) Tycho 1. Relative reflectivity is the ratio of the reflectivity of one area to that of the standard area (in this study an area in the Sea of Serenity).

Figure IID 19. (from McCord et al., 1972)

to the telescopic measurements even though they differ by  $\approx 10^6$  in area of surface measured. Shown in Figure IID-20 are reflectance spectra for representative laboratory soils relative to MS-2 (from Adams et al., 1977). In order to form these spectral ratios, the laboratory spectra were transformed to the same spectral bandpasses as used for the telescopic observations. It is clear that the laboratory spectra for soils contain the same subtle features observed in the purely telescopic data of Figure IID-19.

Currently spectra for close to 400 lunar areas exist. Relative reflectance spectra for many of these areas have been used extensively in the analysis and applications described in Sections III and IV.

b. Spectral Imagery

The early relative reflectance studies also indicated there were certain regions of the spectrum that were critical in characterizing various surface units. The wavelengths .38, .40, .57, .77, .95 and  $1.06\mu\text{m}$  were chosen for two-dimensional spectral mapping (e.g., McCord et al., 1976). Three particularly useful wavelengths are .40, .57, and  $.95\mu\text{m}$  (Charette et al., 1974).

If an image is obtained at  $.40\mu\text{m}$  for a given lunar region and a similar image at  $.57\mu\text{m}$ , the two images can be processed and digitally divided using computer processing techniques. The resulting  $.40/.57\mu\text{m}$  ratio image

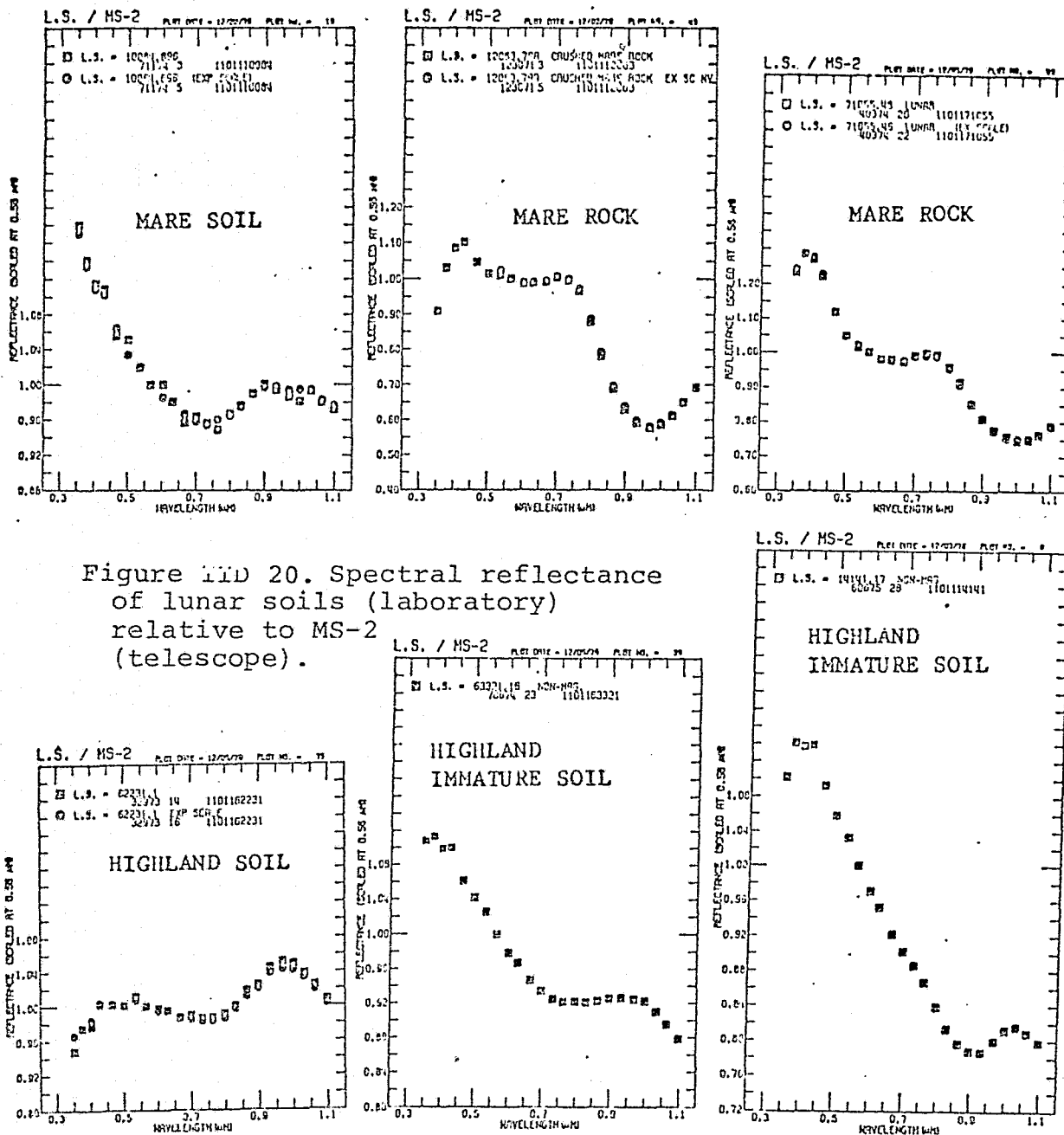


Figure IID 20. Spectral reflectance of lunar soils (laboratory) relative to MS-2 (telescope).

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contains quantified spatial information on continuum slope variations and can be used to "map  $\text{TiO}_2$  content" for mare surface (Charette et al., 1974; McCord et al., 1976; Pieters and McCord, 1976). Since the spectral variations are nevertheless very small, the ratio images must be computer enhanced to display the data. Shown in Figure IID-21 are mosaics of vidicon data for the Flamsteed area of the moon (McCord et al., 1976b). Figure IID-21a is a slightly enhanced  $.57\mu\text{m}$  image and is comparable to a normal albedo image. Figures IID-21-b and c are both  $.40/.57\mu\text{m}$  ratio images, each enhanced to bring out different details in the maria. Note the distinct nature of the 'bluest' flow is only evident in the strongly processed image of Figure IID-21c. This Flamsteed region is discussed in further detail in Section III E2.

The ratio  $.95/.57\mu\text{m}$  ratio image is particularly useful for identifying and examining the areal extent of immature surfaces. This spectral parameter is mentioned in greater detail in Section IV.

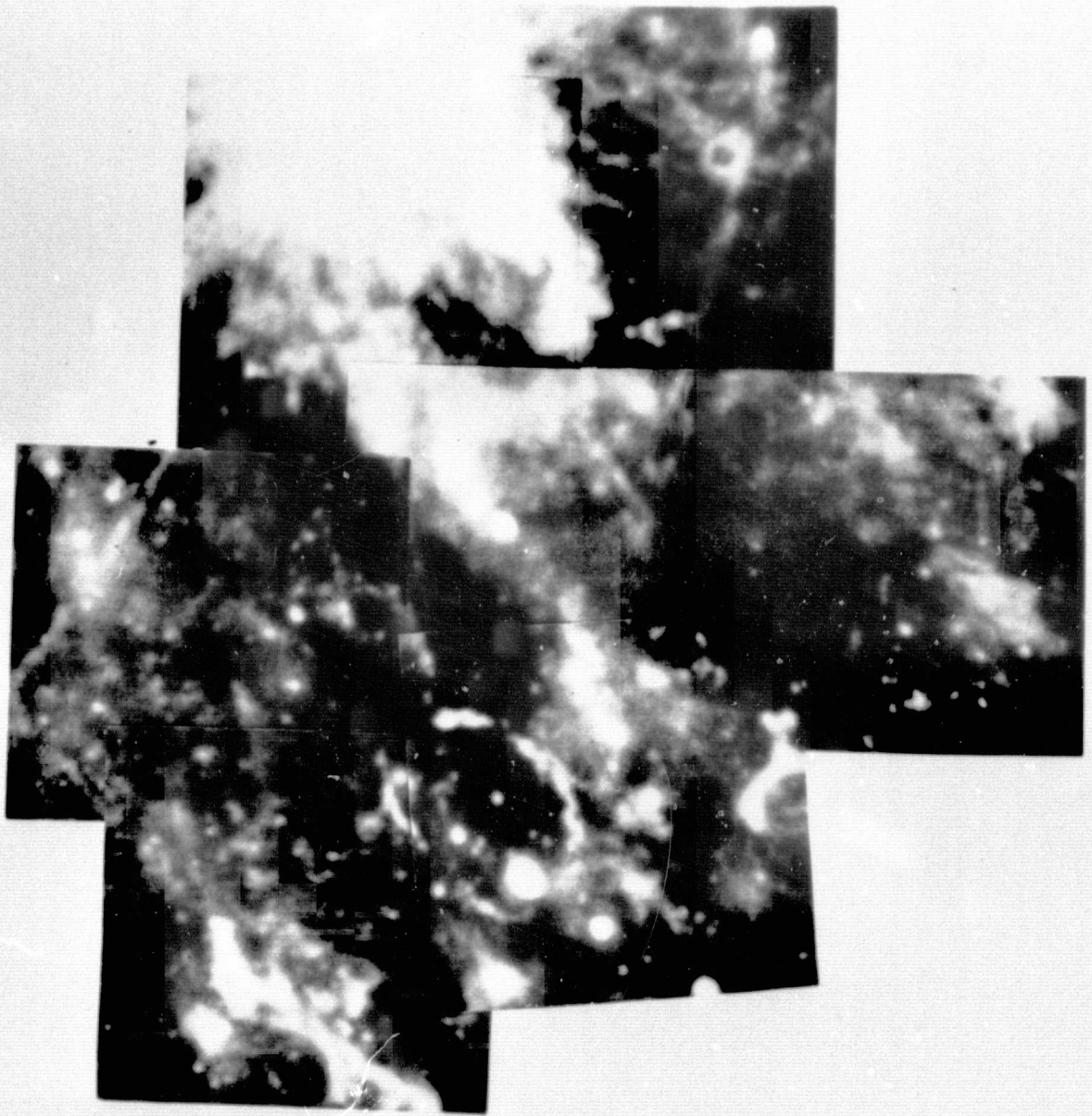


Figure IID 21a. Vidicon image mosaic ( $\lambda = .57\mu\text{m}$ ) of the Flamsteed region of Oceanus Procelliarum slightly contrast enhanced to bring out details in the mare.

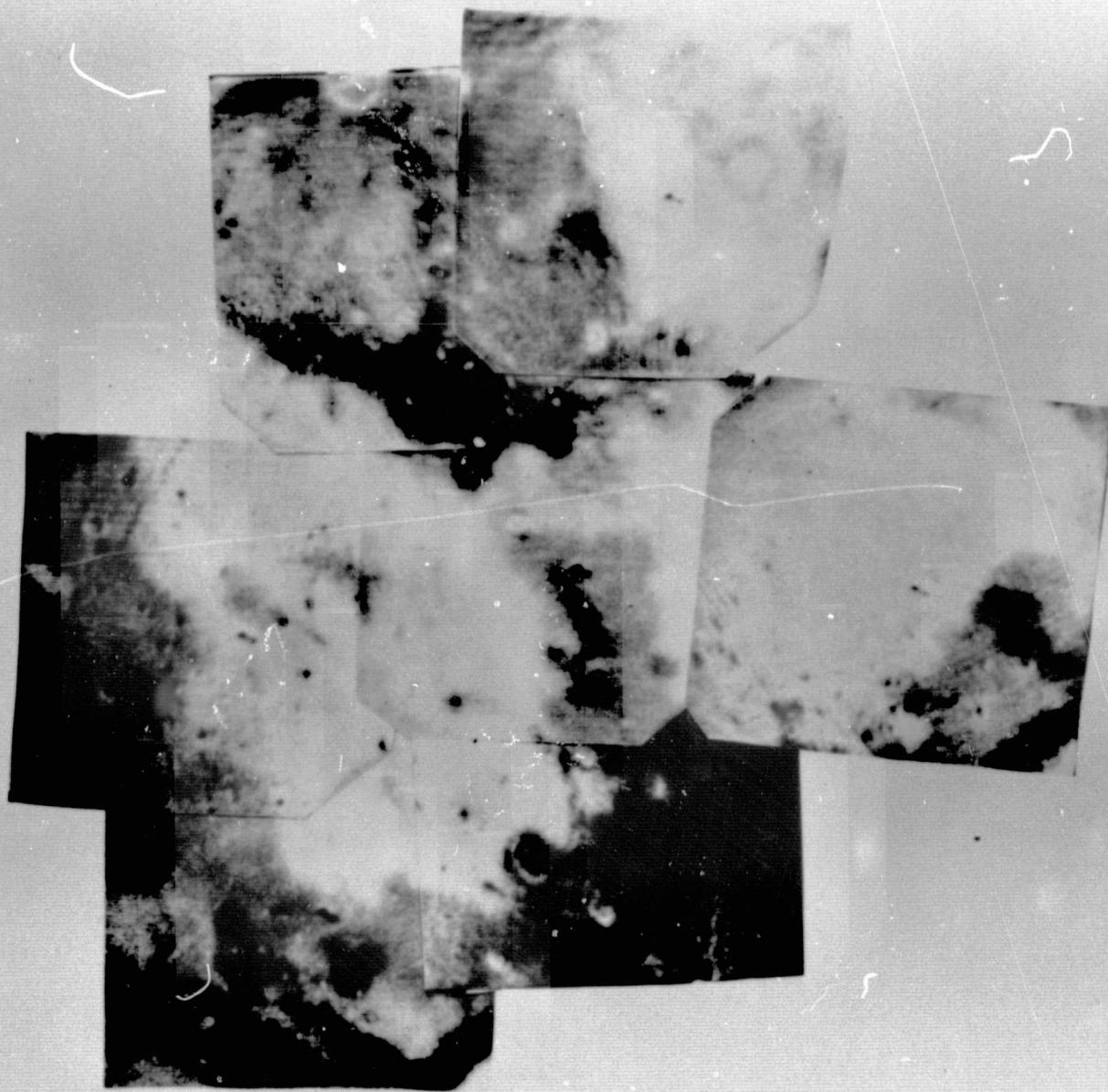


Figure IID 21b. Vidicon spectral ratio image mosaic (.40/.57  $\mu\text{m}$ ) of the same region shown in Figure IID-21a. The images have been contrast enhanced dark to light to show a 12% color difference. Bright indicates relatively bluer (higher  $\text{TiO}_2$  content).

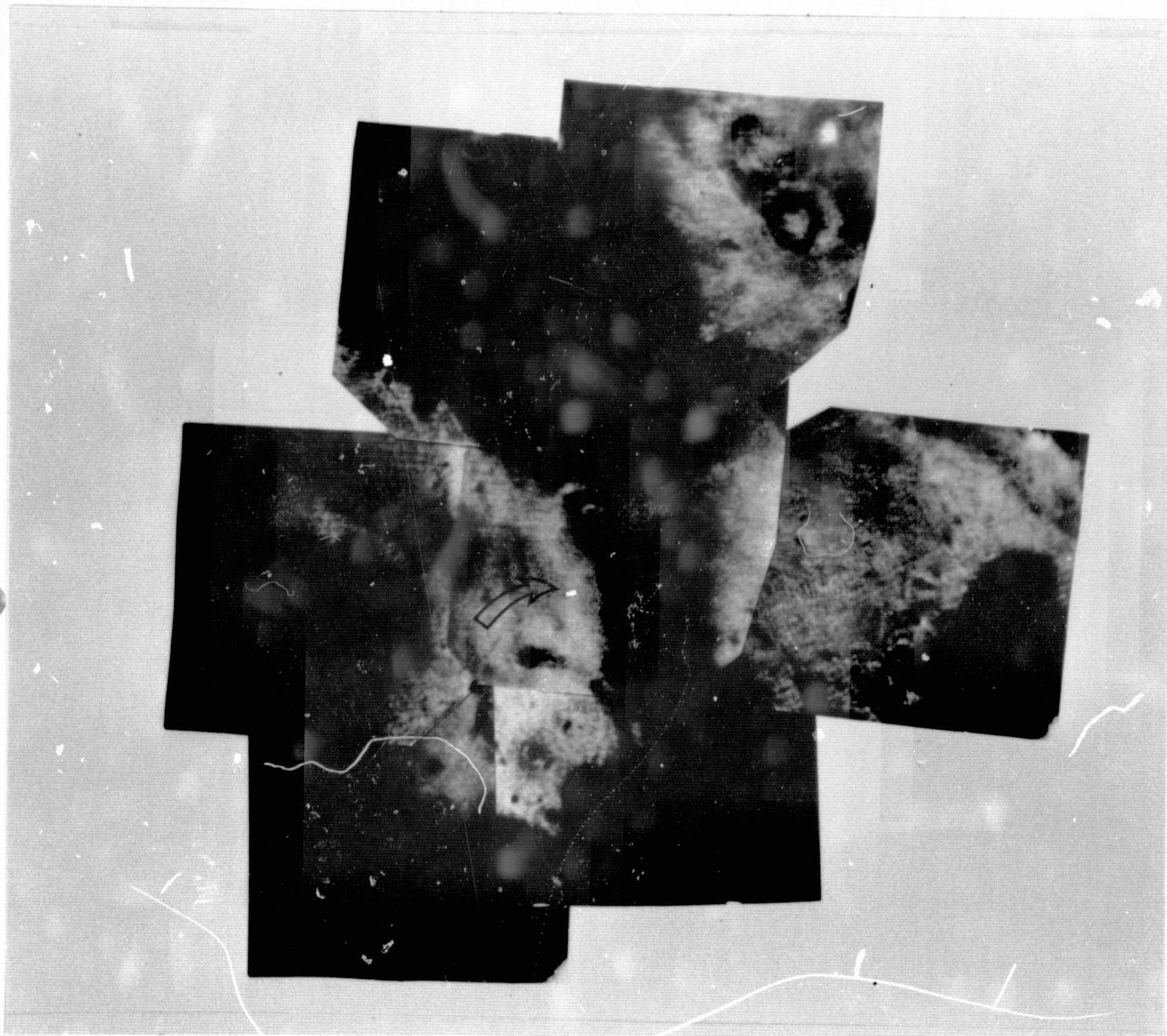


Figure IID 21c. Same vidicon ratio image mosaic as Figure II 21b contrast enhanced further to show the bluest unit (indicated by arrow). This is likely to be the last basalt flow of the region and contains the highest content of  $TiO_2$ .

