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**WORKSHOP ON
GEOLOGY OF THE APOLLO 17 LANDING SITE**

Edited by

G. Ryder, H. H. Schmitt, and P. D. Spudis

Held at
Houston, Texas

December 2-4, 1992

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Cover: Station 4 at Taurus-Littrow, Apollo 17 landing site. The rover is on the rim of Shorty Crater, an impact crater 110 m in diameter, which has excavated dark material (volcanic pyroclastic glasses) from beneath the light mantle (highlands debris). LM Pilot Jack Schmitt prepares to study, sample, and characterize the dark ejecta deposits of this crater.

Primary

PREFACE

This volume contains papers that have been accepted for presentation at the Workshop on Geology of the Apollo 17 Landing Site, December 2–4, 1992, in Houston, Texas. The Program Committee consisted of G. Ryder (*Lunar and Planetary Institute*), H. H. Schmitt (*Consultant, Apollo 17 Lunar Module Pilot*), and P. D. Spudis (*Lunar and Planetary Institute*).

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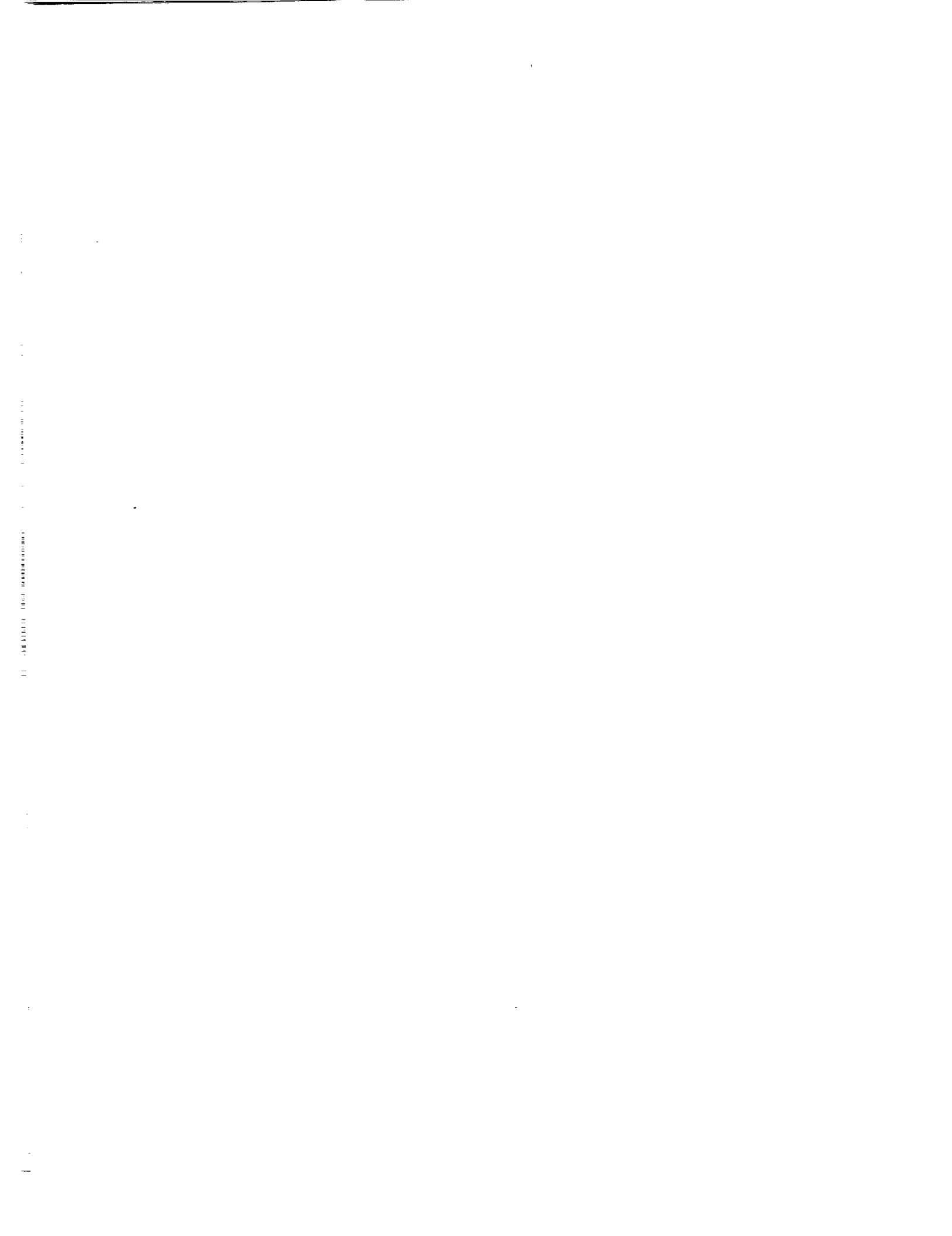
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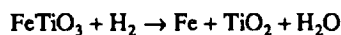
LUNAR RESOURCES—OXYGEN FROM ROCKS AND SOIL.
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Introduction: Permanent human habitation of the Moon will require the use of local resources because of the extremely high cost of transporting materials from the Earth. One of the first lunar resources to be exploited will be oxygen, for both life support and rocket propulsion [1]. Projected demand for oxygen at a lunar base ranges from tons to hundreds of tons per year.

Among the major-element oxides in lunar materials, the iron-oxygen bond is most easily broken. Some oxygen production proposals concentrate on releasing only oxygen bound to iron. Studies of the TiO₂ system, however, have shown that a portion of the oxygen normally required to make TiO₂ can also be released under highly reducing conditions. Some of the Ti⁴⁺ is reduced to a lower oxidation state, presumably Ti³⁺.

Essentially all the iron and titanium oxide in the lunar regolith is concentrated in three mineral groups: metal oxides (ilmenite and rutile), pyroxene, and olivine. Studies of terrestrial minerals show that the high-temperature reduction of iron oxide in ilmenite is rapid and essentially complete [2]. Pyroxene, olivine, and rutile can also be partially reduced. Compared to ilmenite, the reduction of these minerals is slower and less efficient [3].

The "baseline" process for lunar oxygen production involves reduction of ilmenite [2]. A typical reaction, using hydrogen as a reducing agent, is



The water is decomposed to yield oxygen and the hydrogen is recycled as a reactant.

Most scenarios for oxygen production envision reaction of the lunar soil rather than the rocks. Lunar soil samples show evidence of limited natural reduction in the form of iron metal blebs on and within grains of agglutinate glass [1]. If the impact-driven maturing process has subjected the ilmenite or other minerals in lunar soil to significant reduction, mature soils could be undesirable feedstocks for an oxygen plant.

Experimental: Lunar and terrestrial basalt. The first set of hydrogen reduction experiments to use actual lunar material has recently been completed. The sample, 70035, is a coarse-grained vesicular basalt containing 18.46 wt% FeO and 12.97 wt% TiO₂ [4]. The mineralogy includes pyroxene, ilmenite, plagioclase, and minor olivine. The sample was crushed to a grain size of <500 μm.

The crushed basalt was reduced with hydrogen in seven tests at temperatures of 900°–1050°C and pressures of 1–10 atm for 30–60 min [2]. A capacitance probe, measuring the dew point of the gas stream, was used to follow reaction progress.

Experiments have also been conducted using a terrestrial basalt similar to some lunar mare samples. Minnesota Lunar Simulant (MLS-1) contains 13.29 wt% FeO, 2.96 wt% Fe₂O₃, and 6.56 wt% TiO₂ [5]. The major minerals include plagioclase, pyroxene, olivine, ilmenite, and magnetite. The rock was ground and sieved, and experiments were run on the <74- and 500–1168-μm fractions. Experiments were also conducted on <74-μm powders of olivine, pyroxene, synthetic ilmenite, and TiO₂ [3].

The terrestrial rock and mineral samples were reduced with flowing hydrogen at 1100°C in a microbalance furnace, with reaction progress monitored by weight loss. Experiments were run at at-

mospheric pressure for durations of 3–4 hr. Solid samples from both sets of experiments were analyzed by Mossbauer spectroscopy, petrographic microscopy, SEM, TEM, and XRD.

Mature lunar soil. Apollo 17 soil 78221 was examined for evidence of natural reduction in the lunar environment. This sample was chosen based on its high maturity level (I_h/FeO = 93.0). The FeO content is 11.68 wt% and the TiO₂ content is 3.84 wt% [4]. A polished thin section of the 90–150-μm size fraction was analyzed by petrographic microscopy and SEM. **END**

Results: Lunar and terrestrial basalt. Reduction of the lunar basalt samples was extremely rapid, with major evolution of water occurring within minutes after the introduction of hydrogen. Mossbauer analysis of the sample reduced at the lowest temperature, 900°C, showed complete reduction of the Fe²⁺ in ilmenite to iron metal.

All the reduced ilmenite examined by SEM displayed phase separation and reduction throughout. Samples reduced at 1000° and 1050°C showed evidence of iron migration to grain surfaces, while the single 900°C sample did not show such evidence.

One olivine crystal in the sample reduced at 1050°C displayed submicrometer iron blebs at the surface and along interior cracks, indicating partial reduction. Other olivine and pyroxene grains in the same sample showed no such evidence. No sign of olivine nor pyroxene reduction was found in samples run at 1000° or 900°C. High-resolution TEM photographs of ilmenite from the 1050°C experiment showed reduction of the TiO₂ to Ti₆O₁₁.

The reduction of MLS-1 basalt by hydrogen at 1100°C was rapid, with over 90% of the weight loss occurring in the first 10 min. The XRD data showed that the reduction of iron oxide in ilmenite and magnetite was essentially complete. The ilmenite displayed 1–20-μm blebs and stringers of iron metal in a titanium oxide matrix. Iron metal was also concentrated on ilmenite grain surfaces. Most ilmenite, even crystals located hundreds of micrometers from grain edges, showed this texture.

Pyroxene and olivine reduced in hydrogen and examined by SEM displayed evidence of very limited reduction. Submicrometer iron blebs were sparsely scattered over surfaces and along internal fractures. Analysis of crystal interiors demonstrated, however, that most of the iron in these minerals was not reduced.

Reduction of TiO₂ was also observed. After three hours at 1100°C the material was completely reduced to Ti₄O₇. The Ti₄O₇ was also a major phase, along with iron metal, in synthetic ilmenite reduced under the same conditions.

Mature lunar soil. Soil sample 78221 was searched for evidence of ilmenite and olivine reduction. The results were negative for the size range examined (90–150 μm). Isolated ilmenite and olivine crystals were texturally and chemically homogeneous to the resolution limit of the SEM. The same was true of crystals incorporated within agglutinates, even though the agglutinate glass contained submicrometer blebs of metallic iron. Thus, even in this very mature soil, oxygen has apparently not been released from ilmenite by natural processes.

Output from a Lunar Oxygen Plant: These first reduction tests with actual lunar material allow estimation of the output from an oxygen plant. The FeO in ilmenite was totally reduced in the 900°C test, and reduction appeared to be complete at higher temperatures also. The oxygen yield from this reduction is 10.5% of the ilmenite mass, or 20.0% of the mass of TiO₂ in this mineral.

The TiO₂ formed by ilmenite reduction, as well as that found as rutile in lunar basalts, can itself be reduced to Ti₆O₁₁ at 1050°C and to Ti₄O₇ at 1100°C. This process yields a maximum of 5.3% of the TiO₂ mass as oxygen.

Small degrees of FeO reduction occur in terrestrial olivine and pyroxene at 1100°C and in lunar olivine at 1050°C. No evidence of

reduction was detected at lower temperatures. We consider these contributions to the oxygen yield to be negligible.

The predicted output from a lunar oxygen plant thus depends on the ilmenite and TiO_2 contents of the soil. In most lunar soils almost all the TiO_2 is incorporated in ilmenite [1]. The maximum oxygen yield therefore will equal 20% of the TiO_2 content if only ilmenite is reduced, and 25% if further conversion to Ti_4O_7 occurs. Lunar soil 78221 contains 3.84 wt% TiO_2 . The maximum predicted oxygen output from a plant using this feedstock is just under 1% of the total input mass. The output from a high-Ti soil such as 75061, with 18.02 wt% FeO and 10.38 wt% TiO_2 [4], is 2.6%.

Concentration or beneficiation of ilmenite would increase the process yield, but not the overall yield. An output of 2.6% means that 38 tons of lunar soil would be required to produce one ton of oxygen. By terrestrial standards this is a small amount of feedstock. A single medium-sized dump truck can hold 40 tons and can be loaded in under 10 min with a front-end loader [6].

Acknowledgments: The reduction experiments were performed at Carbotek with Shimizu Corporation funding. Samples were analyzed at the NASA Johnson Space Center with NASA funding.

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NORTH MASSIF LITHOLOGIES AND CHEMICAL COMPOSITIONS VIEWED FROM 2–4-mm PARTICLES OF SOIL SAMPLE 76503. Kaylynn M. Bishop, Bradley L. Jolliff, Randy L. Korotev, and Larry A. Haskin, Department of Earth and Planetary Sciences and McDonnell Center for the Space Sciences, Washington University, St. Louis MO 63130, USA.

In this work, we identify the lithologic and compositional components of soil 76503 based on INAA of 243 2–4-mm particles and 72 thin sections from these and associated 1–2-mm particles (76502) [1]. We present a statistical distribution of the major compositional types as the first step of a detailed comparative study of the North and South Massifs. The soil sample was collected well away from any boulder and is more representative of typical North Massif material than any single large rock or boulder sample. So far, our examination of the 76503 particles has provided a better definition of precursor igneous lithologies and their petrogenetic relationships [2]. It has enabled us to refine the nature of mixing components for the North Massif <1-mm fines [3]. It has confirmed the differences in lithologies and their proportions between materials of the North and South Massifs; e.g., the North Massif is distinguished by the absence of a 72275-type KREEP component, the abundance of a highly magnesian igneous component, and the absence of certain types of melt compositions found in the South Massif samples.

Results: On the basis of chemical compositions and binocular microscope observations, sample 76503 comprises 30 wt% dark glassy-matrix breccias, regolith breccias, and agglutinates; 29% highland igneous fragments and granulitic breccias; 24% noritic melt breccias; 13% high-Ti mare basalt; 1.5% orange glass regolith breccias and vitrophyre, 0.4% (1 particle) VLT basaltic breccia, and 2% unclassified.

Impact melt lithologies (noritic breccias) are rich in incompatible trace elements (ITE) (Fig. 1) and include very fine-grained crystalline and poikilitic impact-melt breccias, glassy matrix breccias, and regolith breccias and agglutinates that include only impact melt breccia lithologies. The latter may have developed in the regolith higher on the North Massif or prior to the introduction of mare materials into the soil. On the basis of Sc, Cr, Sm, and Eu concentrations, noritic melt lithologies from 76503 and matrices from station 6 and 7 boulders differ significantly from those of stations 2 and 3, except boulder 2, station 2. Among particles from sample 76503, evidence of more than one melt group is lacking (Fig. 2). Most of the melt breccias are tightly clustered compositionally and fall within the field of North Massif melt breccia compositions defined by analyses from the literature (Fig. 2). Those melt breccias having compositions outside this field contain clasts of highland material having low concentrations of ITEs; thus their compositions are displaced toward those of highland igneous lithologies and granulitic breccias.

Highland lithologies that have low ITE concentrations include fragments of shocked and unshocked anorthositic troctolite, anorthositic norite, gabbroic anorthosite, and granulitic breccias of generally anorthositic-norite or anorthositic-gabbro compositions. Coarse single crystals or clumps of several crystals of plagioclase are common in the 2–4-mm range. These are compositionally very similar to plagioclase in 76535 troctolite [4]; however, we believe these, and perhaps 76535 also, are members of a more anorthositic body [2]. We find no igneous particles whose compositions suggest affinity to ferroan-anorthositic suite igneous rocks. Granulitic breccias are generally more pyroxene rich than the samples having igneous textures, and, although they have low ITE concentrations, many are substantially contaminated by meteoritic siderophile elements.

Observations and Implications: Below, we summarize some important features of the distribution of lithologies and compositions of particles in 76503 by comparison to the model distribution of components determined for station 6 <1-mm soil by [3]. Several of these features distinguish this soil from soils of the South Massif. (1) The mass-weighted average composition of the regolith breccias and agglutinates is very similar to the average composition of the station 6 <1-mm fines [3] (Fig. 2). (2) The proportions of components that have been used to model the station 6 soil [3] are similar to the proportions of groups we find in sample 76503 (i.e., the regolith breccias and agglutinates can be well accounted for as a mixture of observed mare basalt and orange glass fragments, noritic melt breccias, and ITE-poor highland lithologies).

The <1-mm fines can be modeled as 51% highlands [36% anorthositic norite and 14% MG component (norite/troctolite mix)], 21% noritic breccia, 21% mare basalt, and 6% orange glass, whereas the proportions of fragments in sample 76503 are 43% highlands, 34.5% noritic breccias, 19% mare basalt, 2% orange glass, and 0.6% VLT basalt (by mass on an agglutinate/breccia-free basis). (3) The proportion of noritic breccias in 76503 exceeds that determined as a mixing component in <1-mm fines by [3]; however, we have included in our particle count noritic breccias whose compositions are skewed toward ITE-poor highland compositions (see Fig 1). Therefore, a portion of the "MG" and "AN" highland components of [3] is taken up in our proportion of noritic breccias. This portion consists of mineral and lithic clasts that, on average, have a composition similar to magnesian granulite or magnesian anorthositic norite [2]. (4) More orange glass was found in the fines model than in the 2–4-mm particles because orange glass particles have a mean size of 40 μm [5] and so concentrate in the finer soil fractions. Particles with orange glass composition in sample 76503 were orange/black glass regolith breccias, not large, individual glass spheres. (5) Magnesian troctolitic

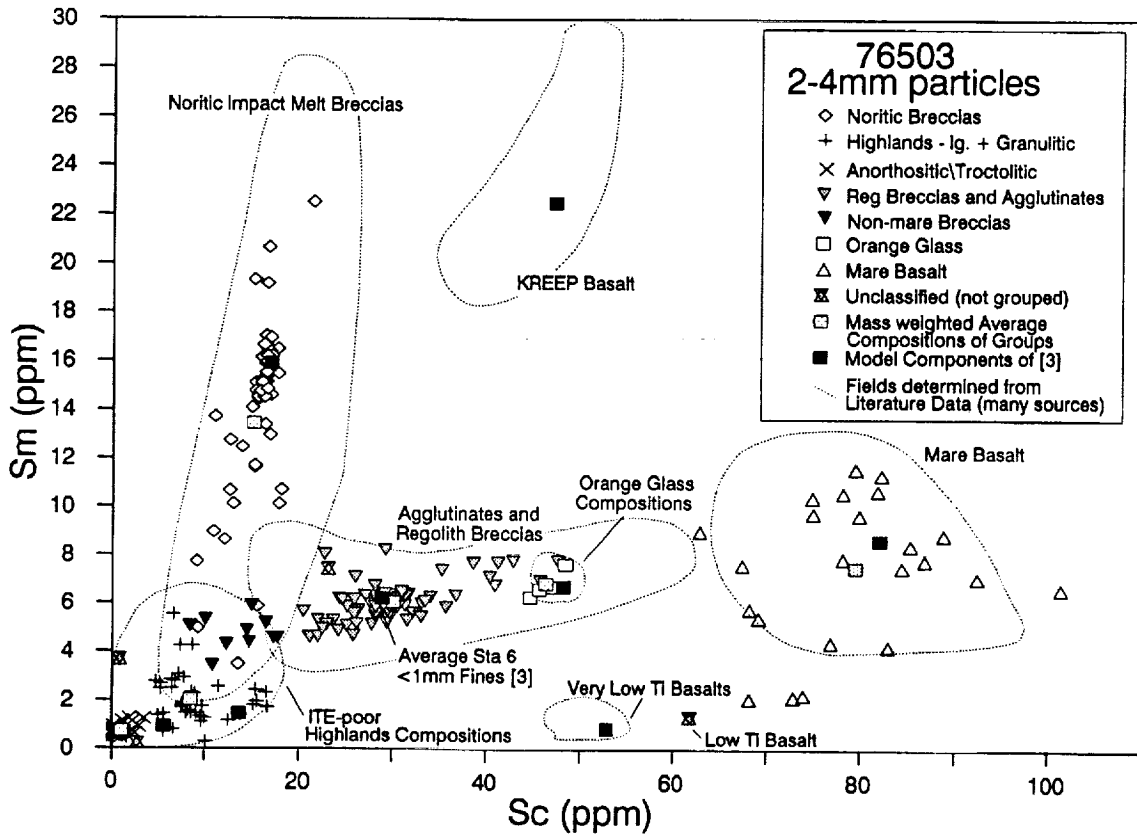


Fig. 1.

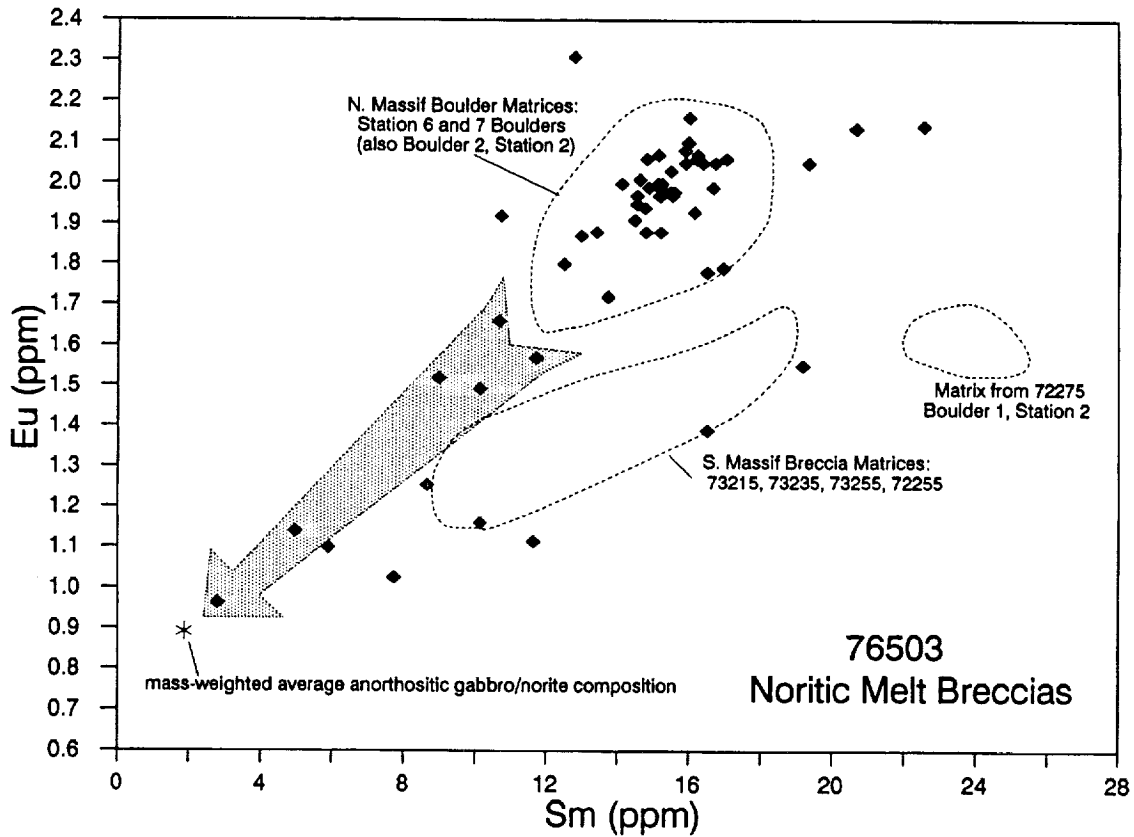


Fig. 2.

anorthosite appears to be the dominant lithology of the "MG" component and granulitic breccias, the dominant lithology of the "AN" component of [3]. The abundance of the Mg-rich component coupled with the absence of a KREEP component distinguish North Massif soils from South Massif soils.

Acknowledgments: Funding was through NASA grant NAG 9-56.

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REFINING THE GRANULITE SUITE. Janet A. Cushing, G. Jeffrey Taylor, Marc D. Norman, and Klaus Keil, Planetary Geosciences, Department of Geology and Geophysics, University of Hawaii at Manoa, 2525 Correa Rd., Honolulu HI 96822, USA.

Early studies of rocks retrieved from the Moon during the Apollo missions defined a group of rocks as granulites or "granulitic impactites" [1,2]. This included rocks with cataclastic, granulitic, and

poikilitic or poikiloblastic textures. Bickel and Warner [3] showed that the "granulites" have bulk compositions that fall into the two major pristine rock groups: the Mg-suite and ferroan anorthosites. Lindstrom and Lindstrom [4] further divided the granulites into four groups based on compositional distinction (Table 1). All these rocks have high contents of siderophile elements, indicating meteoritic contamination and indicating that impacts played a role in their origin. The conventional wisdom for the formation of the granulite suite involves post-"Apollonian" metamorphism of polymict breccias at near-solidus temperatures and low pressures, and for a relatively short period of time [2,5]. Nevertheless, some authors have drawn attention to the igneous appearance of some members of the granulite suite, such as 77017 and 67955 [6].

Petrographic studies indicate that the textures of "granulitic breccias" are significantly varied so as to redefine the granulitic suite into at least two distinct groups. The first group consists of rocks that have true granulitic textures: polygonal to rounded, equant grains that are annealed and have triple junctions with small dispersions from the average 120°. The second group of rocks have poikilitic or poikiloblastic textures, with subhedral to euhedral plagioclase and/or olivine grains enclosed in pyroxene oikocrysts. In some instances, the relationship between the minerals resembles an orthocumulate texture. The rocks

TABLE 1. Classification and data for the granulite suite.

Rock	Comp. Group [4]*	Texture	Equilibrated Minerals?	Mineral Comps.	Ref.	T(°C) (Kretz Ca)
60035	—	poik a	no	—		n/a
67215	sf	poik a	no	—		n/a
67415	sm	poik a	yes	n/a		n/a
67955	sm	poik a	yes	Fo ₇₆₋₈₀ En ₇₈ Wo _{3.1} En ₄₉ Wo ₄₂ An ₉₂₋₉₇	11	1097
76230	mm	poik a	yes	n/a		n/a
76235	mm	poik a	yes	n/a		n/a
77017	mf	poik a	yes	Fo ₆₁ En ₆₂ Wo _{8.5} En ₄₆ Wo ₃₇ An ₉₅	10,15	1165
72559	sm	poik b	yes	Fo ₈₁ En ₈₀ Wo _{3.7} En ₄₈ Wo ₄₄ An ₈₈₋₉₅	12	1031
78527	sm	poik b	yes	Fo ₇₇ En ₇₆ Wo _{4.1} En ₄₈ Wo ₄₂ An ₉₅	12	1089
15418	sf	gran	yes	Fo ₅₃ En ₆₅ An ₉₇	13	n/a
67915	—	gran	no	—		n/a
78155	mf	gran	yes	Fo ₆₂ En ₆₁ Wo ₉ En ₄₈ Wo ₃₀ An ₉₅	14	1247
79215	mm	gran	yes	Fo ₇₃ En ₇₅ Wo _{2.1} En ₄₇ Wo ₄₁ An ₉₃	1	1070

* sf: strongly ferroan; mf: moderately ferroan; sm: strongly magnesian; mm: moderately magnesian.

in this group range from (1) those with irregularly shaped pyroxene oikocrysts of variable size (generally 0.5–2 mm) to (2) those rocks in which the oikocrysts are subrounded, considerably smaller (<0.4 mm), and less pervasive than in the former. Also, the overall texture in type 2 rocks begins to take on a granulitic appearance. Both the granulitic and the poikilitic rocks have plagioclase crystals with small, round, mafic inclusions, though this is less common in the granulitic rocks. Pyroxene compositions taken from the literature were used to obtain equilibration temperatures. Both the Kretz-Ca [7] and Lindsley and Anderson graphical thermometer [8] methods yield similar temperatures, which range from ~1030° to 1240°C for the two major rock types, with a clustering around 1100°C (Table 1).

Rocks previously thought of as granulites may have formed in more than one way. Samples with a true granulitic texture appear to be metamorphosed polymict breccias, as many authors have argued. If the coarse-grained poikilitic samples are also metamorphic, their larger grain sizes suggest more intense metamorphism, but they have fewer clearly metamorphic triple junctions. In fact, in coarse-grained samples, such as 77017 and 67955, many plagioclase and olivine crystals are subhedral to euhedral. The poikilitic rocks might have formed from melts and then cooled at a range of rates after crystallization to account for observed textural variations among them [16]. The high siderophile elements and rare large, angular, plagioclase grains suggest that melting was more likely to have been caused by impact than endogenous igneous processes. If they actually formed by impact, the similar subsolidus histories of granulitic and poikilitic samples, as implied by two-pyroxene temperatures, suggest similar stratigraphic locations in pre-Imbrium crater deposits; perhaps the granulitic breccias were clasts in impact melts [9]. If grain growth in the granulitic rocks was due to solid-state coarsening (Ostwald ripening) controlled by diffusion at 1100°C at a rate equivalent to that in olivine, then the observed olivine and plagioclase grain sizes (excluding the poikilitic crystals) can be obtained in about 10⁴ yr. This is also a sufficient amount of time (at 1100°C) to equilibrate olivine and pyroxene.

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MARE VOLCANISM IN THE TAURUS-LITTROW REGION.
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Introduction: The products of mare volcanism at Taurus-Littrow occur in the form of crystalline basalts and volcanic glass beads. Both categories of sample define a compositionally diverse, but petro-

genetically unrelated, suite of magmas derived by partial melting of a heterogeneous, differentiated mantle beneath the region of the Apollo 17 landing site. This abstract is a brief review of what is known and what is not known about mare volcanism at this location on the Moon.

Mare Basalts: The Taurus-Littrow valley lies within a graben that is radially oriented to the Serenitatis Basin [1]. The valley is thought to contain a vertical sequence of mare volcanics up to about 1400 m in total thickness [1]. Although impact gardening has produced a regolith overlying these mare basalts that averages about 14 m thick in the Taurus-Littrow valley [1], individual cratering events may have excavated basalts from depths of about 100 m below the surface [1]. If correct, this would have excavated only about 7% of the estimated thickness of mare basalts. Consequently, about 90% of the basaltic units comprising this section were not sampled. Petrogeologic and petrologic constraints on mare volcanics from other areas of the Moon suggest that individual flows are usually less than 30 m thick [2–4]. If applicable to this site, it suggests that the Taurus-Littrow valley contains at least 40–50 individual flow units. Regardless of the correct number, sample analysis initially identified three compositionally [e.g., 5] and isotopically [6,7] distinct high-Ti lavas. This is similar to the number of distinct flow units expected from excavation to depths of 100 m within a sequence of lava flows having an average thickness of about 30 m. Recently, the number of compositionally identifiable flow units has been raised to four [8,9]. These high-Ti lavas are designated types A, B1, B2, and C. In addition, a fifth variety (type D) of high-Ti mare basalt has been reported [10,11] from a drive tube at Van Serg Crater. The ages of these individual flow units are as follows: type A = 3.75 ± 0.02 Ga [9], type B1/B2 = 3.69 ± 0.02 Ga [9], and type C = 3.72 ± 0.07 Ga [7,9]. No isotopic data presently exist on the type D basalt. Therefore, the duration of mare volcanism represented by these uppermost basaltic flows is probably less than, or about equal to, 120 Ma. Comparison of this extrusion rate of 100 m/120 Ma (i.e., about 1 m/Ma) with the time-averaged extrusion rate of mare lavas at Taurus-Littrow (i.e., 1400 m extruded between the time of graben formation by the Serenitatis impact at 3.87 Ga [e.g., 12–14] and the last flow at 3.69 Ga; 1400 m/180 Ma = 8 m/Ma) suggests that volcanic activity was higher during the early part of its volcanic history. Although the 1400-m thickness must include not only lava flows but also landslide deposits from the adjacent mountains, the general conclusion of an initially high volcanic activity seems inescapable.

In sequence of sampled abundance, type B1/B2 basalts are more abundant than type A basalts [8], and type A basalts are more abundant than the type C basalts [8], which have thus far been identified only at station 4. In addition to high-Ti lavas, very low-Ti (VLT) mare basalts also occur as fragments in the Apollo 17 drill core [15–17]. These VLT samples may be exotic pieces of mare basalt ballistically transported from elsewhere within Mare Serenitatis. No radiometric ages have yet been acquired on the Apollo 17 VLT basalts.

Chemical and isotopic data from types A, B1, and C are compatible with petrogenetic models involving closed system fractionation of observed phenocryst phases during emplacement of these magmas [9]. In contrast, type B2 basalts contain variable ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd initials [9] that require more complex processes, such as extensive interaction with, and assimilation of, the lunar crust [9]. The recent subdivision of the type B mare basalts into two distinct groups [8,9] lessens the need for appealing to differing percentages of partial melting in order to account for the range of trace element abundances and ratios noted by earlier investigators [e.g., 18,19].

All investigators are in agreement that the Apollo 17 mare basalts were produced by partial melting of differentiated source regions within the lunar mantle [e.g., 18–20]. These regions are generally regarded as being cumulate + intercumulate components [e.g., 21,22] produced by crystal/liquid fractionation of a global magma ocean during the first 200 Ma of the Moon's history. The original cumulate layers may have been stirred by solid-state convection subsequent to this global differentiation event [e.g., 23–25] generating a complex, heterogeneous mantle by the time that mare volcanism began. In addition, metasomatic redistribution of alkalis by halogen-rich fluids within the mantle source-regions of type A and C basalts at perhaps 4.0 Ga [9] has been suggested in order to account for (1) the enhanced Rb/Sr ratios but (2) apparently normal abundances and isotopic ratios among Sr and the rare earth elements.

The liquidus phase relations of type B2 (70215) and type C (74275) basalts have been experimentally determined [26–29]. If these compositions are assumed to be primary liquids (i.e., compositionally unmodified since leaving their mantle source regions), then they were derived from depths of about 120 km [26,28] and 250 km [27] respectively. If fractionation occurred during emplacement, then these are minimum depths for crystalline residues consisting of more than one mineral phase. However, the presence within 74275 of a dunite inclusion [30], which appears to be a xenolith [29–31], indicates that 74275 is not a magmatic composition [29,31]. Consequently, the significance of the experimental results on 74275 has been questioned [29].

Volcanic Glasses: Six varieties of volcanic glass have been identified and analyzed [e.g., 22,32–35]. These glasses are commonly thought to be the product of fire fountaining [e.g., 33,36,37]. In sequence of increasing TiO₂ abundance (from 0.66% to 9.3% by weight), they have the following designations: VLT, Green, Yellow, Orange I, 74220-type, and Orange II [35]. Since these samples are glass, they unambiguously represent liquid compositions. Since they have high Mg and Ni abundances (i.e., compatible elements that are readily removable by crystal/liquid fractionation) compared to the Apollo 17 mare basalts, these volcanic glasses may be samples of primary magma from the lunar mantle [27,35]. The liquidus phase relations of Apollo 17 VLT [38] and 74220 [27] glasses have been determined, and suggest that their source regions were located at depths of about 400 km, if the residuum consisted of more than one mineral phase. However, the physical requirements involved in the ascent of primary magma from 400 km appear daunting, and other, perhaps more realistic, interpretations of the experimental data have been explored [39,40]. Nevertheless, compositionally heterogeneous mantle source regions are generally agreed to be required [e.g., 27,35,39,40].

Volatiles associated with the 74220 glasses have been extensively studied [e.g., 41–54]. The source of these volatiles within the Moon remains largely unknown [e.g., 53,55]. Superb examples of sublimates occurring on the surfaces of the orange/black volcanic glasses have been discussed in detail [e.g., 33,37,43,51]. At present, there are 25 elements known to be concentrated on or near the surfaces of these volcanic glasses: B, C, N, F, Na, S, Cl, K, Cu, Zn, Ga, Ge, Se, Br, Ag, Cd, In, Sb, Te, I, Au, Hg, Tl, Pb, and Bi (refer to [55] for a complete listing of data sources). The importance of a hot plume of gas during the eruption in affecting the cooling rates of individual droplets during the fire fountaining has been experimentally constrained [e.g., 56]. The physical conditions that prevailed during the eruption of these glasses have been derived from sample investigations [33,36,37,43,51,54,56,57], photogeology [58,59], spectral reflectance [60–63], and models [64]. Although the principal gas phase is currently thought

to have been CO [e.g., 65], a halogen-rich vapor associated with CO was the most likely carrier phase for most of the elements listed earlier in this paragraph [e.g., 54,66].

Radiometric ages for the Apollo 17 volcanic glasses have been reported for the 74220 type only. Since the range of published ages for the 74220-type glass is beyond the stated uncertainties of the methods, the current data need to be evaluated critically: 3.54 ± 0.05 Ga by Ar-Ar [67], 3.66 ± 0.06 Ga by Ar-Ar [68], 3.66 ± 0.03 Ga by Ar-Ar [69], 3.60 ± 0.04 Ga by Ar-Ar [70], and 3.48 ± 0.03 Ga by Pb-Pb [71]. This published range of ages suggests that the 74220 glasses are younger than the high-Ti mare basalts at the Apollo 17 site.

Eruptive Stratigraphy: The eruptive stratigraphy among the sampled volcanic units is only partially resolvable using current constraints. Isotopic data suggest that the type B1/B2 basalts are younger than the type A basalts [9]. The radiometric ages reported for the 74220 volcanic orange glass, as well as photogeologic and field constraints [1], suggest that the eruption of the Apollo 17 orange/black volcanic glass was the last major volcanic event at Taurus-Littrow.

Questions: What caused the variations in ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd initials [9] among mare basalts belonging to type B2? Are multiple flow units represented, or was this an isotopically heterogeneous magma?

What is the origin of the dunite inclusion within type C mare basalt 74275?

Did volatile-driven metasomatism occur in the lunar mantle during the production of some mare volcanics [9,72], or is this process merely a model-dependent artifact? If required, how does this affect the commonly held notion of a volatile-poor Moon?

Did thermal erosion occur during emplacement of mare basalts? If so, how would this have affected their chemical and isotopic compositions?

Do the crystalline VLT mare basalts at Apollo 17 come from the higher-albedo mare basalts exposed on the interior of Mare Serenitatis? What are the crystallization ages of the Apollo 17 VLT basalts?

What are the eruption ages and isotopic systematics of the six picritic magmas represented by the volcanic glasses at Taurus-Littrow?

From what depths were the mare magmas derived?

What were the physical processes involved in the production of mare magmas? Which was more important, isobaric melting or polybaric melting?

Why do the mare basalts and the picritic glasses appear to have no direct petrogenetic relationship [73,74]?

What is the volcanic stratigraphy at Taurus-Littrow?

Why do the volcanic glasses span a wider range of compositions at the Apollo 17 site (and all other Apollo sites) than the crystalline mare basalts?

What allowed the picritic magmas (represented by volcanic glasses) to ascend from the lunar mantle without significant crystal/liquid fractionation, whereas the low-Mg magmas (represented by mare basalts) seem to have been significantly modified by near-surface differentiation?

What is the source and composition of the indigenous volatiles associated with mare volcanism?

How many more compositional varieties of mare basalt and picritic glass remain to be identified in the Apollo 17 sample collection?

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ABUNDANCES OF SODIUM, SULFUR, AND POTASSIUM IN LUNAR VOLCANIC GLASSES: EVIDENCE FOR VOLATILE LOSS DURING ERUPTION. J. W. Delano¹ and J. McGuire², ¹Department of Geological Sciences, State University of New York, Albany NY 12222, USA, ²Department of Earth and Space Sciences, State University of New York, Stony Brook NY 11794, USA.

Introduction: Six varieties of lunar volcanic glass are known to occur within the Apollo 17 sample collection [1]. Investigations have shown that 25 volatile elements (B, C, N, F, Na, S, Cl, K, Cu, Zn, Ga, Ge, Se, Br, Ag, Cd, In, Sb, Te, I, Au, Hg, Tl, Pb, Bi) are known to be concentrated on the exterior surfaces of individual volcanic glass spheres [e.g., 2-10]. Since bulk analyses (e.g., INAA, XRF) of volcanic glass provide an integrated abundance of an element on and within the glass spherules, other methods must be relied on to determine the interior abundance of an element. The interior abundance of an element within a volcanic glass sphere establishes the abundance of that element in the melt at the time of quench. The current study is part of a comprehensive attempt to measure the abundance of three volatile elements (Na, S, and K) within representative spheres of the 25 varieties of lunar volcanic glass currently known to exist at the Apollo landing sites. Comparison of the measured abundances of these elements within the interiors of individual glasses with (1) bulk analyses and (2) crystalline mare basalts will furnish new constraints on the geochemical behavior of volatile elements during lunar mare volcanism. **END**

Analytical Procedures: Polished thin sections of lunar regolith and regolith breccias that expose the interiors of volcanic glass spheres were analyzed by electron microprobe (JEOL 733 Superprobe). The acceleration potential was 15 kV and the cup current was 50.0 nanoamps. A spot size of 20 μm was typically used. Counting times per analysis consisted of 300 s on the peak position of each element, as well as 150 s on each of the two background positions. Under most circumstances, each lunar glass was analyzed two or three times in order to check for chemical homogeneity and to improve analytical precision. The nominal detection limits for Na, S, and K for these operating conditions were 50 ppm, 32 ppm, and 37 ppm respectively. Reproducibility of multiple analyses on individual volcanic glass spherules were observed to be within a factor of 2 of those values. In order to check for the possible mobility of these elements within the glass during the analysis, which is commonly observed in terrestrial glasses, a natural lunar glass from Apollo 17 containing 3410 ppm Na, 900 ppm S, and 1220 ppm K was repeatedly analyzed at 60-s intervals, at the same spot location, for a total integrated beam exposure time of 1380 s using a cup current of 50 nanoamps. Within the counting statistics, no mobility of Na, S, or K was observed. This convinced us that mobility of these elements is not a problem in these dry, picritic lunar glasses during electron microprobe analyses. This study also attempted to measure the abundances of F and Cl within lunar volcanic glasses. Although the nominal detection limits for F and Cl were 200 ppm and 70 ppm respectively, neither of these two elements were detected. The higher detection limit for F is due to interferences on both sides of the F K-alpha peak from Fe L-alpha and a second-order Mg K-beta.

Results: Twenty-one volcanic glass spheres from 74220, 128 have been analyzed for S. Eighteen of those same glasses were also analyzed for Na and K. Although these spheres have been previously reported to be chemically homogeneous [11-14] among the major elements (Si, Ti, Al, Cr, Fe, Mg, Ca), the abundances of Na, S, and K have been observed in this study to display ranges among spherules

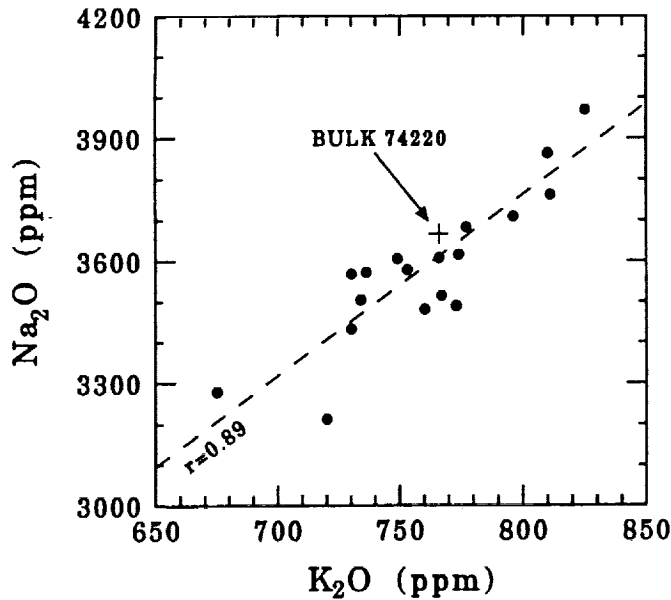


Fig. 1. Abundances of Na and K in 18 glass spherules (solid circles) from 74220,128. A 20–25% variation is observed among these glasses, whereas individual spherules appear to be homogeneous within analytical uncertainty. The bulk composition [15–20] of 74220 is shown by the + symbol. The linear least-squares fit to the data has a correlation coefficient of 0.89.

of 24%, 97%, and 22% respectively. Intrasample variations appear to be within analytical precision (i.e., individual spheres are homogeneous). For Na and K, Fig. 1 shows that the mean abundance of these elements among the 18 glasses are close to that of the bulk 74220 soil [15–20]. Figure 1 also shows that a strong correlation exists between Na and K in the glass spherules from 74220,128. In contrast, the abundance of S within the 74220,128 glasses is markedly different from the bulk S abundance of 820 ppm [21]. The interior abundances of S measured in 21 glasses from 74220,128 range from 148 ppm to 292 ppm with a mean value of 218 ppm. This mean value is in good agreement with a predicted value of about 187 ppm S made by Butler and Meyer [8]. The fact that the abundances of S within the glass are 2.8–5.5 times less than the measured abundance of S in the bulk 74220 underscores the importance of surface-correlated S [e.g., 8,21] on these volcanic glasses. With the exception of two 74220 spherules among the 18 spherules analyzed in this study, there is a good correlation between Na and S abundances. At the present time, we have no reason to doubt the veracity of the two aberrant points.

Our analyses (Fig. 2) show that the interior abundances of S in 74220, and other varieties of volcanic glass from Apollo 17 and other Apollo sites, are markedly lower than the abundances measured in the crystalline mare basalts. Although it has been previously noted that the bulk 74220 soil is depleted in S compared to the other high-Ti mare volcanics at Apollo 17 [e.g., 21], this depletion is even more marked when the interior abundances within the glasses are compared with the mare basalts. For example, the correlation between Ti and S noted by [22,23] would predict an abundance of 1600 ppm S in the 74220 magma. In contrast, the bulk soil abundance (820 ppm [21]) is about a factor of 2 less than that value, whereas the individual spherules have interior abundances of S with depletion factors ranging from 5.5 to 10.8. While we have no reason to doubt the correlation between Ti and S among the crystalline mare basalts [22,23], it is apparent from Fig. 2 that little, if any, correlation exists between these

two elements among the varieties of lunar volcanic glass analyzed in the current investigation.

Discussion: The 20–25% range in Na abundances observed in the present study among the 74220,128 glasses is nearly identical to the range reported by Hughes et al. [15] in their INAA data on individual 74220 spheres. If one assumes that the 74220 magma was chemically homogeneous prior to eruption, then the range of Na, S, and K abundances observed among individual spheres is indicative of fractional volatile loss from the droplets during fire fountaining and prior to quench. Each sphere would have had its own temperature/time history during the eruption that would be reflected in the fraction of volatile loss. Specifically, those spheres having the lowest Na, S, and K abundances would have remained hotter longer than those having higher Na, S, and K abundances. If this view is correct, then diffusion profiles, especially for S, might be predicted to occur within individual glass spherules (i.e., decreasing abundances from center to edge of spheres). Although we have made several, high-precision profiles on lunar volcanic glasses, no unambiguous diffusion gradients have yet been observed. Instead, complex profiles, such as those reported by [24] for the Apollo 15 volcanic green glasses, appear to be common. Work continues.

The range and highly correlated nature of Na and K among the 74220 glasses implies that Na and K, like S, were being fractionally lost during the volcanic eruption that produced this mist of fine melt droplets. It is important to note that both Na and K have been observed to be significantly concentrated on the surfaces of these glasses [6,8]. Consequently, our analyses provide additional evidence that fractional volatilization of Na and K occurred during the lunar mare volcanism associated with the origin of the 74220 glasses. Volatiliza-

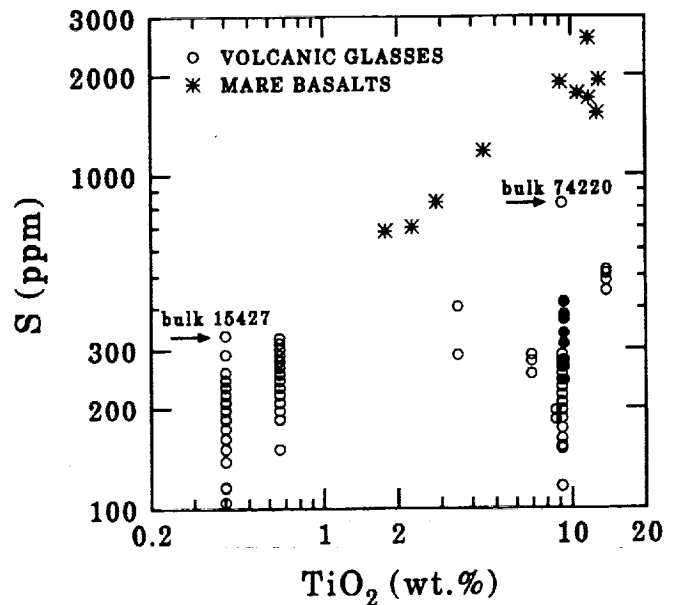


Fig. 2. Comparison of Ti and S abundances within lunar volcanic glasses (circles; this study) and crystalline mare basalts (stars). The abundances of S in bulk 15427 [21] and in bulk 74220 [21] are highlighted by arrows. The varieties of volcanic glass shown in this figure are the following (in sequence of increasing Ti abundance): Apollo 15 green, Apollo 17 VLT, Apollo 15 yellow, Apollo 17 yellow, Apollo 17 orange I, 74220 glasses, Apollo 17 orange II, and Apollo 15 red. Each circle represents an analysis of an individual glass sphere. Note that the volcanic glasses within a single magmatic suite display a large range in S abundances (see text for discussion).

tion of Na, S, and K during lunar volcanism was a subject of some interest during the early days of lunar sample analysis [e.g., 25,26].

The marked depletion of S in lunar volcanic glasses compared to the crystalline mare basalts has been attributed [21] to either (1) strong degassing of magmas during the fire fountaining that produced the volcanic glasses compared to the quiescent fissure eruptions that produced the mare basalts or (2) major differences in the S abundance within the mantle source regions. While our data strongly indicate that the former mechanism certainly contributed to this difference between mare basalts and volcanic glasses, additional constraints [27] suggest that the picritic magmas that produced the volcanic glasses were generated from different source regions than most of the mare basalts.

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RESOURCE AVAILABILITY AT TAURUS-LITTROW. Larry A. Haskin and R. O. Colson, Department of Earth and Planetary Sciences and McDonnell Center for the Space Sciences, Washington University, St. Louis MO 63130, USA.

What constitutes a lunar resource? Presumably, anything available on or from the Moon (e.g., [1]) is a resource. Our knowledge of the Moon's resources stands to be enhanced markedly through further exploration. A resource is not necessarily an ore, i.e., an economical source of a desired product. For commercial lunar extraction or manufacture, we require a resource, a need for the product, and reason to believe a lunar source is economical, i.e., that for some time period costs of mining, processing, fabrication, and delivery can be done more cheaply than from Earth or another extraterrestrial source.

During the next few decades, a lunar source is likely to be economical only for materials used in large quantities on the Moon or in near-Earth space, where the Moon's low gravity may offer an edge over Earth in transportation cost (e.g., [2]). Even so, lunar soils are potential ores for key products (fuel, metal, construction materials). Also, there are two plausible exceptions. One is souvenirs; the other is ^3He , plentiful on the Moon but not on Earth, proposed as fuel for commercial fusion reactors [3].

Long-duration experience in reduced, but nonzero, gravity is a "product" that may be cheaper to obtain on the Moon than elsewhere. This product, research, includes experience with physiological ef-

fects of long-term exposure to low gravity, problems of living, means of exploring, and coping with another environment. It includes experience in mining, extraction, and manufacturing. The Moon, because it is nearby, is probably the most economical extraterrestrial site for this research, especially if astronauts as well as automated equipment are involved. It is not clear that the experience in mining, extraction, production, handling, and storage on a planetary surface can be adequately simulated on Earth or in orbit. Mining and manufacturing can probably be done more cheaply with astronauts to assist in set-up, adjustment, maintenance, and repair of equipment rather than in wholly automated mode.

The complexity of carrying out even simple tasks on the Moon is a taxing overhead on proposed mining and manufacturing processes. The Moon lacks abundant and cheap air, water, and fossil fuels. It has no known common ores and no developed infrastructure. Differences between lunar and terrestrial conditions are so great that simple transplantation of terrestrial technologies is unworkable [2]. Economical use of lunar materials requires invention, demonstration, and development of methods appropriate to the Moon. We contend that lunar technologies for the next few decades must be simple and highly automated or teleoperated, and require minimal tending by astronauts. Processes requiring astronaut attention during nominal operation or for frequent repairs and maintenance will be regarded as labor intensive, and lunar labor will be expensive. Full automation is not obviously an inexpensive substitute, however; the successful Viking landers were a costly means of doing simple laboratory tasks inflexibly.

Early lunar technologies [2] will probably use a common lunar material as ore. They will be robust to minor fluctuations in feedstock composition and will not require appreciable feedstock beneficiation (see Beneficiation in references) such as rock grinding or mineral concentration. Technologies using unprocessed soil and indifferent to its composition will have the advantage. Nevertheless, the size and grade of the ore body must be confirmed for even the most indiscriminate process. Simple uses such as heaping unprocessed lunar soil for thermal insulation or radiation shielding onto a habitat require that we know the depth of the regolith, the size distributions of its soils, the locations of large boulders, and the ease of excavation. Costs of detailed site surveys trade against restrictions on site selection and conservative engineering design to accommodate unknown conditions of a poorly explored site.

Given the above considerations, we consider briefly some abundant lunar materials, their proposed uses, and technologies for their preparation, with particular attention to the Taurus-Littrow site. Relatively few papers are referenced in this text; references to two bibliographies [4,5] and to numerous, mainly more recent papers grouped according to subject area are compiled at the end of this abstract.

The Taurus-Littrow site has its fair share of the most common lunar resources, which are clinging soil; high vacuum; extreme temperatures ($\sim -170^\circ$ to 120°C); half-time sunlight; long, hot days ($\sim 14 \times 24$ hours); and long, cold nights [1]. With experience, we will come to view them as assets rather than just as obstacles. Harnessing the Sun's energy requires reflectors, pointable mirrors, photovoltaic and other devices for converting radiation to electrical energy, and heat engines. Vacuum is handy in casting molten materials. Proposed uses for unprocessed soil include thermal insulation, radiation shielding, and spacecraft ballast. Exactly how to use the soil, dig it up, and pile it on remain research topics.

Converting soil to structural material with adequate mechanical strength adds complexity. Proposed forms include bricks [6-8], cast basalt products (see Glass and Ceramics), glasses (see Glass and

Ceramics), cement, and concrete [9] (see Concrete). Sintering with solar ovens and with microwaves has been proposed [7,10,11]. Molten soil can be cast in the same manner that basalt is cast on Earth to produce pipes. Quenched rapidly, it will yield glass. Items drawn, spun, and cast from molten soil may become the main structural materials of near-Earth space. Glasses produced and used in a water-free vacuum will be somewhat stronger than their counterparts exposed to water [12–15]. Concrete may prove useful on the Moon; it need contain as little as ~1.5 wt% hydrogen. Cohesive lunar rock can serve as aggregate, but gravel appears to be scarce in the regolith [1] and is probably too valuable to use this way. Destructive distillation of lunar soil at very high temperature produces refractory materials that, on fine grinding, might make decent cement. Distillation, grinding, collection of aggregate, mixing, forming, and pouring make concrete use a multistep, probably labor-intensive process, however.

The Apollo 17 site has a good range of soil compositions [1,16], although it lacks the highly feldspathic soils typical of most lunar highlands. Major elemental constituents of lunar soils can be extracted chemically. At the Taurus-Littrow site, typical soil concentrations in weight percent are as follows: Si, 18–21; Ti, 0.8–6; Al, 3–5.5; Fe, 6.5–14; O, ~44 [1,16].

Oxygen production has received the most attention because O₂ makes up the bulk of H₂-O₂ propellant. Oxygen is also crucial to life support. Proposed extraction methods include reduction of ilmenite or glass by H₂, C, or CO (e.g., [17–22]; see Oxygen and Water), chemical dissolution of soil (e.g., [23–25]), electrolysis of molten soil, with or without added electrolyte (e.g., [26–29]), and destructive distillation of soil in a solar furnace (e.g., [31,32]). Gas reduction and electrolytic methods are understood best because the most experiments have been done on them. Early gas reduction studies centered on ilmenite, which is abundant (up to ~25 wt% in basalts) at the Taurus-Littrow site. Cumulus deposits of ilmenite may occur. Compositions of most Taurus-Littrow soils are also within the range (12–22% Al₂O₃) required by electrolytic methods.

Some less abundant materials are possible ores. Lunar volcanic ash has low overall lunar abundance, but is concentrated in relatively few locations (e.g., [33]). The orange soil ash from station 4 indicates that the Taurus-Littrow site may have useful deposits. A deposit would need to be proven by three-dimensional survey before a process using it was selected.

Abundant but dilute materials may be considered as ores if extraction of a product is simple. Solar-wind-implanted elements (SWIE) H, N, C, and noble gases can be extracted by heating soils to >700°C (e.g., [1,34–41]). Concentrations in mature Apollo soils (H and N, ~50 µg/g; C, ~100 µg/g; He, ~10 µg/g) are sufficient that a small operation by terrestrial mining standards could supply life-support needs of a substantial lunar base and enough H₂ for propellant for all operations in near-Earth space [34]. If ³He is ever mined for export to Earth, the other SWIE will be abundant byproducts; their potential future value demands that these materials be collected and stored if this can be done at low cost.

High concentrations of SWIE are favored by the presence of crystalline ilmenite (e.g., [42]), small grain size (e.g., [43]), and high soil maturity (average duration of exposure to solar wind). Unfortunately (in this context), soils of high maturity and small grain size tend to have low proportions of surviving minerals, including ilmenite. The soils collected from the Taurus-Littrow site that have the greatest maturity and highest SWIE are not the high-TiO₂ mare soils (e.g., [16,39,44–46]). Most sampled mare soils were collected near relatively fresh craters, however, and may be less mature than average Taurus-Littrow mare soils. Concentrations of SWIE and maturities at

the tops of cores and drive tubes are not indicative of conditions a few centimeters lower (e.g., [39,44–46]. Thus, ore certification is essential for efficient mining of SWIE. The Taurus-Littrow site has some mature soils with competitively high SWIE concentrations (i.e., >50 µg/g H).

The minor elements P and S may be present in high enough concentrations in KREEPY materials and basalts that those materials can be considered seriously as potential ores if economical methods for their extraction can be developed [1,47]. If so, the Taurus-Littrow site is favorable for S; typical soil concentrations are 0.03–0.1 wt% [1,16]. Higher concentrations, typically, <0.1 wt%, are found in high-Ti mare basalts. Ores of higher grade may occur, however, based on evidence for S mobilization on the Moon as well as theoretical reasons to expect it [48]. Small quantities of H₂S are released when SWIE elements are vaporized from soils [49]. The Apollo 17 orange soil is probably a poor ore despite its sulfide coatings; the bulk S concentration of the orange soil is only ~0.07 wt% [1]. Taurus-Littrow soils are 0.015–0.03 wt% P, several times lower than Fra Mauro soils [1]. Vapor mobilization may have produced P ores [48]. It may be a good economic gamble to search for concentrated ores of P and S rather than to use common rocks or soils.

There may be ores of less abundant elements, owing to the extensive chemical fractionations known to have occurred during formation and evolution of the Moon's crust. This is speculative, prediction of sites of deposition of such ores is speculative, and survival of specialized ores through meteoroid impacts is speculative, so we dare not rely on their presence until we evaluate this through further exploration and sampling. Lunar volcanic ash, and especially the orange glass at the Taurus-Littrow site, is sometimes suggested as a source for relatively volatile elements, as these coat the individual spherules and may be easy to extract. However, except for Zn, bulk concentrations of vapor-mobilized elements (VAPS [1]) are not especially high relative to soil concentrations at several landing sites.

The principal materials required for survival and basic human activity are present on the Moon, if one chooses to imagine oneself a lunar pioneer and chemist [50]. The common materials of construction and the general style of life will differ in interesting ways from what we are used to. The Taurus-Littrow site has adequate supplies of the most common lunar resources.

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A SPECTRAL SURVEY OF THE SERENITATIS BASIN REGION OF THE MOON. B. R. Hawke¹, C. A. Peterson¹, P. G. Lucey¹, D. T. Blewett¹, J. F. Bell III², and P. D. Spudis³, ¹Planetary Geosciences, SOEST, University of Hawaii, Honolulu HI 96822, USA, ²NASA Ames Research Center, Moffett Field CA 94035, USA, ³Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston TX 77058, USA.

Introduction: The returned lunar samples contain much petrological and compositional information for small, select portions of the surface of the Moon. Many of the highland samples returned from the Apollo landing sites are somehow related to lunar multiringed impact basins, either as primary ejecta or reworked local material. These highland samples cannot always be related to a basin directly, and it is extremely difficult to characterize the composition and lithology of basin ejecta based on a small number of returned samples. However, the Apollo orbital geochemistry instruments and Earth-based remote sensing techniques provide compositional data for large portions of the lunar surface, including basin ejecta deposits. These data provide a regional framework within which to interpret the significance of the returned samples as well as information concerning the stratigraphy of the lunar crust [1].

In recent years, we have utilized the Apollo orbital geochemistry datasets and Earth-based spectral reflectance data to investigate the composition of highland units associated with lunar multiring basins. These include Imbrium [2], Orientale [3,4], and Nectaris [1,5] Basins. We have also analyzed a large number of near-IR reflectance spectra and multispectral images in an attempt to answer a variety of questions concerning the Serenitatis Basin. These questions include (1) What is the composition of highland units in the region and how do these compositions vary as a function of position around and distance from Serenitatis? (2) What was the crustal stratigraphy of the Serenitatis preimpact target site? (3) How do the Apollo 17 samples relate to geologic units in the surrounding highlands? (4) What is the nature and origin of light plains deposits in the region? (5) Do cryptomare occur in the Serenitatis region? The purpose of this paper is to present the preliminary results of our analyses of spectral data obtained for the Serenitatis Basin region.

Method: Near-infrared reflectance spectra (0.6-2.5 μm) were obtained at the 2.24-m telescope of the Mauna Kea Observatory during a series of observing runs using the Planetary Geosciences indium antimonide spectrometer. The lunar standard area at the Apollo 16 landing site was frequently observed during the course of each evening, and these observations were used to monitor atmospheric extinction throughout each night. Extinction corrections were made using the methods described by McCord and Clark [6], and spectral analyses were made using the techniques described by McCord et al. [7]. The multispectral images used in this study were described by McCord et al. [8] and Bell and Hawke [9].

Results and Discussion: *Taurus-Littrow region.* Spectra were obtained for various portions of the North and South Massifs at the Apollo 17 site. Analysis indicates that they exhibit "1- μm " bands that are centered at 0.92 μm and have band strengths of ~6%. These characteristics indicate a feldspar-bearing mineral assemblage with a

mafic component dominated by low-Ca orthopyroxene. The areas for which these spectra were obtained contain abundant anorthositic norite.

Spectra were also collected for various areas of Littrow Crater. The dominant lithology is anorthositic norite. However, spectra for a small, fresh crater on the north rim of Littrow as well as the floor of Littrow have "1- μm " bands centered at slightly longer wavelengths (~0.94 μm) than the spectra obtained for the Apollo 17 massifs. The highland-rich soil in the floor of Littrow may contain minor amounts of dark mantle material of pyroclastic origin.

Both localized (LDMD) and regional dark mantle deposits (RDMD) of pyroclastic origin occur in the Taurus-Littrow region. The Taurus-Littrow RDMD occurs west of the Apollo 17 site, and the thickest portion covers an area of 4000 km² [10,11]. We have obtained near-IR spectra for several portions of this RDMD. The spectra exhibit very steep continuum slopes and broad, shallow 1.0- μm absorption bands. Band analysis suggests that olivine is the most abundant mafic mineral. Previous studies have demonstrated that the black spheres from the rim of Shorty Crater are the characteristic ingredients and dominate the spectra of the Taurus-Littrow RDMD [11-14]. These dark pyroclastic spheres, which contain abundant ilmenite and olivine, are the quench-crystallized chemical equivalents of the homogeneous Apollo 17 orange glasses. This RDMD exhibits very weak echoes on 3.8-cm radar images; a lack of surface scatterers (fragments 1-50 cm in size) is believed to be responsible for the low radar returns [11,12]. Previous studies of lunar RDMD have suggested that they would be excellent sites for lunar outposts and that the pyroclastic debris could be a source of several valuable resources [15].

The dark floor deposit in Vitruvius Crater was mapped as an LDMD by Wilhelms and McCauley [16]. The spectrum obtained for this deposit falls in LDMD group 2 as defined by Hawke et al. [10]. These spectra closely resemble those obtained for mature mare.

Plinius-Dawes region. The preliminary results of analyses of CCD imaging and spectroscopy data obtained for the Plinius-Dawes region was recently presented by Bell and Hawke [9]. Dawes (18 km) has a symmetrical band centered at 1.0 μm that is deeper than that of the surrounding mare, indicative of a lesser degree of soil maturity. In addition, Dawes has little or no highland signature in its six-color spectrum. In contrast, Plinius Crater (43 km) spectra exhibit more complex 1- μm bands. Spectra for the southwest wall and southwest floor have bands that appear rather symmetrical and are centered near 1.0 μm . The bands are not as deep as those exhibited by the Dawes spectra or those of surrounding mature mare surfaces. The 1- μm bands of the Plinius central peak and northeast wall spectra are clearly composed of two features: a short-wavelength component centered near 0.92 μm and a longer wavelength component centered near 1.0 μm . The 0.92- μm feature is weaker than the 1.0- μm band in the northeast wall spectrum, suggesting that this is primarily mare material with a minor highland component. However, the 0.92- μm band is much more prominent in the central peak spectrum, suggesting that the peak is primarily noritic highland material with minor mare contamination [9]. Our linear mixing model studies of the six-color CCD images verify and extend the spectral interpretations. The Plinius central peak and several regions of the floor clearly have a substantial highland signature, and small outliers of highland material can be seen in the northwest ejecta. In addition, several regions of the floor and south-southwest rim exhibit a spectral signature more like that of the low-Ti Serenitatis mare than the Tranquillitatis basalts.