III. LUNAR BASALT TYPES I: SOIL SPECTRA (.3 to 1.1 $\mu$ m)

The material presented in Sections III and IV constitute much of the scientific output of this thesis. These sections are largely in the form of manuscripts that were prepared as the thesis progressed. The content concerns the classification of remotely-obtained lunar reflectance spectra and the applications of spectral reflectance data to the characterization and distribution of lunar mare basalt types. The material presented in the two previous sections provides the detailed background on the nature of basalts and spectral reflectance techniques that cannot be included in a (page-limited) published manuscript.

Section III includes discussion of lunar mare soil spectra presented in three parts. The spectral characterization of lunar mare basalt types is presented first followed by two applications of the spectral information: a discussion of the Luna 24 landing area and a preliminary report concerning a major unsampled basalt type in the western maria.

Section IV includes a discussion of lunar crater spectra for both mare and highland regions.

The major conclusions that can be drawn from the currently available spectral reflectance data are summarized in Section V.

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## **Characterization of lunar mare basalt types: I. A remote sensing study using reflection spectroscopy of surface soils**

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**Abstract**—Telescopic reflection spectra of mature mare surfaces are used to identify and characterize major basalt types on the frontside of the moon. The spectra are classified according to (1) continuum slope and (2) near-infrared features. This study indicates that there are major lunar basalt types that are unlikely to have been sampled during the landing missions. Regions of basalt exist in the western maria with a  $\text{TiO}_2$  content comparable to that of Apollo 11 but with infrared characteristics that indicate a distinctly different composition. Samples from two landing sites, Apollo 12 and Luna 16, may contain fragments of a nearby basalt unit compositionally different from the dominant basalt type of the landing area.

### INTRODUCTION

LUNAR MARE BASALTS play an important role in understanding the evolution of the crust and mantle, although they account for less than 1% of a 60 km thick lunar crust (Head, 1975). The creation of models of lunar basalt petrogenesis (e.g., Green *et al.*, 1975; Hubbard and Minear, 1975; O'Hara *et al.*, 1975; Walker *et al.*, 1975; Kesson, 1975) requires a detailed characterization of the geochemistry of the basalts as well as a description of the sequence and location of emplacement. Mare basalt samples returned from five different areas have been studied in detail and several general basalt classifications have been proposed (e.g., Taylor, 1975; Rhodes *et al.*, 1975). Remote sensing techniques can then be used to characterize unexplored lunar areas which, of course, is most of the lunar surface. In this paper we present a classification of lunar mare basalt types derived through the analysis of telescopic reflectance spectra of mature mare surface areas.

With the exception of fresh craters and areas of high relief, it is the glassy agglutinate-rich mature soils that are observed by all geochemical remote sensing techniques. Mature mare surfaces contain in excess of 70% agglutinates (e.g., Charette and Adams, 1975). A valid classification of lunar basalt types nevertheless can be derived from reflectance spectra of mature surfaces if the following assumptions are accepted: (1) soils are derived largely from the local dominant surface basalt type, (2) the processes and effects of soil maturation are the same on all maria, and (3) during the 2.5 to 4 b.y. since emplacement, all mare surfaces

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studied (uppermost 1 cm) have reached an equivalent maturity.

Given the above assumptions and knowing the sensitivity of spectral reflectance to geochemistry (e.g., Adams 1974, 1975; McCord and Adams, 1973), it is likely that differences in spectral reflectance observed remotely between mare areas are due to differences in geochemistry of the dominant surface basalt type of the region. The lunar basalt types discussed in the following sections are derived from spectra of mature areas: no craters or highland regions are included in this study. The classification is based on two sets of parameters. The first concerns the slope of the reflectance spectrum in the blue and ultraviolet which is directly related to the percent  $\text{TiO}_2$  in the surface soil (Charette *et al.*, 1974). The second set of parameters concerns the near-infrared portion of the spectrum which is strongly affected by electronic absorptions in mafic minerals and glass.

#### DATA DESCRIPTION

Telescopic spectra of about 400 small (5–20 km diameter) lunar areas have been obtained using a filter photometer covering a spectral range of .3 to  $1.06 \mu\text{m}$ . A description of the equipment and observing techniques is given in McCord *et al.* 1972. Over 100 of these areas are in mature mare regions and are included in the study presented here. A few spectra obtained prior to 1970 over a comparable spectral range were also included in this study (McCord and Johnson, 1969; Johnson and Soderblom, 1969; McCord *et al.*, 1969). Many of these mare areas were chosen in order to characterize units evident in multi-spectral maps. The location of these mare areas are indicated by over-sized symbols in the frontispiece map. More precise coordinates of these small areas can be obtained from the authors on request. Not shown on the map are the more than 30 areas in Mare Humorum discussed previously by Pieters *et al.* (1975).

Since the spectral reflectance differences between mature soils are small, we have analyzed the telescopic data in the form of relative reflectance spectra. To form a relative reflectance spectrum,  $\bar{R}(\lambda)$ , the spectrum of an area is divided by the spectrum of a standard mare area, MS-2 (18.7°N, 21.5°E). The result is scaled to unity at  $0.57 \mu\text{m}$  to eliminate the effects of albedo differences. This procedure removes most of a steeply sloped continuum (increasing reflectance towards longer wavelengths) and allows the subtle spectral differences to be analyzed in detail. Relative reflectance spectra are shown in Fig. 1 for many of the areas included in this study.

Many mare areas were observed during more than one night as a check for consistency of the measurements. Two mare areas, one in the east and one in the west of the lunar frontside, were observed more than 20 times through four lunations to determine if the phase of the moon during an observation affected the relative reflectance spectrum sufficiently to cause a misclassification. The results of these multiple observations indicated that although small phase effects on differential color do occur (see McCord, 1969), they are fairly well defined and do not change the general spectral character of a relative reflectance spectrum.

## CLASSIFICATION OF REFLECTANCE SPECTRA

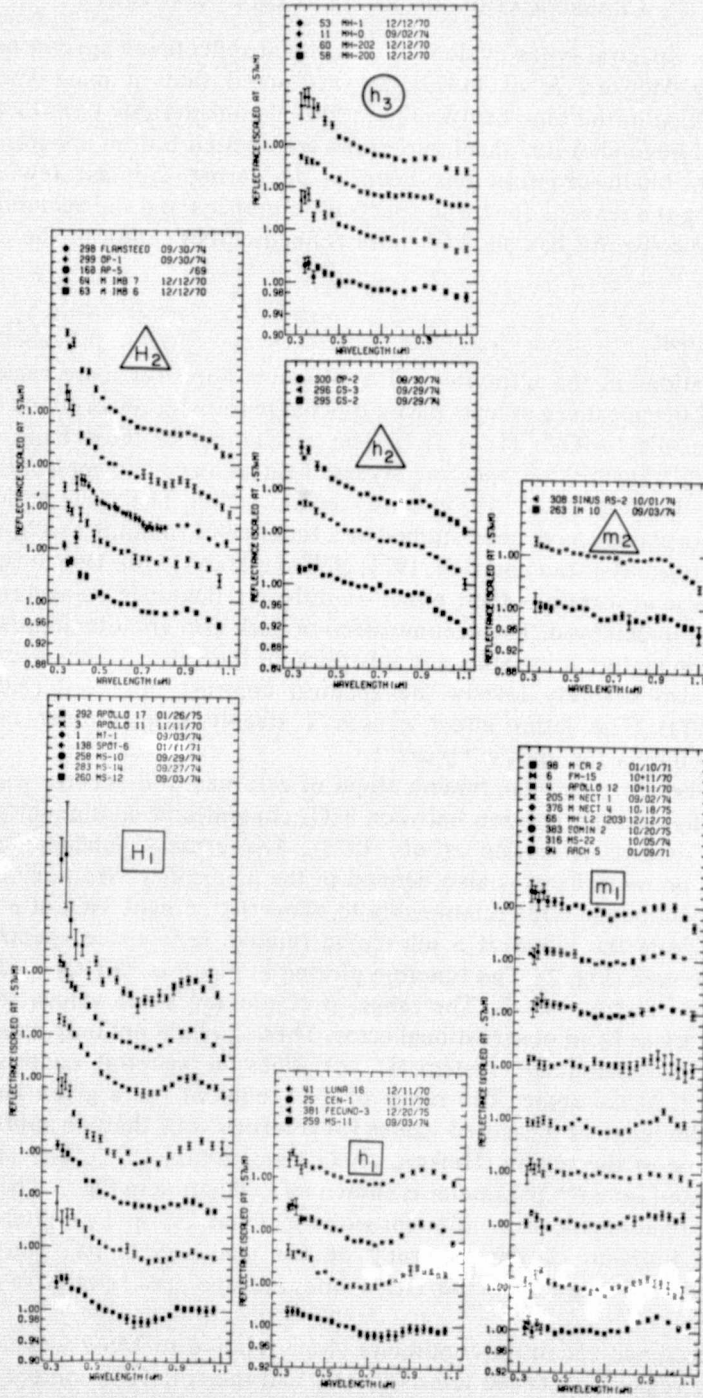
The major spectral types evident in most lunar reflectance spectra have been described by McCord *et al.* (1972). It was noted that in mare spectra the continuum slope in the blue below  $.57 \mu\text{m}$  varied considerably ( $\pm 8\text{--}10\%$ ). In the near-infrared beyond  $.8 \mu\text{m}$  small variations were noted but no apparent pattern was evident. Much progress has been made during the last few years in understanding the reasons for these spectral differences and the accumulation of new telescopic spectra has allowed more refined classification of the spectra.

A.  $\text{TiO}_2$  content

The variations in the ultraviolet of continuum slope for mare regions were suspected of being caused at least partly by charge transfer absorptions involving titanium (e.g.,  $\text{Fe}^{+2} \rightarrow \text{Ti}^{+4}$ ,  $\text{Ti}^{+3} \rightarrow \text{Ti}^{+4}$ ). The assignment of these bands is being derived using transmission spectra, crystal field theory, and molecular orbital theory (e.g., Loeffler *et al.*, 1975; Burns *et al.*, 1976). The systematics of the absorptions in glasses have been studied as a function of titanium and iron content and oxygen fugacity (Mao and Bell, 1973; Bell and Mao, 1976). Due to the intense nature of these absorptions their effect on diffuse reflectance measurements are only partially understood. The accumulation of dark iron and titanium-rich glassy agglutinates in mature mare soils not only darkens the soils (Adams and McCord, 1973), but also severely lowers the spectral contrast observed (Adams and McCord, 1971). The latter effect causes a steeply sloped lunar reflectance spectrum to flatten (or appear "bluer").

In the laboratory, the continuum slope of returned soil spectra was studied systematically and the relation between  $\text{TiO}_2$  content and continuum slope was empirically defined (Charette *et al.*, 1974). The critical dependence of this relationship on maturity was also defined in the laboratory. We have converted this  $\text{TiO}_2$ —continuum slope relationship to allow the percent  $\text{TiO}_2$  of a soil to be determined from the slope of a telescopic relative reflectance spectrum for a mature mare area (Fig. 2). The function plotted in Fig. 2 is:  $\% \text{TiO}_2 = ae^{bR}$ , where  $a = 8.12 \times 10^{-7}$  and  $b = 14.88$ . The range of continuum slope values for a given landing site arises from observational error. These include not only measurement errors but also the small systematic variations in spectral contrast due to differences in phase angle. The range of  $\text{TiO}_2$  content for a given landing site material is the range of measured values for returned soils that are approximately representative of the region (Heiken, 1974; Vinogradov, 1971). The most likely value estimated for each parameter is shown by a solid line in Fig. 2. The accuracy of the  $\text{TiO}_2$  relationship to continuum slope is about 1% in  $\text{TiO}_2$  content and is primarily a function of the accuracy of the telescopic data. The accuracy associated with distinguishing the  $\text{TiO}_2$  content of one area relative to another is closer to  $\frac{1}{2}\%$ .

The reflectance spectrum continuum slope relative to MS-2 was determined for all mare regions observed telescopically and the soil  $\text{TiO}_2$  content estimated



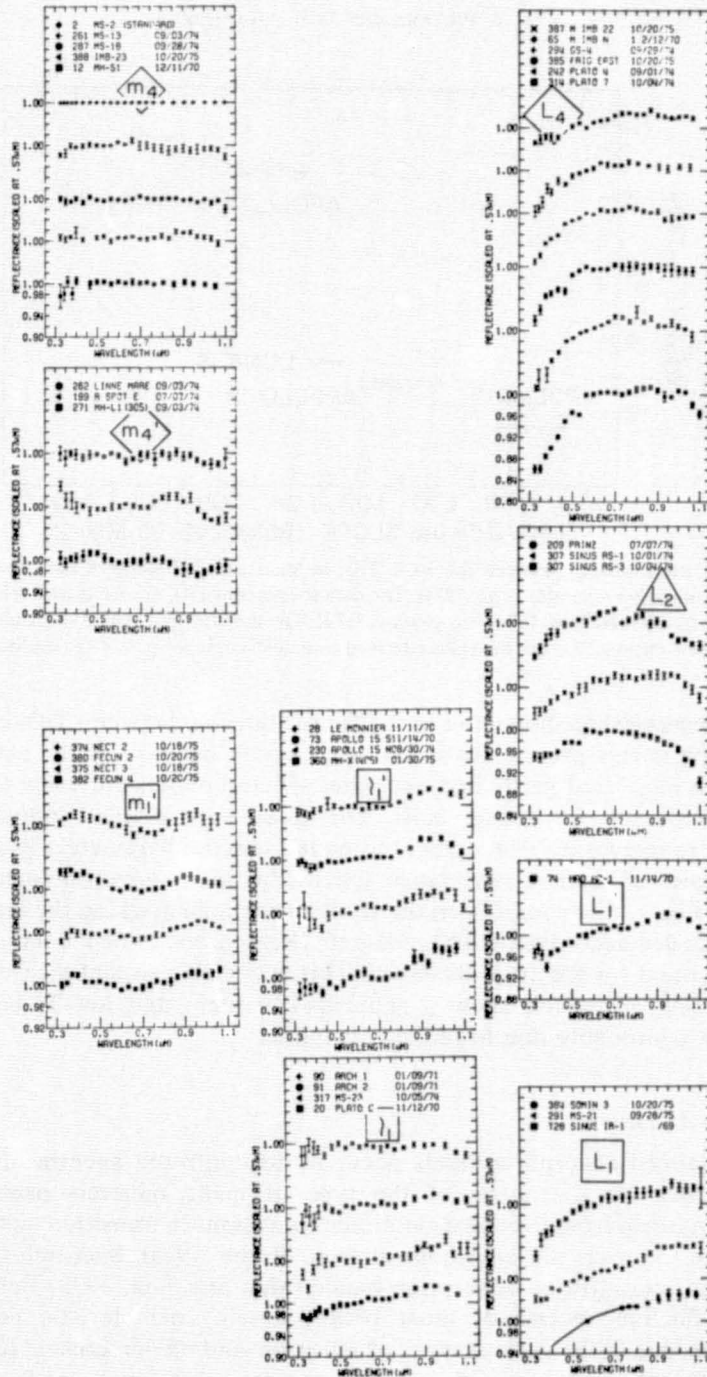


Fig. 1. Relative reflectance spectra of mature mare regions. All spectra are relative to MS-2. The spectra have been characterized and grouped according to the parameters listed in Table 1.

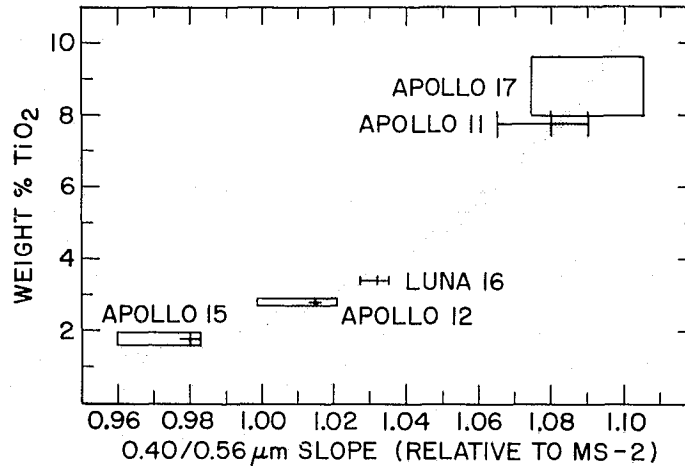


Fig. 2. Relationship between the wt.% TiO<sub>2</sub> in lunar soil and slope of the reflectance spectrum (after Charette *et al.*, 1974). The data for percent TiO<sub>2</sub> are measured values for returned soils (Heiken, 1974; Vinogradov, 1971). The data for reflectance spectrum slope between .40 and .57 m are the values derived from telescopic spectra of the landing sites.

from the relationship shown in Fig. 2. The relationship between TiO<sub>2</sub> content and curve slope is less precise for low-Ti soils partly because of the nature of the function in Fig. 2 and partly because other spectral effects currently under study may be important for redder soils. The areas were divided into five groups according to increasing TiO<sub>2</sub> content using the spectral parameters listed in Table 1a. Examples of relative reflectance spectra for many areas in each group are shown in Fig. 1. The symbols on the frontispiece map marking the areas studied are color coded according to TiO<sub>2</sub> content. There seems to be a continuous range of TiO<sub>2</sub> content for the areas observed. This is contrary to that measured for the returned samples which show a gap between high- and low-Ti basalts. This difference is probably due to limited sampling.

### B. Infrared features

Well-defined absorption bands occur in near infrared spectra of rocks and minerals and are diagnostic of the type of mafic minerals present. These absorptions arise from *d*-orbital electronic transitions in transition metal ions (Fe, Ti, Cr, etc.) in a crystal environment (e.g., Burns, 1970). Iron-rich glasses also have diagnostic infrared absorption bands (Mao and Bell, 1973; Bell and Mao, 1976). Reflection spectra of lunar rocks contain characteristic near-infrared absorptions due to the various types of pyroxene and olivine present (e.g., Adams 1974, 1975).

The lunar soils, however, are a combination of mineral fragments, glass, and agglutinates. A mature soil contains in excess of 70% agglutinates, the mineralogy of which is poorly known. The agglutinates can be described as glass-welded

Table 1. Spectral parameters used for distinguishing mare basalt types.