Montes Haemus region. The Haemus Mountains form one segment of the mare-bounding ring of the Serenitatis Basin. Menelaus Crater straddles the mare-highlands border, and the interior and

exterior deposits are very heterogeneous in composition. The dominant highland material associated with the interior of Menelaus is noritic anorthosite [17]. However, more pyroxene-rich highland debris also occurs in the Menelaus area. No deposits of pure anorthosite have been identified in the Serenitatis Basin region. The most plagioclase-rich material in the region was exposed by Menelaus.

Small craters in the Haemus region generally expose anorthositic norites. This material was derived from relatively shallow depths and is probably dominated by Imbrium Basin ejecta. Previous studies have shown that Imbrium ejecta on the backslope west of the Haemus region is dominated by norite and anorthositic norite [2].

The Sulpicius Gallus RDMD covers ~6000 km² in the western portion of the Haemus region. Spectral studies indicate that this deposit is a mixture of orange glass and black spheres [11].

Terrain north and east of Serenitatis. Band analysis of the limited number of spectra available indicates that mare basalts are exposed in the walls of Burg Crater and that the Burg peak has a mafic assemblage dominated by high-Ca pyroxene. Anorthositic norites are exposed in the peak of Eudoxus Crater as well as the walls and peak of Romer. Dark-haloed impact craters south of Hercules and southeast of Posidonius have excavated mare basalts from beneath highland-rich surface units. At least small areas of cryptomare appear to exist northeast of Serenitatis Basin.

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THE SERENITATIS BASIN AND THE TAURUS-LITTROW HIGHLANDS: GEOLOGICAL CONTEXT AND HISTORY.

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Introduction: The Apollo 17 mission was targeted to land at the southeastern edge of the Serenitatis Basin, one of a number of large impact basins on the Moon. The choice of the landing site was influenced heavily by the desire to obtain detailed information about large impact basins [1] by constructing a composite "geological traverse" radial to a basin [2] (Fig. 1), with the Apollo 16 site (Descartes) representing the most distal regions, the Apollo 14 site (Fra Mauro) the intermediate textured ejecta unit, the Apollo 15 site (Hadley-Apennine) the basin topographic rim, and the Apollo 17 site (Taurus-Littrow) within the basin interior. The remarkable geologic exploration of the Taurus-Littrow Valley by astronauts Harrison Schmitt and Eugene Cernan provided fundamental information about processes associated with impact basin formation and evolution [3]. This information, further analysis of returned samples and other data, and subsequent exploration of the Moon, have raised additional questions that can potentially be answered from the site data. The Apollo 17 site is thus a keystone to the understanding of the geology of impact basins in general and basin interiors specifically. In this contribution, the geologic setting of the Apollo 17 site is reviewed, the

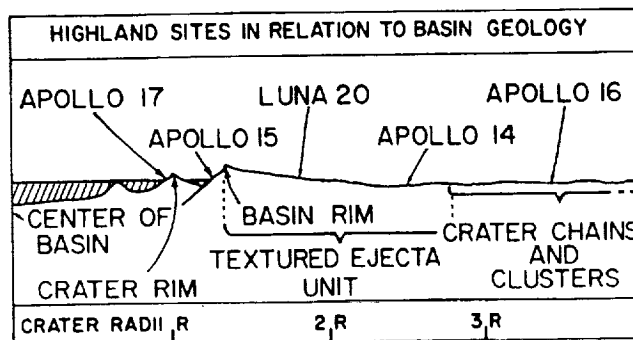


Fig. 1. Composite cross section across lunar impact basin showing relationship of Apollo landing sites to basin geology. From [2].

implications for the formation of basins from Apollo 17 results are assessed, and unanswered questions potentially addressable with existing and new data are outlined.

Geological Setting of the Serenitatis Basin: Mare Serenitatis is paradoxically one of the most clearly defined circular maria, but the basin structure itself is so indistinct, primarily due to modification by the post-Serenitatis Imbrium Basin, that some early basin studies did not even discuss it [4]. In fact, Serenitatis appears to be one of the more recent lunar basins. Stratigraphically, Serenitatis is one of 12 basins that have been assigned a Nectarian age [5]. It lies within the younger group of these basins and is immediately predated by Crisium, Humorum, and Humboldtianum, and immediately postdated by Hertzprung. In terms of its regional setting, several pre-Nectarian and Imbrian-aged basins have had a major influence on its history.

Pre-Serenitatis basin geology. Serenitatis lies within the ancient pre-Nectarian Procellarum Basin, which may be as large as 3200 km diameter [5,6]. The Serenitatis impact would have occurred astride the first ring, extending westward into the basin interior and eastward to the vicinity of the second ring. The Vitruvius Front [7], a topographic scarp occurring in the area to the east of the basin, may be related to the Procellarum Basin [5]. This structure, a long, irregular, but generally north-trending scarp, occasionally rises over 2 km about the surrounding terrain and is associated with a plateau to the east. It does not appear to be directly associated with the Serenitatis Basin [5-7] and may be a remnant of the second ring (excavation rim?) of the Procellarum Basin [5,6]. If so, the Apollo 17 site would lie just within the excavation cavity and Procellarum deposits there would be dominated by impact melts and deep ejecta emplaced on a substrate thinned considerably by impact excavation.

One of the most obvious basins forming in the vicinity of the target area prior to the Serenitatis event is the 800-km-diameter pre-Nectarian Tranquillitatis Basin. The outer ring of this structure actually intersects the Serenitatis Basin in the vicinity of the Apollo 17 site and thus early site geology may be influenced by geologic structure and deposits similar to those seen along the Cordillera ring of the Orientale Basin, or the Apennine ring of the Imbrium Basin at the Apollo 15 site. Crisium is interpreted to have just predated the Serenitatis event [5]. There is no clear evidence of Crisium secondaries near the eastern Serenitatis basin deposits and some Serenitatis secondary craters are interpreted to lie on Crisium ejecta [5]. In addition to these major basins, a number of impact craters in the sub-basin size range must have occurred subsequent to Procellarum and Tranquillitatis. Continuing documentation of the presence of cryptomaria in pre-Oriente times [8-11] suggests that many ancient basins may have been flooded by early mare basalts, or KREEP

basalts. Thus, the substrate in the pre-Serenitatis region must have been dominated by a wide variety of impact crater and basin deposits with the likelihood of some cryptomaria in the vicinity as well.

Formation of the Serenitatis Basin. The irregular outline of the Serenitatis Basin and the characteristics of the mascon gravity anomaly are suggestive of the possibility of two nearly superposed impact basins, with the larger one to the south [12-15]. An alternate possibility is a simultaneous impact as has been suggested for the Humboldtianum Basin [5,16]. In any case, the main (southern) basin is defined by the Montes Haemus and the Littrow massifs [15]. Although there has been considerable debate, many interpret this ring as the edge of the cavity of excavation on the basis of similarities of this ring to the Outer Rook Ring of the Orientale Basin [17]. The morphologic similarity of the Sculptured Hills east of Mare Serenitatis and the Montes Rook Formation between the Outer Rook ring and Cordillera ring have been cited as further evidence in support of this conclusion [13-15]. According to this reconstruction, the deposits of

the massifs in the vicinity of the Taurus-Littrow Valley should be composed of uplifted pre-Serenitatis upper crustal rocks overlain by Serenitatis ejecta, and modified by postbasin events. Because of the proximity of the site to the basin rim, impact melt rocks and associated breccias should be dominant.

Post-Serenitatis basin geology. The formation of the adjacent Imbrium Basin had a major effect on the Serenitatis Basin (Fig. 2), primarily due to intersection of the Imbrium ring with the northwest part of the basin and ejecta emplacement across the basin. The outer edge of the Fra Mauro Formation probably extends through central Serenitatis (Fig. 2). If the characteristics of Orientale Basin ejecta can be taken as a guide, then emplacement of ejecta from Imbrium surely heavily modified basin floor topography and caused ponding of the combined ejecta and locally excavated and eroded material in low-lying areas. Estimates of the amount of Imbrium ejecta at the Apollo 17 site based on simple assumptions of radial thickness decay [18] are in the range of 100 m, although the ballistic emplacement of ejecta at

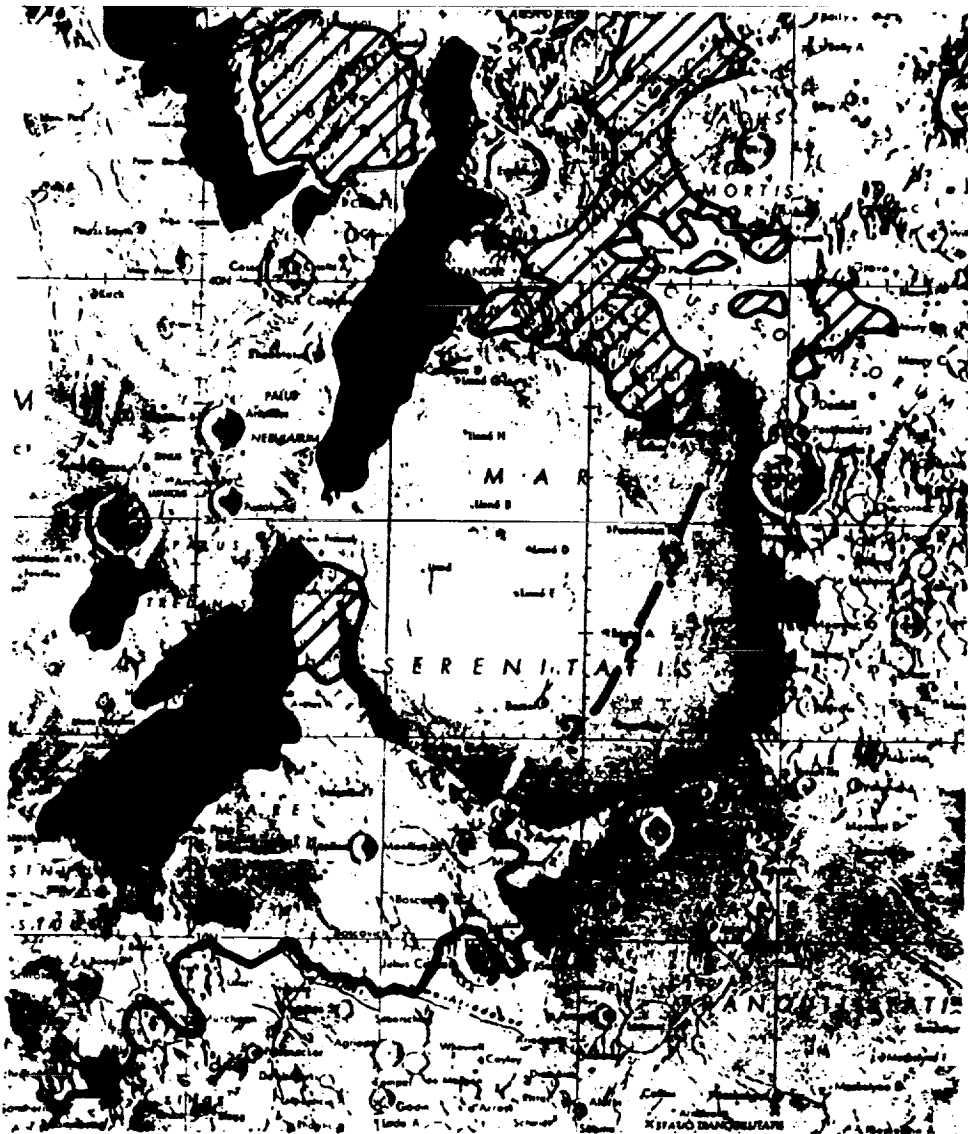


Fig. 2. Influence of the Imbrium Basin event on Serenitatis. Black pattern is Imbrium massifs and Apennine facies, patterned areas is Alpes Formation of Imbrium age. Solid black line represents outer edge of continuous Fra Mauro Formation, the textured ejecta deposit of Imbrium. This line is extended under Mare Serenitatis along an average range from the rim of the Imbrium Basin. From [15].

this range [19] would result in the fragmentation and mixing of Imbrium ejecta with local material. Thus, any Imbrium materials in the vicinity of the Apollo 17 site should have had a mode of emplacement similar to Imbrium ejecta at the Apollo 14 or 16 site. In addition to Imbrium, several post-Serenitatis sub-basin-sized craters are seen in highlands east of Serenitatis [15]. Although none directly intersect the landing site, their presence indicates that ejecta from their formation may be emplaced in the site area. One enigmatic aspect of several of these craters, such as Littrow, is that although they appear to postdate the Serenitatis event, their rims are characterized by textures similar to the Sculptured Hills. This would imply that the formation of the Sculptured Hills texture postdates the Serenitatis event, and thus casts into doubt the interpretation of the Sculptured Hills as a Montes Rook analogue formed simultaneously with Serenitatis [20]. On the other hand, it may be that the seismic effects of the Imbrium Basin event, which must have been significant [21], caused movement that modified both the Sculptured Hills and superposed crater deposits.

Geological Setting of the Apollo 17 Site: Apollo 17 surface exploration provided an excellent sampling of the different massifs and deposits in the Taurus Littrow Valley [22]. The sample materials consist of fragment-laden impact melts showing a diversity of textures (poikilitic, ophitic, subophitic, aphanitic) and a rather narrow range of chemical composition [23,24]. These characteristics have been generally interpreted to indicate that the majority of the rocks were derived from cooled melt rocks containing lithic clasts (primarily granulite breccias and Mg-suite cumulates) and representing Serenitatis ejecta and impact melt. Although the characteristics of the sample suite support this view, the outline of events in the Serenitatis region in addition to the Serenitatis event itself (e.g., Procellarum, Tranquillitatis, Imbrium, post-Serenitatis craters, etc.) all suggest that there may be other candidates for many of the materials sampled [20]. In addition, recent work on the formation of impact basins and new multispectral imaging data on basins [11,24] have provided a basis for the reinterpretation of the Apollo 17 site region. In December 1992 the Galileo spacecraft will obtain multispectral images of the north polar nearside region, including the Serenitatis Basin, and these data will provide an excellent framework for reanalysis of the Serenitatis Basin. Among the important questions to be addressed are (1) What is the role of Procellarum and Tranquillitatis Basin ring deposits at the Apollo 17 site? (2) What are the correct ring assignments for the Serenitatis Basin? (3) How many impact events are recorded in breccias? (4) How can we provide a better link of sample data to impact crater environments? (5) What is the true age of the Sculptured Hills and what represents them in the sample collection?

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N 98-918 7/2/99
 THE ANCIENT LUNAR CRUST, APOLLO 17 REGION.
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Introduction: The Apollo 17 highland collection is dominated by fragment-laden melt rocks, generally thought to represent impact melt from the Serenitatis basin-forming impact. Fortunately for our understanding of the lunar crust, the melt rocks contain unmelted clasts of preexisting rocks. Similar ancient rocks are also found in the regolith; most are probably clasts eroded out of melt rocks.

The ancient rocks can be divided into groups by age, composition, and history. Oldest are plutonic igneous rocks, representing the magmatic components of the ancient crust. Younger are granulitic breccias, which are thoroughly recrystallized rocks of diverse parent-ages. Youngest are KREEPY basalts and felsites, products of relatively evolved magmas. Some characteristics of each group are given below.

Plutonic Igneous Rocks: All large Apollo 17 samples of plutonic igneous rocks are members of the Mg-suite. They are troctolites, norites, one gabbronorite, and one dunite. Detailed investigations of smaller samples (breccia clasts, coarse fines, and rake samples) have identified additional samples of the first three of these rock types, plus several samples of ferroan anorthosites.

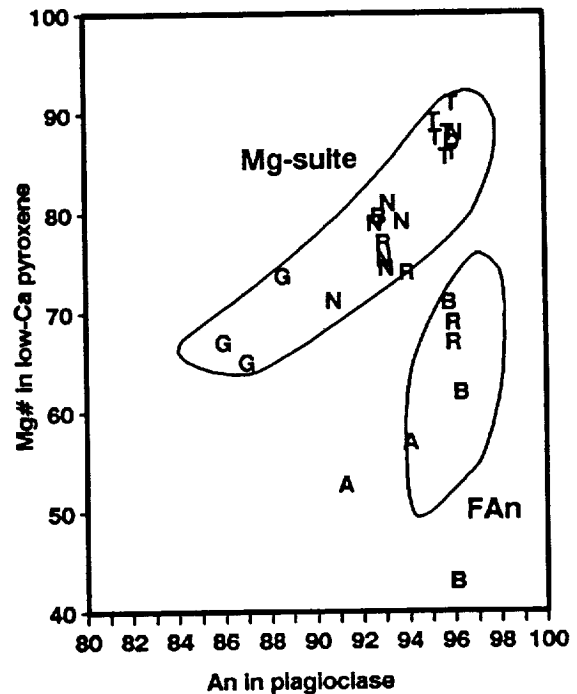


Fig. 1. Content of An in plagioclase vs. mg# in low-Ca pyroxene in some Apollo 17 ancient crustal rocks. Symbols are as follows: T, troctolite; D, dunite; N, norite; G, gabbronorite; A, anorthosite with relict igneous texture; B, anorthosite or norite with granulitic texture; R, granulitic breccia. Lines enclose the fields of compositional variation of minerals in all lunar Mg-suite rocks and in ferroan-anorthosite-suite rocks from sites other than Apollo 17.

Mg-suite rocks. The troctolites are mineralogically fairly simple. Most consist almost entirely of olivine and plagioclase, with only traces of other minerals (typically low-Ca pyroxene and chrome spinel) [1–8]. The largest troctolite, 76535 [1,2], also contains trace augite, whitlockite, apatite, baddeleyite, and K-feldspar. These rocks have high mg# ($100 \times \text{molar Mg}/[\text{Mg} + \text{Fe}]$) in their mafic minerals and high An ($100 \times \text{molar Ca}/[\text{Ca} + \text{Na} + \text{K}]$) in their plagioclases (Fig. 1). Mineral compositions are homogeneous in individual samples and fairly homogeneous in the group as a whole (Fig. 1). Most samples have cataclastic texture, but several have relict cumulate textures. Relict grains and monomineralic granulated patches are generally about 1–2 mm across. The coarse grain size and homogeneous mineral compositions suggest that these rocks formed deep within the crust; estimated depth of origin of 76535 is 10 to 30 km [2]. Estimates of crystallization age of 76535, from isotopic studies, range from 4.24 to 4.51 Ga [9–12].

The norites are mineralogically more diverse than the troctolites. Major minerals are plagioclase and low-Ca pyroxene, and most samples contain minor augite; many also have traces of a silica mineral, K-feldspar, phosphate minerals, chrome spinel, ilmenite, armalcolite, rutile, zircon, and baddeleyite [3,4,13–21]. Mineral compositions show a trend of increasing mg# in mafic minerals with increasing An in plagioclase (Fig. 1). In individual samples, mineral compositions are fairly homogeneous. All the norites have cataclastic texture, but some also have relict cumulate texture. Relict grain size is like that in the troctolites. One sample, 78527, has been granulated and recrystallized [14], but its mineralogy suggests that it is monomict. Depth of origin of the norites is difficult to determine quantitatively, but textures and mineral compositions indicate somewhat less reequilibration than in troctolite 76535, suggesting a shallower origin. Estimates of crystallization age, based on isotopic studies of several rocks, range from 4.33 to 4.43 Ga [22–26].

The gabbro norites (including one gabbro) contain more abundant augite relative to low-Ca pyroxene than the norites, and their assemblage of trace minerals is simpler [3,5,6,13,17]. Major minerals are plagioclase, low-Ca pyroxene, and augite; trace constituents are ilmenite, a silica mineral, K-feldspar, and chrome spinel. The pyroxenes average slightly more iron rich, and the plagioclases significantly more sodic, than in the norites (Fig. 1). The largest gabbro norite sample has cataclastic texture, but the smaller samples have relict igneous textures. Relict grain size is like that in troctolites and norites. As for the norites, quantitative determination of depth of origin is not possible, but textures and mineral compositions suggest a shallower origin than for troctolite 76535. Crystallization age of the only dated sample, 73255,27,45, is 4.23 Ga [26].

The dunite (72415-7) is mineralogically similar to troctolite 76535, except that olivine is by far the dominant mineral, K-feldspar and baddeleyite are absent, and a trace of armalcolite is present [1,27,28]. Compositions of the major minerals are similar to those in the troctolites (Fig. 1). The rock texture indicates a complex history [27]. The dunite was originally a cumulate, and vestiges of relict texture remain; zoning in the olivine has been interpreted as relict igneous variation [28]. After the dunite formed, it was subjected to multiple episodes of impact-induced shock and granulation, followed by recrystallization. The complex history of this rock obscures information on its depth of origin, but, if the olivine zoning is indeed primary, then the rock crystallized at a shallow depth [28]. Crystallization age, based on Rb-Sr studies, is 4.45 Ga [29].

Ferroan-anorthosite-suite rocks. The Apollo 17 anorthosites are fairly uniform modally but vary in compositions of their major minerals. Plagioclase makes up >93% by volume of most samples; low-Ca pyroxene, augite, and olivine are the most common minor

constituents [5,21,30–32]. Mg# in mafic minerals spans a broad range and is lower than in most Mg-suite rocks (Fig. 1). One sample (76504,18) has unusually sodic plagioclase [5], and another (72275,350) has unusually iron-rich low-Ca pyroxene [31] (Fig. 1). Minerals in most samples are somewhat less homogeneous than in Mg-suite rocks. Textures of the rocks are also diverse. Three appear to be granulated coarse-grained plutonic rocks with little contamination from other rock types (74114,5; 72464,17; 76504,18) [5,30,32]. The others have had more complex histories: Two have been strongly shocked, three have been thoroughly recrystallized, and one is granulated and mixed with the surrounding breccia. In some of the complex samples, there may be unrecognized contamination by Mg-suite rocks, or the rocks may be genomict. Relict grain size, where present, is like that in Mg-suite rocks. There is no information on depth of origin or crystallization age, but data on Sr isotopic ratios in anorthosites from other sites indicate that such rocks formed very early in lunar history.

Granulitic Breccias: Numerous samples of granulitic breccias have been found at the Apollo 17 site. They range from relatively large samples, found in the regolith, to small clasts in breccias and fines. Typically, plagioclase content is 70–80% by volume and mafic minerals make up most of the remainder; commonly low-Ca pyroxene, olivine, and augite are all present. Mineral compositions in some granulitic breccias are like those in Mg-suite rocks; in others, they are like those in ferroan anorthosites (Fig. 1). The minerals are generally very homogeneous. Textures of the granulitic breccias range from coarse-grained poikiloblastic (e.g., 77017 [33]) to fine-grained granoblastic (e.g., 79215 [34], 78155 [35]) and indicate thorough recrystallization and reequilibration. Studies of granulitic breccias have established that such rocks can form by recrystallization of varied parents, including polymict breccias [34–36], fragment-laden melt rocks [37,38], and granulated monomict igneous rocks [39,40]. Samples 79215 and 78155 clearly formed from polymict precursors [34,35]. The precursor of 77017 may have been a monomict cumulate [33]; whether or not this interpretation is correct, mineral compositions in 77017 and 78155 indicate that the dominant components were ferroan-anorthosite-suite rocks, with little or no contribution from Mg-suite rocks. Recrystallization, in 78155 and three clasts from 73215, took place at about 4.25 Ga [41,42].

KREEPy Basalt: KREEPy basalt occurs at the Apollo 17 site only as granulated clasts making up the matrix of boulder 1 at station 2. Major minerals are plagioclase and clinopyroxene (ranging from magnesian pigeonite to ferroaugite); minor constituents are chrome spinel, a silica mineral, olivine, ilmenite, phosphate minerals, K-feldspar, zircon, and Si-rich glass [15,43,44]. The absence of meteoritic contamination and xenocrysts, and the presence of subophitic-intersertal texture and a broad compositional variation in the pyroxene, establish that these basalts are extrusive igneous rocks, not impact melts [43–45]. Crystallization age is about 4 Ga [25]. Chemically and mineralogically, these basalts are intermediate between high-alumina mare basalts and Apollo 15 KREEP basalts [43]. Possible explanations for this intermediate character are (1) mixing of a mare-type and a KREEP magma, (2) contamination of a mare magma with KREEP, or (3) partial melting of a distinctive source region with intermediate characteristics. The last of these alternatives may be the most likely [43].

Felsites: Felsites (fine-grained igneous rocks of granitic composition) are found as small clasts, in small numbers, at Apollo 17. Most are clasts in breccias from boulder 1 at station 2, South Massif [46], and from station 3 [36,47]. Major minerals in the felsites are quartz and Ba-K feldspar; minor minerals are Ca-K-rich ternary plagioclase, Ca-Fe-rich pyroxene, Fe-rich olivine, ilmenite, and

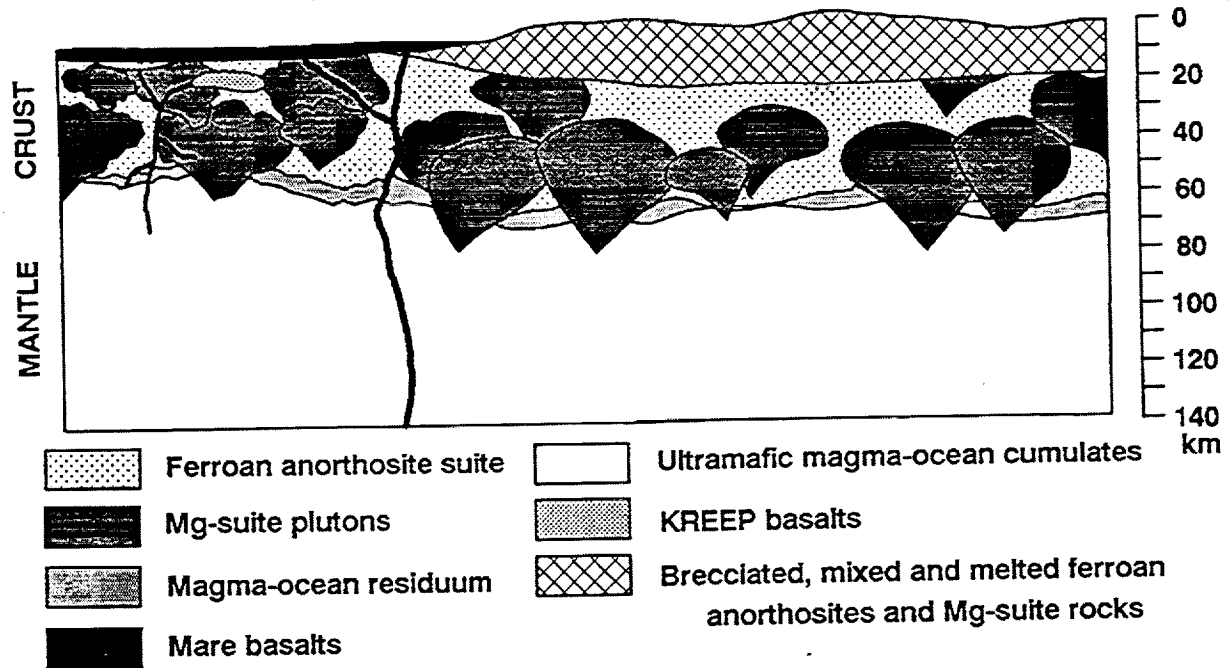


Fig. 2. Hypothetical cross section through the lunar crust and upper mantle.

phosphate minerals [36,46,47]. Many show relict igneous textures, most commonly micrographic or vermicular intergrowths of the quartz and feldspars. Many clasts have been partly melted. The relationship of felsites to other highland lithologies is not well understood. The evolved characteristics of these rocks suggest that their parent magmas could have formed by remelting within Mg-suite plutons, or by differentiation of KREEP basalt magmas; their relatively young crystallization age of about 4 Ga [48] indicates that they cannot be direct differentiates of Mg-suite parent magmas.

Origin and Evolution of the Lunar Crust: The results of lunar sample studies by numerous investigators have led to a hypothesis [49,50] for the evolution of the lunar crust and mantle. Figure 2 shows the structure of the outer 140 km of the Moon implied by this hypothesis, which is as follows. Differentiation of a primordial magma ocean several hundred kilometers deep formed a crust of ferroan anorthosite and related rocks (by plagioclase flotation) and an ultramafic mantle (by mafic-mineral sinking); the residual magma was rich in K, REE, and P and concentrated at or near the crust-mantle boundary. Partial melting near the end of magma-ocean crystallization, either in magma-ocean cumulates or in underlying undifferentiated mantle, produced mafic magmas that rose and intruded the primordial crust, forming large plutons of Mg-suite rocks. Impacts at the lunar surface granulated, mixed, and melted the crustal rocks, forming a near-surface layer of impact-modified rocks, including granulitic breccias. Later, more evolved magmas formed, perhaps by assimilation of the evolved parts of Mg-suite plutons by mare-basalt magmas, perhaps by remelting in Mg-suite plutons; these magmas formed KREEPy basalts and felsites. Locally, large impacts stripped away the top layers of the crust and formed multiring basins and fragment-laden impact melts. Subsequent melting in the magma-ocean cumulates produced mare-basalt magmas, which were extruded on the lunar surface and crystallized as mare basalts.

Ferroan anorthosites are rare at the Apollo 17 site, a fact noted many years ago. The most widely accepted explanation for the rarity of these rocks [4] is that the Serenitatis impact removed the upper part of the crust, which is dominated by ferroan anorthosite, from the

region, and the samples are mostly rocks from the lower crust, where Mg-suite plutons probably dominate (see Fig. 2). It is also possible [32] that the layer of anorthosite formed by the primordial differentiation was thinner in the Apollo 17 region than elsewhere on the Moon. These two explanations are not mutually exclusive; both could be true.

Comparison of the Apollo 17 ferroan anorthosites to ferroan anorthosites found elsewhere on the Moon can potentially yield information on the nature of the primordial differentiation. The Apollo 17 anorthosites are modally and mineralogically (Fig. 1) much like anorthosites from elsewhere on the Moon. Differences are (1) relatively high contents of incompatible elements [30-32], (2) an unusually broad range of mineral compositions (Fig. 1), and (3) common development of granulitic texture. The significance of these differences is as yet unclear.

One of the greatest uncertainties in interpreting the history of the lunar crust concerns the nature of the precursors of granulitic breccias. Samples of these rocks have been variously shown to represent recrystallized polymict breccias, impact-melt rocks, and monomict igneous rocks. Those whose precursors are monomict igneous rocks are potentially the most important. Most large samples of granulitic breccias that have been exhaustively studied have been shown to be polymict, but small regions in such rocks are clearly monomict; thus, small granulitic breccia clasts could indeed represent monomict igneous rocks. Detailed studies of granulitic breccias to identify monomict samples could contribute significantly to our understanding of lunar magmatic processes.

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NS 937/1879500 P. 2
THE APOLLO 17 REGION: A COMPOSITIONAL OVERVIEW. R. Jaumann and G. Neukum, DLR, Institute for Planetary Exploration, Berlin/Oberpfaffenhofen, Germany.

— Apollo 17 is located at a mare/highland boundary where the surface shows significant compositional heterogeneities. The composition of surface materials is estimated by analyzing their spectral/chemical correlations. Based on this spectral/chemical analysis the chemical and normative mineralogical composition of two highland units and three mare units has been estimated.

Introduction: The purpose of this investigation is to use spectral/chemical studies in order to determine the composition of Apollo 17 geologic units. The landing site itself and its surroundings are expected to expose different geologic settings ranging from basaltic materials over highland components to pyroclastic deposits [1–3]. Thus, Apollo 17 is an ideal place to study the mechanisms of mare volcanism, interaction of mare and highland materials, and pyroclastic activities. However, the understanding of such processes requires detailed knowledge about the chemical and mineralogical composition of the geologic units studied. Although the composition of the Apollo 17 landing site is known from the analysis of returned samples,

TABLE 1. Chemical composition of the materials exposed at the local surface points (Fig. 1).

Location	FeO	TiO ₂	Al ₂ O ₃	MgO
Ap17	16.8	7.9	12.9	11.0
Ap17 (L)	16.8	8.3	11.8	10.1
T5	19.8	8.9	9.8	8.8
S2	17.1	4.8	10.6	10.7
S3	17.1	7.4	11.7	11.0
S4	16.7	6.3	11.5	11.5
V2	11.2	1.0	20.6	6.3
L1	14.1	2.1	15.0	9.6
L2	14.0	2.2	14.2	11.2
σ	<1.7	<2.0	<2.5	<2.0

Ap17 (L) summarizes the results of the chemical analysis of Apollo 17 samples. All other data are derived from the spectral/chemical analysis of remote sensed measurements. The dimension is wt% and the error (σ) indicates an upper limit for accuracy of the analysis.

only little compositional information can be provided from direct sample studies for the surrounding areas of this complex geologic structure. So far, data of the Apollo 17 Geochemical Orbiter Experiments provide information on the concentration of Fe, Ti, and radioactive elements as well as on Al/Si and Mg/Si concentration ratios [4,5]. This information, however, is restricted to low ground resolutions, which are not sufficient for detailed mineralogical analyses. On the other hand, when we consider spectroscopic measurements of the Apollo 17 area and combine these measurements with the spectral/chemical correlations of Apollo 17 samples, it is feasible to establish a compositional analysis of the Apollo 17 area that is calibrated to the spectral/chemical evidence of the Apollo 17 landing site [6].