1a	$\bar{R}_{.40/.56}$	% TiO <sub>2</sub> (soil)	Class	Map color
	< .98	(< 1.5)	L-low	Red
	.98-1.00	1.5-2.0	l-med. low	Orange
	1.00-1.025	2-3	m-medium	Yellow
	1.025-1.06	3-5.5	h-med. high	Green
	> 1.05	> 5	H-high	Blue

1b	Type	Map symbol	Description of relative reflectance infrared feature
	1	□	Bump concave downward (symmetric around .95 $\mu\text{m}$ )
	1'	▣	Same as 1 plus a slope change at .35-.40 $\mu\text{m}$
	2	△	Linear decrease in reflectance beyond .90 $\mu\text{m}$
	3	○	Small bump near .90 $\mu\text{m}$ with a constant or increased reflectance beyond 1.0 $\mu\text{m}$
	4	◇	Featureless (same as MS-2)
	4'	◊	Small to large dip concave upward
	U	☆	Unclassified or unclear

aggregates of mineral fragments and devitrified soil material. Finely dispersed grains of Fe<sup>o</sup> are also found in agglutinates (e.g., Agrell *et al.*, 1970). The strong absorption bands evident in spectra of fresh basalt are no longer present in the spectrum of mature soils. The weak features observed in soil spectra in the near infrared are a subdued combination of a dominant glass band (Fe<sup>+2</sup>) and bands arising from the few remaining pyroxene and olivine mineral fragments (Charette and Adams, 1976; Charette *et al.*, 1976). Although differences which are known to have geochemical significance occur in the infrared spectra of mare surfaces, quantitative interpretations of these variations are difficult without further investigations. An important step toward such a refined interpretation of soil spectra is the recognition of infrared spectral features that are associated with regional basalts.

A sufficient number of mare spectra exist now to identify systematics that occur in the near-infrared spectra for many areas. A classification scheme for mare spectra was derived using the infrared spectral parameters listed in Table 1b. With additional data this classification probably could be subdivided. The location of these areas is indicated on the frontispiece map; each type is indicated with a different shape of symbol. Examples of the relative reflectance for many areas in each type is shown in Fig. 1.

#### DISCUSSION

This first characterization of global mare basalt surfaces according to their spectral features reveals some otherwise unobtainable geochemical information

for areas beyond the landing sites. Some interesting results immediately evident are discussed below.

#### A. Western high-Ti regions

Regions of basalt with a  $\text{TiO}_2$  content comparable to that at Apollo 11 and 17 occur in the western portion of the moon. However, the infrared classification of the spectra indicates that the western high-Ti basalts ( $H_2$ ) are compositionally distinct from those in the east ( $H_1$ ): the  $1 \mu\text{m}$  ( $\text{Fe}^{+2}$ ) absorption is stronger for the western basalts (Fig. 3). A quantitative interpretation of this distinction is being derived through laboratory studies on lunar soils (Charette and Adams, 1976). Furthermore, the areal extent of the younger (Boyce, 1975) western high-Ti basalts seems to be much less than that of the eastern basalts. The Flamsteed Region is the type-area for these distinct western high-Ti basalts. A few of the Imbrium basalts may have the same high- $\text{TiO}_2$  content, but the vast majority of extensive Imbrium and Procellarum "blue" basalts related to those at Flamsteed are apparently only medium high in titanium content. Global spectral imagery, such as the frontispiece by Soderblom and Boyce (1976), graphically displays this small areal extent of the high-Ti (as opposed to moderately high-Ti) western basalts.

#### B. Unsampled basalt types

The spectra of all mare landing sites are classified as type 1 (see Fig. 1 and frontispiece map). It is thus unlikely that any significant amount of samples has been returned from basalt types 2, 3, and 4. Of these unsampled basalts, there are at least three specific types that cover a large lunar surface area and occur in several locations: (1) the Flamsteed/Imbrium high- to medium-high Ti basalts ( $H_2$ ,  $h_2$ ,  $m_2$ ); (2) the Humorum/S. Procellarum medium-high Ti basalts ( $h_3$ ) (Pieters *et al.*, 1975); and (3) the low-Ti basalts (very red) of Frigoris and Imbrium ( $L_4$ ).

#### C. Minor components in returned samples

Since impact processes cause a redistribution of surface material, it may be possible that some of the minor components of returned samples are from unsampled regions. The major "contaminant" at Apollo 15 and 17 is the material from the surrounding highland. The Apollo 11 site is also slightly contaminated by highland material, but spectral images (e.g., Soderblom, 1970; Johnson *et al.*, 1975) indicate that within a radius of 100 km there are no other major mare units. However, there are distinctly different basaltic units within 100 km of the Apollo 12 and Luna 16 landing sites.

For the Apollo 12 landing area it was shown using crater degradation studies (Soderblom and Lebofsky, 1972) that there exist two neighboring units of differing age; the landing site is located in the older unit. Spectral data support this distinction and characterize the younger unit (a few kilometers from the landing

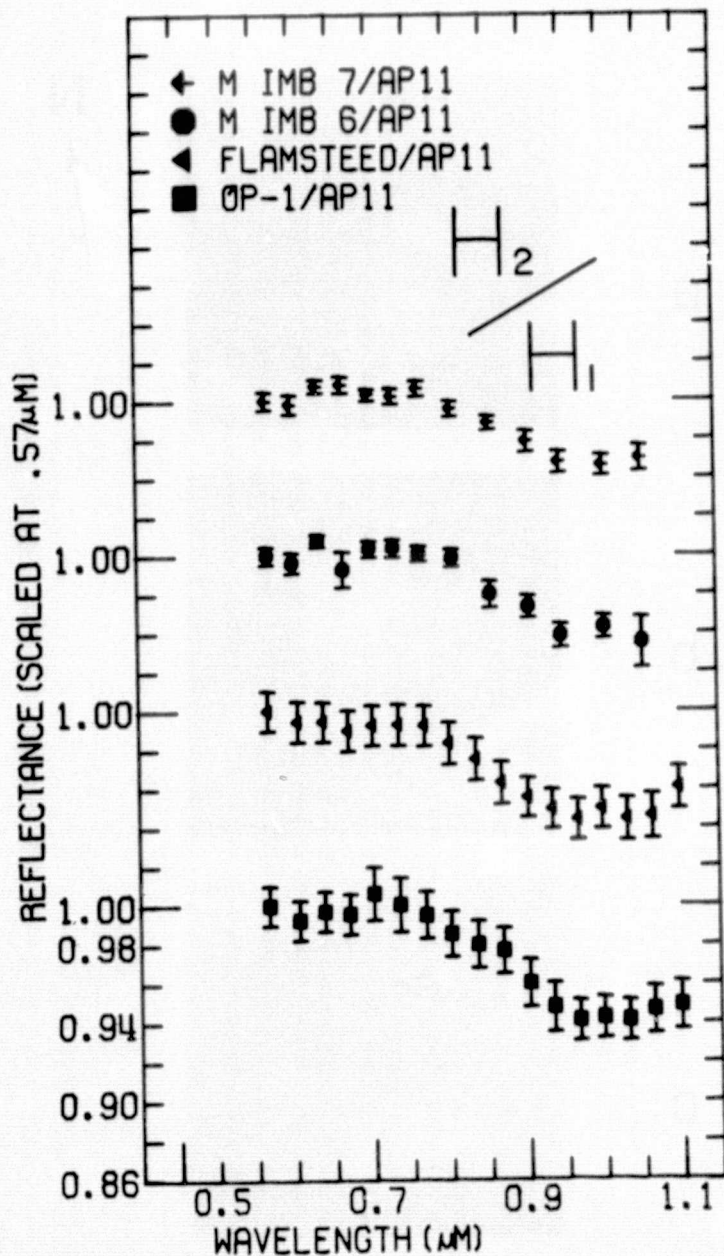


Fig. 3. Reflectance spectra of type  $H_2$  basalts relative to Apollo 11 ( $H_1$ ). Since the slope of the continuum is approximately the same for spectra of these two basalt types, their relative reflectance spectra are featureless below about  $.7\mu\text{m}$ . The difference in the  $1\mu\text{m}$  absorption feature between spectra of  $H_2$  and  $H_1$  basalt types is clearly evident in these relative reflectance spectra.

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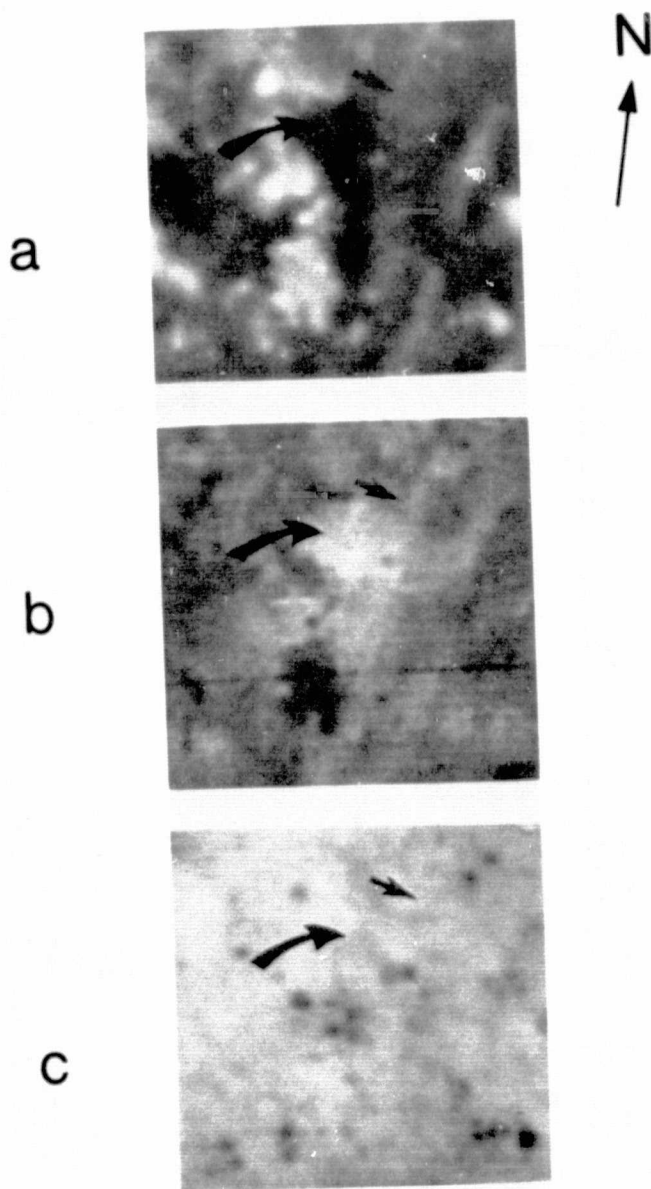


Fig. 4. Vidicon spectral images of the Apollo 12 landing area. The landing site is indicated with a short arrow. Area S III B is indicated with a long arrow. Field of view is about 150 km. Images were obtained digitally for the same area through a sequence of filters with a silicon vidicon detector and processed using standard image processing techniques (McCord *et al.*, 1976). (a) Image taken through a  $.57 \mu\text{m}$  filter. (b) Ratio image:  $.40/.57 \mu\text{m}$  contrast enhanced. (c) Ratio image:  $.97/.57 \mu\text{m}$  contrast enhanced.

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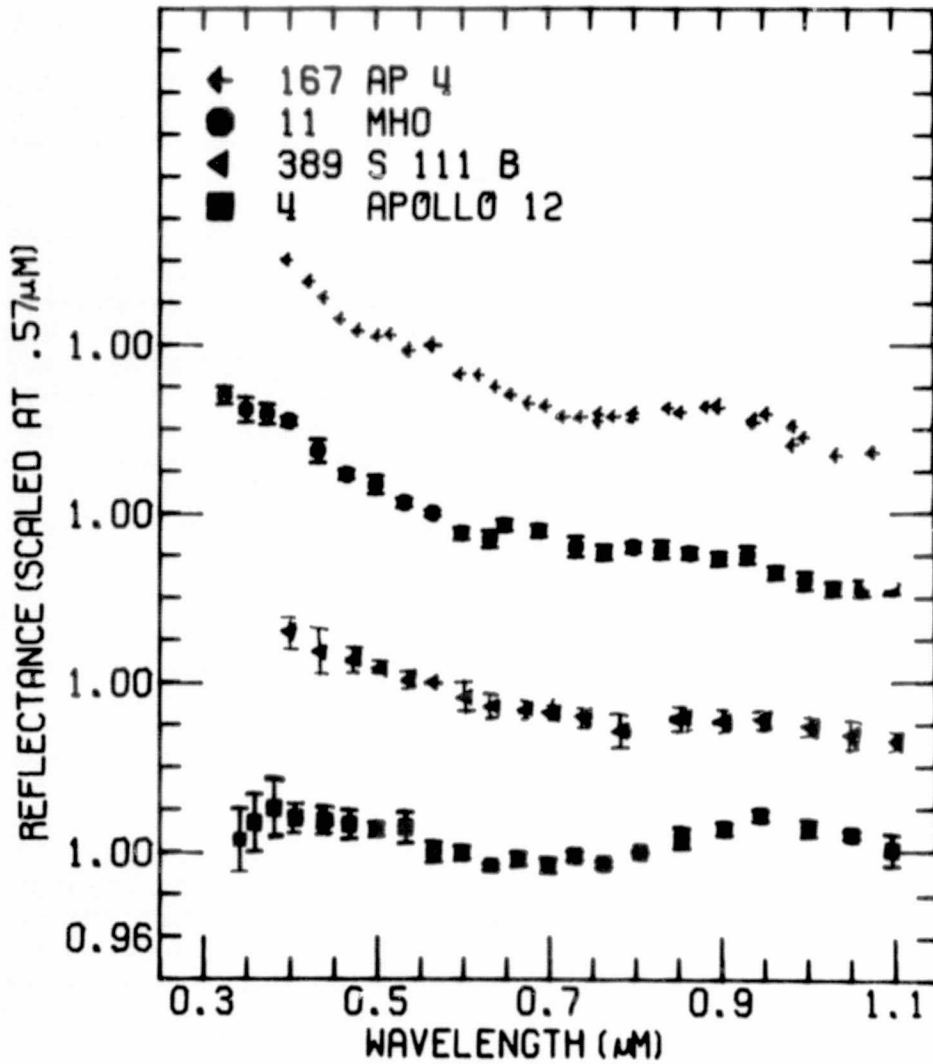


Fig. 5. Reflectance spectra relative to MS-2 for mare areas in the Apollo 12 region (Apollo 12 and S III B), in Mare Humorum (MHO), and in southern Procellarum (AP 4).

site) as a basalt with a slightly higher  $\text{TiO}_2$  content. Figure 4 contains high resolution spectral vidicon images (McCord *et al.*, 1976) of the Apollo 12 region; the field of view is about 150 km. The  $.40/.57 \mu\text{m}$  ratio image allows the mare soil  $\text{TiO}_2$  content to be mapped. Note the small tongue of higher titanium material (brighter) a few kilometers to the east of the landing site. This corresponds closely to the unit independently mapped by Soderblom and Lebofsky (1972). Although no spectra exist for this small unit, a spectrum exists for the area marked S III B (Johnson and Soderblom, 1969) which seems to be an extension of the material

near Apollo 12. The relative reflectance spectrum for area S III B is shown in Fig. 5 along with spectra of the Apollo 12 landing site and two examples of basalt type  $h_3$ . We have tentatively assigned the area S III B to the basalt type  $h_3$ . If future spectra confirm this classification, then some of the Apollo 12 ilmenite basalts (e.g., Rhodes *et al.*, 1975) may be derived from this nearby unit and thus may well be samples of a major "unsampled" basaltic unit.

Spectral images of Mare Fecunditatis indicate that the Luna 16 landing site is in a rather small unit (approximately 150 km in diameter) surrounded by basalts with a lower  $\text{TiO}_2$  content. The Luna 16 unit can be seen as a bluish area on the frontispiece spectral map by Soderblom and Boyce (1976) and as a distinct area on the color difference map by Barbarshov (1973). Unpublished higher spatial resolution spectral images of the region by McCord and by Whitaker further substantiate the distinction between two basalt types in close proximity in the Luna 16 region. Although there is a difference in  $\text{TiO}_2$  content between these two units, the spectra are both classified as type 1 (Fig. 1, Frontispiece map). The boundaries seen in the multispectral images match closely those seen in the independently derived age unit maps of Boyce (1976); Luna 16 is located in the older of the two units.

#### CONCLUSION

For the near future remote sensing techniques are the only means to determine the geochemistry of unsampled lunar regions. The data presented here have been used to show that a variety of basalt types exist on the frontside of the moon. Many basalt types covering a major portion of the maria have not been sampled. The spectral reflectance measurements of surface soil presented here have allowed the %  $\text{TiO}_2$  to be discerned for the major basalt types. It may also soon be possible to estimate the FeO content from such soil spectra (Charette and Adams, 1976). Spectra of fresh mare craters will further define the geochemistry of the unsampled basalts by allowing components of the mineralogy to be identified (e.g., composition of pyroxene and olivine). It is, however, not until data from orbital geochemical experiments are available that remote sensing techniques will be fully utilized in lunar exploration.

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REGIONAL BASALT TYPES IN THE LUNA 24 LANDING AREA  
AS DERIVED FROM REMOTE OBSERVATIONS

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**Abstract.** The Soviet spacecraft Luna 24 landed in Mare Crisium and returned samples that are expected to be much like the low titanium basalts from Luna 16 and Apollo 12. This conclusion is based on earth-based spectral reflectance measurements and multispectral imagery of the Mare Crisium region and uses a background of laboratory measurements of the spectral properties of Lunar soils. These data are used to describe the regional context and composition of the Crisium basaltic units. The returned sample may also contain minor components of a high-titanium basalt and a very low titanium basalt as well as highland material.

A classification and characterization of surface mare basalt types has been derived from the remote observations of the spectral reflectance of mature mare soils (Pieters and McCord, 1976; Adams and Pieters, 1976). Multispectral ratio images allow the areal distribution of mare units to then be mapped (McCord et. al., 1976a). Using our experience with the spectral properties of lunar soils derived in the laboratory, we have examined both kinds of remotely obtained spectral information for the region containing the Luna 24 landing site. We describe here the basalt types most likely to have been sampled and place the samples in a regional context.

The Soviet Luna 24 spacecraft landed in Mare Crisium (12° 45' N, 62° 12' E) and returned a 2m core sample to earth (Aviation Week, Aug. 18, 1976, p. 18).

Figure 1 is a sketch map of the region showing the location of the landing site. The geology of the region has been described by Olson and Wilhelms (1974) as extensive dark plains material (basaltic). About 10 km from the landing site is a small region mapped by Olson and Wilhelms as dark mantling material primarily on the basis of its lower albedo. The land-

ing site is indicated in the vidicon image mosaics (McCord et. al., 1976b) of Figure 2 by a small arrow. A relative reflectance spectrum (.3 to 1.1 $\mu$ m) for the region M Cr-2, indicated with a large arrow, was included in the mare basalt classification scheme and is shown in Figure 3 (from Pieters and McCord, 1976). Useful spectra for three additional areas in Mare Crisium were obtained previously (McCord, 1968) over a more limited spectral range (.4 to .8 $\mu$ m).