Compositional Studies: Compositional information can be derived from an analysis of the spectral/chemical correlations of lunar surface materials. For this purpose the spectral variations of lunar samples are identified and compared with the variations of the concentration of chemical constituents [6]. Such a sample-based analysis provides parameters that can be used to interpret the spectral variations in remotely sensed spectroscopic data. By applying the technique to Apollo 17 soil samples, the concentration of some chemical key elements like Fe, Ti, Al, and Mg can be estimated from spectral measurements. When the technique is transported to remotely

TABLE 2. Normative mineralogical composition of the materials exposed at the local surface points (Fig. 1).

Location	OI	Px	Pl	Il
Ap17	13.8	34.1	36.5	15.6
T5	3.8	49.6	27.3	19.2
S2	8.4	50.1	30.0	11.4
S3	10.0	41.5	33.3	15.2
S4	11.2	43.1	32.5	13.2
V2	15.2	19.0	61.3	4.4
L1	13.6	35.4	44.3	6.6
L2	15.9	37.3	41.3	5.6
σ	<1.5	<2.0	<1.5	<2.0

The dimension is wt% and the error (σ) indicates an upper limit for the accuracy of the analysis. OI = olivine, Px = pyroxene, Pl = plagioclase, Il = ilmenite.

sensed spectral data the accuracy is within 10% compared with the result of the chemical analysis of Apollo 17 samples (Table 1) or the results of the Apollo 17 Geochemical Orbiter Experiments [6]. Normative mineralogical information (Table 2) can be derived by comparing the calculated chemical concentrations with the concentration of Apollo 17 normative minerals (mineral phases of samples 74255 and 72395 [7]) [8]. Based on this approach the composition of the Apollo 17 surrounding areas can be determined relative to the ground-truth of the Apollo 17 landing site. The database consists of spectral/chemical parameters derived from the analysis of Apollo 17 soil samples, which transform spectral measurements into compositionally interpretable information, and of telescopic spectra obtained in the visible and infrared wavelength range [6,8]. The visible to near-infrared spectra (0.35–1.0 μm) are taken by using a CCD camera covering a regionally extended area, whereas the infrared spectra (0.7–2.5 μm) are taken at local surface points (Fig. 1) by using a CVF spectrometer. The ground resolution of the spectral measurements is 1–2 km. When the characteristics of spectra taken at the local surface points (Fig. 1) are used as end members, the distribution of compositional units can be estimated from the areally extended CCD spectral measurements.

Geological Interpretation: The geological interpretation of surface units is based on the chemical (Table 1) and mineralogical (Table 2) results, as well as on photogeological studies. Due to this interpretation, five geological units are identified (Fig. 1).

(1) Anorthositic highland materials (a) constitute an Al-rich unit that has an Fe/Al ratio < 0.6 and is characterized by its high content of plagioclase. The anorthositic highland materials are defined spectrally and chemically at the surface point V2 (Fig. 1). The unit is distributed in the form of small patches north and south of Apollo 17. Photogeological studies [1] indicate a pre-Nectarian age. (2) Mafic highland materials (n) are relatively Al-rich, but with an Fe/Al ratio ≈ 1 and a significant pyroxene component indicating the mafic characteristics of the unit. In terms of highland rocks these materials are comparable with gabbros (norites). The spectral/chemical reference for this unit is defined at the surface points L1 and L2 (Fig. 1).

The mafic materials dominate the highlands surrounding Apollo 17. The age is assumed to be pre-Nectarian to Nectarian. The mare units of the Apollo 17 region are attributed to the Imbrian system [1,2]. In general, basalts are classified by their titanium content. Based on the TiO_2 concentrations normal mare basalts ($\text{TiO}_2 < 5 \text{ wt}\%$) can be distinguished from titanium basalts ($\text{TiO}_2 > 5 \text{ wt}\%$). (3) The spectral/chemical characteristics of a typical mare basalt (mb) as exposed at the surface point S2 (Fig. 1) can be compared with the characteristics of a pigeonite basalt defined by a relatively enriched pyroxene component. Mare basalts dominate the southwestern part of the mare/highland boundary between Mare Serenitatis and the Taurus Littrow mountains and also fill the valleys west of Apollo 17. (4) Titanium basalts (tb) as exposed at the surface points Ap17 and T5 (Fig. 1) are characterized by high TiO_2 concentrations and thus by a high ilmenite content. Titanium basalts are mainly exposed at Apollo 17 and south of the landing site toward Mare Tranquillitatis. (5) The materials exposed on the Mare Serenitatis shelf area west to the Apollo 17 landing site (S3 and S4 in Fig. 1) exhibit a spectral/chemical characteristic similar to that of the titanium basalts. However, this unit shows a fine texture, is smoother than the titanium basalt unit, and is enriched in the agglutinate and glass component [3]. The photogeologic interpretation indicates that this unit is mainly composed of pyroclastic materials (dm). The present study shows that the investigation of spectral properties is a useful tool to interpret the composition of lunar surface materials.

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N93-187930/ P-4
NEOKREEP: A NEW LUNAR COMPONENT AT APOLLO 17.
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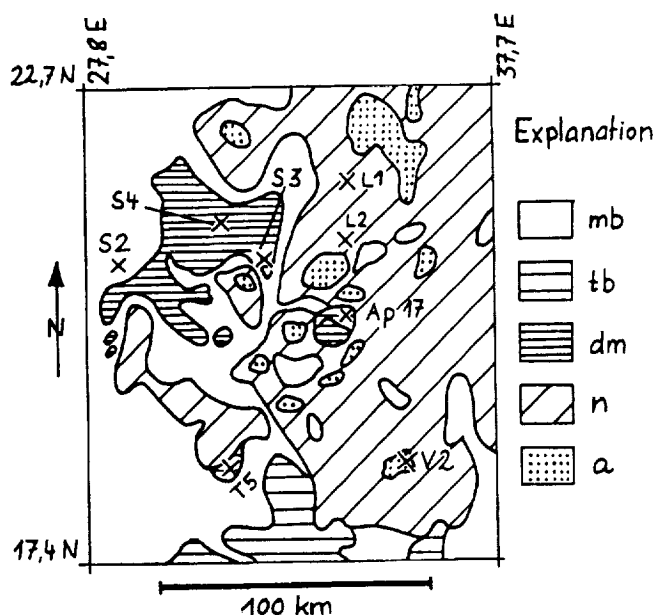


Fig. 1. Geologic map of the Apollo 17 area. (X) indicates the position of local surface points. For detailed description of the units see text.

Introduction: The Apollo 11 (Mare Tranquillitatis) and Apollo 17 (Mare Serenitatis) landing sites are important as the only sources of high-Ti basalt visited by the Apollo missions. The lunar high-Ti basalts ($> 6\% \text{ TiO}_2$; [1]) have no volumetrically comparable analogues among terrestrial basalts and require the presence of ilmenite in the source region, probably representing cumulates produced late in the crystallization of the lunar magma ocean [2].

Six principal groups of high-Ti basalts have been described, three from each of the two sites (see the review by [1]). The three groups of high-Ti basalts at the Apollo 17 site (termed "types") are all low-K ($< 2000 \text{ ppm K}$). These types are A, B, and C, with the type B basalts recently subdivided into varieties 1 and 2 (Fig. 1). It was shown by Neal et al. [3] that the type A basalts and both type B varieties can be generated through fractionation of observed phenocryst phases from a single magma. An evaluation of isotopic data [4] indicates an age of $3.75 \pm 0.02 \text{ Ga}$ for the type A basalts and $3.69 \pm 0.02 \text{ Ga}$ for the type B1/B2 basalts. The scarce type C basalts (only six examples) from Apollo 17 are more primitive than types A and B, having elevated MgO and Cr contents. However, the REE and alkali element (e.g., Rb) abundances in type C basalts are elevated relative to the type B basalts [5]. Neal et al. [5] suggest that the parent to the type C basalts was metasomatized by an alkali-rich fluid. Only two type C samples have been dated, yielding ages ranging from 3.63 ± 0.14 to $3.75 \pm 0.08 \text{ Ga}$.

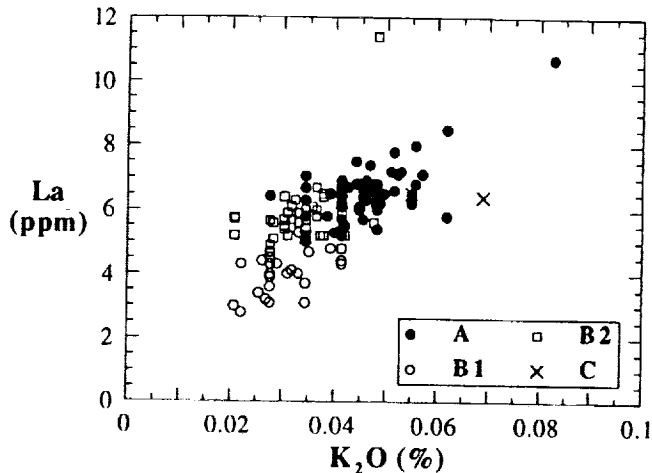


Fig. 1. Apollo 17 high-Ti basalts.

Clearly these cannot be accurately placed into the stratigraphy represented by the types A and B basalts. A single example of another type of basalt was described by Ryder [6], who classified it as type D. This basalt has lower REE than type B basalts, but is richer in Mg and Cr and poorer in Ca and Ti.

Three groups of basalts also exist at the Apollo 11 site, divided into groups A, B, and D, with the low-K group B basalts subdivided into varieties 1, 2, and 3 (Fig. 2). None of the Apollo 11 basalt groups correspond to the types present at Apollo 17. Group B3 basalts are the most primitive at the Apollo 11 site and form a continuum of compositions with the B1 basalts (Fig. 2). The ages for these basalts, 3.71 ± 0.02 Ga and 3.67 ± 0.02 Ga for B3 and B1, respectively, can be considered coeval. Group B2 basalts are also low-K basalts, but are enriched in many incompatible trace elements, particularly the rare earth elements. The basalts of group B2 are the oldest at Apollo 11, with an age of 3.84 ± 0.02 Ga. Only three group D basalts (also low-K) have been described. Recent studies of the high-Ti basalts of Apollo 11 (e.g., [7-9]) have shown that all the basalts of groups B and D could have been generated through periodic melting of a single source region coupled with some fractional crystallization of the resulting magmas.

Apollo 11 Group A Basalts: An Anomalous High-Ti Group: The group A basalts of Apollo 11 differ in many respects from other

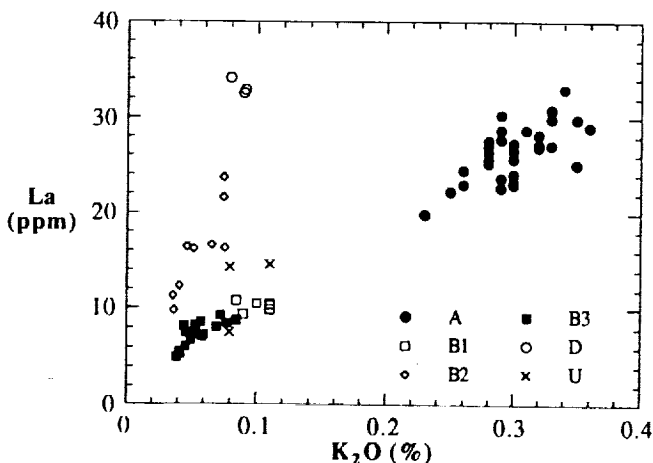


Fig. 2. Apollo 11 high-Ti basalts.

TABLE 1. Fractionation calculations.

	A17 OG	30% xtallzd	avg A11-A
SiO ₂	38.6	40.6	40.7
TiO ₂	8.8	11.0	11.0
Al ₂ O ₃	6.3	8.15	8.20
Cr ₂ O ₃	0.75	0.12	0.33
FeO	22.0	21.7	20.2
MnO	0.27	0.27	0.23
MgO	14.4	7.06	8.01
CaO	7.7	10.4	10.5
Na ₂ O	0.36	0.50	0.52
K ₂ O	0.09	0.12	0.30
P ₂ O ₅	0.06	0.08	0.19
Total	99.33	100.00	100.18
La	6.2	8.8	26.7
Ce	19.0	26.9	80.1
Nd	17.8	25.2	65.6
Sm	6.5	9.2	21.0
Eu	1.8	2.55	2.26
Gd	8.5	12.1	(26.8)
Tb	1.65	2.34	4.63
Dy	9.4	13.3	30.9
Er	5.1	7.2	18.9
Yb	4.43	6.13	17.0
Lu	0.61	0.86	2.46

Oxides are in %, REE in ppm.

high-Ti basalts of the region. Chemically, they are the only high-K (>2000 ppm K; [1]) variety of high-Ti basalt and are enriched in incompatible trace elements relative to other basalts from both the Apollo 11 and Apollo 17 sites (Fig. 2). In addition, group A basalts are the youngest of all high-Ti basalts, with an age of 3.56 ± 0.02 Ga [8]. The cluster of compositions is consistent with the Apollo 11 group A basalts representing a single flow (e.g., [7,10]). Papanastassiou et al. [11] have also indicated, based particularly on relatively young Rb-Sr model ages (3.8-3.9 Ga), the uniqueness of these basalts.

A model for the formation of the group A basalts was presented by Jerde et al. [7], wherein the Apollo 17 orange volcanic glass is the parent liquid. Fractionation of this composition, coupled with the assimilation of some incompatible-element-rich material, gave compositions akin to those of the Apollo 11 group A basalt population. It was postulated that this liquid was erupted near the Apollo 17 site and flowed into the Tranquillitatis Basin to the Apollo 11 site. Photographic and spectral evidence exists for a single flow unit spanning the distance between the two sites [12,13].

Evidence of KREEPy Contamination: Incompatible trace elements in group A basalts were modeled by Jerde et al. [7] using bulk partition coefficients calculated from the mineral assemblages indicated by fractionation calculations. It is evident from the results (Table 1) that the incompatible-element abundances in group A basalts are much too high (e.g., La = 26.5 ppm) to be explained by simple Rayleigh fractionation of Apollo 17 orange volcanic glass, which only leads to La = 8.8 ppm after 30% fractionation. The fact that K, P, and the REE are low in the calculated liquids suggests involvement of a KREEP component to obtain the basalts observed.

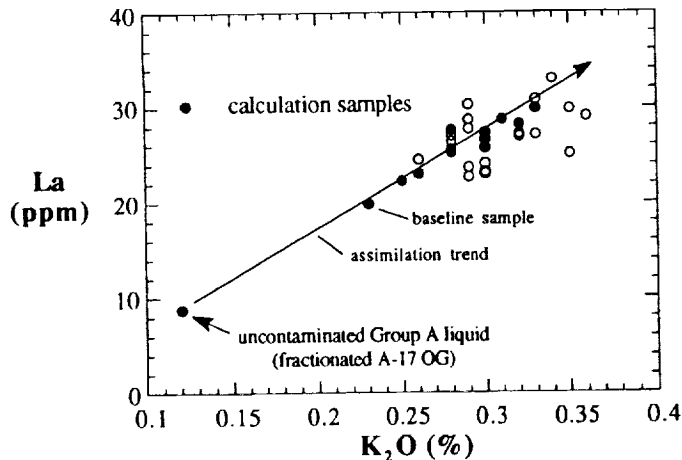


Fig. 3. Apollo 11 group A basalts.

In order to match the trace elements, ~20% of this component (the high-K KREEP of Warren [14]) must be added, which would imply that the Apollo 11 group A basalts are a young type of KREEP basalt. However, while this works chemically, isotopic considerations preclude it.

Addition of 20% of an old KREEP component (~4.4 Ga with $\epsilon_{Nd} = -3$ at 3.56 Ga) would overwhelm the LILE in the magma, yielding basalts with ϵ_{Nd} values <0 . However, ϵ_{Nd} values of the group A basalts of Apollo 11 fall in the range of +3 to +4. Therefore, any KREEPY assimilant could not have had an ϵ_{Nd} much less than +2 to +3. This KREEPY component must have been generated after complete crystallization of the LMO and must represent the late-stage crystallization products of a magma melted from the depleted "adcumulate" mantle. Assuming that this melt, indeed, was originally derived from the depleted cumulate mantle and that it crystallized with a $^{147}\text{Sm}/^{144}\text{Nd}$ greater than, but similar to, KREEP (=0.168), an age of 4.15 Ga can be calculated. This calculated model age is similar to the measured age (4.08 ± 0.07 Ga) of a KREEP basalt from Apollo 17 [15]. Furthermore, Shih et al. [15] suggested that other KREEPY rocks (including basalts, granites, troctolites, and norites) from various landing sites may also be cogenetic with this Apollo 17 KREEP basalt. This could further indicate that a Moon-wide melting event occurred at this time, which fused only the most evolved rocks. This event would have occurred after crystallization of the lunar magma ocean but prior to initiation of high-Ti basalt volcanism. This post-LMO KREEPY

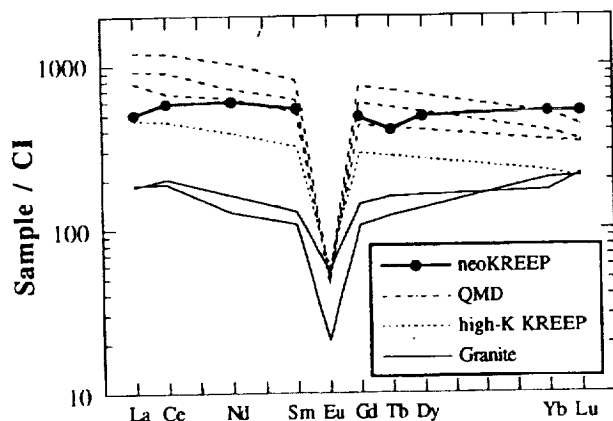


Fig. 4. Evolved lunar rocks.

component has been dubbed neoKREEP because of its relatively young age compared to postulated urKREEP (dregs of the LMO).

An Incompatible Trace-Element Composition for NeoKREEP: To determine a composition for neoKREEP, a set of 12 samples was chosen from the suite of group A basalts. These samples fall on or near a line from the liquid composition generated from 30% fractionation of the Apollo 17 orange volcanic glass through the field of group A basalt compositions (Fig. 3). It is assumed that the array of sample compositions along this line represents varying degrees of assimilation of the neoKREEP component.

For each element i , the expression for the composition in a sample is given as

$$(1-x)L_i + xN_i = A_i$$

where x is the fractional amount of assimilation, L_i is the concentration of element i in the original liquid, N_i is the concentration of the same element in neoKREEP, and A_i is the concentration of this element in one of the group A basalt samples. This relation can be rearranged to

$$N_i = (A_i - L_i + xL_i)/x$$

providing an expression for the neoKREEP composition as a function of the amount of assimilation. For each sample and element, a series of N_i values were generated for various amounts of assimilation (x). A best-fit curve was then determined through the points. The actual amounts of assimilation, of course, are different for each sample, resulting in a new assimilation variable for each equation. To eliminate this problem, all the samples were related to a common baseline sample, showing the least assimilation (deviation from the original liquid). For the rare earth elements (REE), sample 10004,224 was defined as the baseline sample because it has the lowest concentration of REE among the 12 samples used. The assimilations in the other 11 samples were then given in terms relative to the baseline assimilation. This procedure provided 12 equations for the neoKREEP composition as a function of the baseline assimilation. Eleven pairs of simultaneous equations were then solved (each sample coupled with the baseline sample) for each element, yielding 11 values for the composition of neoKREEP and the baseline assimi-

TABLE 2. neoKREEP composition.

		σ	x_{base}	σ
La	164	10	7.59	0.55
Ce	504	34	7.58	0.56
Nd	377	21	7.64	0.49
Sm	110	6	7.63	0.53
Eu				
Gd	134	8	6.59	1.10
Tb	21.0	1.1	7.47	0.51
Dy	167	8	7.54	0.40
Yb	114	7	7.98	0.56
Lu	17.6	0.9	8.20	0.54
Hf	96.0	6.5	9.06	0.71
Ba	1375	128	6.17	0.75
			avg. x_{base}	7.59
				0.18

All concentrations are in ppm.

lation. These 11 values were then averaged. The relations between the assimilation of each sample and the baseline assimilation were then used to determine the amounts of assimilation necessary in each sample. The results for some REE and other trace elements are given in Table 2. Approximately 7.5% assimilation of neoKREEP is required to produce the baseline group A basalt, and the entire array of group A compositions can be generated by 7.5–15% assimilation of neoKREEP with the composition in Table 2.

Comparison of NeoKREEP with Known Lunar Materials: The neoKREEP composition given in Table 2 and shown in Fig. 4 is more evolved than the postulated high-K KREEP composition given by Warren [14], but is not as evolved as materials such as the quartz monzodiorites from Apollo 15 (e.g., [16]). In addition, the chondrite-normalized REE slope is less than that of both high-K KREEP and the quartz monzodiorites, more akin to some lunar granites (e.g., 73215c, 14321, 1027 in [17]). Warren et al. [17] noted the possibility that some lunar granites may be the late-stage remnants of younger, smaller intrusions formed through partial melting of the deep interior, which is essentially what we envisage as the origin of neoKREEP. Internal isochrons for granite 14321, 1027 give ages of 4.09 ± 0.11 Ga and 4.11 ± 0.20 Ga from Rb-Sr and Sm-Nd, respectively [18], coinciding with our estimated age of neoKREEP of 4.15 Ga. Our neoKREEP composition is enriched in the REE over that of lunar granite. If neoKREEP represents some "pure" differentiate, perhaps the lunar granites reflect some dilution of similar material in the same way that KREEP basalts represent dilution of the "pure" differentiate or KREEP.

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N937/1879/302
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POSSIBLE PETROGENETIC ASSOCIATIONS AMONG IGNEOUS COMPONENTS IN NORTH MASSIF SOILS: EVIDENCE IN 2–4-mm SOIL PARTICLES FROM 76503.

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Studies of Apollo 17 highland igneous rocks and clasts in breccias from the North and South Massifs have described magnesian troctolite, norite, anorthositic gabbro, dunite, spinel cataclastites (e.g., [1,2]), and granulitic lithologies that may have noritic anorthosite or anorthositic norite/gabbro as igneous precursors [3], and have speculated on possible petrogenetic relationships among these rock types. Mineral compositions and relative proportions of plagioclase and plagioclase-olivine particles in sample 76503 indicate that the precursor lithology of those particles was troctolitic anorthosite, not troctolite. Mineral and chemical compositions of more pyroxene-rich, magnesian breccias and granulites in 76503 indicate that their precursor lithology was anorthositic norite/gabbro. The combination of mineral compo-

sitions and whole-rock trace-element compositional trends supports a genetic relationship among these two groups as would result from differentiation of a single pluton.

View from the Soil: Although highland igneous lithologies in Apollo 17 materials have been described previously, the proportions of different igneous lithologies present in the massifs, their frequency of association, and how they are related are not well known. In this abstract, we consider the proportions of, and associations among, the igneous lithologies found in a North Massif soil (76503), which may represent those of the North Massif or a major part of it. Soil 76500 was collected far from any boulders [4] that might have added components to the soil. We assume for this exercise that the proportions of lithologies in the 2–4-mm fraction (76503) reasonably represent those of their igneous precursors. Some soil particles are polymict impact-melt breccias; we also seek to constrain the composition of igneous materials that were incorporated into them.

Based on a geochemical survey of 243 2–4-mm particles, soil at the base of North Massif comprises mainly impact melt breccias, regolith or glassy breccias, admixed high-Ti mare basalt and orange glass fragments, and surviving fragments of highland igneous lithologies [5]. Some 25 particles retain highland-igneous textures or have compositions indistinguishable from those that do. A subset of these igneous fragments comprises coarse single crystals or aggregates that have unshocked igneous or shocked, relict igneous textures. Some 40 particles are polymict breccias or are recrystallized and have granulitic or polygonal textures. The polymict particles are dominated by igneous lithologies that are more pyroxene rich (anorthositic norite/gabbro) than those of the first group; they include two magnesian and one ferroan granulitic breccia groups (see [3]).

On the basis of a mixing-model calculation on the composition of <1-mm soil, [6] concluded that there must be some 15% of a high-Mg' (Mg/Mg + Fe) and low-ITE (incompatible trace element) component in addition to 36% of an anorthositic-norite component (roughly equivalent to gabbroic-anorthosite component of [7]). This is of special interest because the South Massif soils do not require the high-Mg' component in similar mixing models [6]. High-Mg' lithologies, however, are present in samples from both massifs.

The proportions of highland lithologies after removal of mare basalt and orange glass components and of regolith breccias and agglutinates whose compositions reflect mare-basalt admixture are as follows: 39 wt% noritic impact melt breccias, 18% magnesian granulitic breccias, 5% ferroan granulitic breccias, 9% other breccia lithologies that have generally anorthositic-norite compositions, and 29% troctolitic and noritic anorthositic fragments, including ~5% coarse, unshocked plagioclase crystals.

Olivine-Plagioclase Lithology: Trace-element and mineral compositions indicate that the coarse-grained plagioclase crystals found in the 2–4-mm size fraction and the fragments of coarse plagioclase and olivine \pm orthopyroxene are closely related to those of 76535 troctolite [8,9], 76335 troctolitic anorthosite [10], and a troctolitic clast in 76255 [11] (Fig. 1). The mass weighted mean composition of this group of fragments, however, is substantially more anorthositic (~85–90% plagioclase) than bulk 76535, suggesting that these derive from a more anorthositic body of rock, such as 76335. We have not found coarse olivine fragments in troctolitic proportion, nor is olivine present in sufficient abundance in finer soil fractions [12] to indicate a parent of troctolitic bulk composition. The proportion of troctolitic-anorthosite component in 76503 (19% on a regolith-breccia-free basis) is greater than expected from previous investigations. This component is part of the high-Mg' component of [6] and appears to be more prevalent in the local North Massif soil than in the breccia boulders from higher on the Massif and more prevalent than in the

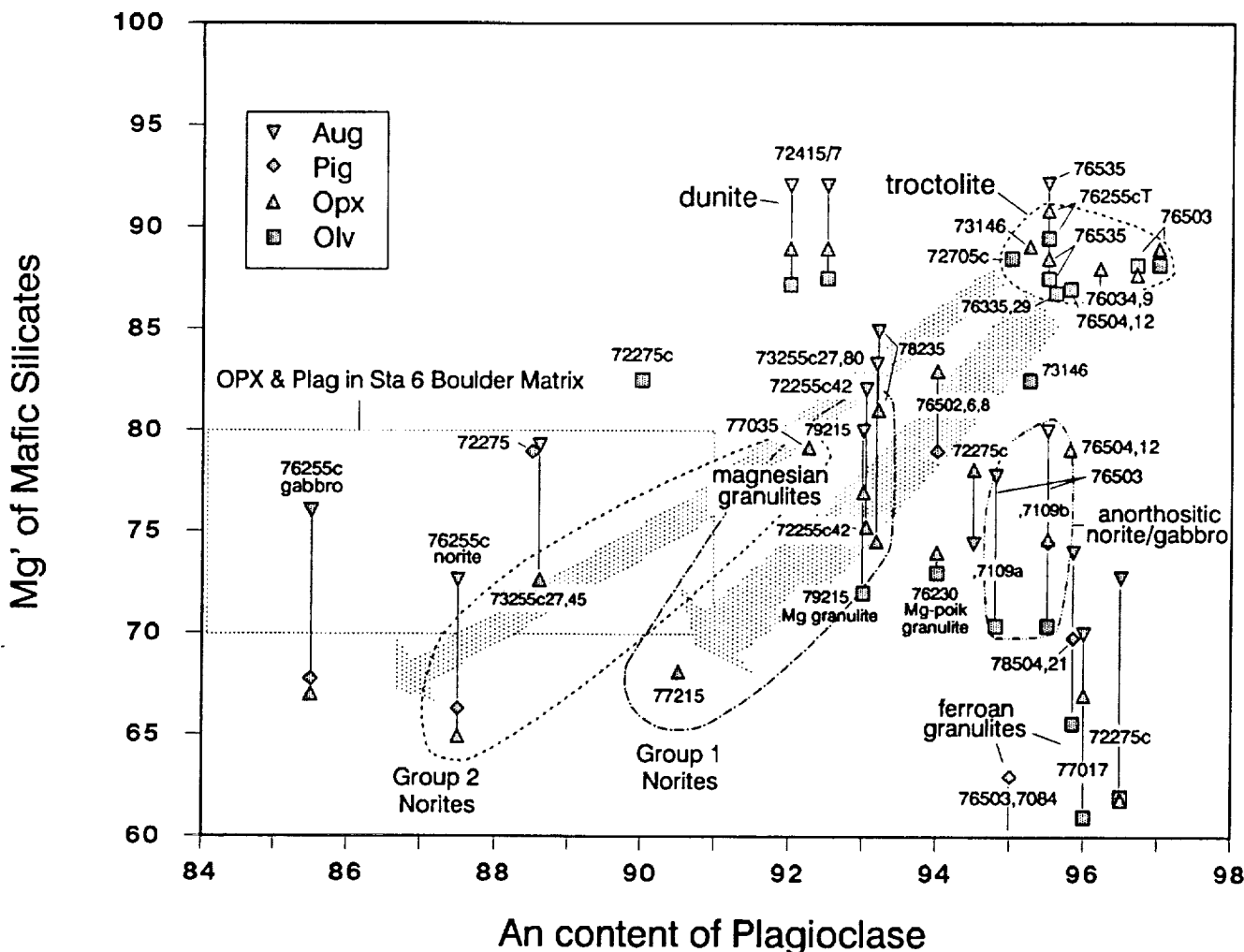


Fig. 1. Plagioclase and mafic silicates, Apollo 17 highlands.

sampled materials from South Massif. More pyroxene-rich lithologies do occur among fragments in 76503; however, in almost all cases, they have textures altered by impact and thermal annealing.

Granulites: Granulitic breccias and other impact-modified rocks of similar composition constitute 25% of the highland lithologies in 76503. Compositionally, both ferroan and magnesian groups as defined by [3] are present (Fig. 2); these compositions correspond to anorthositic norite/gabbro. Mineral compositions of the magnesian granulites are most similar to those of group 1 norites as defined by [13] (Fig. 1). Some of the moderately magnesian granulitic breccias have mineral compositions displaced toward those of the ferroan-anorthositic suite (Fig. 1). One such sample, 76503,7109, contains compositionally similar clasts with granulitic texture and relict igneous (anorthositic gabbro) texture.

Crystallization of a melt parental to the troctolitic anorthositic could potentially produce the observed mineral compositions and trace-element chemistry of the magnesian granulites and group 1 norites (Fig. 1, lower stippled arrow). The common occurrence of clasts of troctolite and gabbro with a group 2 norite in 76255 [10] suggests that these may also be related (Fig. 1, upper stippled arrow); however, we support the argument, on the basis of trace-element concentrations and mineral compositions, that group 1 and 2 norites cannot be related by means of a common magma (see [2,12]).

Constraints from Polymict Noritic Breccias: The mass-weighted mean composition of noritic melt breccias in 76503 has lower ITE concentrations than the average matrix composition of station 6 melt breccias (Fig. 2); however, in calculating the mean composition for 76503 noritic melt breccias, we have included particles that obviously contain highland igneous clasts or components. The distribution of mixed compositions suggests that the highland igneous component(s) of the noritic breccias is (are) on average similar to anorthositic norite/gabbro, i.e., magnesian granulitic breccias, and not to the troctolitic-anorthositic component that we find as actual coarse fragments in 76503. We suggest that the anorthositic norite/gabbro precursors are widespread components, as suggested by [3], and that the troctolitic anorthositic lithology, although possibly petrogenetically related, is more restricted in occurrence, perhaps excavated only by impacts reaching deep plutonic levels.

Acknowledgments: Our funding is from NASA grant NAG 9-56.

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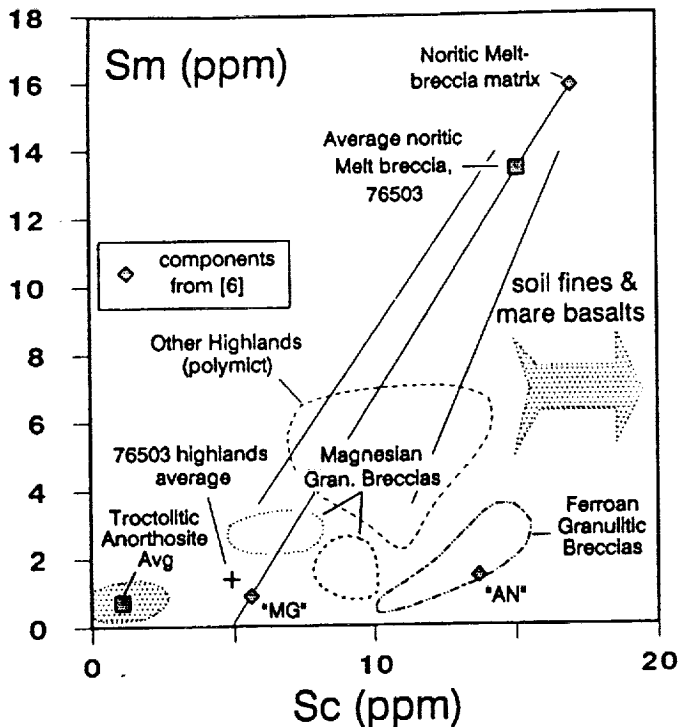


Fig. 2. Sample 76503, highlands.