The .40/.56 $\mu$ m spectral reflectance ratio has been shown to be sensitive to the TiO<sub>2</sub> content of mature mare soils and an estimation of the TiO<sub>2</sub> content for the region M Cr-2 was made (Charette et. al., 1974). This relationship between .40/.56 $\mu$ m reflectance ratio and soil TiO<sub>2</sub> con-

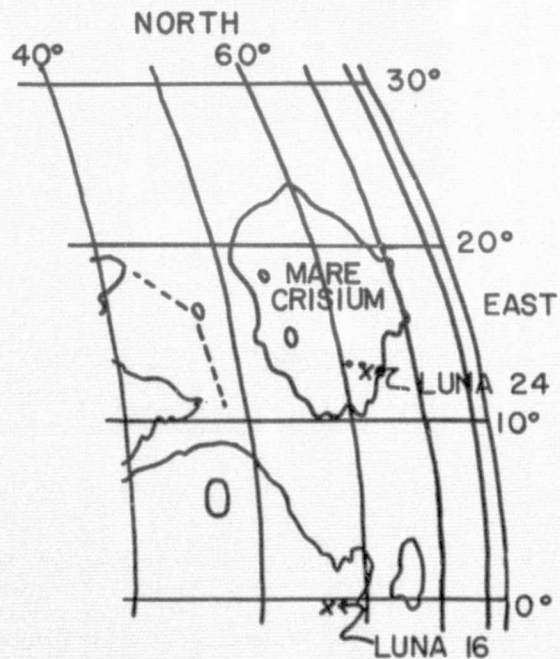


Figure 1. Sketch map of the eastern limb of the moon showing the location of the Soviet Luna 24 landing site.

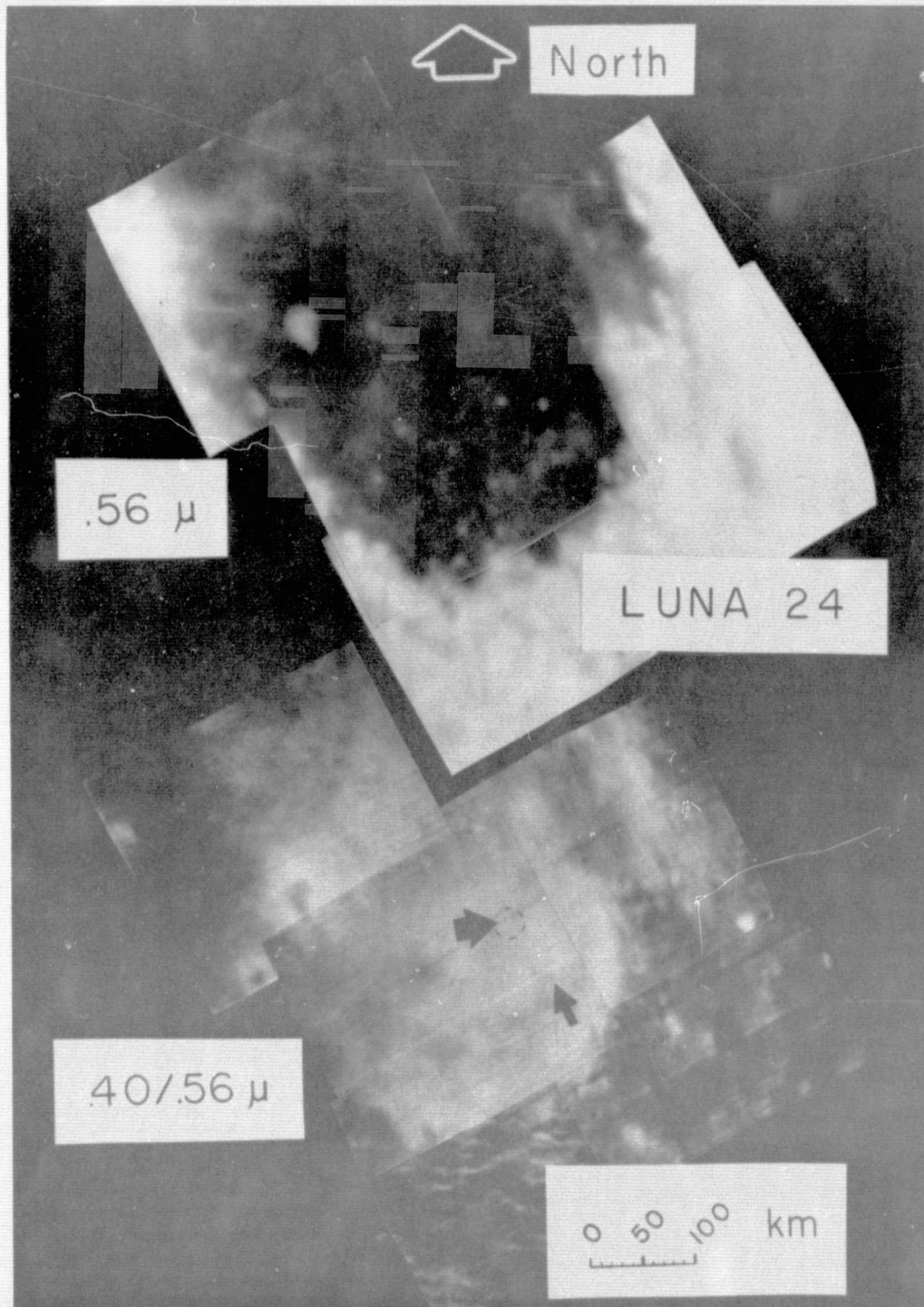


Figure 2. Vidicon image mosaics of southern Mare Crisium. TOP: .56 $\mu$ m image slightly contrast enhanced to bring out features in the mare. The Luna 24 landing site is indicated by a small arrow. The large arrow indicates the region for which a spectrum was obtained (M Cr-2). BOTTOM: .40/.56 $\mu$ m ratio image contrast enhanced with the grey tone variation dark to light indicating a 12% difference in the .40/.56 $\mu$ m ratio.

tent is shown in Figure 4 for mature mare soils. Figure 2-bottom is a contrast enhanced .40/.56 $\mu$ m ratio image mosaic of the landing site region with the grey tone variation dark to bright spanning a 12%

difference in the .40/.56 $\mu$ m ratio.

From these data the following conclusions can be drawn:

1) A major unit filling central Mare Crisium is composed of low-titanium ba-

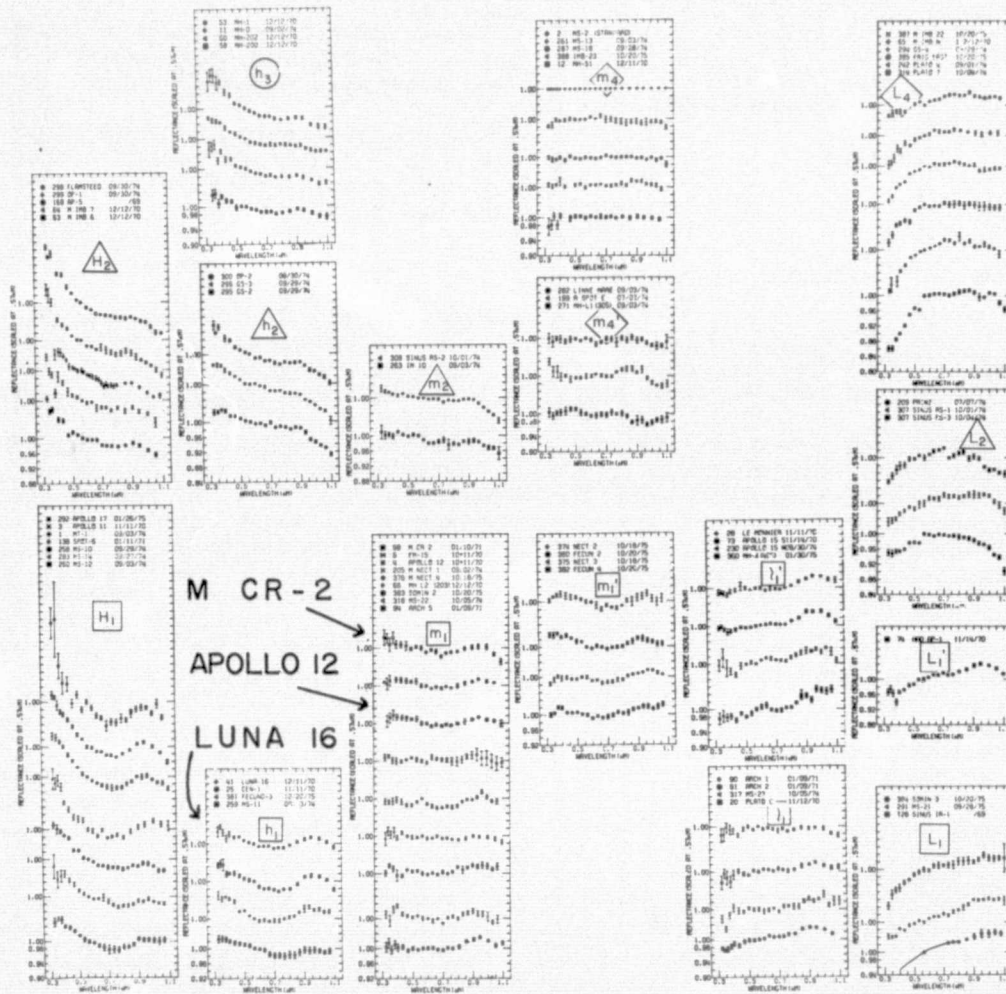


Figure 3. Relative reflectance spectra of mature mare regions. All spectra are relative to the standard area MS-2. These spectra are composed of reflectance ratios for 25 filters between .33 and 1.1μm scaled to unity at .56μm. The intervals marked along the vertical axis are .02 in reflectance ratio. The horizontal intervals are 500Å (or .05μm). The spectra have been characterized and grouped according to TiO<sub>2</sub> content and the nature of infrared spectral features. The spectra indicate the Mare Crisium low-titanium basalt is comparable to that at Luna 16 and to a lesser extent to that at Apollo 12.

salts much like those found by Luna 16 and to a lesser extent by Apollo 12. This conclusion is made from noting the shape of the spectrum for M Cr-2 (Figure 3) and comparing it to spectra of other basalt types observed. The vidicon ratio image mosaic (Figure 2) indicates regional continuity between the telescopic site M Cr-2 and the Luna 24 landing site.

2) The Luna 24 landing site itself is in an inhomogeneous area. The returned core could have sampled either or both of two low titanium basalt units which differ in (soil) TiO<sub>2</sub> content by about 2%. From the vidicon spectral ratio data of Figure 2-bottom and the TiO<sub>2</sub> relationship of Figure 4, the weight % TiO<sub>2</sub> expected for these Mare Crisium soils is 2-4%.

3) The Eratosthenian crater Picard,

200 km to the west of the Luna 24 site, has apparently excavated higher titanium material (Figure 2), assuming the ejecta soil surrounding the crater has matured to the same extent as the background mare. If the stratigraphy evident at Picard extends to the Luna 24 region, it is likely that the low-titanium surface basalt at Luna 24 is underlain by high titanium material.

4) 100 km to the north of the landing site there exists a low titanium (<1.5%) basalt unit that is distinct from that at the Luna 24 landing site. The spectral boundaries for this unit (upper left in Figure 2-bottom) do not correspond to albedo variations (Figure 2-top). From the vidicon spectral ratio data of Figure 2 and the TiO<sub>2</sub> relationship of Figure 4, this low-titanium basalt is shown to be

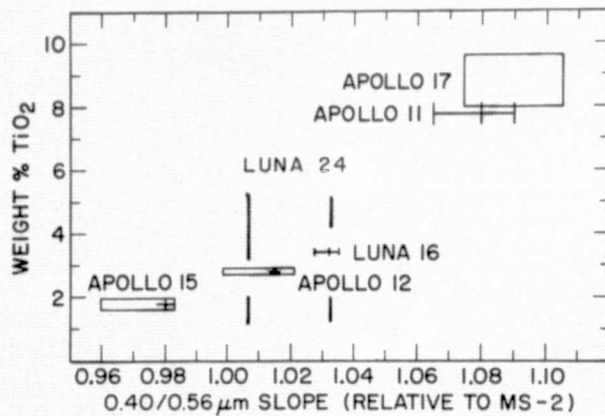


Figure 4. Relationship between the wt%  $\text{TiO}_2$  in the lunar mare soil and the slope of the reflectance spectrum (.40/.56  $\mu\text{m}$  ratio) relative to the slope of the spectrum for the standard area MS-2 (after Charette et al., 1974). This relationship for telescopic data was first substantiated through laboratory reflectance measurements of lunar samples. The data for percent  $\text{TiO}_2$  are measured values for returned soils. The data for reflectance spectrum slope are the values derived from telescopic spectra of the landing sites (from Pieters and McCord, 1976). Reflectance values for soils from the major low-titanium basalts in Mare Crisium near Luna 24 are derived from the vidicon spectral ratio data calibrated with the spectrum for M Cr-2.

comparable to those at Apollo 15 in  $\text{TiO}_2$  content. Since no spectra exist for this unit, however, further geochemical distinction cannot be made.

Thus, the basalt types returned by Luna 24 are likely to be the familiar low-titanium basalts of Luna 16 and Apollo 12. The returned sample may also include some components of a high titanium basalt, a very low titanium basalt, and highland material. Although the spectral data presented here cannot confirm the existence of mantling material in the region, such material would be expected to be evidenced by a component of glass droplets (Heiken et al., 1974; Adams et al., 1974). The degree to which these additional components are included in the core is dependent on the local cratering history and the continuity of underlying stratigraphy.

The return of Luna 24 will certainly increase our knowledge of the known lunar basalt types, but there still remain major and extensive unsampled mare units.

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UPDATE: Luna 24 information

During the 8th Lunar Science Conference, the following information concerning Luna 24 was presented:

(1) Crater degradation age determinations by J. Boyce (USGS, Flagstaff) indicated two major age units in the Mare Crisium area. The surface basalt in the region of the Luna 24 landing site is estimated to be of an age comparable to Apollo 15 basalts. Older basalts are exposed in an ill-defined region to the west of the landing site and also in an area to the northeast. These older basalts are expected to be slightly younger than Apollo 11 basalts. A small northeast section of Mare Crisium (far from the landing site) appears to be the youngest unit.

(2) Butler et al. (JSC) described the geological setting of the Luna 24 region as derived from orbital photographs. Of particular interest is the 6.5 km crater Fahrenheit (16 km from the landing site). This crater penetrates to 1200m with 5 diffuse bands being identified in the crater wall. The Luna 24 region also contains abundant secondary craters associated with a ray pattern of the distant (1200 km away) crater Giordano Bruno.

(3) Preliminary descriptions of the returned core sample made by Barsukov et al. (USSR) as best interpreted include:

a. The core drilled to a nominal depth of 225 cm. However, the upper ~60 cm were not recovered in the returned sample.

b. Most of the sample consists of various layers of immature regolith with a high content of coarse grained rock fragments.

c. Chemical analysis of some of the fragments classified them as very low titanium (~1%) mare basalts. Their aluminum content is high and KREEP component low.

These new data allow the following possible interpretations of the Luna 24 spectral information (Pieters et al., 1976):

I. The Luna 24 core sampled only the surface unit seen by remote sensing techniques. If the very low titanium basalts are representative of this surface unit, then the calibration of %  $TiO_2$  as derived from continuum slope is inaccurate for very low titanium basalts. Due to the nature of the calibration curve, such an inaccuracy would not be greatly surprising for unsampled low Ti basalts. (This has been pointed out in all papers discussing the nature of this calibration.)

II. The Luna 24 core is composed largely of a (subsurface) unit distinct from the general surface unit in southeast Mare Crisium. Supporting this interpretation

is the relationship between the age units of Boyce and the spectral units of Pieters et al. and the fact that the core seems to be largely a sample of relatively fresh ejecta. The Boyce 'old' age unit is roughly spatially comparable to the very low titanium basalt to the north-east defined by the spectral images. Furthermore, the spectrally mottled region extending to the west of the landing site is again comparable to the 'old' unit defined in that area. These combined data indicate the sequence of mare fill for Mare Crisium may have been: (a) early high-Ti basalt (excavated by Picard) followed by or contemporaneous with (b) a very low titanium basalt (Boyce's 'old' unit; spectrally red) and a later (c) medium low-Ti basalt (2-4%  $\text{TiO}_2$ ) emplaced about 3.3 b.y. ago. This medium low-Ti surface unit is perhaps thin especially to the west of the landing site. The freshness of the sampled material indicates it could easily be largely ejecta excavated (by Farenheit?) from a proposed very low-Ti subsurface unit (b).

The basic problem in understanding the spectral information is determining what the Luna 24 core actually sampled. Analysis of the core material itself should distinguish which of the above possibilities are more likely correct:

If the dominant basalt type throughout the core is the very low-Ti basalt and it is found to be comparable in age to Apollo 15 basalts (~3.3 b.y.), then one would conclude this very low-Ti basalt is likely to be the surface unit and interpretation I is probable. The spectral calibration of low-Ti basalts would have to be reevaluated. A significant component, however, of unrelated basalt types of higher  $TiO_2$  in the core would weaken this interpretation.

On the other hand, if the very low-Ti basalts are found to be relatively old, interpretation II would be more probable; the sample would be interpreted largely as ejecta from a subsurface unit. Some component of younger medium low-Ti material would be expected somewhere in the core.

Potentially, the Luna 24 core could contain one of the richest assortments of mare basalt types returned to date.

2. Synthesis study of the Flamsteed region of Oceanus Procellarum

Full moon spectral imagery (e.g., Barbarshov, 1973) indicates that only a small portion of the western 'blue' basalts are comparable in continuum slope to those of the east (e.g., Apollo 11). This is confirmed by the filter photometry measurements (Pieters and McCord, 1976) that further show the western high-Ti basalts to be distinct and compositionally different from those sampled at Apollo 11 and Apollo 17. A synthesis study was undertaken to better understand the regional geology and define the geochemistry for a case area of western high-Ti basalts--the Flamsteed area of Oceanus Procellarum. Representative spectra for this region were included in the previous basalt classification study ( $H_2$ ,  $h_2$ ,  $h_3$ ). Vidicon spectral images (shown in Figure IID-21) were used to map the extent of these units. The preliminary results from this study were presented at the 8th Lunar Science Conference. New data are presented here to update the following published abstract. A more extensive discussion of the Flamsteed region will be prepared for publication during summer, 1977.

GEOLOGY AND GEOCHEMISTRY OF THE FLAMSTEED REGION OF OCEANUS PROCELLARUM: A PRELIMINARY REPORT BASED ON REMOTE SENSING AND LUNAR SAMPLE INFORMATION. C. Pieters<sup>1</sup>, J. B. Adams<sup>2</sup>, R. Bryan<sup>3</sup>, J. W. Head<sup>3</sup>, T. B. McCord<sup>1,4</sup>, S. Zisk<sup>5</sup>. <sup>1</sup>Dept. of Earth and Planetary Sci.; M.I.T., Camb., Ma. 02139; <sup>2</sup>Dept Geol. Sci., Univ. Wash., Seattle; <sup>3</sup>Dept. Geol. Sci., Brown Univ., Prov., R.I. 02912; <sup>4</sup>Univ. Hawaii at Manoa, Honolulu, 96821; <sup>5</sup>NEROC Haystack Observ., Westford, Ma. 01886.

The Flamsteed region of Oceanus Procellarum displays a number of mare and highland geologic units which are important in the geologic framework and volcanic history of the western maria. Fig. 1 shows the distribution of major units in the Flamsteed region and was derived using remote sensing techniques and lunar sample information. The principal techniques (1) used in defining unit boundaries included a) vidicon spectral ratio images, b) earth-based photographs (albedo), c) orbital photography. Other techniques essential for unit characterization and interpretation include a) reflectance spectra (0.3-1.1  $\mu\text{m}$ ), b) radar backscatter maps, c) radar topography, d) spectra and compositional analysis of returned samples from the Apollo sites. Relative ages were obtained by superposition relationships and crater degradation techniques (2).

On the basis of unit characteristics, distribution, and superposition relationships, the following history is outlined: A cratered uplands existed in the region prior to mare emplacement and consisted of normal feldspathic highland crust (U) and local development of a spectrally red unit similar to other "red spots" ( $U_x$ ) (3,4,5). Earliest exposed mare material is a series of low-Ti basalts (L<sub>1,m</sub>) which flooded the upland topography. Mare ridges developed on these units and the crater Flamsteed formed prior to the emplacement of later basalts. The last major phase of volcanism was characterized by moderately high-Ti basalts ( $H_2, h_2, h_3$ ) spreading as relatively thin flows over preexisting maria. Flow emplacement is generally controlled by preexisting local and regional topography including mare ridges developed on earlier maria.

The geochemical characterization of these units is based primarily on spectral reflectance measurements (.33-1.06  $\mu\text{m}$ ) of small (~10 km) regions within the units and similar laboratory reflectance measurements (.35-2.5  $\mu\text{m}$ ) of returned lunar soil samples. The spectrally defined basaltic units are characterized with respect to other basalt types for the frontside of the moon (6). Although a number of units can be distinguished from each other with this data, characterization of the mineralogy is incomplete until spectra to 2.5  $\mu\text{m}$  can be obtained for fresh craters within each unit.

- References (1) Head, J.W. et al. (1976) Lunar Science VII, p. 357.  
 (2) Boyce, J. (1976) Proc. Lunar Sci. Conf. VII, p. 2719-2728. (3) Malin, M. (1974) Earth, Planet. Sci. Lett. 21, p. 331-341. (4) Pieters, C. et al. (1975) Proc. Lunar Sci. Conf. VI, p. 2689-2710. (5) Wood, C.A. and J.W. Head (1975) Origins of Mare Basalts, LSI, p. 189. (6) Pieters, C. and T.B. McCord (1976) Proc. Lunar Sci. Conf. VII, p. 2677-2690

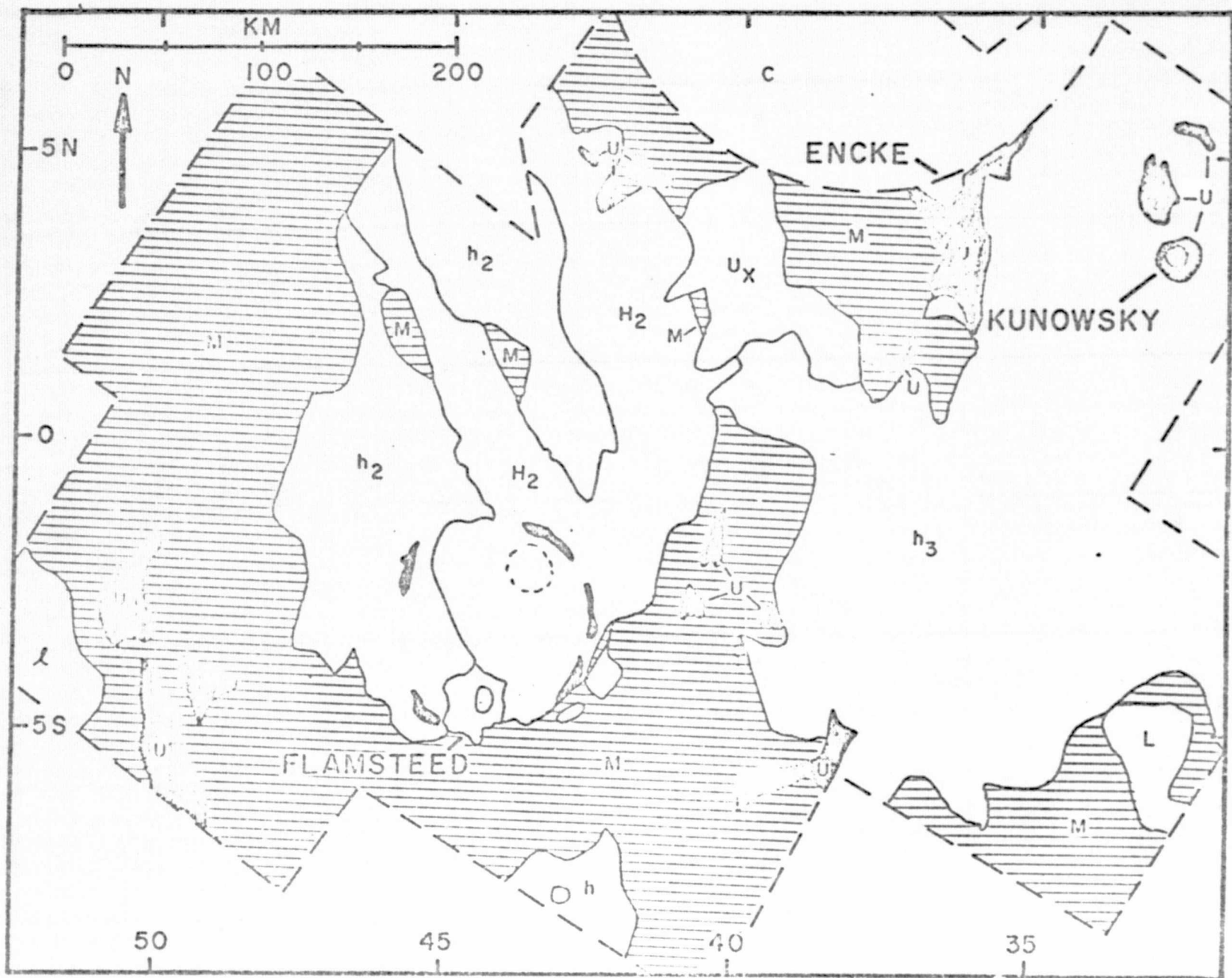


FIG. 263

REPRODUCIBILITY OF THE ORIGINAL PAGE IS POOR

<u>SYMBOL</u>	<u>DESCRIPTION</u>	<u>INTERPRETATION</u>
C	High albedo; high frequency of rays and satellitic craters surrounding Kepler.	Kepler ejecta; underlying material significantly affected by cratering event.
<u>Mare</u>	Low albedo; type area NE interior of Flamsteed P indicated with dashed circle; continuum slope (.40/.56 $\mu$ m) relatively blue; 1 $\mu$ m absorption feature ( $Fe^{+2}$ ) stronger and broader than Tranquillity soils. Several associated sinuous rilles.	Titaniferous basalt. A young basalt with a soil composition that ranges from 5-8.5% $TiO_2$ ; probably different from older eastern high-Ti basalts in mineralogy; more glass-rich than basalts of Mare Humorum ( $h_3$ ); related to Imbrium high-Ti basalts. Unsampled.
$H_2$	Spectrally much like $H_2$ but with continuum slope slightly (1-2%) less blue.	Titaniferous basalt. A young medium-high-Ti basalt with soil $TiO_2$ 3-5%; related to $H_2$ in mineralogy. Unsampled.
$h_2$	Continuum slope equally blue as $h_2$ but distinct from $H_2$ and $h_2$ in infrared spectral character; 1 $\mu$ m feature ( $Fe^{+2}$ ) strong but less broad.	Titaniferous basalt. A young medium-high Ti basalt with soil $TiO_2$ 3-5%; related to medium-high-Ti basalts of NE Mare Humorum (4). Unsampled.
$h_3$	Continuum slope comparable to Apollo 12; most extensive unit of the region (no spectra exist to allow further investigation).	Low-Ti basalt. Soil composition 2-3% $TiO_2$ ; older than $H_2$ , $h_2$ , and $h_3$ basalts. Unclassified.
m	Continuum slope slightly red (comparable to Apollo 15 landing site); not an extensive surface unit in this region.	Low-Ti basalt. Soil composition 1.5-2% $TiO_2$ . Unclassified.
1	High albedo; Relatively red continuum slope; 1 $\mu$ m feature stronger than for other low-Ti units of the region. Extensive development of mare ridges.	Low-Ti basalts. Soil composition possibly less than 1.5% $TiO_2$ . Unsampled.
$L_4$	Moderately high albedo; continuum slope relatively red. Topographically higher than mare.	Crater remnants and associated uplands. Mostly feldspathic islands surrounded by mare material.
<u>Upland</u>	Moderately high albedo; upland plains and crater remnants; strong UV absorption causing very red continuum slope; regionally high.	Undesignated highlands. Suspected of being related to other uniquely red highland regions, some of which may be of volcanic origin.
U		
$U_x$		

UPDATE: Characterization of the  $1\mu\text{m}$  spectral differences between the eastern and western high-Ti basalts

Good quality infrared spectra ( $.65$  to  $2.15\mu\text{m}$ ) were obtained by McCord (1977) for representative lunar regions. Shown in Figure IIIE-1 (McCord, unpublished) are spectra for mature soil regions: Apollo 11 ( $H_1$  basalt), Flamsteed (the case area for  $H_2$  basalts), OP-4 ( $h_3$  basalt to the east of Flamsteed), MH-0 (case area for  $h_3$  basalt in Mare Humorum), and the standard MS-2 ( $m_4$  basalt) in Mare Serenitatis. These spectra have been further processed in order to examine the nature of the  $1\mu\text{m}$  absorption. An iron continuum was fitted to the data (Figure IIIE-2) and removed using the techniques of Gaffey (1977) described in Section IID-1b for laboratory samples. The residual absorption features for the five regions are shown in Figure IIIE-3. The background discussion of Section IID-1 can be used to describe the observed spectral features. The following spectral characteristics should be noted: (1) the lack of a second pyroxene feature at  $2\mu\text{m}$  for Apollo 11 and Flamsteed, (2) the relative strength of the  $1\mu\text{m}$  feature for Flamsteed relative to Apollo 11, and (3) the longer wavelength position of the  $1\mu\text{m}$  minimum for Apollo 11 and Flamsteed with respect to MH-0 and MS-2. The relative

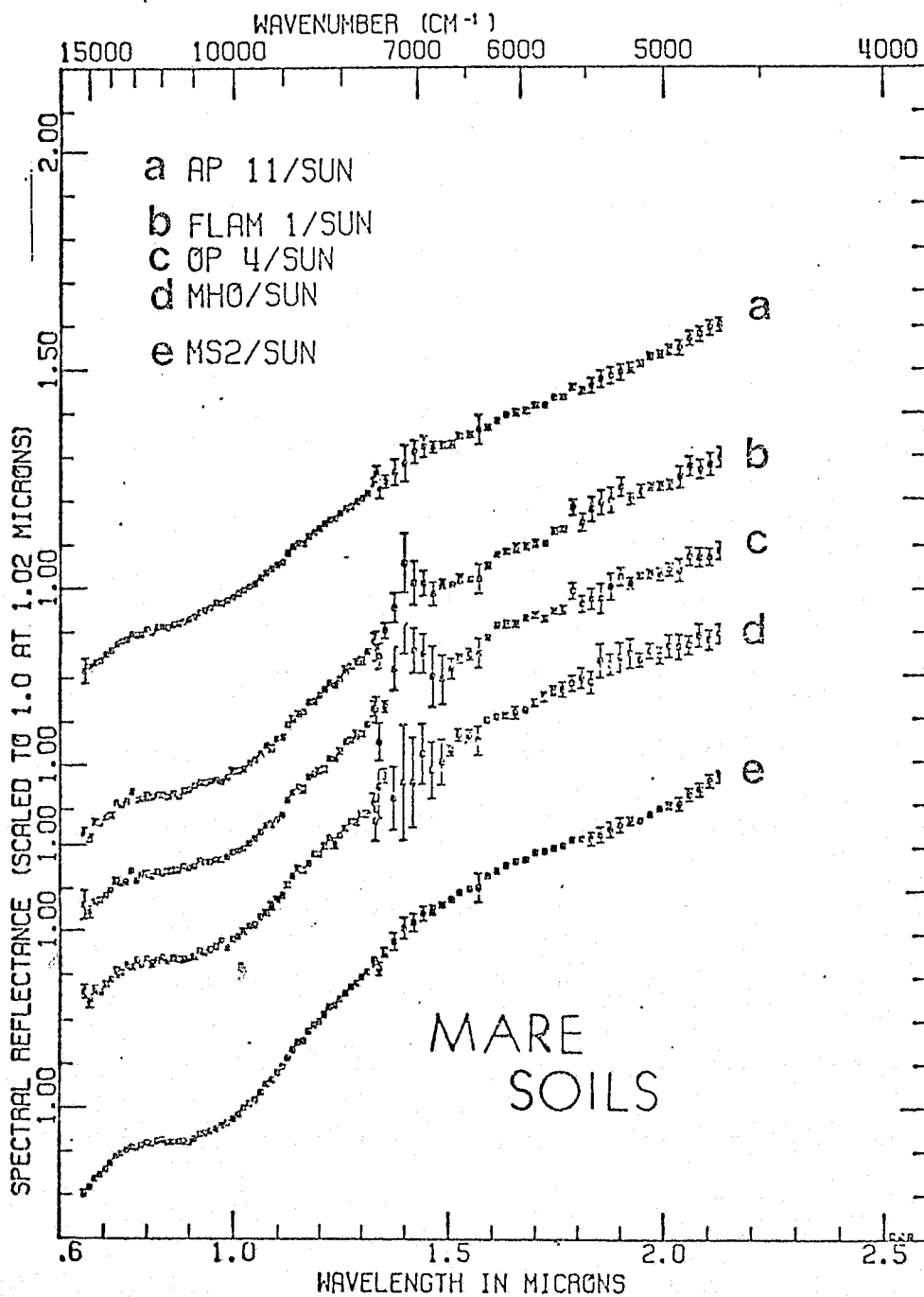
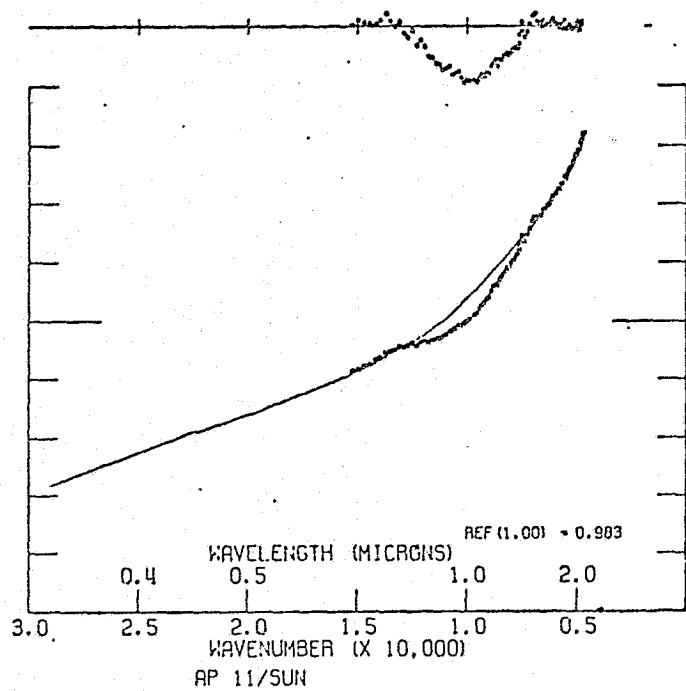
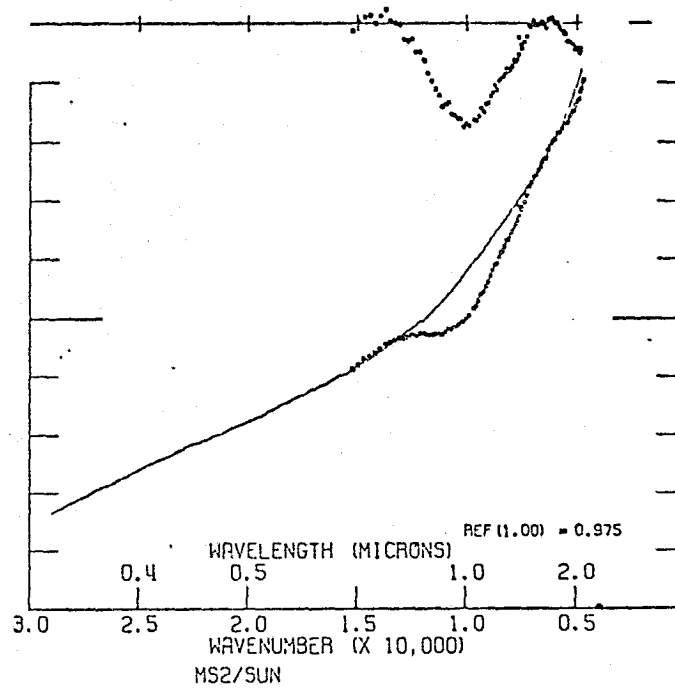


Figure IIIE 1. Recent telescopic spectra for lunar mare soils, .65 to 2.1 $\mu$ m. (McCord, 1977)



CONTINUUM = METALLIC IRON - MOD 2



CONTINUUM = METALLIC IRON - MOD 2

Figure III E 2. An iron continuum is fit to the telescopic spectra. The residual absorption features in the telescopic spectra are shown above the spectra.