Kremser D. T. (1992) *Proc. LPS, Vol. 22*, 275-301. [7] Rhodes J. M. et al. (1974) *Proc. LSC 5th*, 1097-1117. [8] Dymek R. F. et al. (1975) *Proc. LSC 7th*, 301-341. [9] Haskin L. A. et al. (1974) *Proc. LSC 5th*, 1213-1225. [10] Warren P. H. and Wasson J. T. (1978) *Proc. LPSC 9th*, 185-217. [11] Warner R. D. et al. (1976) *Proc. LSC 7th*, 2233-2250. [12] Heiken G. and McKay D. S. (1974) *LSC V*, 319-321. [13] James O. B. and McGee J. J. (1979) *Proc. LPSC 10th*, 713-743.

N 93-9187/94303

THE APOLLO 17 REGOLITH. Randy L. Korotev, Department of Earth and Planetary Sciences, Washington University, St. Louis MO 63130, USA.

Among Apollo landing sites, Apollo 17 provides the best opportunity to study the efficiency of formation and evolution of regolith by impacts, both large and small. The mare-highlands interface is crucial to this endeavor, but the Light Mantle avalanche and presence of fine-grained pyroclastics offer additional constraints. Compositional variation among soils from different locations and depths provides a means to quantify the extent of mixing by larger impacts. Because of their variety and complex history, Apollo 17 soils have been important in establishing agglutinate abundance, mean grain size, and abundance of fine-grained iron metal (as measured by I_p/FeO) as simple index of maturity (relative extent of reworking by micrometeorite impact at the surface) [7.9].

Surface Soils: Both the composition and modal petrography of the surface soils vary significantly across the site, and these variations are related in a reasonable way to the site geology. Soils from the valley floor are dominated by mare basalt, but even the most Fe-rich soils (stations 1, 5, LRV 12) contain ~15% highland material, and this proportion is greater in soils closer to the massifs (Fig. 1). Soil from the South Massif (stations 2, 2A, 3) contains only a small amount of

mare basalt (<4%) and one regolith breccia from station 3 (73131) appears to consist entirely of highland material. Soils from the North Massif/Sculptured Hills area contain a larger proportion of mare material than those from the South Massif, and that proportion increases to the east from stations 6 to 7 to 8 [16,4,5].

Mass-balance models have been successful at quantifying the compositional variation in terms of differences in proportions of components representing major lithologies at the site. Early models used four components, two of mare affinity and two of highlands affinity: high-Ti mare basalt (HT), orange/black pyroclastic glass (OG), noritic impact-melt breccia (NB), and anorthositic norite (gabbro) (AN) [16,6]. An important observation of this early work was that the greater concentrations of incompatible trace elements (ITEs, e.g., Sm; Fig. 1) at the South Massif indicate that the NB:AN ratio (~1:1) is greater there than that at the North Massif (~1:2) [16]. This suggested that the massifs, which were assumed to be structurally similar, had a high proportion of noritic melt (NB, deposited by the Serenitatis impact) at the top, and that the lower slopes were dominated by "anorthositic norite" (AN, actually a potpourri of prebasin crustal lithologies such as granulitic breccias and anorthositic troctolites and norites) [16,5] (Fig. 2). The high NB:AN ratio of the Light Mantle soils reflects their derivation from the upper slope

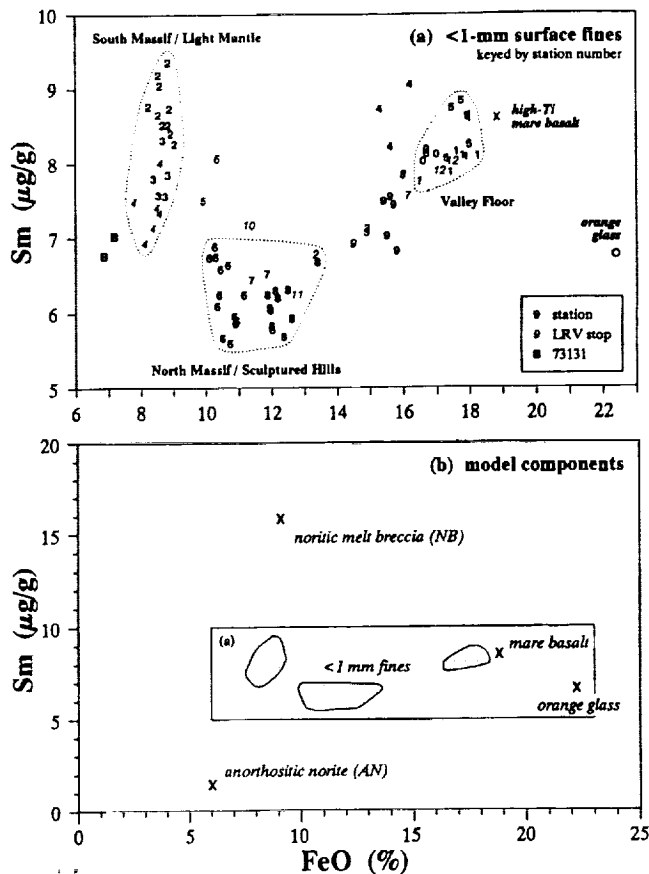


Fig. 1. (a) Compositional variation in Apollo 17 surface soils, from [5]. Sample 73131 is a regolith breccia from station 3. (b) Comparison of compositions of soils with the four major components of mass balance models [5]; inset box shows range of (a). No valley floor soil is devoid of highland material. The South Massif soils contain the least mare basalt and have a greater abundance of noritic melt breccia than the North Massif soil.

of the South Massif, and the low abundance of mare basalt indicates insignificant mixing with underlying or adjacent mare material since the avalanche. The North Massif soils derive primarily from mass wasting from the lower slopes, with mixing of basalt at their interface, hence the greater abundance of mare material [16,4].

Recent refinement of the model, based on a complete set of major- and trace-element data for all surface soils, indicates that more than four components are required to account for the compositional variation [5]. The basalt component of the valley floor soils consistently has a lower average TiO_2 concentration than typical high-Ti mare basalt (HT), suggesting the presence of a low-Ti basalt component in the <1-mm fines [5]. When modeled as VLT (very-low-Ti) basalt, ~7% of the total basalt component (HT+VLT) of soils from the valley floor is VLT basalt on average. This proportion is surprisingly high considering that the only VLT basalts found at Apollo 17 are small soil fragments [e.g., 23]. Perhaps VLT basalt is an old basalt that existed primarily as regolith at the time of eruption of the high-Ti basalts. Modeling also indicates that soils from the North Massif (but not South Massif) contain high-Mg/Fe lithologies not represented by any of the model components [16,5]. The inclusion of an additional component to represent troctolites and norites such as those found at stations 6 and 7 significantly improves model fits for soils from the North Massif area [5]. An improvement in fit for soils from station 2 is achieved by inclusion of a small amount (~4%) of a component of KREEP basalt [5,18]. With these refinements, the new model requires a smaller proportion of orange glass component to achieve mass balance in most soils than is indicated by previous models, and model results [5] now agree well with results of modal petrography for proportions of orange glass in the <1-mm fines [4]. Some orange glass occurs in most soils; away from station 4, the highest abundances are found at LRV stops 3 and 7 (~26%) [5].

Cores: Of the six cores taken on the mission, two are unopened (70012 at LM and 73001/2 at station 3, Lara Crater) and one has not been well studied (76001); the three well-studied cores are all from the valley floor. The deep drill core (DDC: 70001-9; 2.9 m) exhibits substantial variations in composition, lithology, and maturity with depth [6,22,13]. Three to five petrographically distinct units occur in the DDC although all were probably deposited in one event or two closely spaced events during the formation of the Central Cluster of craters [13,21,22]. Between 22 and 71 cm is a unit of coarse-grained,

immature soil dominated by basalt fragments [4,22]. An unusual concentration of siliceous, high-ITE glass occurs between 224 and 256 cm depth [6,22].

The core at Van Serg Crater (79001/2; 0.47 m) is unusual in that the top 8.5 cm is very mature and rich in both total N and cosmogenic ^{15}N . This soil is interpreted as an old soil that received extensive (~2 Ga) near-surface exposure, was buried, and then excavated by the Van Serg impact ~1.6 Ma ago [20]. The bottom of the core is enriched in a high-ITE component that may be noritic melt breccia (Fig. 3) [12]. An unusual variant of high-Ti basalt was encountered at ~42 cm depth [17].

The Shorty Crater core (74001/2; 0.67 m depth) is unique in being composed mainly of orange and black glass droplets having a mean grain size of ~40 μm . There is little variation in modal petrography and composition with depth [1,8], except that the top ~5 cm of soil has undergone *in situ* reworking by micrometeorite impact and addition of basalt and highland material, most of which is probably of local origin [8,11]. The material below 5 cm depth has received practically no surface exposure [2,8,11]. A five-stage model has been proposed [11]: accumulation as a volcanic ash deposit ~3.6 Ga ago, shallow burial for a short time, deep burial for most of the last 3.6 Ga, excavation by Shorty Crater impact 10–15 Ma ago, and *in situ* reworking of surface material since then.

The North Massif core (76001; 0.31 m) appears to be the best example in the lunar collection of continuous accumulation through downslope mass wasting. The core material is mature throughout, but the surface exposure was probably received upslope. Soil at the top (0–20 cm depth) has a larger proportion of noritic melt breccia as well as mare basalt than soil at the bottom (20–31 cm). This is taken as support of the model [16] that the massifs are capped with material rich in noritic melt [14,10,15].

Gray Soil from Station 4: The gray soil (74240, 74260) associated with the orange soils (74220) at station 4 is unusual in being rich in volatiles, containing abundant ropy glass particles, and being highly immature. Petrographic data and the mass-balance model indicate that it has a high NB/AN ratio and a low proportion of orange glass (~5%) [4,5]. The gray soil may contain a large component of old regolith, one that developed by mixing of basalt and underlying highland material prior to the pyroclastic eruptions. It appears to have been in close proximity to, but undergone negligible mixing with, a

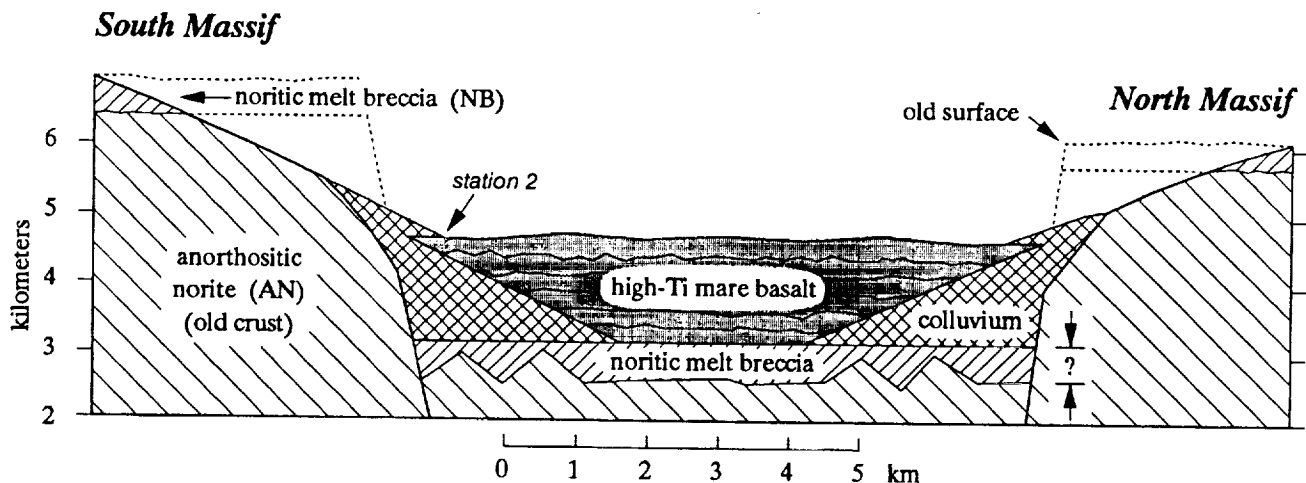


Fig. 2. Cross section of the Taurus-Littrow valley, after [24], with modifications of [16] (no vertical exaggeration). If a unit of noritic melt breccia underlies the basalt, it is probably not as discrete as pictured here, but occurs largely as a regolith mixed with the underlying anorthositic-norite-rich older crustal material.

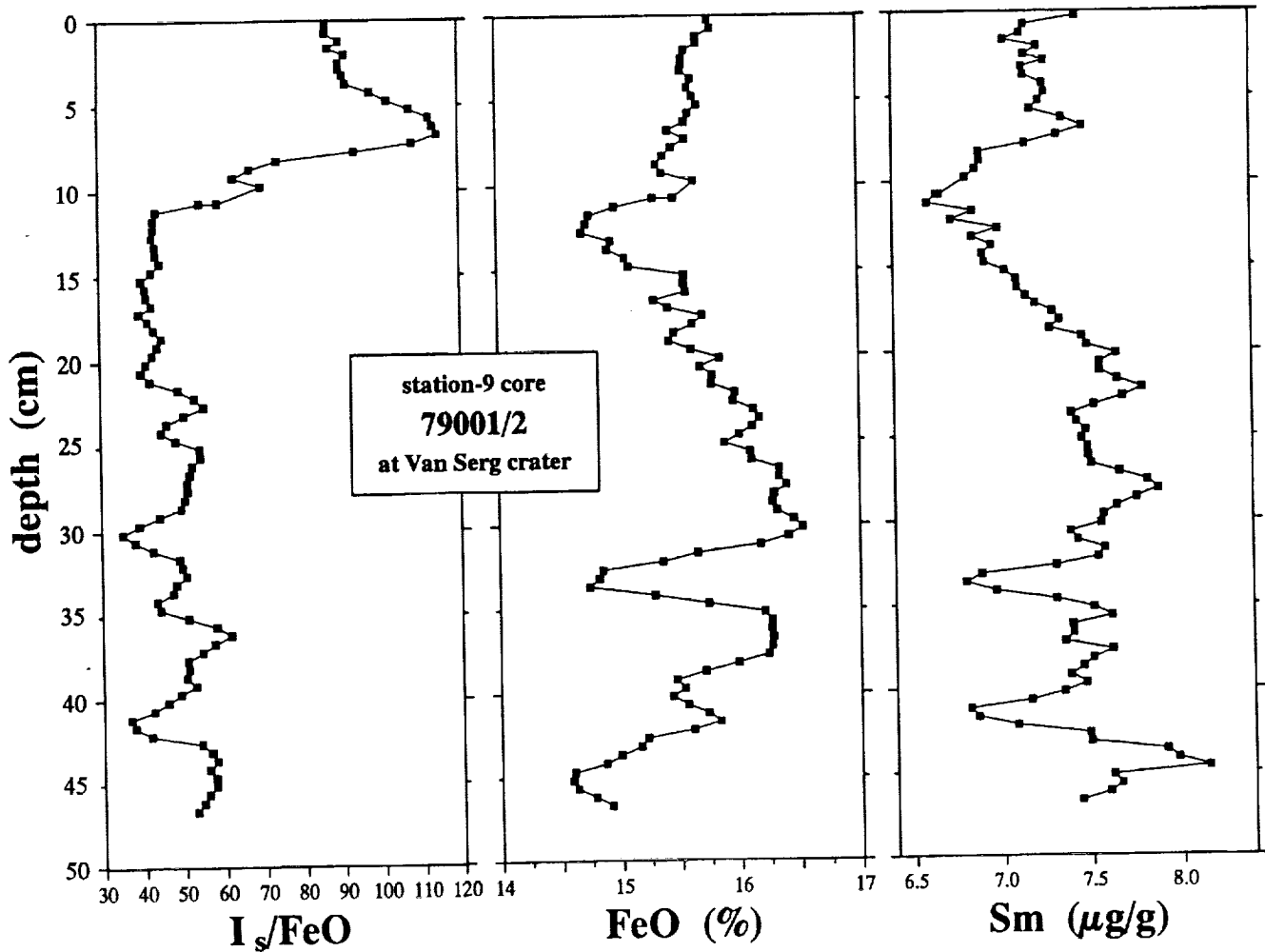


Fig. 3. Variation in maturity (I_s/FeO) and FeO and Sm concentration with depth in double drive tube 79001/2 [12]. For clarity, the data have been subjected to a three-point smooth (running average).

warm deposit of pyroclastics [5]. This soil has had little surface exposure prior to excavation by the Shorty Crater event.

Components with Unknown Sources: The regolith contains several rare components, the source of which is not known. Ropy glass particles are common in the gray soil of station 4 [3]. The composition of the ropy glasses is entirely consistent with local derivation as the interior glass is similar in composition to mare-free regolith breccia 73131 (above) [5]. However, a local source crater as large as that believed necessary to produce ropy glasses has not been identified. The high-silica, high-ITE glasses from the DDC (above) have been suggested as a possible Tycho component as the glasses do not resemble any lithic component found at the site or elsewhere [22]. The source of the volcanic glasses with 14% TiO_2 in regolith breccia 74246 is not known [19]. Likewise, VLT basalt is a regolith component with no known local source.

Important Points: Although the evidence of lateral transport of material is undeniable in the Light Mantle deposit, Tycho rays, and station 6 boulders, vertical mixing has also been important in bringing highland regolith to the valley floor [5,15,16,19]. The highland component of the soils from the center of the valley floor has a high NB/AN ratio. This component probably derives not from the tops of the massifs, but from a unit of similar composition underneath the basalt flows, as the valley was formed by block faulting [5,24]

(Fig. 2). The lateral and vertical variations in composition and petrography of the Apollo 17 regolith indicate that impact mixing, even over a time span of >3.5 Ga, is not overwhelmingly efficient. Thus, we are still able to correlate the major components of specific soils with the geologic formations from which they derive.

Future Work: Information about the early history of the site probably lies undiscovered in the gray soil at station 4. Systematic characterization of coarse fines from the massifs should reveal important differences between these bodies and provide data complementary to boulder studies. Mass balance models for soils from stations 1 and 5 suggest the presence of unidentified (mare?) components [5].

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N9439/1880004 P-3

GEOCHEMISTRY OF HASP, VLT, AND OTHER GLASSES FROM DOUBLE DRIVE TUBE 79001/2. D. J. Lindstrom¹, S. J. Wentworth², R. R. Martinez², and D. S. McKay¹, ¹NASA Johnson Space Center, Houston TX 77058, USA, ²Lockheed Engineering and Sciences Co., 2400 NASA Road 1, Houston TX 77058, USA.

Background: The Apollo 17 double drive tube 79001/2 (station 9, Van Serg Crater) is distinctive because of its extreme maturity and abundance and variety of glass clasts. It contains mare glasses of both high Ti and very low Ti (VLT) compositions, and highland glasses of all compositions common in lunar regolith samples: highland basalt (feldspathic; Al₂O₃ > 23 wt%), KREEP (Al₂O₃ < 23 wt%, K₂O > 0.25 wt%), and low-K Fra Mauro (LKFM; Al₂O₃ < 23 wt%, K₂O < 0.25 wt%). It also contains rare specimens of high-alumina, silica-poor (HASP) and ultra Mg¹ glasses. HASP glasses [I] contain insufficient SiO₂ to permit the calculation of a standard norm, and are thought to

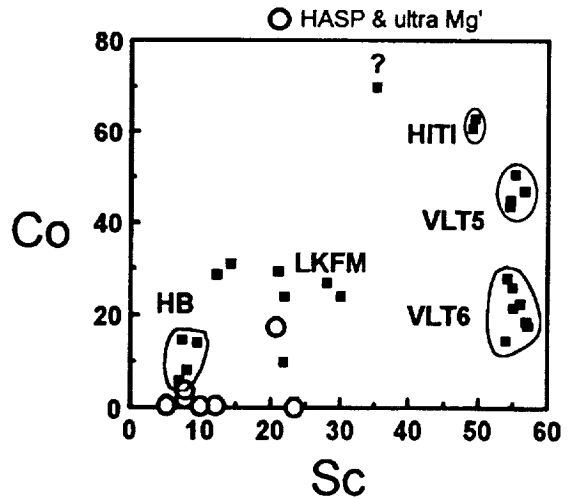


Fig. 1.

be the product of volatilization during impact melting. They have been studied by electron microprobe major-element analysis techniques but have not previously been analyzed for trace elements.

Samples and Methods: The samples analyzed for this study were polished grain mounts of the 90-150-µm fraction of four sieved samples from the 79001/2 core (depth range 2.3-11.5 cm). 80 glasses were analyzed by SEM/EDS and electron microprobe, and a subset of 33 of the glasses, representing a wide range of compositional types, was chosen for high-sensitivity INAA [2]. A microdrilling device removed disks (mostly 50-100 µm diameter, weighing ~0.1-0.5 µg) for INAA. Preliminary data reported here are based only on short counts done within two weeks of irradiation.

Results: Almost half the 80 glasses analyzed by electron microprobe are highland compositions, mostly with compositions ranging from LKFM to KREEP. Seven LKFM glasses were shown by INAA to have typically moderate Sc, Cr, and Co contents and a considerable range in REE (Table 1), and are not discussed further here. Of the more interesting 13 highland samples, five samples are classed as highland basalts. Six samples have HASP compositions (Table 2),

TABLE 1. Average compositions of common glass types in 79002.

(Wt%)	n=	H B	LKFM	VLT5	VLT6	High Ti
		5	23	7	21	16
SiO ₂		44.65	46.76	46.80	46.39	38.73
TiO ₂		0.24	1.15	0.86	0.87	8.87
Al ₂ O ₃		26.31	19.00	10.83	12.25	6.00
Cr ₂ O ₃		0.11	0.24	0.62	0.52	0.66
FeO		4.25	9.27	17.13	17.65	22.37
MnO		0.06	0.13	0.25	0.27	0.27
MgO		8.43	10.35	12.05	10.54	13.90
CaO		15.37	12.15	9.85	10.67	7.50
Na ₂ O		0.12	0.41	0.19	0.16	0.43
K ₂ O		0.03	0.11	0.06	0.02	0.10
P ₂ O ₅		0.02	0.05	0.05	0.03	0.06
Total		99.62	99.67	98.71	99.39	98.98
CaO/Al ₂ O ₃		0.58	0.64	0.92	0.87	1.27
Mg/Mg + Fe		0.77	0.67	0.56	0.52	0.52
(µg/g)	n=	4	7	4	7	2
Sc		8	21	55	55	49
Cr		740	1650	5050	3430	4450
Co		10	25	46	21	62
La		4	4-45	1.1	2.5	6.0
Sm		1.7	2-20	0.9	1.6	6.4
Yb		1.6	1.8-16	1.4	2	4.3

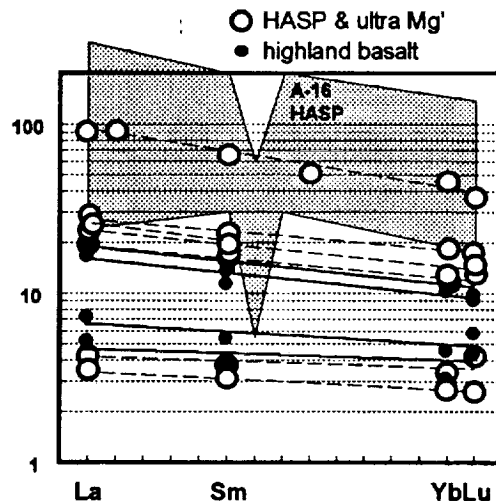


Fig. 2.

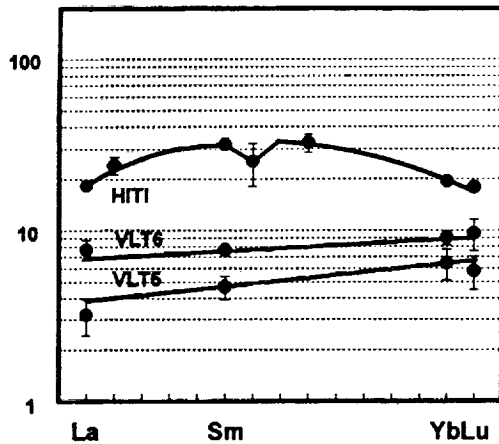


Fig. 3.

and four of the six also fall into the ultra Mg' (atomic Mg/Mg + Fe > 0.90) category. Two HASP glasses have more normal Mg' values, while two additional glasses have non-HASP, ultra Mg' compositions. INAA trace-element results on these 13 particles do not correlate strongly with the major element categories. Most of these glasses are low in ferromagnesian elements (Tables 1 and 2, Fig. 1), and all but one (,146-10) have REE contents below the range observed by [3] in Apollo 16 ultra Mg' glasses (Fig. 2). Except for ,146-10, the REE patterns appear to form two groups, those with 3-7 \times chondritic La and those with 16-30 \times . Whether the gap between

them is real or merely due to the small numbers of samples analyzed is not clear. The important point is that ultra Mg' and HASP samples occur in both groups, as do highland basalts, which are the best candidates for precursors of most of these HASP and ultra Mg' glasses.

Mare glasses comprise the other half of the samples microprobed. Sixteen of these have high-Ti compositions, and just two of them were analyzed by INAA, showing the typical bow-shaped REE patterns (Fig. 3). VLT glasses were more numerous and included 6 VLT5 and 21 VLT6 glasses (the "mare 5" and "mare 6" varieties of [4]). These glasses form tight compositional groups, with a clear separation between the two groups in both major- and trace-element data. The VLT5 group glasses contain almost twice the Co, higher MgO, lower Al₂O₃, and about 1.5 \times lower REE than the VLT6 group. Previous analyses of Apollo 17 VLT lithic clasts [5] found only the VLT5 trace-element pattern. The VLT6 glasses cluster very tightly compositionally (e.g., 21 analyses for Al₂O₃ ranged from 12.01% to 12.67%), suggesting that this may be a pyroclastic deposit.

Conclusions: Much can be learned from small glass samples by first using SEM/microprobe analyses of grain mounts to scan a large number of particles to find those with interesting compositions, then microdrilling to obtain samples for high-sensitivity INAA for trace elements. This technique has shown that five out of seven of the HASP and ultra Mg' glasses found in these 79002 soils are most likely to be formed from the highland basalts. While cobalt follows iron in being lost, presumably due to the extreme impact heating that produces these glasses, characteristic abundances of Sc, Cr, REE, and other refractory elements remain in the melt. These five glasses are all quite low in REE and Sc, and most closely resemble the noritic anorthosite composition of some of the lunar meteorites [e.g., 6].

TABLE 2. Compositions of HASP and ultra Mg' glasses in 79002.

(Wt%)	Ultra Mg HASP				Non-HASP Ultra Mg		Other HASP	
	,143 23	,143 33	,145 25	,146 33	,146 4	,146 10	,142 16	,143 31
SiO ₂	43.01	41.21	40.68	36.29	43.03	52.41	37.64	40.23
TiO ₂	0.28	0.23	0.23	0.23	0.21	1.50	1.30	0.38
Al ₂ O ₃	28.76	29.39	34.06	32.85	32.17	20.14	25.69	27.84
Cr ₂ O ₃	0.15	0.12	0.05	0.09	0.08	0.24	0.10	0.13
FeO	1.09	2.03	0.65	0.68	0.62	1.95	7.88	4.46
MnO	0.03	0.04	0.01	0.01	0.00	0.16	0.11	0.07
MgO	10.50	11.15	3.91	11.06	4.08	9.64	10.76	10.71
CaO	16.18	16.06	20.23	18.17	18.81	12.85	15.63	16.70
Na ₂ O	0.12	0.01	0.01	0.01	0.01	0.24	0.01	0.04
K ₂ O	0.01	0.01	0.00	0.00	0.01	0.12	0.00	0.01
P ₂ O ₅	0.04	0.03	0.00	0.01	0.00	0.01	0.01	0.03
Total	100.27	100.28	99.85	99.42	99.01	99.31	99.16	100.60
CaO/Al ₂ O ₃	0.56	0.55	0.59	0.55	0.58	0.64	0.61	0.60
Mg/Mg Fe	0.95	0.91	0.91	0.97	0.92	0.90	0.71	0.81
Sc	8	8	12	—	10	23	21	5
Cr	870	890	460	—	650	1400	750	360
Co	2	3	<2	—	<2	<4	17	<12
La	8	7	1.4	—	1.2	29	9	6
Sm	3.3	2.9	0.8	—	0.6	13	4.6	2.9
Yb	2.8	2.5	0.8	—	0.6	10	4.0	<5

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P. 1 N 93-1880 B05
THE TAURUS-LITTROW DARK MANTLE, LIGHT MANTLE, CRATER CLUSTER, AND SCARP. Baerbel K. Lucchitta, U.S. Geological Survey, Flagstaff AZ 86001, USA.

The Taurus-Littrow landing site is on the floor of a grabenlike valley that is radial to the Serenitatis Basin; the valley is gently inclined to the east. It is bordered by steep-sided massifs that rise 2 km above the valley floor and form part of the Serenitatis Basin rim [1,2]. The valley floor is exceptionally dark, and in one place it is overlain by a light-colored mantle apparently derived from the massif to the south. The floor is also peppered by overlapping craters. A scarp, up on the west, transects the valley floor and enters the highlands to the north.

Before the mission, the valley floor was interpreted to be covered by a dark mantle that also covers part of the highlands along the southeastern rim of the Serenitatis Basin [3-5]. The dark unit has an albedo as low as 0.79, which is darker than the outer dark mare ring in the Serenitatis Basin [6]. The dark mantle was generally interpreted to be a young pyroclastic deposit. The reasons for the young age assignment were (1) a dearth of small craters when compared with crater densities on surrounding mare surfaces and (2) the observation that the dark mantle appears to cover young craters and the fresh-looking scarp. Postmission analysis established that the dark material in the landing area is mostly composed of old mare regolith averaging about 14 m in thickness and containing about 5% dark beads [7]. These beads are associated with orange beads in the ejecta of a small crater. The generally accepted postmission interpretation of the dark mantle is that it is a deposit of dark and orange beads erupted from fire fountains soon after the mare basalts were emplaced [7], and that the reworking of the beads into the regolith gave it the low albedo [1,2]. The presence of orange and dark layers on mare and highland material at the base of the regolith is also supported by detailed studies of the Sulpicius Gallus Formation on the west side of the Serenitatis Basin, where many outcrops give clues to the stratigraphic relations [8]. There the dark mantle covers mare and highlands with deposits on the order of 50 m thick. The reason for the paucity of small craters in the dark mantle is now attributed to the unconsolidated, friable nature of the mantle; small craters are eroded much more rapidly by mass wasting and impact gardening than in the adjacent hard-surfaced, younger lavas, which have thin regoliths [9].

The controversy concerning the light mantle mostly centers on whether it is a simple avalanche deposit or whether it was emplaced largely by impact (whereby secondary projectiles may have hit the top of the massif from which the light mantle is derived and dislodged the material). A detailed study of secondary impact craters of Tycho over the entire lunar nearside showed that Tycho secondaries have distinctive characteristics [10]: they are crater clusters with sharp, fresh-looking surface textures; the surface downrange from secondary craters displays a braided pattern with V-shaped grooves and ridges; acute angles of the Vs as well as the trend of many ridges, together with the overall cluster trend, point in the general direction of Tycho. An analysis of the light mantle, the crater cluster on top of the massif, and the crater cluster on the valley floor showed that the Tycho secondary characteristics are present. Thus, the light mantle is not a

true avalanche but an impact-propelled feature, and the crater cluster on the valley floor is composed of Tycho secondaries. The scarp on the valley floor displays the typical characteristics of a mare-type wrinkle ridge. Wrinkle ridges in the Serenitatis region were studied by several authors [11] and were generally considered to be thrust faults. A detailed study of the scarp on the valley floor and its continuation into the adjacent highlands showed that it has characteristics of high-angle normal and high-angle reverse faults [12]. It is likely that this scarp is the surface expression of minor postmare adjustments along high-angle faults that reactivated older faults bordering basin-rim massifs.

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MORPHOLOGY AND COMPOSITION OF CONDENSATES ON APOLLO 17 ORANGE AND BLACK GLASS. David S. McKay¹ and Sue J. Wentworth², ¹NASA Johnson Space Center, Houston TX 77058, USA, ²Lockheed, 2400 NASA Road 1, Houston TX 77058, USA.

Background: Lunar soil sample 74220 and core samples 74001/2 consist mainly of orange glass droplets, droplet fragments, and their crystallized equivalents. These samples are now generally accepted to be pyroclastic ejecta from early lunar volcanic eruptions [1,2]. It has been known since early examination of these samples that they contain surface coatings and material rich in volatile condensable phases, including S, Zn, F, Cl, and many volatile metals. Meyer [3] summarizes the voluminous published chemical data and calculates the volatile enrichment ratios for most of the surface condensates.

The volatiles associated with these orange and black glasses (and the Apollo 15 green glasses) may provide important clues in understanding the differentiation and volcanic history of the Moon. In addition, condensable volatiles can be mobilized and concentrated by volcanic processes.