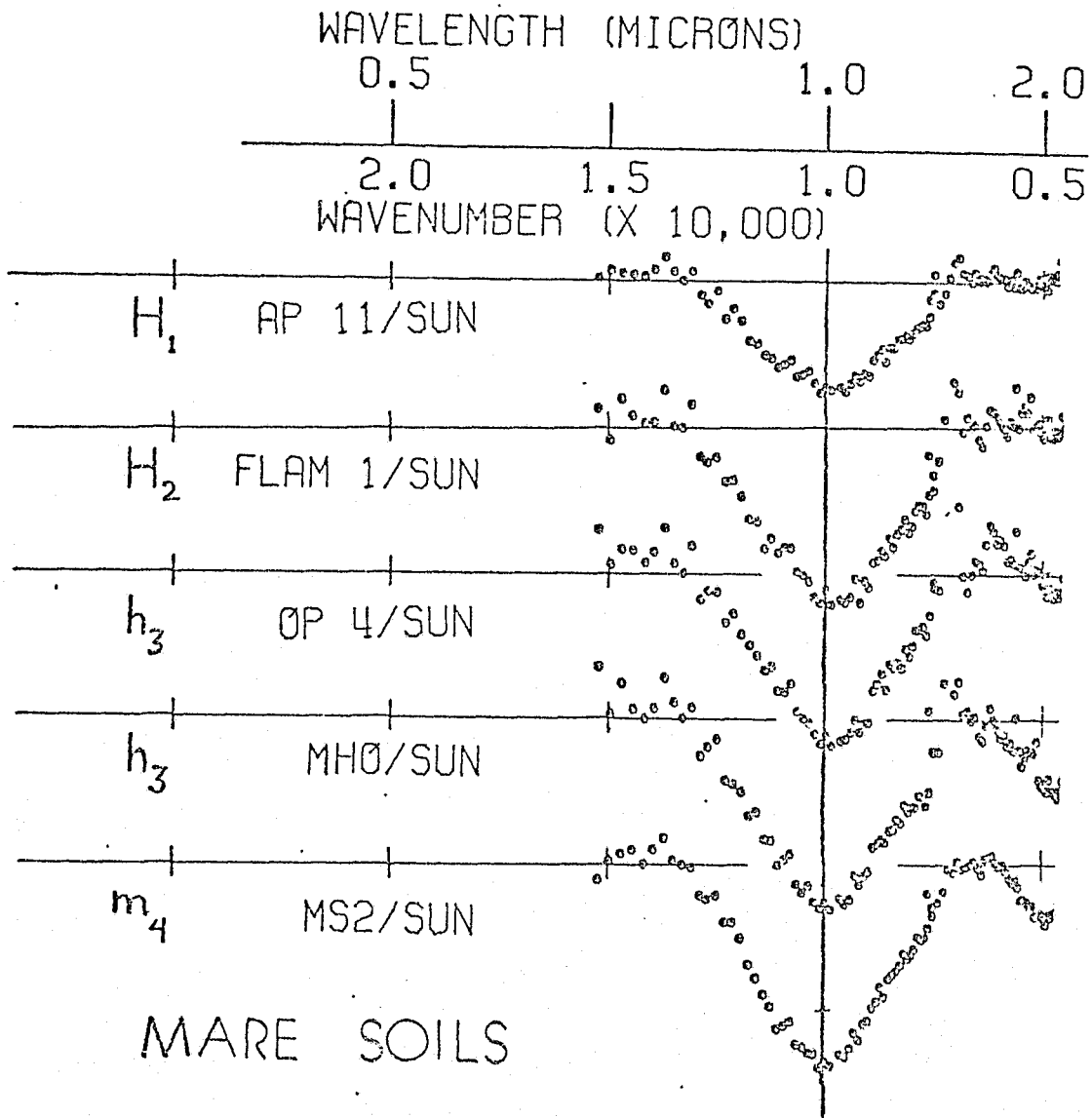


Figure IIIE 3. Residual absorption features for the spectra of figure 1 after an iron continuum has been removed.

strength of the glass and pyroxene bands derived from these spectra for the various basalt types are summarized in Table TIIIE-1. The strength of the continuum absorption in this table is derived from the visible spectra.

The spectral distinction between Apollo 11 ( $H_1$ ) basalts and Flamsteed ( $H_2$ ) basalts concerns the strength of the glass ( $Fe^{2+}$ ) band; Flamsteed spectrum has a stronger band than Apollo 11. The spectral distinction between the two western high-Ti basalts, represented by Flamsteed ( $H_2$ ) and MH-O ( $h_3$ ), is a stronger pyroxene feature in the spectrum for  $h_3$  basalts than for  $H_2$  basalts.

Table TIIIE-1.

Absorption Strengths in basalt type spectra

<u>H<sub>2</sub> (Flamsteed)</u>	<u>h<sub>3</sub> (Humorum)</u>	<u>H<sub>1</sub> (Apollo 11)</u>	<u>m<sub>4</sub> (MS-2)</u>	<u>Soil Absorption</u>
Strong	Strong	Weak	Strong	Glass Band
Weak	Strong	Weak	Strong	Pyroxene Band
Strong	Strong	Strong	Weak	Continuum Ab- sorption

IV. LUNAR BASALT TYPES II: SPECTRA OF CRATERS  
(.3 to 1.1 $\mu$ m)

CHARACTERIZATION OF LUNAR MARE BASALT TYPES-II:  
SPECTRAL CLASSIFICATION OF FRESH MARE CRATERS

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Abstract

Telescopic reflectance spectra (.3 - 1.1 $\mu$ m) of fresh craters are presented and classified according to the spectral features observed. These spectra are the closest lunar surface analogue to laboratory spectra obtained for lunar rock powders. Mineral absorption features can be identified in these crater spectra and interpreted using returned lunar samples. Classification of lunar telescopic spectra indicates that most (>80%) of the lunar surface is composed of a finite number of discrete and describable geochemical units. Spectra of craters to 2.5 $\mu$ m are required for specific mineralogical determinations from remote observation.

In a previous paper, telescopic spectra of mature agglutinate-rich mare surfaces were used to develop a classification of mare basalt types (Pieters and McCord, 1976). It was assumed that as regional basalt surfaces weathered to soil they maintained geochemical identity, although most original mineralogical character was lost. Lunar mare regions were characterized in the previous study by soil  $\text{TiO}_2$  content as derived from continuum slope and by near-infrared spectral features. The data indicate that a variety of specific basalt types exists on the frontside of the moon, the majority of which have not been sampled in any of the Apollo or Luna landing missions.

Spectra of rock powders contain diagnostic absorption features that can be used to define the mineral constituents present (Adams, 1974, 1975; McCord et al., 1976a). Due to the special character of agglutinate-dominated lunar soils, however, it has been difficult to identify the mineralogy of the original basalts using spectral reflectance techniques. The spectral properties of lunar soils are now better understood and changes in optical properties that occur during soil maturation have been described (Adams and McCord, 1973; Adams and Charette, 1975). The objective of the current spectral reflectance study is to further define and characterize the optical properties of fresh, relatively crystalline lunar surfaces. It is apparent that the best

regional analogue on the lunar surface to crystalline rock powders studied in the laboratory are fresh craters (Adams and McCord, 1972). Additional telescopic spectra of fresh craters have been obtained from .3 to 1.1 $\mu$ m. It is anticipated that spectra to 2.5 $\mu$ m for a few craters will be available in the near future (McCord, 1977). Lunar sample studies are being undertaken in parallel with telescopic observations to define the characteristic spectral features of petrographically defined rock types (Charette and Adams, 1975, 1977). The ultimate objective of such a combined program is to identify from absorption features the major mineral phases (e.g., pyroxene, olivine, plagioclase, ilmenite), their composition when possible, and their relative proportions for a variety of regional basalt types previously described (Pieters and McCord, 1976).

Presented here is the first step of the observational program: a classification and analysis of the currently available spectra (.3 to 1.1 $\mu$ m) for fresh lunar craters. The data are used to further confirm that the majority of the lunar surface is made up of a finite number of rock types; some are distinct while others may be gradational. Although higher spatial resolution and greater spectral coverage are required to identify specific mineralogies, these data show that mineral absorption features do exist

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in fresh crater spectra with systematic variations that can be directly interpreted in terms of mineral composition.

#### A. Observations and Data Description

Telescopic spectra of about 400 small (5-25 km diameter) lunar areas have been obtained using a filter photometer covering the spectral range from .3 to 1.1 $\mu$ m. Based on their location and general morphology, most regions can be easily classified as mare, upland, mare crater or upland crater. The general spectral types of such a classification was discussed by McCord et al. (1972) and Charette et al. (1974). Over 60 fresh crater regions have now been observed telescopically during one or more lunations. Due to observational difficulties (see below), data for  $\sim$ 10 regions were unacceptable and not included in this analysis. A minor amount of useful data was obtained with a vidicon spectrometer/polarimeter (.6 - 1.0 $\mu$ m). The location of the craters used in this study are indicated by oversized symbols in Figure 1.

There are a variety of observational limitations that make reliable spectra of fresh craters difficult to obtain. Most spectrometers measure radiation sequentially as a function of wavelength. The photometer used to obtain the data discussed here requires 2 minutes to cycle through 25 filters. During this time, two unavoidable situations

allow the object to 'wobble' in the optical path up to about 2 second of arc (about 4 km on the lunar surface): normal atmospheric turbulence and imperfect lunar tracking of the telescope. This spatial uncertainty during two minutes generally does not affect data for mature soil regions since the albedo of the region is homogeneous on a scale larger than the observed area. For fresh bright craters, however, the albedo variations are large over small areas and the integrated radiation measured by the spectrometer can vary significantly during the total measurement of the spectrum. These observational difficulties are enhanced for small ( $\lesssim 8$  km) fresh craters. Analysis of spectra for small craters is further burdened by the problem of optical mixing that occurs with low (10 km) spatial resolution--part of the radiation detected is from the crater and part is from the surrounding (darker) material. For the data presented here, observations have generally been repeated 3-5 times. A precision better than 1% is generally required for classification.

Two additional factors influence the interpretation of crater spectra. Craters excavate material that may not be homogeneous with depth. Thin surface units may be vertically mixed by the cratering event. Also, the spectral characteristics of fresh material are known to degrade

rapidly with the development of agglutinate-rich soil (Adams and McCord, 1973). Precisely what the time scale is for such weathering effects to be significant is unknown, but expected to be on the order of a few hundred million years. For large craters with sufficiently steep walls, a mature soil is not allowed to accumulate and a spectrum of relatively fresh material can be obtained on the rim, even though the ejecta itself may have reached equilibrium maturity. Most fresh crater spectra presented here seem to have features of comparable magnitude implying maturation differences are minimal.

Telescopic spectra for a variety of soils and craters are presented in Figure 2. Note the prominent absorption near  $1\mu\text{m}$ ; for soils this feature is largely due to  $\text{Fe}^{2+}$  in agglutinitic glasses (Charette and Adams, 1977), whereas for craters this feature is due to  $\text{Fe}^{2+}$  in mineral fragments (e.g., see Adams, 1975). The wavelength of the absorption in crystalline lunar material is determined by the average pyroxene composition (Adams, 1974). The symmetry of the lunar absorption is dependent on absorptions due to other mineral components such as olivine and plagioclase. Weaker absorption features can often be discerned in laboratory spectra in this spectral range (e.g., continuum inflections in the blue and broad absorptions at .4 - .6 in Ti-rich pyroxenes and ilmenite). In Figure 2, the

details of the specific mineral absorption features contribute small but extremely significant variations to a general red lunar spectrum.

#### B. Classification

In order to distinguish the small variations in absorption features and develop a classification of possible rock types, the spectra for lunar craters have been divided by the spectrum of a standard area in Mare Serenitatis (MS2: 18°40'N, 21°25'E). This procedure does not preserve the absolute wavelength position of absorption features but it does allow small spectral variations to be detected and classified. The relative positions of features between areas is maintained.

The craters included in this study were first generally classified as mare or upland craters according to the type of terrain in which they occur. A few exceptions were made to this original grouping and will be discussed along with the more detailed classification of crater spectral types. Since it is possible that mare craters may have penetrated into crustal material, a classification of the spectra for highland craters needs to be made before mare crater spectra can be distinguished.

Relative reflectance spectra for highland crater spectral types are included in Figure 3. From the available

data, three general spectral types were distinguished according to the following characteristics: UI) The continuum is relatively blue between  $.5 - .7\mu\text{m}$  with an inflection towards the ultraviolet that occurs between  $.4$  and  $.5\mu\text{m}$ ; the  $.9\mu\text{m}$  absorption feature varies in strength and perhaps wavelength position resulting in an inflection between  $.7$  and  $.95\mu\text{m}$ . This group was further subdivided (UIa--UIe) according to the sharpness and wavelength position of the ultraviolet inflection and according to the infrared characteristics. UII) There exists a very prominent and strong absorption band at  $.9\mu\text{m}$  with strong relative absorption towards the ultraviolet. UIII) The  $.9\mu\text{m}$  feature is weak to non-existent although there is strong relative absorption shortwards of  $.45\mu\text{m}$ .

Relative reflectance spectra for two major mare crater spectral types are shown in Figure 4. These two spectral types of mare craters are distinguished by: MI) The existence of two absorption bands, a strong one centered near  $1\mu\text{m}$  and a weaker one near  $.5\mu\text{m}$ ; MII) The existence of one strong band near  $1\mu\text{m}$ . Each of these two types of mare craters can be further subdivided according to the wavelength position of the  $1\mu\text{m}$  feature and possibly the strength of ultraviolet absorption. A third group (MIII) of mare cra-

ters have identical spectral characteristics as those identified as highland crater type UIII. These MIII spectra are shown in the lower right of Figure 3.

A few crater spectra do not fit easily into these mare and highland groups:

(a) The center and rim of Kepler has been observed frequently (Figure 5A). The first whole crater spectrum (1-09-71), although noisy, seemed comparable to most mare craters and Kepler was originally classified as a mare crater (McCord et al., 1972). Since then, additional spectra of varying quality have been obtained which together indicate that Kepler is a fairly unique spectral type. There exists a relative minimum in the spectrum near  $.9\mu\text{m}$  and a plateau between  $1.0$  and  $1.1\mu\text{m}$ . A region within the ejecta from Kepler, however, contained the same spectral character as group UIId and was thus classified as highland material.

(b) Three regions (Lalande, Reiner  $\gamma$ , and Dawes rim) may be mixtures of a variety of mare and highland material. Their spectra (Figure 5b) are more similar to each other than to other craters, however, implying perhaps related mineralogy.

(c) Hesodius B (and perhaps Dawes ejecta) is an old

crater (Eratosthenian) in a blue (high-Ti) mare. Its spectral characteristics (Figure 5c) are consistent with the interpretation that the surface is an intermediate stage during degradation of high-Ti basalt to soil with maturity.

### C. Discussion

Relative reflectance spectra are especially useful for understanding spectral differences seen in two-dimensional imagery. With the exception of MIII and UIII craters, the pyroxene band near  $.95\mu\text{m}$  is stronger in crater spectra than in mature soils. A  $.95/.57\mu\text{m}$  ratio image (cf., McCord et al., 1976b) is thus frequently used as a data source to easily identify immature regions and to estimate their areal extent. Spectral ratio images at other wavelengths, however, do not provide sufficient information for classification of craters. As can be seen in Figures 3-5, there is great variety in the strength of spectral features and the wavelength of spectral inflections. No simple combination of spectral ratio images can distinguish the spectral types discerned in the crater spectra presented here.

It has been shown previously that the general spectral characteristics of fresh craters are essentially the same

as those for rock powders measured in the laboratory (Adams and McCord, 1972). The strength of the features in laboratory measurements, however, are about 4 times as strong as those observed telescopically. The data presented here, together with a recently compiled atlas of similar laboratory data (Adams et al., 1977), reestablishes this link between rock powders and fresh craters and will be discussed in further detail in later publications. The laboratory samples with spectral characteristics that are most comparable to those for fresh craters, both in nature and in strength, are the non-agglutinate separates from immature soils. Either the surface material at all craters observed in this study has matured to some degree, causing the mineral spectral features to become somewhat degraded, or the cratering event itself has created a surface material that is not purely crystalline.

Some spectral characteristics that distinguish the subgroups of craters presented here appear to be gradational in nature. Spectral types may be related to each other as a function of the maturity of the surface as suggested by McCord et al. (1972) and Adams and McCord (1973). Alternatively, the gradations could be due to a mixing of different rock types. The data presented in Figures 3-5

indicate that distinct rock types do exist and suggest that many of the spectral variations are not simple a matter of surface maturity.

A few cases where spectral evidence indicates a large mare crater has excavated crustal material have already been documented (McCord et al., 1972): Aristarchus and Copernicus. (The soil of Copernicus ejecta, however, has already developed to a fairly mature agglutinate-rich state.) Additionally, from the data presented here, one group of smaller mare craters (MIII) have apparently excavated highland material of the same type as that found in UIII craters. MIII craters occur near the edge of the mare on what appears to be thinly flooded basin shelves in Mare Humorum and Serenitatis.

A few spatial systematics can be tentatively identified in these data. All of the regions in highland group UIId and III are on basin rims. The Apollo 16 region (Descartes) contains all of the craters identified as group UIe and a few from UIc. Spectra of very immature lunar soils from Luna 20 and Apollo 16 contain these same spectral characteristics (Doms et al., 1977). From the data available,

the craters Aristarchus (UIa) and Mosting A (UII) are singularities in the remote observations although almost identical spectral characteristics can be identified in a few lunar highland samples (Adams and McCord, 1972, Adams et al., 1977). Identification of lunar samples with spectral characteristics comparable to the craters in group UIb and III has not been made.

The two primary absorption features observed in mare craters have clear compositional implications. The strong absorption features at  $1\mu\text{m}$  is primarily due to clinopyroxenes in the surface material (Adams and McCord, 1972). The wavelength position of this band is a function of the Fe and Ca content of the average pyroxene (Adams, 1974). If the wavelength position of both this band and the accompanying one near  $2\mu\text{m}$  can be measured, then the pyroxenes composition can be determined. The feature in mare crater MI spectra near  $.5\mu\text{m}$  is commonly observed in titanium-rich rocks and is due to either titanium-rich pyroxenes or ilmenite.

The occurrence of type MI and MII craters is generally associated with the  $\text{TiO}_2$  content of the surface unit (as derived from soil spectra): type MI craters occur in regions that are titanium rich and type MII occur in regions

that are low titanium. Halo crater (MIb) occurs in the Copernicus ejecta blanket and is likely to have excavated high titanium basalt.