The Problem: While considerable chemical data exist on these samples, the phases that contain the volatile species are not known; no unequivocal condensate mineral has ever been identified in these samples, with the possible exception of sodium chloride salt crystals [4]. No X-ray diffraction or electron diffraction data exist on any of these phases. More positive information on the phases or mineralogy of the condensates would lead to a better understanding of the nature of the volatile gases from which they formed and the temperature, total pressure, and gas fugacity conditions associated with the eruption. This in turn would lead to a better understanding of lunar extrusive volcanism and of the history of the magma from which the pyroclastics came.

Some basic questions still unanswered include

1. What is the composition of the gas phase of the eruption? Was the major driving gas CO-CO₂ or was it something else?
2. What is the source of the condensable volatiles found on the droplets? Are they all indigenous lunar volatiles? Are any of them,

particularly the siderophiles, from meteorites? Are they from primitive undifferentiated regions at depth?

3. What were the conditions of the eruption? At what point in the eruption were volatiles condensed on droplet surfaces? Were volatiles still mobile after the pyroclastics were deposited and were some of them condensed at that time? Were conditions present that might lead to volatile concentrations in some part of the deposit? Is it likely that such concentrations exist and can they be located?

Approach: We have reviewed many of our existing photomicrographs and energy dispersive analysis (EDXA) of grain surfaces and have reexamined some of our older SEM mounts using an improved EDXA system capable of light-element detection and analysis (oxygen, nitrogen, and carbon).

Results: (1) *Micromounds.* The 74001/2 glasses have micromound coatings (Fig. 1), which were first described and named by Heiken et al. [1] and appear to be characteristic of other lunar volcanic glasses such as the Apollo 15 green glasses. Individual micromounds range in size from ~20–300 Å (e.g., [4]). The micromound coatings commonly contain scrapes and gouges that suggest that the glasses were abraded while the micromound coatings were still soft, i.e., during fire fountaining. The sizes and densities of micromounds vary from glass to glass. In some cases, they form relatively large masses of micromounds (Fig. 2). (2) *Shrinkage cracks and vesicles.* Very dense continuous coatings have also been found; some of the continuous coatings contain cracks (Fig. 3) and vesicles that may have resulted from shrinkage during cooling, from desiccation, or by evolution of a volatile species [4]. In addition to micromounds, the glass surfaces also contained volatile-rich grains, such as NaCl phases and sulfur-rich forms. Some sulfur peaks were large, suggesting that elemental sulfur might even be present. The volatile-rich phases were found as small euhedral crystals, such as the NaCl in Fig. 4, and as more massive forms, such as the round NaCl(?) puddles in Fig. 5a.

Secular changes. Our current studies of the surface features on the 74002 glasses show quite clearly that the volatile species have been affected by the storage of the samples at ambient conditions since 1978. Figures 5a and 6a are SEM micrographs obtained in 1978;

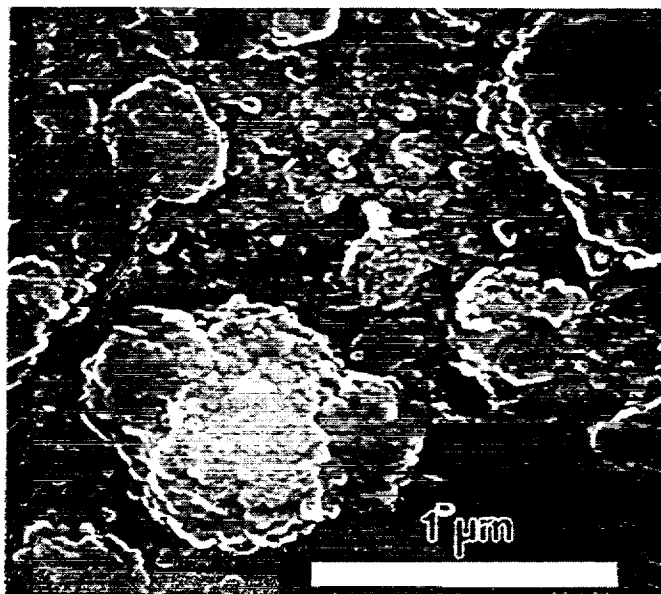


Fig. 2.

the surface deposits identified at that time were Na and Cl in both views (the round masses in Fig. 5a and the 2- μ m-size flat, angular grain at the center of Fig. 6a). Figures 5b and 6b are SEM images of the same two areas obtained in 1992, and they demonstrate that both surfaces have undergone significant changes. In Fig. 5b, the round NaCl masses have disappeared and although the rest of that surface is much like it was, there is some indication that the bumpy micromound surface has been somewhat subdued and otherwise changed. The surface of Fig. 6b shows a much more drastic change from the earlier surface of Fig. 6a: It is now covered by a dark deposit consisting of Na or Zn (the two peaks cannot be readily distinguished) and S. The



Fig. 1.

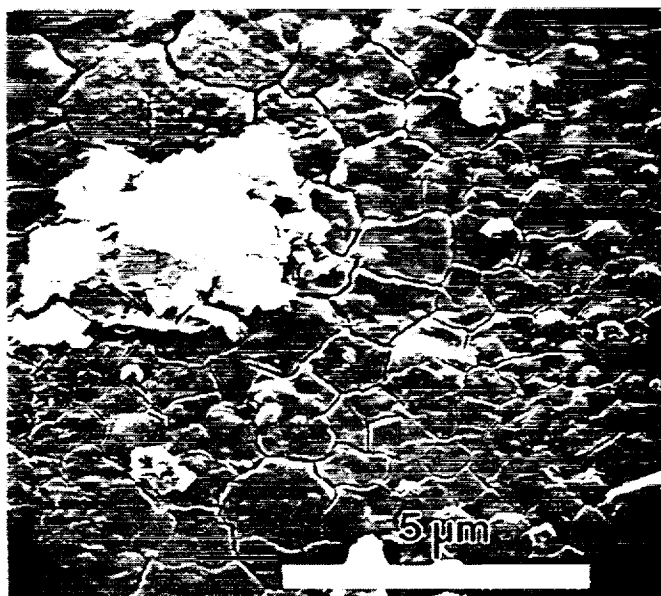


Fig. 3.



Fig. 4.

EDXA spectrum for this new deposit is shown in Fig. 8a and a nearby reference surface spectrum is shown in Fig. 8b. It is not clear whether the original NaCl grain still underlies the new deposit. We know of no simple way that this composition could result from contamination, handling, or reaction with air or moisture; the best explanation seems to be that it represents a remobilization of the existing Zn (\pm Na) and S on the surface of the droplet.

In addition to loss of some of the phases, a softening and merging of some of the micromound textures appears to have occurred, and the surface now appears to be mottled at a coarser scale. It is possible that

some of these changes have resulted from reaction of the condensate phases with the moist Houston air. These samples were originally Au-Pd coated to a thickness of about 5 nm, and were stored in air in plastic boxes since their original studies. No attempt was made to keep them in a controlled atmosphere. It appears that at least some of the condensate phases are relatively unstable under terrestrial conditions. These observations emphasize the potentially fragile nature of the surface condensates and the necessity to protect the samples from atmospheric exposure. On the plus side, this instability may provide clues to the type of phases present.

Carbon on droplet surfaces. Analysis of surface grains and coatings of some droplets indicated the presence of carbon in some of the spectra (Fig. 7a). Because the grains were mounted on carbon planchettes, we originally attributed the carbon signal to stray X-rays from the planchette. However, closer examination and comparison with nearby broken or chipped surfaces under nearly identical geometric conditions showed that in most cases carbon was present on the original droplet surface but not on the nearby chipped surface (Fig. 7b). The chipping apparently occurred on the Moon, but the absence of micromound coatings on many chipped surfaces indicates that chipping probably occurred late in the eruption sequence or during the deposition of the pyroclastic deposit. We interpret this relationship to indicate that the carbon is a real component of the volatile coatings and condensed phases on the surface of the pyroclastic droplets.

It is possible that the driving gas for the pyroclastic eruption was CO. Some graphite remains in the sample, perhaps because of self-reduction displayed during intrinsic fugacity measurements. It is estimated that the orange glass reflects oxygen fugacity conditions about 0.5 to 1.0 log units above typical mare basalts and slightly above the iron-wustite buffer. Mao et al. [5] estimate on the basis of optical spectra that the orange glass was produced in a somewhat more oxidizing environment compared to typical lunar basalts. But Morris et al. [6] report some metallic iron present in all samples of orange and black glass, so the overall oxygen fugacity could not have been

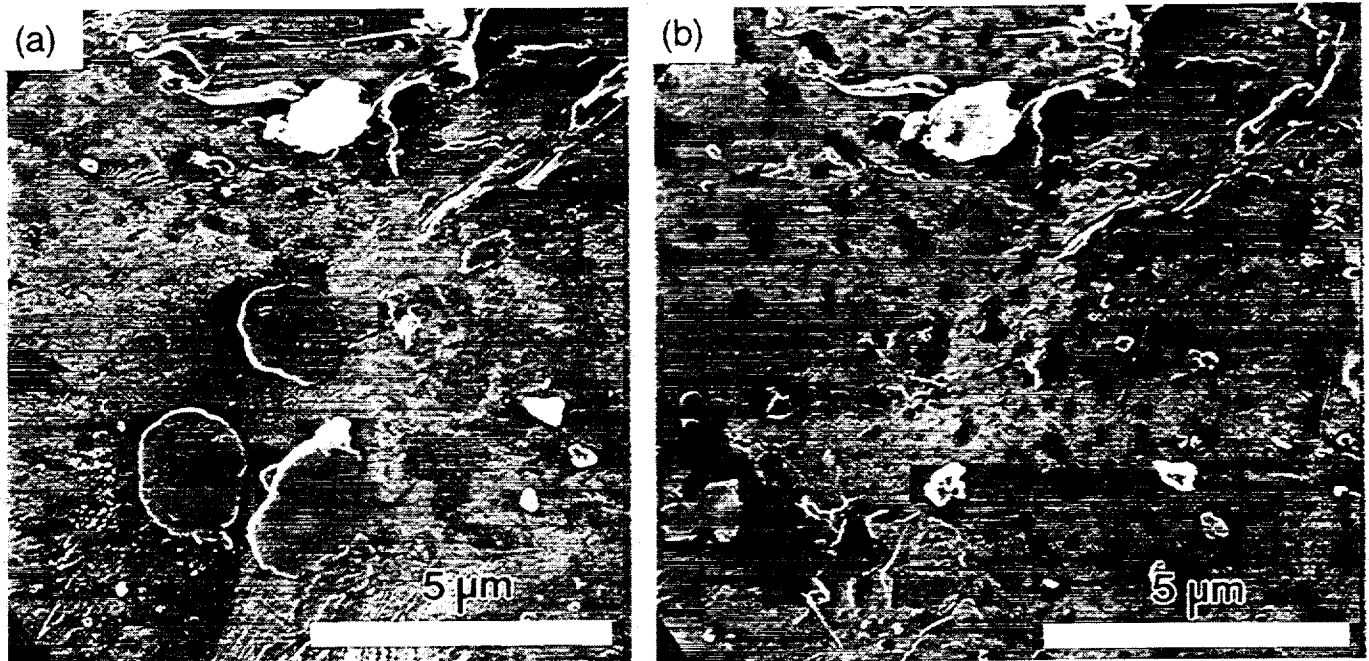


Fig. 5.

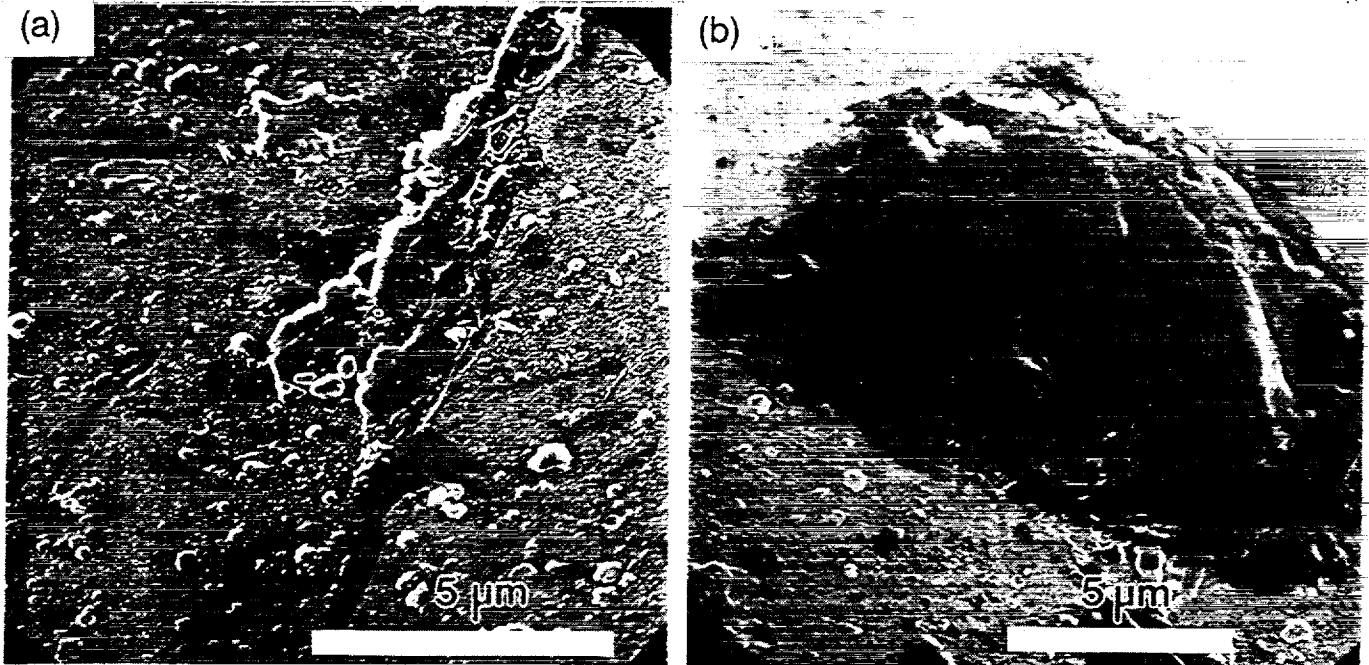


Fig. 6.

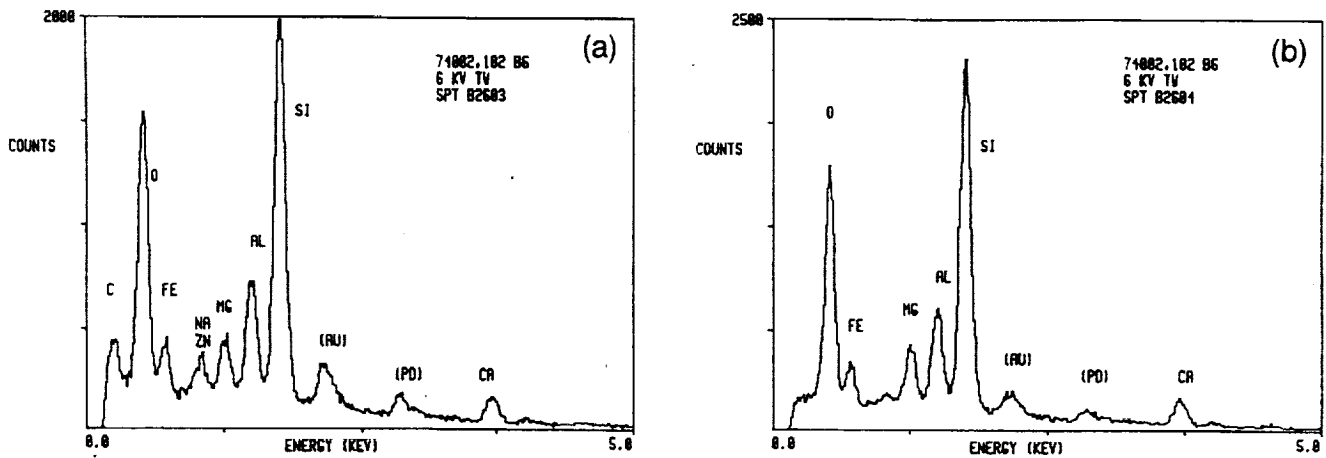


Fig. 7.

significantly above the iron-wustite buffer before and during the period when the glass was quenching. However, it may be that oxygen fugacity changed considerably during the eruption process. Normally CO gas would not leave a condensible trace. However, it is possible that during the rapid changes in pressure and temperature associated with the pyroclastic eruptions, a number of chemical reactions may have taken place that produced a condensible carbon phase. Because of rapidly changing conditions, these reactions would not necessarily reflect equilibrium conditions. An example might be a disproportionation reaction $\text{CO} \rightarrow \text{C} + \text{CO}_2$ that could have deposited carbon (graphite) on droplet surfaces during some phase of the eruption and cooling.

Another possibility is that a metal carbonyl was formed during some late stage of the eruption or after deposition and that this

carbonyl may have decomposed and deposited iron and some carbon on grain surfaces.

Yet another possibility is the formation of carbonates at lower temperatures, e.g., ZnCO_3 , during the late stages of the eruption or after deposition of the ejecta blanket. Such carbonates could form if CO_2 becomes a significant phase in the transient gas environment, produced, for example, by the disproportionation reaction described above. Although the orange and black glass is low in carbon [7], gas release studies indicated that CO_2 evolved at relatively low temperatures ($<400^\circ\text{C}$) and a carbonate phase might be present on grain surfaces.

Sulfur, sulfides, sulfates, or what? The association of sulfur with other volatiles such as Zn has been well documented for these surface coatings (e.g., [8]). Many have assumed the sulfur is present

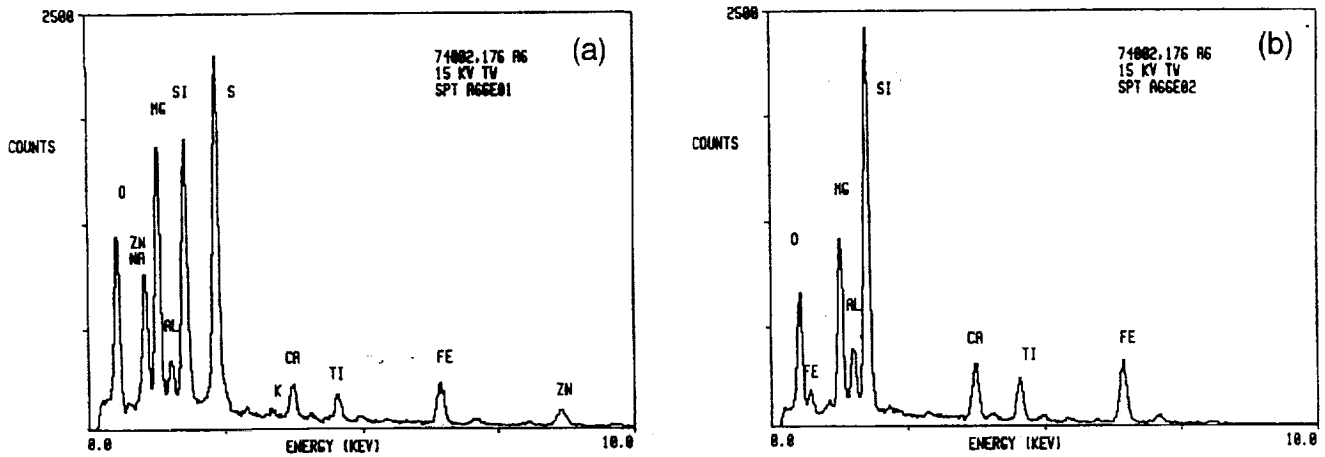


Fig. 8.

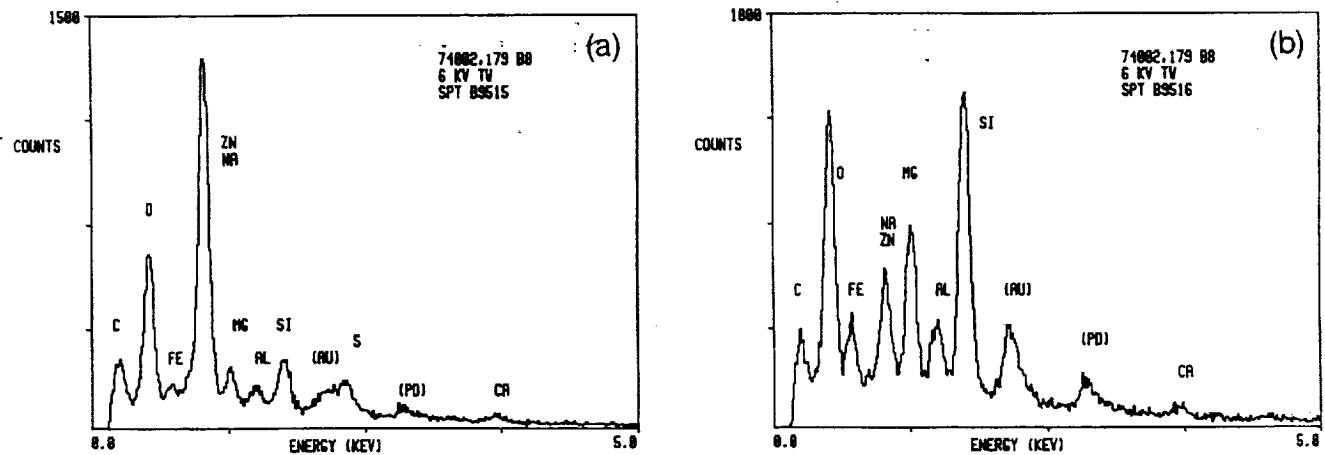


Fig. 9.

as a sulfide, but Clanton et al. [4] suggested that elemental sulfur might be present. Sulfur in combination with a variety of other species could also be present in amorphous vapor deposits. We have now observed two associations not previously reported that may bear on this question. First is the association of oxygen with sulfur when some small surface grains are analyzed. Second is the association, in some cases, of sulfur with both oxygen and carbon. Figure 9a illustrates the spectrum (6KV) from a small surface grain and Fig. 9b is from the adjacent "background" droplet surface. This pair illustrates that for the small grain, Zn (\pm Na), O, and C are all greatly enhanced with respect to major elements such as Si and Mg when compared to the background spectrum. Sulfur is a very minor peak on both spectra. While certainly not conclusive, this relationship suggests that the Zn may be present as an oxide or carbonate rather than a sulfide. Other examples of enhanced oxygen in specific condensate grains relative to the background silicate oxygen value are present. Both oxide and carbonate formation would probably require higher oxygen fugacities than necessary for sulfides or elemental sulfur, but, as discussed above, some increased oxygen fugacity compared to normal mare basalts may be indicated by other data, and late-stage increases in local oxygen fugacity during cooling or after deposition are also possible. However, we cannot rule out that the secular changes described above may have oxidized to some extent the original phases. Cirlin et al. [9] observed oxidized sulfur in XPS spectra of

74001, but proposed that the sulfates resulted from reactions with terrestrial air and water vapor. This can best be checked by analyzing for comparison freshly mounted pristine samples that have been maintained in nitrogen.

Summary and Conclusions: A review of existing data and new observations indicates that volatile condensates on Apollo 17 orange and black glasses are not only complex in chemistry but are complex in morphology and phase composition. New data indicate that carbon appears to be present in variable amounts on droplet surfaces. Zinc may be associated with oxygen and/or carbon in some cases, and a zinc carbonate or a zinc oxide appears to be a possible phase. Sulfates as a possible condensate are suggested by an apparent association of sulfur with oxygen, but verification will require more detailed work. Effects of exposure to air are clearly present and must be more fully analyzed before definitive condensate phase identification can be made. Additional studies of these phases may lead to a better understanding of the details of the eruption and could lead to better models of the gas phase composition and the fate of volatile species. Studies of lunar volcanic volatiles and condensates may also have major implications for lunar resource issues.

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Proc. LSC 4th, 397-412. [6] Morris R. et al. (1978) *Proc. LPSC 9th*, 2033-2048. [7] Gibson E. and Moore G. (1973) *EPSL*, 20, 404-408. [8] Butler P. and Meyer C. (1976) *Proc. LSC 7th*, 1561-1581. [9] Cirlin E. H. et al. (1978) *Proc. LPSC 9th*, 2049-2063.

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GEOLOGY OF THE APOLLO 17 SITE. W. R. Muehlberger, The University of Texas at Austin, Austin TX 78712, USA.

The Apollo 17 landing site was unique in several respects: (1) It was the only site that was not selected from telescopic-based geologic interpretation; interest in the site was generated by the visual observations of Al Worden, Apollo 15 Command Module pilot, who interpreted dark-haloed craters as possible cinder cones. (2) Instead of 20-m-resolution photographs, as was the norm for all earlier missions, this site had Apollo 15 panoramic camera photography coverage that had 2-m resolution. (3) It had a geologist-astronaut aboard who was intimately involved in all stages of planning and mission operation, and was also instrumental in the design of a long-handled sample bag holder that eliminated the need for the crew to dismount before collecting a sample, which then permitted sampling between major stations.

The following summary is mainly verbatim extracts (paragraphs) from reference [1], to which the interested reader is referred for details of site geology, sample description, and geologic synthesis of the site as viewed from studies through 1976.

The two major geologic objectives identified by the NASA Ad Hoc Site Evaluation Committee for this (the last) mission to the Moon were (1) sampling of very old lunar material such as might be found in pre-Imbrian highlands as distant as possible from the Imbrium Basin and (2) sampling of volcanic materials significantly younger than the mare basalts returned from the earlier missions. Photogeologic interpretation had suggested that such young volcanic materials on the Moon were pyroclastic, which would make them attractive not only for extending our knowledge of the Moon's thermal history, but because they might provide a record of volatile materials from the Moon's interior; furthermore, they might contain xenoliths of deep-seated lunar rocks.

Premission Plans: The three traverses were designed to sample, observe, and photograph each of the units recognized in the landing area (listed in order of decreasing priority): highlands (massifs and

Sculptured Hills), dark mantle (interpreted to be young pyroclastic material), and subfloor material (interpreted to have been emplaced as a fluid, or fluidized, material).

Highland materials rim the Taurus-Littrow Valley and thus would be sampled at several stations along the base of the mountain fronts. In addition, large boulders that had rolled down the slopes would be sampled so that they could be restored to their initial position to determine whether there was any internal stratigraphy to the highlands (Serenitatis Basin or older basin ejecta beneath a cap of Imbrium ejecta). The light mantle at the base of the South Massif (an avalanche deposit believed to have been caused by ejecta from Tycho when it impacted the South Massif) was to be sampled at stations at various distances from the base of the massif in the hope that these would be samples representative of different stratigraphic (?) levels on the massif.

Dark mantle material, a veneer of dark material over both valley floor and highland regions, would be collected at several localities on the valley floor. The rims of large craters were places that it was hoped that the contact relations between the younger dark mantle and the older crater rim, wall, and floor materials could be observed. Ambiguous age relations where the boundary between the dark and light mantles appeared to be diffuse led to the possibility that they were, at least in part, deposited concurrently. Shorty and Van Serg Craters were planned for the study and collection of dark mantle material supposedly erupted from volcanic vents or excavated by impacts. Sampling of the dark mantle at different locations would provide information about lateral variation.

Large blocks on the rims of the large craters on the valley floor were interpreted to be subfloor material blasted to the rim by impact. Because Apollo 16 demonstrated that plainslike surfaces could be formed by impact ejecta, the nature and origin of the valley-filling material was in doubt, and awaited sampling to demonstrate that it was basalt as was the surface on which most other lunar landings had been made.

The Mission: The actual traverses closely approximated the planned ones. This was probably the result of fewer parts failures, better photography, which permitted better premission interpretation/planning, and a timeline that had some flexibility. Emory Crater on EVA 1 was deleted, but Steno Crater, one of comparable size but closer to the LM, was sampled. Sherlock Crater, the last planned stop of EVA 3, was also deleted and only a scoop sample using the long-handled sampler was obtained about a crater radius from Sherlock.

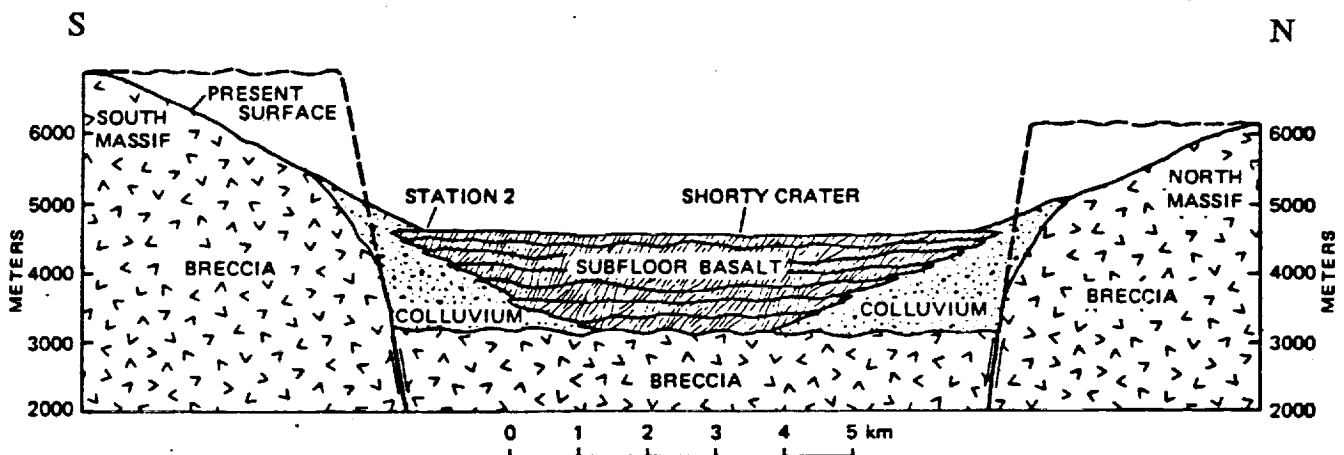


Fig. 1. Relations inferred among major subregolith units. Dark mantle is too thin to draw at this scale. Light mantle extends from the South Massif to just beyond Shorty Crater. No vertical exaggeration. Figure 242 in [1].

The astronauts worked approximately 22 hr on the lunar surface, traversed about 30 km, collected nearly 120 kg of samples, took more than 2200 photographs, and recorded many direct geologic observations (all lunar records!). The lunar surface data, sample results, and geologic interpretation from orbital photographs are the bases for the following geologic synthesis.

Postmission Interpretation: The Taurus-Littrow massifs are interpreted as the upper part of the thick, faulted ejecta deposited on the rim of the transient cavity of the large southern Serenitatis Basin, which was formed about 3.9 to 4.0 b.y. ago by the impact of a planetesimal. The target rocks, predominantly of the dunite-anorthosite-norite-troctolite suite or its metamorphosed equivalents, were fractured, sheared, crushed, and/or melted by the impact. The resulting mixture of crushed rock and melt was transported up and out of the transient cavity and deposited on and beyond its rim. Hot fragmental to partially molten ejecta and relatively cool cataclastic and relict target rocks were intermixed in a melange of lenses, pods, and veins. Crystallization of melts and thermal metamorphism of fine-grained fragmental debris produced breccia composed of rock and mineral fragments in a fine-grained, coherent, crystalline matrix. Such breccia dominates the massif samples.

High-angle faults that bound the massifs were activated during the formation of the basin, so that structural relief of several kilometers was imposed on the ejecta almost as soon as it was deposited. Massive slumping that produced thick wedges of colluvium on the lower massif slopes probably occurred nearly contemporaneously with the faulting. Material of the Sculptured Hills, probably largely cataclastic excavated from the southern Serenitatis Basin by the same impact, was then deposited on and around the massifs.

Basalt, estimated to be about 1400 m thick at the landing site, flooded the Taurus-Littrow graben before approximately 3.7 b.y. ago. The basalt (subfloor basalt) is part of a more extensive unit that was broadly warped and cut by extensional faults before the accumulation in Mare Serenitatis of younger, less deformed basalts that overlap it. A thin volcanic ash unit (dark mantle), probably about 3.5 b.y. old, mantled the subfloor basalt and the nearby highlands. It, too, was subsequently overlapped by the younger basalt of Mare Serenitatis.

In the time since the deposition of the volcanic ash, continued bombardment by primary and secondary projectiles has produced regolith, which is a mechanical mixture of debris derived mainly from the subfloor basalt, the volcanic ash, and the rocks of the nearby massif and Sculptured Hills. The regolith and the underlying volcanic ash form an unconsolidated surficial deposit with an average thickness of about 14 m, sufficiently thick to permit abnormally rapid degradation of the smaller craters, especially those less than 200 m in diameter, so as to create a surface that appears less cratered than the other mare surfaces. Admixed volcanic ash gives the surface a distinctive dark color, which, in combination with the less cratered appearance, led to its interpretation before the mission as a young dark mantling unit.

The uppermost part of the regolith over much of the landing area is basalt-rich ejecta from the clustered craters of the valley floor. Most of the valley floor craters are interpreted as part of a secondary cluster formed by projectiles of ejecta from Tycho, 2200 km to the southwest. When they struck the face of the South Massif, the projectiles mobilized fine-grained regolith material that was deposited on the valley floor as the light mantle. Exposure ages suggest that the swarm of secondary projectiles struck the Taurus-Littrow area about 100 m.y. ago.