It is unfortunate that no spectra for craters within the major unsampled basalt types of the western moon (Pieters and McCord, 1976) were available for this study. A primary reason for this lack of data is the difficulty of locating a crater within the current telescopic resolution ( $\sim 10$  km) that is likely to have excavated only the most recent surface unit. For example, the western high-Ti flows near Flamsteed are thin and many craters seem to have excavated an earlier low-Ti unit (Pieters et al., 1977, in preparation).

#### D. Conclusions.

(1) These additional data strengthen and expand the conclusion of Adams and McCord (1973) that the spectral features evident in spectra of fresh crater are stronger than those for mature soils (Figure 2) and are due to the more crystalline nature of the surface material.

(2) The systematic variations observed in the spectral features of these fresh craters are directly associated with variations in the mineralogy of the surface material and can be understood largely in terms of returned lunar samples. To utilize this information in a quantitative

manner, spectra to  $2.5\mu\text{m}$  are required in order to examine the details of mineral absorption features.

(3) This spectral classification of fresh craters together with the previous classification of mature mare surfaces, an unfinished classification of highland surfaces, and the orbital geochemistry data indicate that most ( $>80\%$ ) of the lunar surface is composed of a finite number of discrete and describable geochemical units. Even though  $1/2$  to  $2/3$  of these units are unsampled, there is some hope of understanding their general mineralogy and geochemistry in the future using remote sensing techniques.

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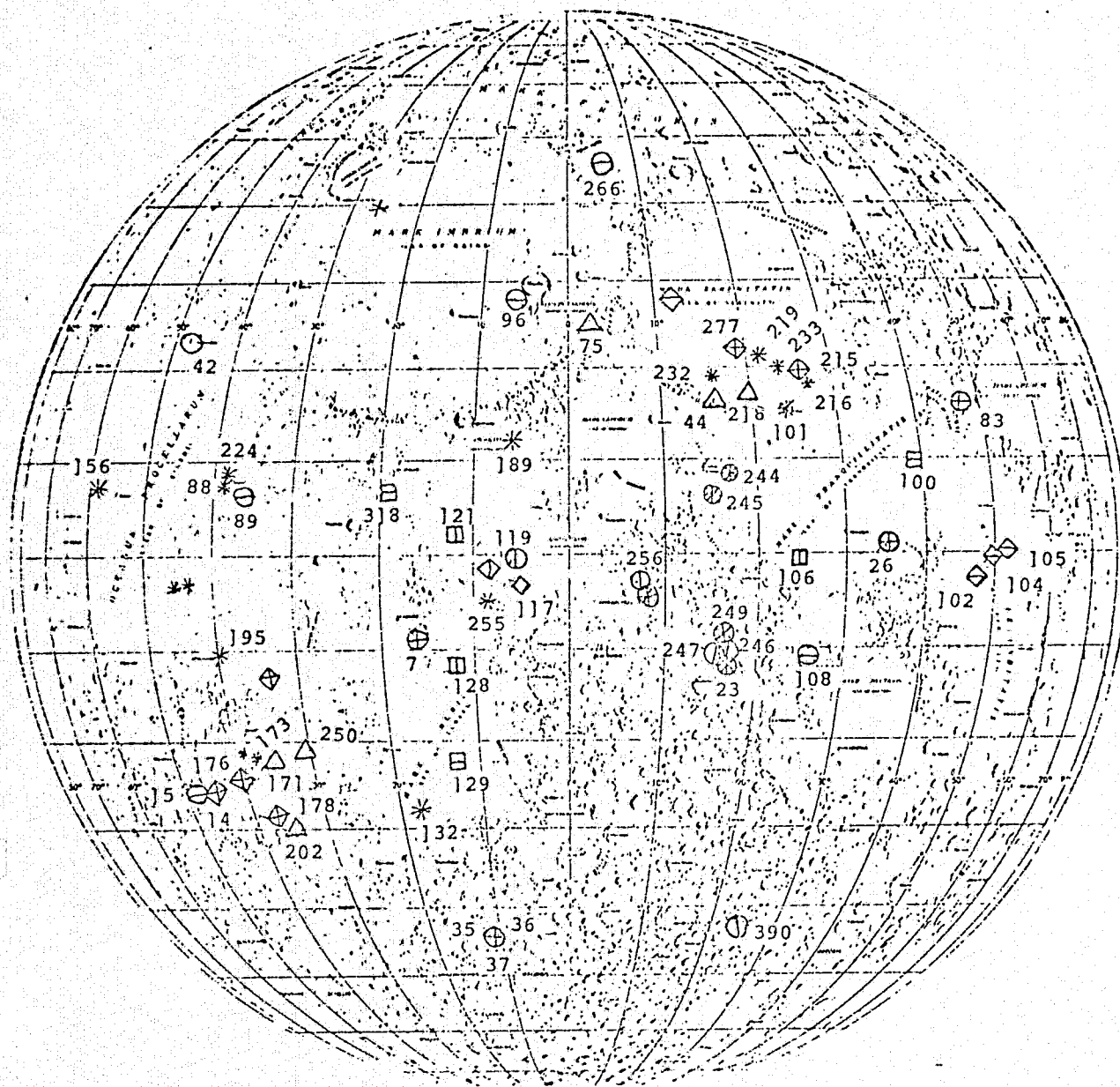
Figure Captions

1. Classification of lunar bright craters based on spectral reflectance measurements. The shape of the oversized symbols indicates the spectral groups discussed in the text. The number associated with each symbol is for identification of the spectra in Figure 2-5.
2. Telescopic reflectance spectra for representative lunar soils and craters.
3. Reflectance spectra relative to MS-2 for highland crater material. Three major groups are distinguished. Group I can be further subdivided (a-e) according to the wavelength position of distinct spectral inflections. Groups UIII and MIII are spectrally equivalent although craters of the former are in the highlands and of the latter are on the edge of a mare.
4. Reflectance spectra relative to MS-2 for mare craters. Group I has two absorption features, one near  $.5\mu\text{m}$  and one near  $1.0\mu\text{m}$ . Group II only has one major feature near  $1.0\mu\text{m}$ . Each group can be further subdivided according to the wavelength position of the  $1.0\mu\text{m}$  feature.
5. Reflectance spectra relative to MS-2 for special crater groups. Kepler center and rim appear to have unique spectral characteristics. LaLande, Reiner Gamma, and

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Dawes are likely to be mixtures of material although they have roughly similar spectra. Spectra for Dawes ejecta and Hesiodus B appear to show significant maturation effects.

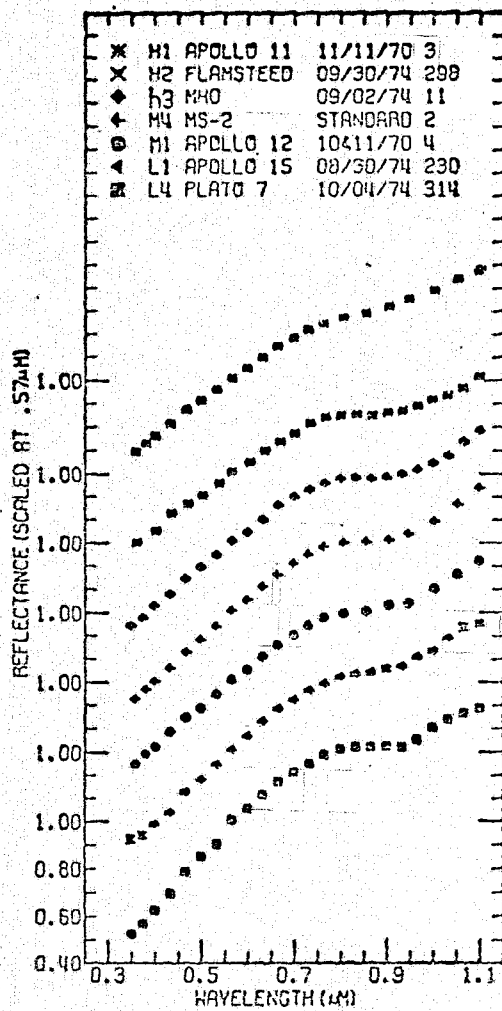
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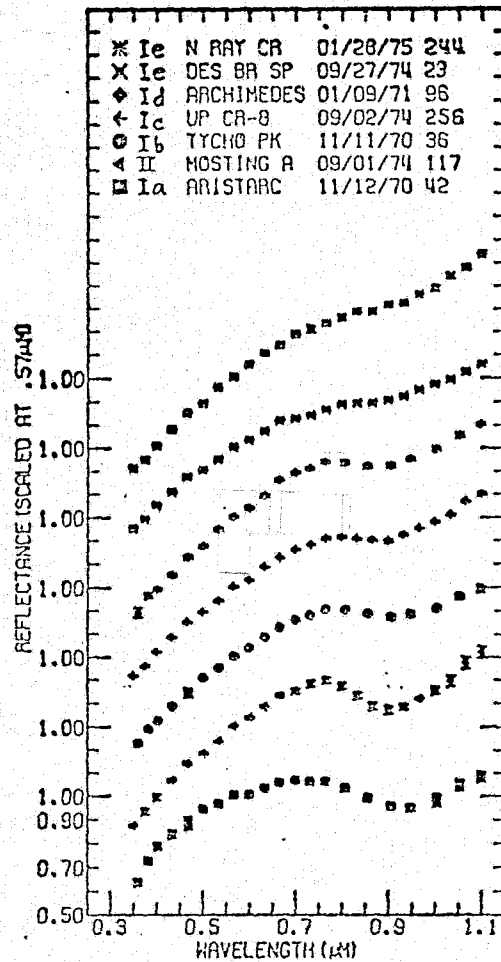
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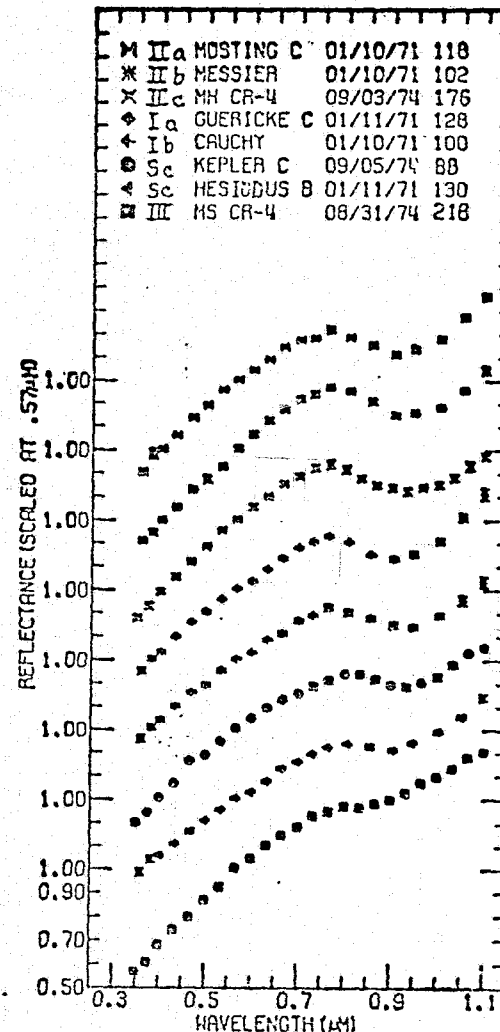
Figure 1.



MARE SOILS



UPLAND CRATERS



MARE CRATERS

Figure 2.

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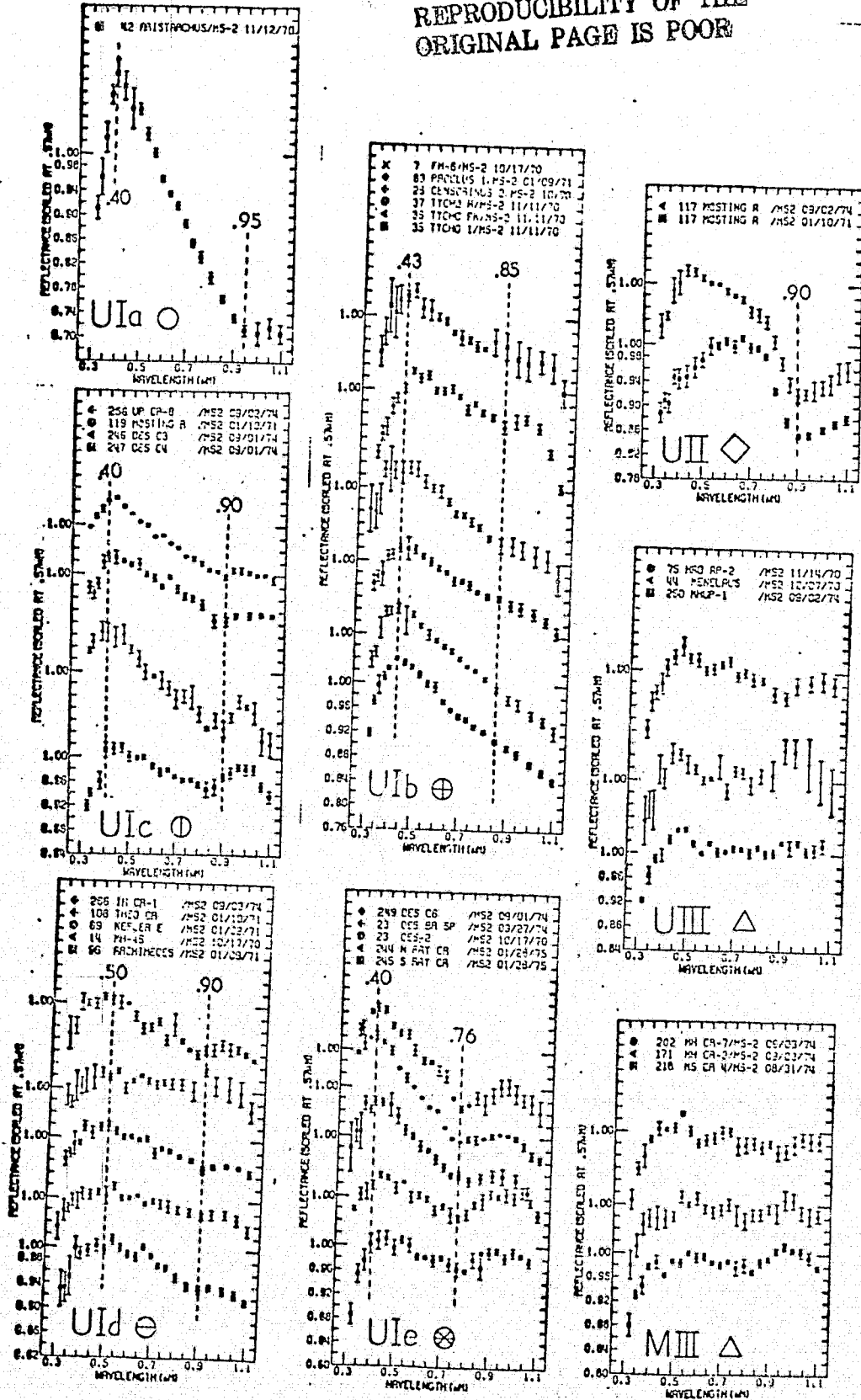


Figure 3.

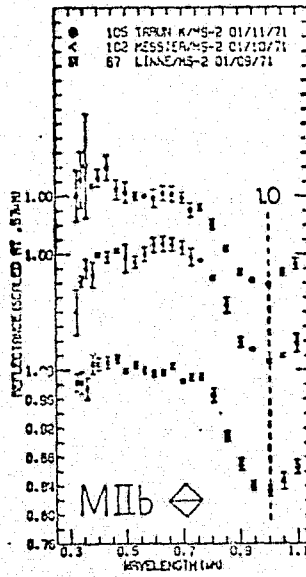
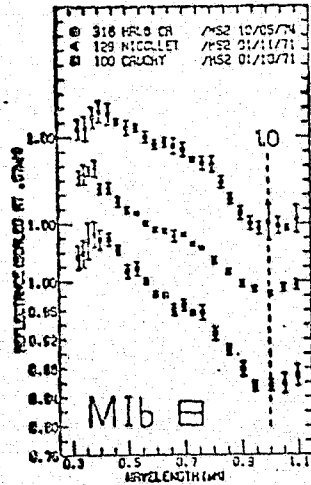
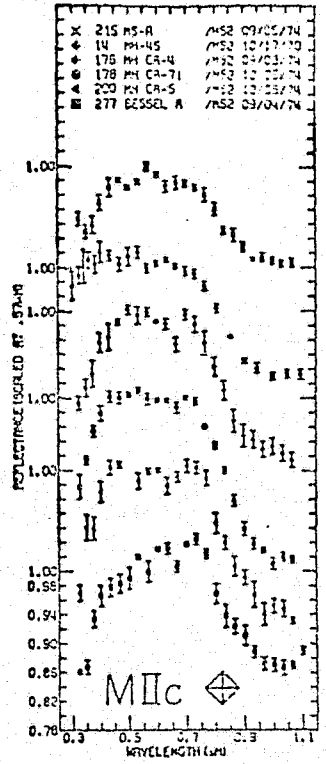
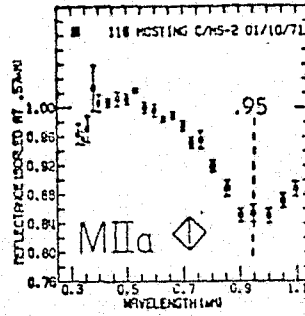
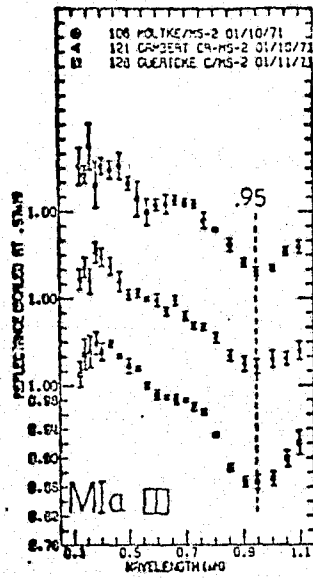


Figure 4.

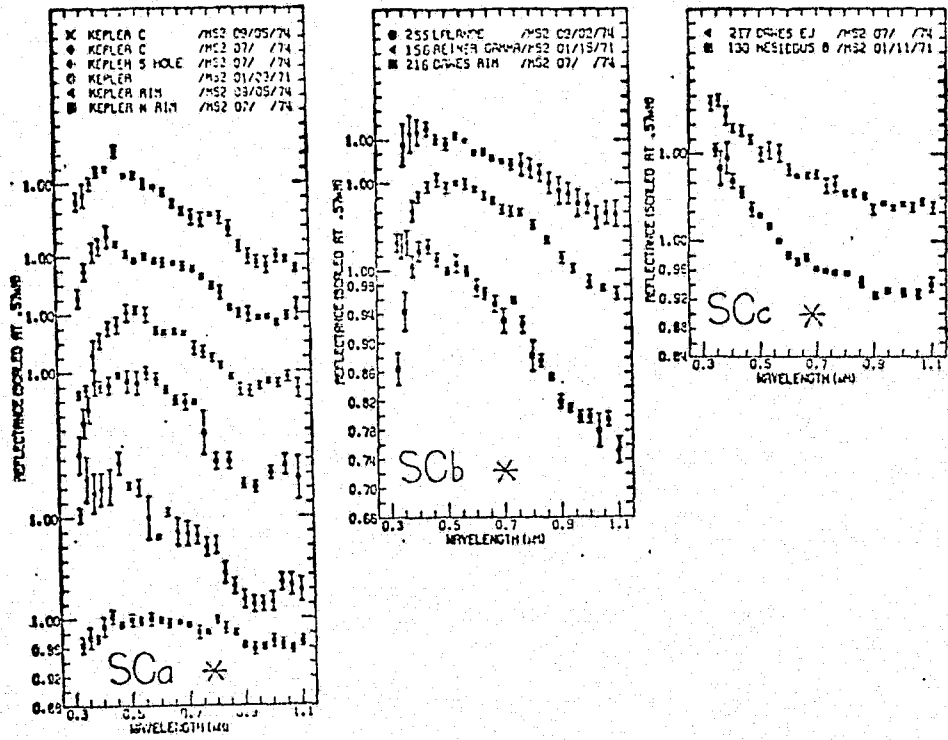


Figure 5.

V. SUMMARY AND SYNTHESIS

## V. SUMMARY AND SYNTHESIS

Samples of basalts from two planets, the earth and the moon, have been studied in detail by petrologists and geochemists in order to understand the geochemistry of the interiors as well as the processes that have been part of planetary evolution. Basalts from these two planets are similar in the general sense that they have crystallized from mafic igneous melts derived by partial melting of the planet's interior. However, it is also apparent there are some fundamental differences in chemistry, age, mode of emplacement, tectonic environment, and by implication, character of the mantles. The Apollo program opened a new era of exploration; the return of the lunar samples allowed the first direct comparison of the earth with another known solar system object and deepened the understanding of our own planetary environment.

The basic nature of the samples from the moon will continue to be studied in terrestrial laboratories to further understand the evolution of the planet. However, until more comprehensive samples are obtained of yet unsampled units, it will be the role of remote sensing specialists to place limits on what can be extrapolated globally from the selective samples in hand returned by Apollo and Luna missions. Near future exploration of the lunar

surface will have to be performed using remote sensing techniques. Spectral reflectance measurements in particular have been used to derive information of surface mineralogy and geochemistry.

## V. SUMMARY AND SYNTHESIS

### A. Conclusions

Spectral reflectance data (spectra and imagery) discussed in the preceding sections allow the following general conclusions to be made concerning mare basalt types on the frontside of the moon.

1. Regional basalt types exist. There are regions up to hundreds or perhaps thousands of square kilometers in extent of essentially uniform surface composition. There are also groups of smaller flows that are widely separated but nevertheless apparently identical in character. The homogeneity of some surface units is evident in various spectral ratio images for extensive regions. The spectral character of a few of these units have been studied in detail, providing information to allow their regional extent to be mapped (e.g., Mare Tranquillitatis, Central Serenitatis, NE Humorum, West Imbrium; see Section IIB). One implication of the observed regional homogeneity is that if surface fractionation processes act to significantly change the surface composition (or spectral character of a unit) then such processes must do so uniformly over extensive areas.

2. Basalt types have been characterized (from soil spectra) in terms of certain useful geochemical characteristics (e.g.,  $\text{TiO}_2$  content) and mineralogical features

(glass and pyroxene components).

3. About 15 spectrally distinct surface units can be identified for mare soils. These represent at least 6 fundamental (extensive) basalt types or related series.

(a) Apollo 12, 15 ( $m_1, l_1$ ) low titanium basalts found in most nearside regions. SAMPLED.

(b) Apollo 17, 11, Luna 16 ( $H_1, h_1$ ) high to medium high titanium basalts found on the eastern moon. Weak pyroxene band, weak glass ( $Fe^{2+}$ ) band. SAMPLED.

(c) Flamsteed, Imbrium ( $H_2, h_2, m_2$ ) high to medium titanium basalts found only on the western moon. Weak pyroxene band, strong glass ( $Fe^{2+}$ ) band. UNSAMPLED. These basalts include the western "blue" basalts with  $TiO_2$  content likely to be distinctly high, comparable to that at Apollo 11, but with different mafic mineralogy or composition.

(d) Mare Humorum ( $h_3$ ) medium high titanium basalts found only in the western maria. Strong pyroxene band, strong glass ( $Fe^{2+}$ ) band. UNSAMPLED.

(e) Serenitatis ( $m_4$ ) low titanium basalts. Strong pyroxene band, strong glass ( $Fe^{2+}$ ) band. UNSAMPLED.

(f) East Imbrium, Frigoris ( $L_2, L_4$ ) (very) low titanium basalts. UNSAMPLED.

4. Although not all regions have been spectrally analyzed, it is clear that the lunar samples represent only a few of the major basalt types on the lunar surface.

It appears that about 2/3 of the lunar maria are not represented in the lunar sample collection returned by U.S. Apollo and Soviet Luna missions.

5. All mare surfaces contain lateral variations of compositionally heterogeneous basalts. Although a few regions contain vast areas of apparently homogeneous basalt (conclusion 1) many mare regions are heterogeneous on the lateral scale of 2-50 kilometers. This heterogeneity is apparent in any detailed study of high resolution spectral images. An example of this lateral heterogeneity occurs at the Apollo 12 landing area. Remote sensing studies show two compositionally distinct units within a few kilometers. Recent examination of the Apollo 12 ilmenite basalts indicates they are unrelated to the rest of the Apollo 12 low-Ti basalts and may indeed represent this second unit. These two basalt types were roughly contemporaneous.

6. Mare fill (even in the mascon basins) may also be stratigraphically heterogeneous. The only spectrally documented case is Mare Crisium for which the likely sequence of mare fill, oldest to youngest, was: a high-Ti

basalt, a (very) low-Ti basalt, and a (medium) low-Ti basalt. Photogeologic evidence and multispectral imagery strongly suggests Mare Serenitatis may have had a somewhat similar sequence of fill (high-Ti followed by low-Ti). The western maria seem to be far more complex.

7. Some basalt types are spectrally gradational suggesting minor variation in geochemistry; three regional series can be identified. Two clear series are evident in the western maria: (a) Flamsteed-Imbrium high to medium high-Ti basalts ( $H_2$ ,  $h_2$ ,  $m_2$ ), and (b) Mare Humorum medium high-Ti basalts ( $h_3$  varying in  $TiO_2$  by about 3%). It is suggested that the basalts for each series are derived from the same source region with minor variation in P, T, chemistry or degree of partial melting. The observed variations could also be due to near surface fractionation. The Flamsteed flows (see Figure IID-21) appear to be individually homogeneous and distinct, whereas the variations in the Humorum basalts are regionally diffuse. Spectral data also indicates there may exist a sequence of basalt types in the eastern maria that range in composition from Apollo 11 ( $H_1$ ) to Luna 16 ( $h_1$ ).

8. Although most mare surfaces are now composed of mature soil, more crystalline material is excavated by fresh craters. Spectra of fresh craters are comparable

to those obtained for laboratory samples of very immature soils and allow the mineralogy of an ejecta unit to be examined. The variations in spectral features for fresh craters are known to be directly related to specific mineral absorption features of minerals. Thus, if spectra to 2.5 $\mu$ m are obtained for fresh craters, components of the mineralogy of the excavated unit can be discerned.

## V. SUMMARY AND SYNTHESIS

## B. Inferences

The spectral information (both spectra and imagery) integrated with other global data (e.g., Ap 15, 16 orbital geochemistry, relative age units derived from crater degradation studies, thermal and geochemical models of the moon, etc.) support the following inferences.

1. The bi-modal relationship observed in both age and  $TiO_2$  content of mare basalt samples is likely to be simply a matter of limited sampling. The spectral data suggest a continuum of  $TiO_2$  content for mare basalts.

When spectral data for  $TiO_2$  are combined with crater degradation age units of Boyce (1976), a more complicated relationship emerges (e.g., Soderbloom and Boyce, 1976).

From the data currently available, an update of the  $TiO_2$  and age relationship of Papike et al. (1976) is estimated and is shown in Figure V-1.

2. It is unlikely that the lunar mantle is zoned in any simple global sense and that the sequence of observed basalts is simply the result of partial melting of progressively deeper regions. Thermal models show that the upper limit of a zone of partial melt in the lunar mantle becomes deeper with time. Any thermal and geochemical model for the source region of mare basalts must also account for

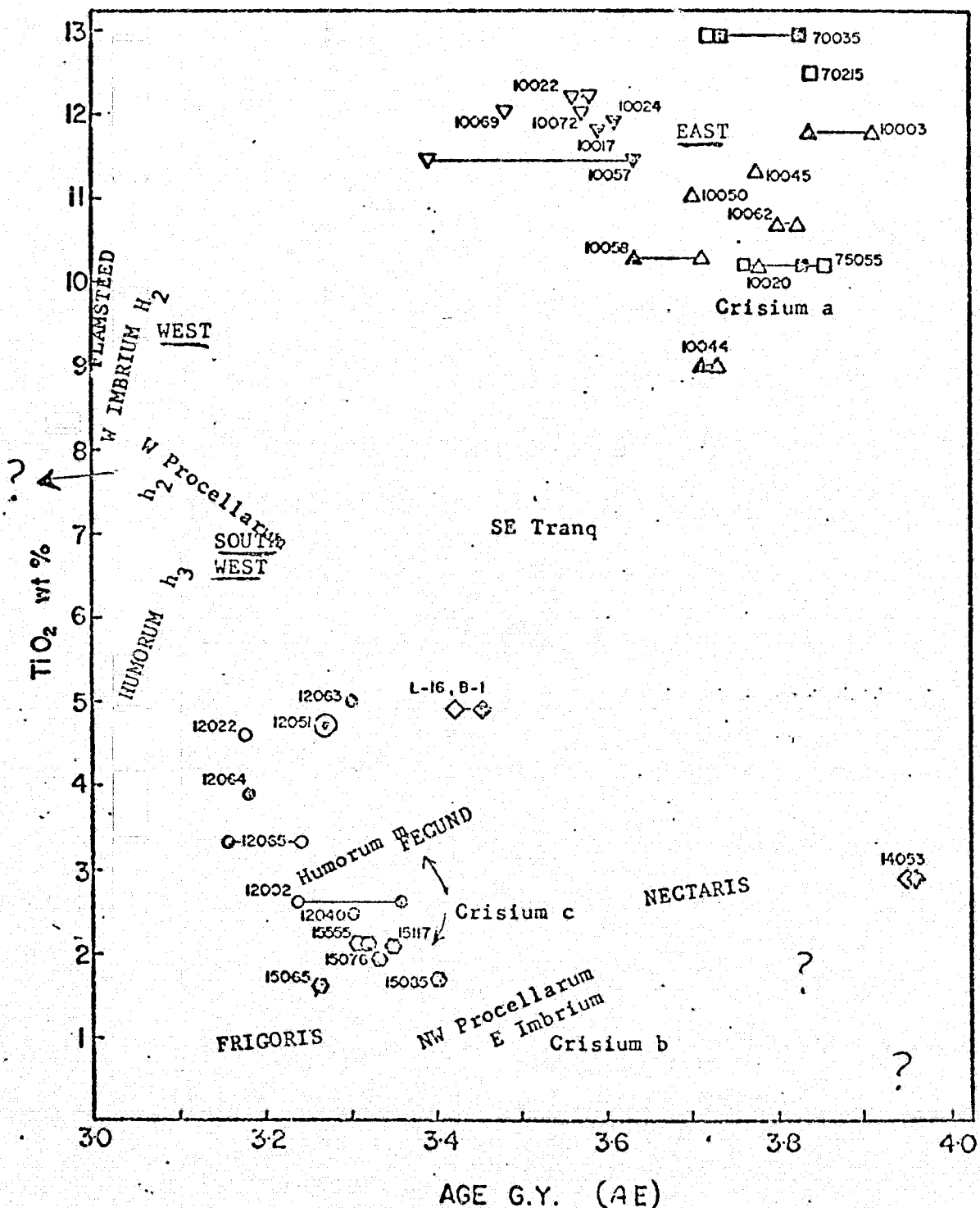


Figure V 1. Estimates of  $TiO_2$  content (from continuum slope) and age (from Boyce, 1976) for various sampled and unsampled mare basalts. (The  $TiO_2$  content estimates are very rough since the data are calibrated for soils.) Lunar sample information was compiled by Papike et al., (1976). Many basalt types appear to occur simultaneously. The pattern, if there is any, suggests the earliest basalts were either very high-Ti or very low-Ti. Around 3.3 AE large amounts of medium low-Ti basalts were produced. There appears to be an increase of  $TiO_2$  with the latest basalts.

the frequently observed heterogeneity of nearly contemporaneous basaltic units both locally and globally. For example, if the same source region depth was tapped for the two Apollo 12 types of basalt, then the lunar mantle must contain significant lateral heterogeneities. An alternative model is that different depths are tapped within the zone of partial melting during any particular lunar era.

3. The lunar nearside maria can be roughly divided into two and perhaps three regions which seem to have evolved independently in terms of both time and geochemistry: activity in the eastern maria apparently ended first; in the western maria last; and in the southwestern maria (Humorum, Nubium) perhaps sometime between the two. Orbital geochemistry indicates a concentration of heat producing elements in the surface rocks of the western regions, perhaps implying a more vigorous radioactive heat source at depth.

## V. SUMMARY AND SYNTHESIS

## C. Recommendations

A few specific and general recommendations for the lunar science community (and NASA) are perhaps justified from these results and inferences.

1. Since many major lunar basalt units have not been sampled directly, the continued search for other distinct basalt types as minor fragmental components of the returned samples could significantly increase the information from the samples. The recent discovery of VLT basalts in the Apollo 17 core shows that this is a fruitful endeavor. It is likely that correlated studies between sample and remote sensing geochemists can identify the type area of such minor components.

2. Caution should be exercised in extrapolating lunar sample information to global geochemistry. The sequence of mare basalt extrusion is locally and globally geochemically variable and complex. This is not to say regional patterns cannot be identified and characterized, but it is important to recognize that only a limited amount of geochemical information is available in the samples.

3. Integrated studies between lunar sample geochemists and remote sensing specialists should be encouraged. Two specific topics need to be addressed:

(a) "Ground Truth" must be rigorously defined. The largest scale observable in the laboratory is measured in centimeters. The smallest scale currently observable with remote sensing techniques is measured in kilometers. The rock types identified in the samples need to be related to regional units observed remotely. Furthermore, the two types of data need to be expressed in the same terms. For example, spectroscopic techniques can determine the average pyroxene composition for a regional rock type (which varies from one rock type to another). "Average pyroxene composition", however, is somewhat a foreign concept to petrographers.

(b) An accurate model of soil genesis needs to be developed that quantitatively links the mineralogy and geochemistry of the regional rock type (s) to that for a mature soil. Most of the lunar surface has developed a regolith and mature soil, and it is such a soil that is observed using remote sensing techniques. On the other hand, most useful laboratory geochemistry is based on crystalline or rock samples.

4. An effective program of lunar surface exploration can and should be undertaken to map the global geochemistry of the moon. The rationale for such a program is twofold:

(a) The scientific understanding of the moon that resulted from the Apollo program has matured to the extent that global data will permit a maximum scientific yield from the huge investment in Apollo. The U.S. scientific expertise that has enthusiastically carried this exploration program will be dispersed if lunar exploration is terminated half-way through the sequence.

(b) If the moon is to be utilized in the future as a scientific and economic resource, the global chemistry must first be mapped and explored. With present technology, such an assessment of lunar materials and resources is relatively easy and can be done at low cost.

A limited amount of further geochemical information can and should be obtained using earth-based telescopes and recently developed spectroscopic instruments. The obvious and perhaps ultimate application of spectral reflectance techniques for remotely determining lunar geochemistry would be done from lunar orbit: continuous high resolution (1/2 km) spectroscopy (.3 - 2.5 $\mu$ m) of surfaces and craters with detailed two-dimensional spectral mapping at selected useful wavelengths.

The spectral reflectance study of lunar basalts presented here has shown that there are significant numbers of unexplored regions and unsolved problems that cannot

be understood simply by a more extensive study of the lunar samples. A rigorous program of lunar exploration is certain to uncover many unexpected and extremely useful facts about the moon. If the moon is to be utilized as a long-range asset, rather than simply an object of interesting beauty, detailed exploration and mapping is an essential endeavor.

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M.S. Planetary Science	- Mass. Inst. of Tech.	1972
Ph.D. Planetary Geochemistry	- Mass. Inst. of Tech.	1977

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January 1970 - August 1971: Undergraduate Research Assistant, M.I.T. Planetary Astronomy Laboratory, Department of Earth and Planetary Sciences

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Synthesis of spectral reflectance data with  
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problems of surface geology and geochemistry

Terrestrial geological applications of remote  
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Stellar calibration (for planetary reflectance)

Spectral polarimetry of planetary surfaces

Orbital geochemical experiments

CARLE PIETERS PUBLICATIONS  
MIT REMOTE SENSING LABORATORY

1. McCord, T.B., M. Charette, T.V. Johnson, L.A. Lebofsky, C. Pieters, J.B. Adams. "Lunar Spectral Types." Journal of Geophysical Research, 77, 1349-1359, (1972).
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5. Chapman, C.R., T.B. McCord, C. Pieters, "Spectrophotometric Study of the Composition of (1685) Toro." Astronomical Journal, 78, 502-505, (1973).
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10. Adams, J.B., C. Pieters, and T.B. McCord, "Orange Glass: Evidence for Regional Deposits of Pyroclastic Origin on the Moon." Proceedings of the Fifth Lunar Science Conference, 1, 171-187, (1974).
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16. Pieters, C., M. Gaffey, C. Chapman, T.B. McCord, "Spectrophotometry of 433 Eros (0.33-1.0 $\mu$ m) and Compositional Implications." Icarus, 28, 105-115, (1976).
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20. McFadden, Lucy, T.B. McCord, Carle Pieters, "Vesta: The First Pyroxene Band from New Spectroscopic Measurements," Icarus, in press (1976).
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25. Pieters, C. "Characterization of Lunar Mare Basalt Types-II: Spectral classification of Fresh Mare Craters." Submitted to Proc. Lunar Sci. Conf. 8th, 1977.