The Lee-Lincoln fault scarp is part of an extensive system of wrinkle ridges and scarps that transect both mare and highland rocks.

The scarp cuts Lara Crater, but the major part of the displacement occurred before the deposition of the light mantle. Small extensional faults cut the surface of the light mantle west of the Lee-Lincoln scarp.

References: [1] Wolfe E. W. et al. (1981) *U.S. Geol. Surv. Prof.*

Paper 1080, 280 pp.

THE APOLLO 17 MARE BASALTS: SERENELY SAMPLING TAURUS-LITTROW. Clive R. Neal¹ and Lawrence A. Taylor², ¹Department of Civil Engineering and Geological Sciences, University of Notre Dame, Notre Dame IN 46556, USA, ²Department of Geological Sciences, University of Tennessee, Knoxville TN 37996, USA.

As we are all aware, the Apollo 17 mission marked the final manned lunar landing of the Apollo program. The lunar module (LM) landed approximately 0.7 km due east of Camelot Crater in the Taurus-Littrow region on the southwestern edge of Mare Serenitatis [1]. Three extravehicular activities (EVAs) were performed, the first concentrating around the LM and including station 1 approximately 1.1 km south-southeast of the LM at the northwestern edge of Steno Crater [1]. The second traversed approximately 8 km west of the LM to include stations 2, 3, 4, and 5, and the third EVA traversed approximately 4.5 km to the northwest of the LM to include stations 6, 7, 8, and 9. This final manned mission returned the largest quantity of lunar rock samples, 110.5 kg/243.7 lb, and included soils, breccias, highland samples, and mare basalts. This abstract concentrates upon the Apollo 17 mare basalt samples.

One hundred and fifty-six basaltic samples were returned weighing 32.19 kg, or approximately one-third the total weight of Apollo 17 samples. The majority of Apollo 17 mare basalts were found at station 1A (75 samples), 22 from station 0, 19 from station 8, 11 from station 5, 10 from station 6, and 9, 4, 3, 2, and 1 from stations 4, 9, 7, 2, and 3 respectively. Note that these statistics include rake samples, but do not include basaltic clasts from breccia samples (e.g., [2]). Practically all these samples have been studied to various degrees, whether it be just a thin-section cut, to rare-gas, magnetic, isotope, and whole-rock analyses.

Petrographic Studies: The Apollo 17 mare basalts were divided into a three-fold classification on the basis of petrography. The petrographic divisions were derived independently by three different studies [3-5] and the differences between the groups defined by each petrographic study are identical. The petrographic groups are

1. *Type 1A: Olivine porphyritic ilmenite basalts.* These are usually fine-grained (typically <1 mm) with a general subvolcanic to variolitic texture, or vitrophyric (e.g., 71157). All type 1A basalts contain olivine and ilmenite phenocrysts (e.g., 71048). Armalcolite, where present, forms cores to ilmenite, is partially rimmed by ilmenite, or is rarely present as discrete grains (up to 0.2 mm; e.g., 71097). Chromite-ulvöspinel is present either as discrete grains (0.1-0.3 mm) or as inclusions in olivine (<0.1 mm). All type 1A basalts contain ilmenites with sawtooth margins, indicative of rapid crystallization (e.g., [3]). Pink pyroxene prisms (≈0.4 mm) and plagioclase laths (≈0.3 mm) are interstitial, sometimes combining to form "bowtie" textures.

2. *Type 1B: Plagioclase-poikilitic ilmenite basalts.* All type 1B basalts are coarse grained (>2 mm), with rare medium-grained examples (e.g., 74287). However, all are olivine poor. In all cases, ilmenite has exsolved both ulvöspinel and rutile, which are present as thin (<0.05 mm wide) lamellae within ilmenite. Armalcolite and discrete chromite-ulvöspinel (0.2-0.5 mm and 0.1-0.2 mm respectively), where present, are inclusions in olivine, pyroxene, and/or

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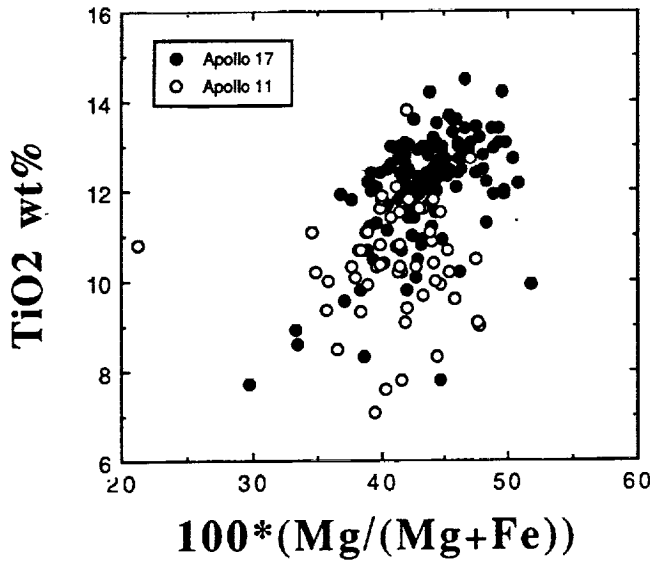


Fig. 1.

plagioclase. Some armalcolites possess a discontinuous ilmenite overgrowth. Where olivine is present, it forms cores to larger pyroxenes or small (up to 0.1 mm) discrete inclusions in plagioclase. It is likely that where olivine is absent, it has been totally resorbed and replaced by pyroxene (e.g., 71047). The petrographic relationships within the type 1B basalts indicate that clinopyroxene (e.g., titanaugite) after olivine. Type 1B basalts commonly contain interstitial SiO_2 , demonstrating the fractionated nature of these basalts.

Type II: Apollo 11, low-K-type basalts. These are olivine-free and have medium-grained (generally 1.5 mm), subophitic textures, similar to the low-K Apollo 11 high-Ti basalts. Subhedral ilmenite is practically the only opaque phase present; it exhibits smooth grain boundaries and contains ulvöspinel and rutile exsolution lamellae (<0.05 mm wide). Interstitial SiO_2 is present (up to 0.6 mm). Unlike the previous groups, type II basalts contain no low-Ca pyroxene.

With regard to mineral species, there is a decrease in modal olivine from type 1A through type II. Ilmenite, armalcolite, and chromite-ulvöspinel tend to be present in similar proportions in both type 1A and type 1B, but type II basalts contain lower modal abundances of these elements. As expected, type 1B basalts generally have higher plagioclase abundances than type 1A; type II basalts have similar plagioclase abundances to type 1B.

Very-low-Ti Basalts: Wentworth et al. [8] described the petrography of Apollo 17 VLT basalts as being granular to subophitic. They consist primarily of pyroxene and plagioclase, with minor olivine and Al-chromite. Pyroxenes range from pigeonite to subcalcic augite, but the majority have pigeonitic compositions. Plagioclase ranges from An_{98} to An_{93} and olivine has a variable composition ($\approx \text{Fo}_{75-58}$).

Whole-Rock Chemistry: Like the Apollo 11 counterparts, 99% of Apollo 17 basalts were originally classified as high Ti [1,7,8], in that they contain >6 wt% TiO_2 , and the remaining 1% are of the very-low-Ti (VLT) subdivision (e.g., [9,10]). However, Apollo 17 high-Ti basalts contain a higher TiO_2 content for a given MG# (Fig. 1), suggesting a higher proportion of Ti-rich minerals in the source region for Apollo 17 high-Ti mare basalts.

Early studies of whole-rock chemistry (e.g., [5,11,12]) subdivided Apollo 17 high-Ti basalts into three groups—A, B, and C. Rhodes et

al. [12] and Warner et al. [11] recognized that data from coarse-grained basalts appeared to plot as clouds about the arrays defined by the fine-grained samples. This scatter was attributed to unrepresentative sampling and were consequently classified as "class U" basalts. It was noted immediately that the three petrographic types did not correspond to the three chemical groups—all three of the former are present in each of the latter.

Apollo 17 type A basalts contain 50–60% higher incompatible trace-element abundances than do the type B variants (Fig. 2), although both possess similar major-element compositions. Neal et al. [13] demonstrated that the type B division contained two subdivisions (B1 and B2) on the basis of REE abundances (Fig. 3). Type C basalts contain higher MgO and Cr_2O_3 abundances than types A, B1, and B2, but are most notable in their enrichment of Rb over the other Apollo 17 high-Ti mare basalts (Fig. 4). The type C basalts have similar incompatible trace-element abundances to the type As. Also, type C basalts contain olivines of the highest Fo content (Fo_{80}) found at Apollo 17, but only five examples have been described [11,12,14], all from station 4 (Shorty Crater). Ryder [15,16] reported the chemistry and petrography of a sample that may be a new variant of Apollo 17 high-Ti basalt (70091,2161), which he classified as type D. This basalt has lower REE abundances, a deeper Eu anomaly (Fig. 2), higher MgO and Cr_2O_3 abundances, and lower CaO and TiO_2 contents than other Apollo 17 high-Ti basalts.

Apollo 17 VLT basalts contain <1 wt% TiO_2 and were first recognized in thin sections from the Apollo 17 deep drill core [9,17], but have since been found in impact melt rocks [18,19] and as basaltic clasts in breccia 73255 and 72235 [2]. However, understanding these basalts is difficult due to small sample size—only a few milligrams of sample have been analyzed for whole-rock chemistry.

The four samples that have been analyzed demonstrate that the VLT basalts contain lower REE abundances than their high-Ti counterparts. In fact, the VLT basalts generally contain lower incompatible trace-element abundances than the high-Ti varieties, but do have slightly higher MG#s (Fig. 1). However, the inherently small sample sizes of these VLT basalts means that interpretation of whole-rock chemistry must proceed with caution.

Isotope Chemistry and Age Dating: Thirty-two published Rb-Sr, Sm-Nd, or ^{40}Ar - ^{39}Ar ages on 17 different Apollo 17 high-Ti basalts have been determined over a period from 1973 to present. Ages range from 3.64 Ga to 3.84 Ga, with the eruption ages of each group

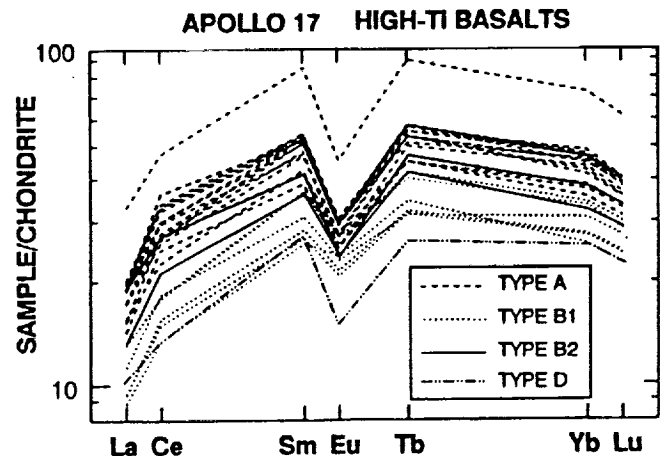


Fig. 2.

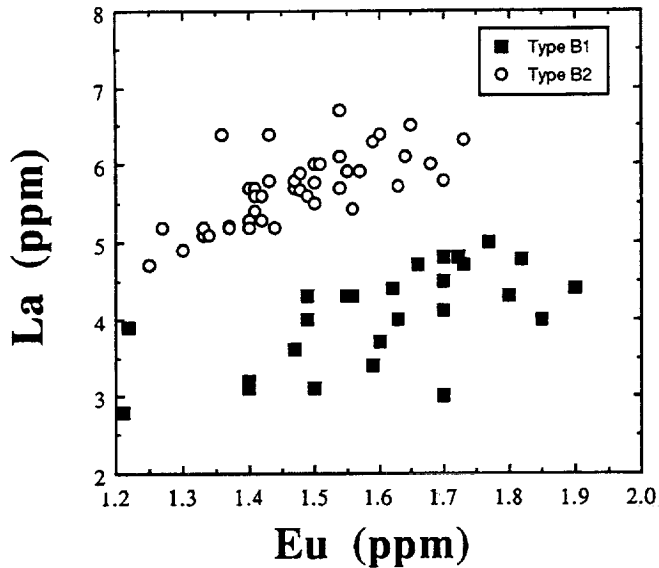


Fig. 3.

being indistinguishable from each other. However, Paces et al. [20] suggested that type A basalts were slightly older (3.75 ± 0.02 Ga) than type B (3.69 ± 0.02 Ga) on the basis of weighted averages of available data. The available age data for type C basalts overlap these two groups and types B1 and B2 are indistinguishable from each other.

An interesting feature of the Apollo 17 high-Ti basalts is that I(Nd) ratios are indistinguishable between the various groups, yet I(Sr) ratios are resolvable between types A, B1, and C [20]. Type B2 basalts appear to have evolved by open-system processes as they exhibit a range in both I(Sr) and I(Nd) [20]. Nyquist et al. [21,22] argued that the range in Rb/Sr between type B (lowest), through type A, to type C (highest) was due to varying degrees of partial melting of slightly heterogeneous sources, with retention of clinopyroxene in the residue. Paces et al. [20] concluded that the Rb/Sr and Sr isotopic heterogeneities were formed at 4.1 Ga by metasomatic processes and clinopyroxene was exhausted during Apollo 17 high-Ti magma genesis. Whatever the process, it is clear that a three-stage model is required to explain the Rb-Sr systematics.

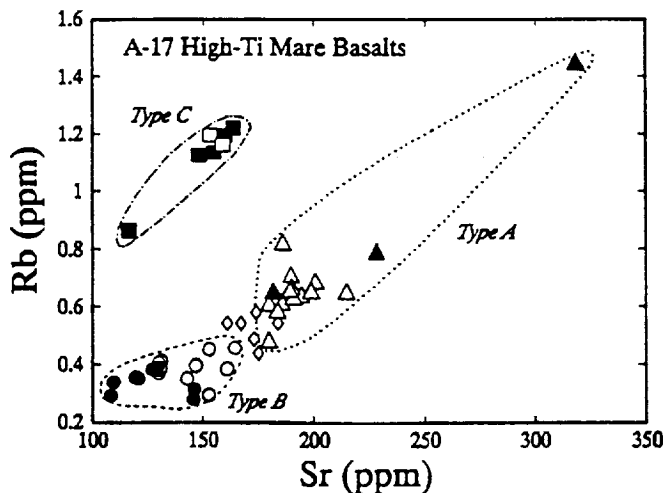


Fig. 4.

Petrogenesis: Rhodes et al. [12] and Warner et al. [11] demonstrated that the type A basalts evolved by closed-system fractional crystallization and proposed that, due to variations of La/Sm ratio, type B basalts could be modeled by varying the degree of partial melting of the source. With the identification of type B1 and B2 basalts [13], it was recognized that on the basis of major and trace elements, both groups could be generated by closed-system fractional crystallization. However, the isotopic study of Paces et al. [20] demonstrated the validity of such a model for the type B1 high-Ti basalts, but identified open-system behavior of the type B2s. Neal et al. [13] also successfully quantified a closed-system fractional crystallization model to the type A basalts. Only five basalts comprise the type C grouping, making quantitative petrogenetic interpretation impossible. Also, the small number of samples and small sample size of the type D and VLT basalts preclude definitive petrogenetic interpretation.

Source Modeling: Apollo 17 high-Ti mare basalts. Source models proposed for these basalts can be divided into those involving (1) early-stage LMO cumulates, (2) late-stage LMO cumulates, and (3) hybridized sources. Isotopic and experimental studies have demonstrated the presence of ilmenite and clinopyroxene in the source region of high-Ti basalts (late-stage LMO cumulates), and the relatively high MG# of these basalts (Fig. 1) requires olivine and/or orthopyroxene (early LMO cumulates). Hughes et al. [23] suggested convective overturn of the LMO cumulate pile in order to account for the chemical composition of the Apollo 17 high-Ti basalt source. Spera [24] stated that LMO-wide overturn was possible, but Snyder et al. [25] considered that the overturn may have been more localized. There are two points of contention about the high-Ti source: (1) Is ilmenite exhausted during partial melting? (2) Is plagioclase in the source?

Apollo 17 VLT mare basalts. Source modeling of the VLT basalts was undertaken by Wentworth et al. [10], who demonstrated that the observed VLT composition could be produced by a 1–2% partial melting of a 90% olivine, 10% orthopyroxene cumulate, assuming that this source crystallized from a LMO with a flat REE profile. These authors further suggested that if this cumulate source formed from a magma with a fractionated REE pattern [e.g., $(La/Lu)_N \approx 2$], the VLT basalt REE pattern could be generated by $\approx 4\%$ partial melting of 70% olivine, 30% orthopyroxene source mineralogy.

Summary: Two types of mare basalt were returned from the Apollo 17 site: (1) the commonly found high-Ti variety and (2) the rare VLT basalts. The high-Ti basalts were most probably derived from a heterogeneous hybrid source of early and late-stage LMO cumulates. Four, possibly five, groups of basalts were erupted. At least two of these groups evolved through closed-system fractional crystallization and at least one group underwent open-system evolution. Types A, B1, and B2 appear to be present from all over the Taurus-Littrow site, whereas type Cs are only found at station 4 (Shorty Crater) and the extent of the type Ds is not known.

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[12] Rhodes J. M. et al. (1976) *Proc. LSC 7th*, 1467–1489. [13] Neal C. R. et al. (1990) *GCA*, 54, 1817–1833. [14] Neal C. R. et al. (1990) *LPSC XXI*, 855–856. [15] Ryder G. (1988) *Eos*, 69, 292. [16] Ryder G. (1990) *Meteoritics*, 25, 249–258. [17] Taylor G. J. et al. (1977) *GRL*, 4, 207–210. [18] Warner R. D. et al. (1978) *Proc. LPSC 9th*, 547–564. [19] Taylor G. J. et al. (1978) *Mare Crisium: The View from Lunar 24*, 357–370. [20] Paces J. B. et al. (1991) *GCA*, 55, 2025–2043. [21] Nyquist L. E. et al. (1976) *Proc. LSC 7th*, 1507–1528. [22] Nyquist L. E. et al. (1979) *Proc. LPSC 10th*, 77–114. [23] Hughes S. S. et al. (1989) *Proc. LPSC 19th*, 175–188. [24] Spera F. J. (1992) *GCA*, 56, 2253–2265. [25] Snyder G. A. et al. (1992) *GCA*, in press.

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USING APOLLO 17 HIGH-TI MARE BASALTS AS WINDOWS TO THE LUNAR MANTLE. Clive R. Neal¹ and Lawrence A. Taylor², ¹Department of Civil Engineering and Geological Sciences, University of Notre Dame, Notre Dame IN 46556, USA, ²Department of Geological Sciences, University of Tennessee, Knoxville TN 37996, USA.

Detailed study of mare basalts and volcanic glasses is critical in our understanding of the lunar mantle because, as yet, no mantle xenoliths have been recognized in the current collection. Primitive endmember, volcanic lunar glass compositions can be considered primary mantle melts because of their glassy, uncrystallized nature, and analyses of such samples yielded important information regarding the mantle source (e.g., [1–3]). The fact that no relationship between the glasses and the crystalline mare basalts has been established suggests that the basalts were derived from a separate source and therefore have the potential to yield further critical information regarding the lunar mantle. By understanding the processes that have occurred during post-magma-generation evolution, parental or possibly primary compositions can be highlighted for further study, possibly leading to source evaluation. However, where it is demonstrable that the composition chosen as the parental/“primary” melt has experienced some post-magma-generation evolution (i.e., a vitrophyre containing olivine/ilmenite/chromite-ulvöspinel phenocrysts), this composition cannot be used for direct calculation of a mantle source. In this case, incompatible trace-element ratios can be utilized to remove the effects of fractional crystallization and allow an evaluation of the source region (e.g., [4,5]).

Apollo 17 Source Modeling: The source composition and location of Apollo 17 high-Ti mare basalts has been under debate for many years and several models have been proposed. The one unifying theme of 99% of these models is that they require a mafic, cumulate mantle source. This is where the similarity ends! Models for titaniferous basalt petrogenesis include those requiring early LMO cumulates of olivine and opx [6], late-stage cumulates [4,7–9], to those requiring a mixture of early and late-stage cumulates [10–15] or hybrid sources [3,16]. Experimental evidence puts the depth of origin of the Apollo 17 high-Ti mare basalts at between 200 and 400 km ([17]; source dominated by olivine and low-Ca pyroxene with ilmenite) to ≤150 km [10–12]. The degree of partial melting ranges from 5% to 20%. It is generally agreed that a Fe-Ti oxide mineral is required in the source of Apollo 17 high-Ti mare basalts.

Determination of the depth of origin is critical in understanding the composition, stratigraphy, and dynamics of the lunar mantle. For example, at depths exceeding 300–350 km within the Moon, the source for the high-Ti basalts would be eclogitic [18] and such a source would not yield the inherently LREE-depleted signature of the Apollo 17 high-Ti mare basalts. Therefore, depths must be shallower than 300 km. Furthermore, there are two points of contention in

defining the source regions of Apollo 17 high-Ti mare basalts: (1) Was ilmenite exhausted or retained in the residue? (2) Was plagioclase present in the source? Shih et al. [13] proposed a source composition of olivine + clinopyroxene + ilmenite ± plagioclase, all of which were retained in the source. On the basis of experimental evidence, Walker et al. [10,11], suggest that ilmenite remains in the residue, and plagioclase, if present, would be exhausted; conversely, Green et al. [17] demonstrate that ilmenite would be exhausted and suggests plagioclase was never present. Hughes et al. [3] conclude that both ilmenite and plagioclase are exhausted from their modeled source composition. Clearly, the nature of the Apollo 17 high-Ti mare basalt source region composition is far from resolved, and the situation is complicated by the recognition of four, possibly five, separate magma types: A, B1, B2, C, and D [19–23].

Recognizing the Apollo 17 High-Ti Mare Basalt Source Region Components: The purpose of this abstract is to use incompatible trace-element ratios to evaluate the nature of the source regions of Apollo 17 type A, B1, B2, and possibly C high-Ti basalts in an attempt to answer the two questions posed above. A database of practically all analyzed Apollo 17 basalts was assembled and the nature of any potential fractionating assemblage evaluated using petrographic observations, coupled with the MAGMAFOX program of Longhi [24]. Samples chosen for processing by the MAGMAFOX program were either fine grained or vitrophyric. The results demonstrate a fractionating sequence dominated by chromite-ulvöspinel, olivine, ilmenite, and pyroxene (low and high Ca). Plagioclase becomes a liquidus phase only after ≈55% crystallization. As demonstrated by Neal et al. [21], over 98% of all types A, B1, and B2 can be modeled by 40% closed-system fractional crystallization of olivine, chromite-ulvöspinel, armalcolite, ilmenite, and augite. This assemblage is generally supported by the MAGMAFOX program, except for armalcolite (calculated as ilmenite), and the program results also include pigeonite. Therefore, because plagioclase is not part of the fractionating assemblage, Sm and Eu can be considered as incompatible elements, as are La, Yb, and Hf. Although ilmenite is fractionating, the Hf crystal/liquid partition coefficient for ilmenite is ≈0.4 [25]. As part of this study, the Yb/Hf ratio is used because although both are incompatible (i.e., $K_d < 1$), these elements will have approximately the same partition coefficient value in the fractionating assemblage described above. Significant changes in this ratio will

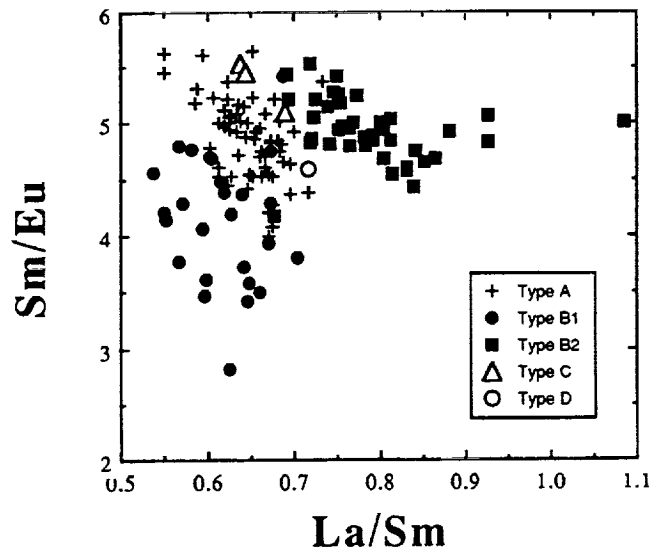


Fig. 1.

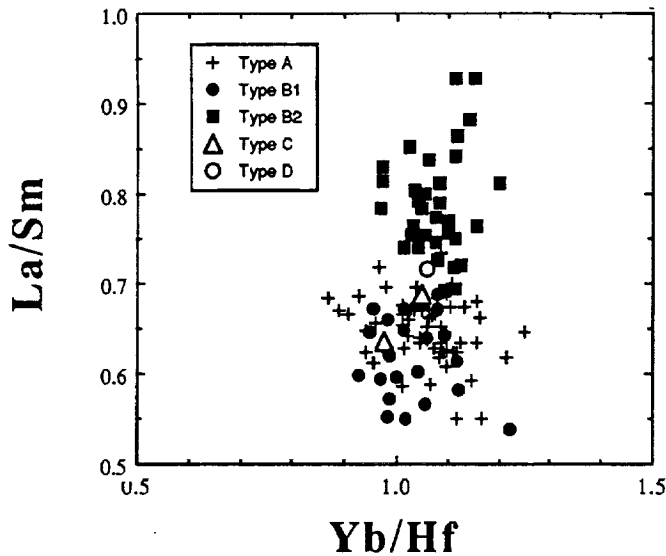


Fig. 2.

signify preferential exhaustion of ilmenite or pyroxene in the source. Other incompatible trace elements, such as Ba, Zr, Nb, etc., have not been used because they have only been analyzed for a few samples in the database.

Results: This study uses three ratios to (1) track the degree of partial melting (La/Sm, which will decrease with increasing degrees of partial melt); (2) examine the role of plagioclase (Sm/Eu); and (3) examine the role of ilmenite (Yb/Hf). Figure 1 is a plot of Sm/Eu against La/Sm. Types A, B1, and B2 all form parallel to subparallel negative correlations. In Fig. 2, Sm/Eu is plotted against Yb/Hf with each group of basalts exhibiting a large range of Sm/Eu and a limited range of Yb/Hf. In Fig. 3, La/Sm is plotted against Yb/Hf. The type B2 basalts have the highest La/Sm ratios, but all Apollo 17 high-Ti mare basalts contain similar Yb/Hf ratios between 0.9 and 1.2.

Discussion: The parallel negative correlations observed in Fig. 1 are interpreted as representing the presence of plagioclase in the

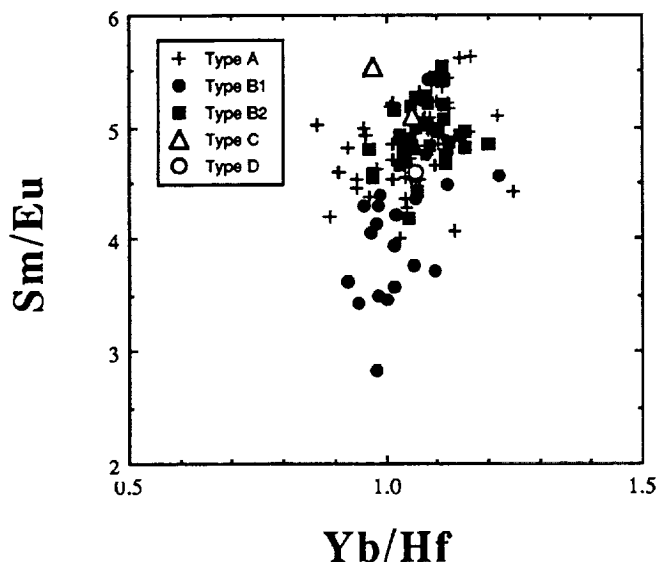


Fig. 3.

source that becomes exhausted with increasing degrees of partial melt. As plagioclase is not part of the fractionating sequence, the range in Sm/Eu can only be accounted for by plagioclase in the source that does not remain in the residue (à la [11]). The type B2 basalts show a decreasing La/Sm ratio at constant Sm/Eu followed by increasing Sm/Eu with decreasing La/Sm. It is tempting to conclude that this trend documents the point at which plagioclase is exhausted in the B2 source. However, Paces et al. [26] demonstrated the open-system behavior of the type B2 basalts; such interpretation of this trend must be undertaken with a degree of caution. This is also supported by Fig. 2, but the restricted range of Yb/Hf suggests that neither pyroxene nor ilmenite is exhausted during partial melting. The small scatter of Yb/Hf is probably due to ilmenite becoming a liquidus phase before pyroxene during crystal fractionation. If ilmenite was exhausted in the source relative to pyroxene, the range in the Yb/Hf ratio would be dramatic (see inset on Fig. 2). Figure 3 affirms the conclusions derived from Figs. 1 and 2, with a range of Sm/Eu and a restricted range in Yb/Hf. The type B2 basalts have higher La/Sm and Yb/Hf ratios relative to the other Apollo 17 high-Ti basalt types, which may be a result of open-system behavior (KREEP/KREEP residue assimilation?).

Summary: The Apollo 17 high-Ti mare basalts are derived from source regions containing plagioclase that was not retained in the residue. Ilmenite appears to remain as a residual phase, but plagioclase is exhausted. The open-system behavior of the type B2 basalts results in slightly higher Yb/Hf and La/Sm ratios. The nature of the added component is not clear, but may be a KREEP derivative or residue. The recognition of plagioclase in the source(s) of these basalts suggests that the location of the source region(s) would be more likely to be <150 km (i.e., closer to the plagioclase-rich crust), which would allow incorporation of plagioclase into the source through incomplete separation of crustal feldspar, supporting the conclusion of [15].

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p. 3

LITHOLOGIES CONTRIBUTING TO THE CLAST POPULATION IN APOLLO 17 LKFM BASALTIC IMPACT MELTS. Marc D. Norman¹, G. Jeffrey Taylor¹, Paul Spudis², and Graham Ryder², ¹Planetary Geosciences, Department of Geology and Geophysics, University of Hawaii, Honolulu HI 96822, USA, ²Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston TX 77058, USA.

LKFM basaltic impact melts are abundant among Apollo lunar samples, especially those from Apollo 15, 16, and 17. They are generally basaltic in composition, but are found exclusively as impact melts [1]. They seem to be related to basins and so could represent the composition of the lower lunar crust [2-4]. They contain lithic clasts that cannot be mixed in any proportion to produce the composition of the melt matrix; components rich in transition elements (Ti, Cr, Sc) and REE are not accounted for [5]. To search for the mysterious cryptic component, we previously investigated the mineral clast population in two Apollo 14 LKFM basaltic impact melts, 15445 and 15455 [6]. The cryptic component was not present in the mineral clast assemblage of these breccias either, but some olivine and pyroxene grains appeared to be from lithologies not represented among identified igneous rocks from the lunar highlands. In addition, none of the mineral clasts could be unambiguously assigned to a ferroan anorthosite source. We have now extended this study to Apollo 17, starting with two LKFM impact melt breccias (76295 and 76315) from the Apollo 17 station 6 boulder. The results show that mineral clasts in these rocks are dominated by Mg-suite sources, but that once again

lithologies not recognized as pristine rocks are represented in the mineral clasts. The compositions and relative abundances of mineral clasts suggest a significant contribution from olivine gabbronorites similar to Iherzolite 67667. Little, if any, ferroan anorthosite was detected. The cryptic component remains enigmatic.

Rocks 76295 and 76315 are fine-grained impact melt breccias that display little reaction between clasts and the melt matrices [7]. We analyzed virtually all olivine, pyroxene, and plagioclase clasts larger than 50 μm in one thin section of each rock. Data were obtained with a new Cameca SX50 electron microprobe using procedures designed to obtain precise analyses for minor elements. The two rocks differ in the relative abundances of pyroxene to olivine, mafics to plagioclase, and mineral compositions (Fig. 1); these results are generally consistent with published data [7]. For both rocks, plagioclase ranges from An72 to An97 with a mode around An94-96 and a tail to sodic values (Fig. 1). Plagioclase ≥ An95 is abundant in 76315 and comparatively uncommon in 76295. Olivine and pyroxene clasts have diverse compositions (Figs. 1-3). 76295 contains less magnesian olivine than does 76315, and the olivine/pyroxene ratio is 0.85 in 76295 compared to 5.2 in 76315. In both samples, the ratio of high-Ca pyroxene to low-Ca pyroxene is surprisingly high, about 0.5.

Minor-element abundances in both olivine and pyroxene (Figs. 2 and 3) are unlike those found in rocks of the ferroan anorthosite suite (FAS), suggesting at most a small contribution from FAS rocks. Ferroan noritic anorthosites such as those found in 67215 and 67016 [9,10] are especially ruled out; any ferroan anorthosite component in the clast population must have been nearly pure anorthosite contrib-

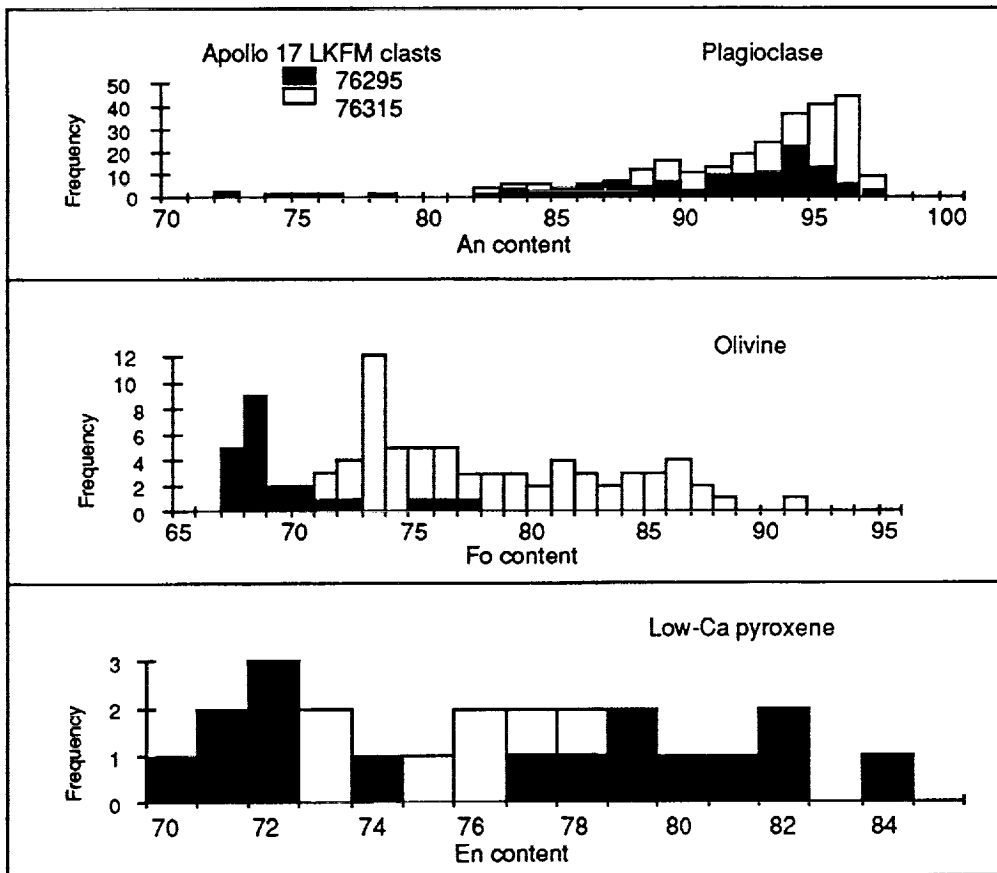


Fig. 1. Histograms of plagioclase, olivine, and low-Ca pyroxene clast compositions in 76295 and 76315.

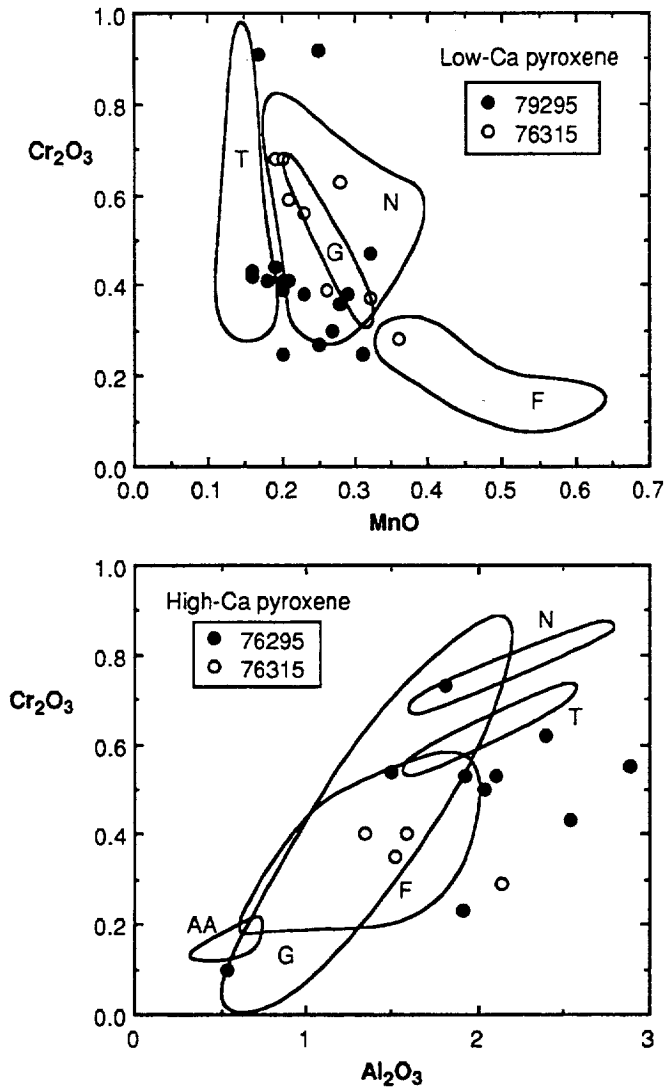


Fig. 2. Minor-element abundances in pyroxene clasts from 76295 and 76315 compared to fields representing compositions observed in pristine highland rocks (data from [13,11]). F = ferroan anorthosite suite, T = Mg-suite troctolites, N = Mg-suite norites, G = gabbronorites, AA = alkali anorthosites.

using only plagioclase. Low-Ca pyroxenes in both 76295 and 76315 have an affinity with Mg-suite norites or gabbronorites, whereas several of the high-Ca pyroxene clasts have minor-element compositions outside the known ranges for pristine highland rocks. The abundance of high-Ca pyroxene clasts, together with their unusual compositions, suggests the presence of poorly sampled gabbroic lithologies in the target stratigraphy of these Apollo 17 LKFM impact melts. That gabbroic rock types may be more abundant in the lunar highlands than is apparent in the sample collection has been suggested by remote sensing studies [8]. Many of the olivine clasts have Cr and Ti contents considerably higher than those observed in pristine highland rocks, including 67667, which has an average olivine composition with 0.049% TiO_2 and 0.13% Cr_2O_3 .

Plagioclase compositions are ambiguous with regard to distinguishing their source lithologies, but considering that mafic phases tend to be dissolved preferentially when incorporated into superheated impact melts, the relatively high abundance of olivine and pyroxene suggests that the clast assemblage is dominated by Mg-suite

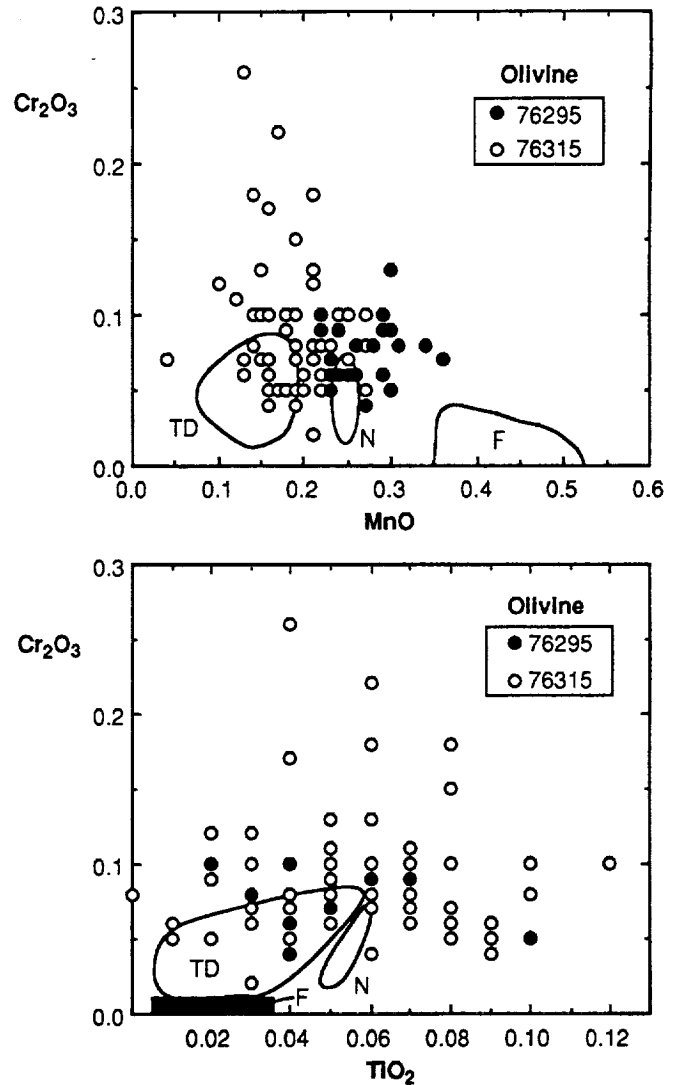


Fig. 3. Minor-element abundances in olivine clasts from 76295 and 76315 compared to fields representing compositions observed in pristine highland rocks. Source of pristine rock data and abbreviations same as Fig. 2 except that TD = troctolites and dunite.

rocks. The abundance of high-Ca pyroxene combined with the relatively Fe-rich olivine and pyroxene compositions and sodic plagioclase suggest the presence of one or more evolved olivine gabbronorite components with overall characteristics generally similar to ilherzolite 67667, but with distinct minor-element abundances in the mafic phases. Although some minor-element abundances are outside the ranges measured for minerals in identified pristine rocks, none have compositions suitable for the cryptic component rich in transition metals.

The clast populations in the two Apollo 17 LKFM melt rocks studied so far are dominated by Mg-suite norites, troctolites, and gabbronorites. Rock 76295 appears to contain a greater proportion of the norite or gabbronorite component whereas mineral clasts in 76315 were derived predominantly from troctolites. Differences in the proportions of clast types is not unusual for impact melt breccias from the same crater [12]. A dearth of ferroan anorthosite was also found in Apollo 15 LKFM breccias [6] and suggests either (1) it was never present in the basin target, (2) the FAS crust was removed before the

Serenitatis Basin impact event either by impact erosion or by assimilation during intense magmatic activity, or (3) impact dynamics during basin formation prevented upper crustal rocks from being included in the excavated impact melt.

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IMPACT GLASSES FROM THE <20- μ m FRACTION OF APOLLO 17 SOILS 72501 AND 78221. John A. Norris¹, Lindsay P. Keller², and David S. McKay², ¹Department of Geology, University of Georgia, Athens GA 30602, USA, ²Code SN, NASA Johnson Space Center, Houston TX 77058, USA.

Introduction: The chemical compositions of microscopic glasses produced during meteoroid impacts on the lunar surface provide information regarding the various fractionation processes that accompany these events. To learn more about these fractionation processes, we studied the compositions of submicrometer glass spheres from two Apollo 17 sampling sites using electron microscopy. The majority of the analyzed glasses show evidence for varying degrees of impact-induced chemical fractionation. Among these are HASP glasses (high-Al, Si-poor), which are believed to represent the refractory residuum left after the loss of volatile elements (e.g., Si, Fe, Na) from the precursor material [1–3]. In addition to HASP-type glasses, we also observed a group of VRAP glasses (volatile-rich, Al-poor) that represent condensates of vaporized volatile constituents, and are complementary to the HASP compositions [3]. High-Ti glasses were also found during the course of this study, and are documented here for the first time.

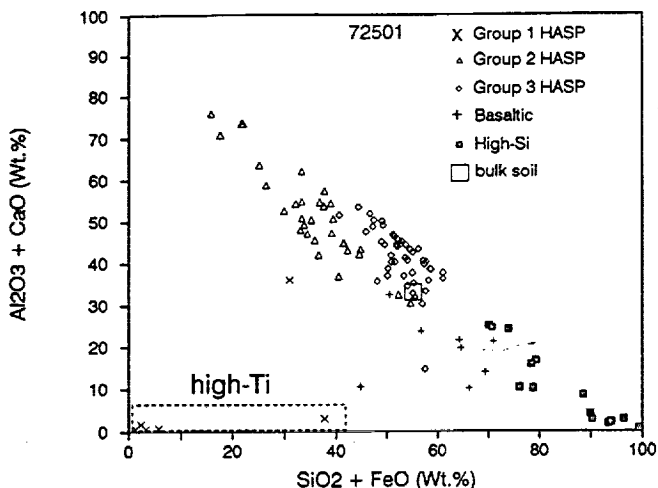


Fig. 1. Compositions of submicrometer glasses in soil 72501.

TABLE 1. Average EDX analyses for subgroups of impact glasses in soils 72501 and 78221.

	1a	1b	2	3	4	5*
MgO	7.0	4.0	9.0	6.0	14.0	3.1
Al ₂ O ₃	53.0	4.0	34.0	27.0	14.0	5.2
SiO ₂	10.0	7.5	30.0	49.0	46.0	65.0
CaO	23.0	1.5	18.0	11.0	9.0	2.1
TiO ₂	3.3	64.0	3.0	1.0	3.0	0.6
FeO	3.7	19.0	6.0	6.0	14.0	20.0

* High-Si glasses contain <2 wt.% each of Na₂O, K₂O, SO₃, P₂O₅, MnO, and Cr₂O₃.

1a = group 1 HASP; 1b = group 1, high-Ti; 2 = group 2 HASP; 3 = group 3 HASP; 4 = basaltic glasses, and 5 = high-Si glasses.

Experimental: Samples from the <20- μ m size fractions of two Apollo 17 soil samples (72501 from station 2 at the base of the South Massif, and 78221 from station 8 at the base of the Sculptured Hills) were embedded in low viscosity epoxy and cut into thin sections (~80–100 nm thick) using diamond-knife ultramicrotomy. The thin sections were analyzed using a PGT energy-dispersive X-ray (EDX) spectrometer with a JEOL 100CX TEM. EDX analyses were obtained for 107 spheres from 72501 and 115 from 78221. The apparent diameters of these spheres in thin section were typically between 100 and 400 nm, although the true diameter of any actual sphere may have been slightly larger. The relative errors associated with the EDX analyses were estimated by analyzing a grossular standard. The relative errors for Al, Si, Ca, and Fe are ~5%. These relative errors increase significantly for concentrations <5 wt%.

The glass compositions were initially divided into a "high-Si" group (SiO₂ > 60 wt%) or a "low-Si" group (SiO₂ < 60 wt%). For the "low-Si" compositions a standard CIPW norm was calculated. However, many of these glasses (e.g., HASP) contain insufficient Si to be used with this method. For these Si-poor compositions, a new normalization scheme was developed using three groups of progressively Si-deficient normative minerals. Group 1 HASP compositions contain insufficient Si to calculate any normative anorthite (AN). Instead, the group 1 normative mineralogy includes gehlenite (GH) + spinel (SP) + Ca aluminates. Group 2 HASP compositions consist of normative GH + AN + SP + olivine + Ca aluminates. The group 3

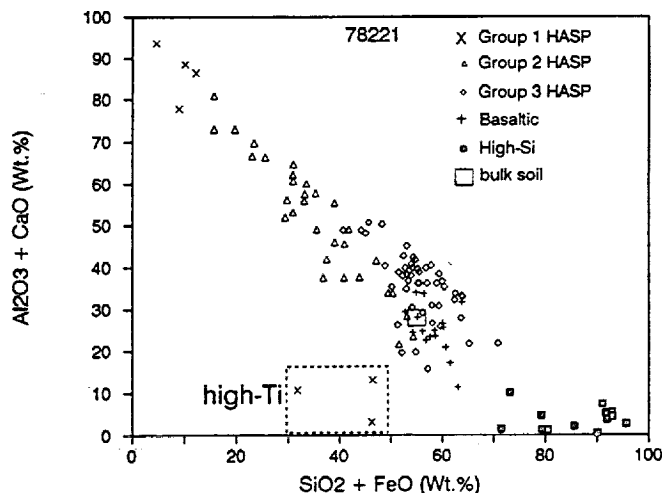


Fig. 2. Compositions of submicrometer glasses in soil 78221.

HASP compositions are cordierite- or mullite-normative, and contain excess Al_2O_3 and SiO_2 after using all of the Ca, Na, and K to make normative feldspar.

Results and Discussion: The compositions of the 107 glasses analyzed from sample 72501 and the 115 from 78221 are plotted in Figs. 1 and 2. The compositions range from nearly 95% refractory components to others composed entirely of more volatile components. HASP compositions comprise ~75% of the total glasses analyzed in each sample. The group 1 HASP glasses comprise ~6% of the total analyses from both 72501 and 78221. The group 1 HASP compositions are the most refractory and the most volatile-depleted of all the analyzed glasses. A cluster of high-Ti glasses were included within group 1 and plot about the origin in Fig. 1. The group 1 high-Ti glasses contain little Ca or Al, but nonetheless have undergone extreme volatile loss in the form of nearly complete removal of Fe from what was once ilmenite (Table 1). The other group 1 members are probably derived from mostly anorthositic material that has lost most of its original SiO_2 content. All but one of the 72501 group 1 HASPs are high-Ti glasses, whereas 78221 contains a larger proportion of the Ca- and Al-rich members of this group (Table 1).

The group 2 HASP glasses span a much wider compositional range in both samples, and comprise 29% of the analyses from 72501 and 27% of those from 78221. These have undergone a lesser degree of volatile loss than those of group 1, and generally retain significant amounts of SiO_2 and FeO, even though considerable amounts of these components have been lost. The average composition of group 2 glasses (in wt%) from both samples is given in Table 1. The similarity in both the relative proportions and the average compositions of the group 2 HASP glasses in these two samples reflects the similarities in bulk soil composition and soil maturity at these two sites. We note that glasses fractionated to the extent of the group 1 and 2 HASP compositions seem to occur only among the finest size fractions.

The group 3 HASP glasses have the least fractionated compositions. The glasses of this compositional type in both samples comprise a larger proportion of the analyzed population than does any other single group (44% of 72501 and nearly 42% of 78221), and the average composition of this group is very similar in both samples (Table 1).

The high-Si glasses are comprised of the relatively volatile elements and are compositionally complementary to the HASP glasses. These compositions make up 13% of the 72501 analyses and 11% of the 78221 population. The compositions of these glasses in both samples span the range from nearly pure SiO_2 , sometimes with other associated volatiles such as Na, K, P, and S, to other Si-rich compositions with high Fe concentrations (Table 1). It is believed that the compositions of this group represent the recondensation of impact-generated vapors [3]. As with most of the HASP glasses, these Si-rich compositions are only found among the finest size fractions of lunar soils. This suggests that such extreme fractionations only occur at sizes where the surface-area-to-volume ratio is high enough to allow the degree of melting and vaporization required to produce these unusual compositions.

In both samples, the glasses of basaltic composition constitute a relatively small group, being just over 8% of the analyses from 72501 and nearly 14% of those from 78221. These compositions display little or no observable volatile loss. The average composition of this group (in wt%) from both samples is given in Table 1.

Conclusions: We found that the majority of the analyzed glasses in both soils either have refractory compositions resulting from volatile loss, or are volatile-rich condensates of impact-generated vapors. In both samples, the three HASP groups comprise ~75% or more of the total number of analyzed glasses. The HASP glasses are

derived from the bulk soil and from the feldspathic component of the soils through the loss of major amounts of Si and lesser quantities of Fe and alkalis. The Si-rich glasses rival the number of the unfractionated basaltic glasses. The pronounced fractionations (volatilization and condensation) that occur in the submicrometer size range result from the large surface-area-to-volume ratio of the glasses.

This is the first report of high-Ti glasses from lunar soils. Considering all high-Ti glasses from both samples together, a trend is observed that begins with Fe-Ti-rich compositions and extends to glasses that consist of nearly pure TiO_2 . We conclude that these compositions originate by the loss of Fe from ilmenite, which is the dominant Ti-rich oxide at the Apollo 17 site. These high-Ti glasses are one of the few types of impact glasses derived from a specific mineral constituent of lunar soils.

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N 93-188083/2 P-4
ISOTOPIC AGES AND CHARACTERISTICS OF ANCIENT (PRE-SERENITATIS) CRUSTAL ROCKS AT APOLLO 17. W. R. Premo and M. Tatsumoto, U.S. Geological Survey, MS 963, Box 25046, Federal Center, Denver CO 80225, USA.

Problems with the Isotopic Systematics in Lunar Samples: Four different decay schemes, K-Ar, Rb-Sr, Sm-Nd, and U-Pb, have predominantly been used for age determinations on lunar rocks. These radiometric systems are particularly useful because they have long half-lives that are on the same scale as most lunar rock ages, which is necessary in order to obtain the most precise ages [1]. However, none of these systems is without problems when applied to lunar samples. In order for any single radiometric system to yield a primary crystallization age, it must remain closed from the time of crystallization until the present, without addition or loss of either parent or daughter isotopes. After many years of isotopic work, investigators have come to realize that most lunar samples (nearly all ancient highland samples) have been metamorphosed and their isotopic systematics disturbed [2]. Many studies report completely reset and partially reset Ar-Ar ages— ^{40}Ar loss readily occurs during impact metamorphism. However, some Ar-Ar ages agree quite well with other radiometric ages from the same sample, illustrating that Ar is not lost in some lunar samples. Reports of disturbances in the Rb-Sr and U-Pb systems [2-4] as well as the Sm-Nd system [3] are also prevalent in the lunar literature. Disturbed Rb-Sr isochron ages are normally explained by either mobility of both elements or mixing due to brecciation. The Sm-Nd system appears to be the most retentive, although some problems have also been noted. U-Pb and Th-Pb isochron ages typically date metamorphic events [5]. Fortunately, there are two U-Pb systems that can be compared simultaneously in a concordia diagram to "look through" disturbances. This attribute of the U-Pb systems is most desirable when working with lunar samples, and helps to identify both the age of the rock and the age of the disturbance. A major drawback of this approach is the necessity for initial Pb corrections in order to calculate radiogenic Pb/U ratios [5,6]. Initial Pb compositions are typically defined by the y-intercepts on U-Pb and Th-Pb isochron diagrams, but because most U-Pb and Th-Pb isochrons for lunar samples are disturbed, the initial Pb values are undefined and therefore must be assumed. This situation has been confronted with norite 78235 [5]. A possible solution to the problem is to use an age (hopefully accurate), perhaps determined using one of the other dating techniques, and back calculate what the initial Pb values must be in order to produce the same age with the two U-Pb

systems. Whereas this procedure relies on an accurate age for the rock, the resulting initial Pb information can be used to characterize the source magma much the same as initial Sr and Nd values, and can have important implications for models of the petrogenesis of lunar magma sources through time.

Because the various isotopic systems are disturbed differently during metamorphism, most samples yield conflicting ages when using the different dating schemes on the same sample. Whereas the discrepancy in ages is usually interpreted as a result of open system behavior due to metamorphic disturbances that characterize lunar rocks, other factors may cause age disparity as well, including either the misuse of basic assumptions regarding the isotopic techniques (e.g., U-Pb), or a lack of understanding of lunar petrogenesis (isotopic closure ages vs. ages of crystallization), or a combination of these factors. One possibility that cannot be dismissed is whether the disparity in ages reflects very slow cooling and crystallizing of the parent magma, particularly at depth in the lunar crust, and therefore the differences in isotopic closure temperatures of the various radiometric systems rather than the age of crystallization of the rock [3]. The isotope data from samples collected at the Apollo 17 site have many of these problems, so that at present investigators have only a very limited reliable (as well as precise) isotope database to work from. Interpretations on the magmatic ages and origins of at least the ancient (pre-Serenitatis) rocks are therefore tentative at best.

Ancient Crustal Ages at the Apollo 17 Site: Samples collected mainly from large boulders ejected from the surrounding regions onto the Taurus-Littrow valley floor include some of the earliest-formed deep crustal cumulates, including the so-called "Mg-suite" of predominantly norites, gabbro-norites, troctolites, and dun-

ites. Most of the samples are brecciated such that individual lithologies and monomict clasts are not large enough or accessible enough for most isotopic age work. Consequently, there are only a handful of reliable radiometric ages from this suite of rocks. The following discussion will contain only those data from pristine, monomict, Apollo 17 highland samples. A more complete list of the data and discussion is given in [2]. Figure 1 compares the best of the radiometric ages (32 analyses) from 14 ancient, pristine Apollo 17 samples—some samples have as many as five different age determinations (e.g., troctolite 76535). The data are all from [2,6,7]. Perhaps the most important feature is that Ar-Ar ages are in almost every case younger than ages determined using the other three systems for the same sample. Some of the Ar-Ar ages are completely reset to the estimated age of the Serenitatis impact at ~3.95 Ga (e.g., 72255 and 77215), whereas others probably show partial resetting. Troctolite 76535 is perhaps the best preserved judging from the general concordance of the isotopic systems—three Ar-Ar ages cluster around 4.16 to 4.27 Ga, the older age agreeing generally with two published Pb-Pb isochron ages [6,8] and a Sm-Nd isochron age [9]. In contrast, norite 78235/6 illustrates the problems that exist in some lunar samples. Out of the four different dating schemes, concordance might exist between two of them. There is even a discrepancy from two results using the Sm-Nd system, suggesting isotopic disequilibrium is possible even on a small scale in this sample. If we assume the Ar-Ar results are basically disturbed (except perhaps 78155 and 76535) and use only the results from the other three dating techniques, the oldest rocks are probably the Mg-suite norite 78236 (assuming 4.425 Ga age) and dunite 72417 at ~4.45 Ga. These rocks are then followed by norite 77215 (~4.37 Ga), troctolite 76535 (~4.26 Ga), gabbro-norite 73255 (~4.23 Ga), and

RADIOMETRIC AGES OF APOLLO-17 ROCKS

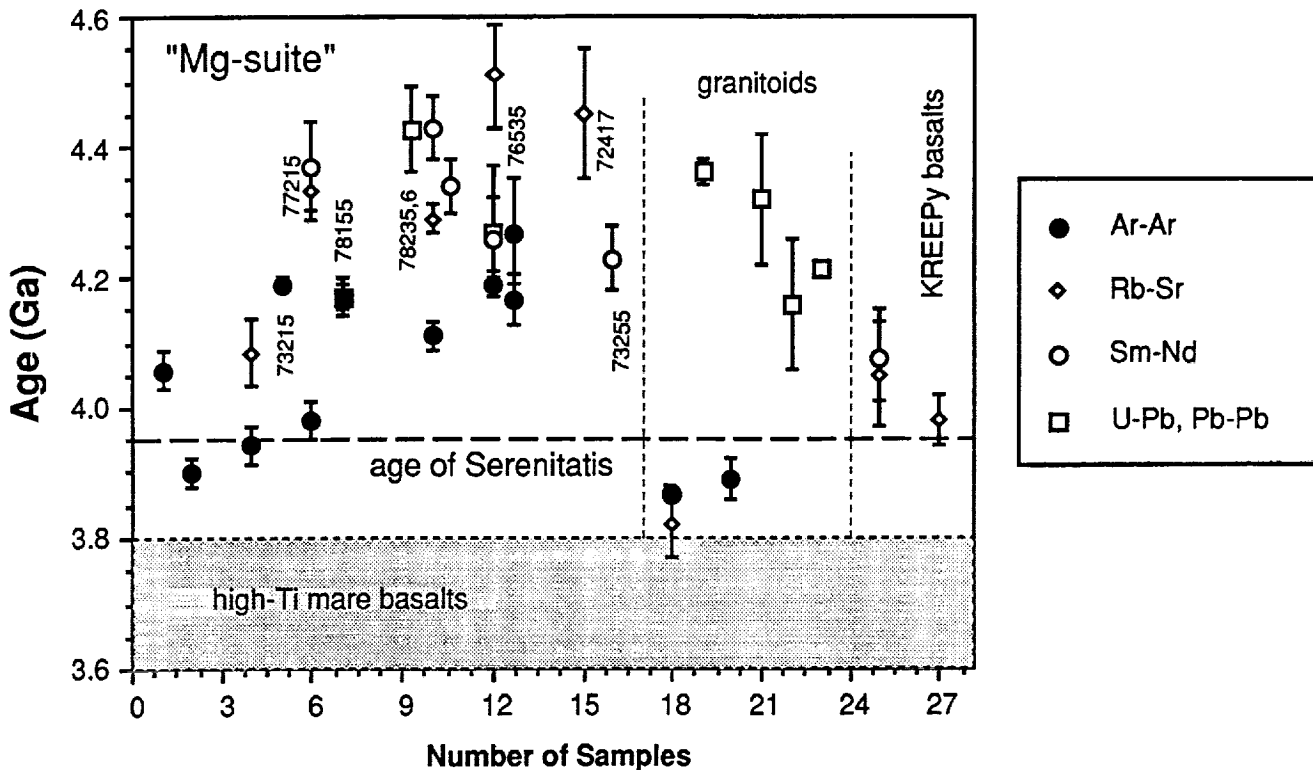


Fig. 1.

norites 73215 and 78155 (~4.17 Ga), although the latter may not be primary ages. Accepting these minimum estimates, Mg-suite formation appears to have continued from ~4.45 to ~4.15 Ga. However, depending on the investigator's interpretation of the isotopic age data, it is also possible to consider a much smaller magmatic interval. For example, we might question the Rb-Sr isochron age for dunite 72417, largely dependent on the mineral olivine, previously noted to be highly altered and possibly isotopically unreliable or disturbed [9] in some of these samples. A "best guess" Pb-Pb age of ~4.37 Ga was reported by [10] for 72415. Conflicting Sm-Nd isochron ages of 4.43

and 4.34 Ga are not supported by other isotopic age results (Fig. 1), resulting in some uncertainty of the true age of norite 78236. If we accept the younger age for this norite and assume that 73215 and 78155 are probably older than 4.17 Ga, then we find that most of the noritic-troctolitic (and probably dunitic) ages lie within a range of ~140 m.y. (4.37 to 4.23 Ga).

Pb-Pb analyses on U-Th-bearing minerals from granitic clasts separated from some of the breccias yield minimum ages between 4.36 and 4.16 Ga, indicating that granitic magmatism occurred contemporaneously with Mg-suite plutonism (Fig. 1), and all the episodes thus far discussed apparently predate KREEP basalt volcanism between 4.1 and 4.0 Ga. These results are interesting because concordant age results from the same sample, basalt 72275, using two different dating techniques, predate estimates for the Serenitatis impact event at ~3.95 Ga (Fig. 1). The isotopic systematics of these basalts were apparently unaffected by the basin-forming event. The KREEP basalts are also apparently at least ~150 m.y. older than the volumetrically dominant high-Ti mare basalts of the Apollo 17 site (Fig. 1). In any case, it should be obvious that there are too few reliable radiometric ages for any of the major ancient igneous rock types at Apollo 17 to feel confident about making any concluding remarks regarding their actual magmatic age durations.

Isotopic Characteristics of Ancient Apollo 17 Rocks: Implications for Their Petrogenesis: Assuming most of the ages given for pristine A-17 rocks are accurate, we can construct Nd, Sr, and Pb isotope evolution curves for the source(s) of these rocks (Fig. 2). With the exception of 77215, all ancient A-17 rocks plot between Nd isotopic evolution curves for the A-17 high-Ti basalts starting from a chondritic value at 4.56 Ga and a KREEP curve that is initiated at ~4.45 Ga (Fig. 2a). Several of the Mg-suite rocks, including troctolite 76535 and norite 15445, lie along the KREEP line, implying their derivation from KREEP-like sources. Depending on which age one refers for norite 78235/6, it may either be part of an initial pile of mafic cumulates settling out of a lunar magma ocean (Nd isotopes similar to early plagioclase float, anorthosite 60025) or part of a later stack of cumulates, crystallizing from reservoirs with variable $^{147}\text{Sm}/^{144}\text{Nd}$ ~0.2. Younger gabbro-norites, 67667 and 73255, may have formed similarly. Sources for the high-Ti mare basalts appear to have remained undisturbed with a $^{147}\text{Sm}/^{144}\text{Nd}$ ~0.25; however, these sources may have mixed isotopically with KREEP sources to produce sources for the younger plutonic rocks. The Sr isotope data do not help clarify the Nd data; however, the adherence of the Sr data to the BABI line indicates that all sources maintained an extremely low Rb/Sr value (Fig. 2b). The Sr evolution line for the high-Ti basalts does include within error some of the older A-17 plutonics as well as other ancient highland rocks, including anorthosites 67075 and 60025, dunite 72417, and norite 78236. These rocks all appear to have formed early, perhaps during differentiation of the lunar magma ocean. Troctolite 76535 lies off this line, just as it does on the Nd diagram. The Pb isotope data is too sparse and imprecise to make confident generalizations; however, again the coherence of the data between anorthosite 60025 and norite 78236 would suggest that Mg-suite rocks and anorthosites were both forming simultaneously during differentiation of the lunar magma ocean (Fig. 2c). The initial $^{206}\text{Pb}/^{204}\text{Pb}$ data is normalized to Canyon Diablo troilite ($\delta = 1$). A large uncertainty on the 67075 data leaves this analysis in question, although an unpublished Sm-Nd internal isochron age of 4455 ± 140 Ma indicates that it was also formed early. An interesting line of ascent is indicated by anorthosite 60025, norite 78235/6, and troctolite 76535, suggesting derivation from magma sources that are progressively evolving to higher $^{238}\text{U}/^{204}\text{Pb}$ and corresponding initial $^{206}\text{Pb}/^{204}\text{Pb}$ values.

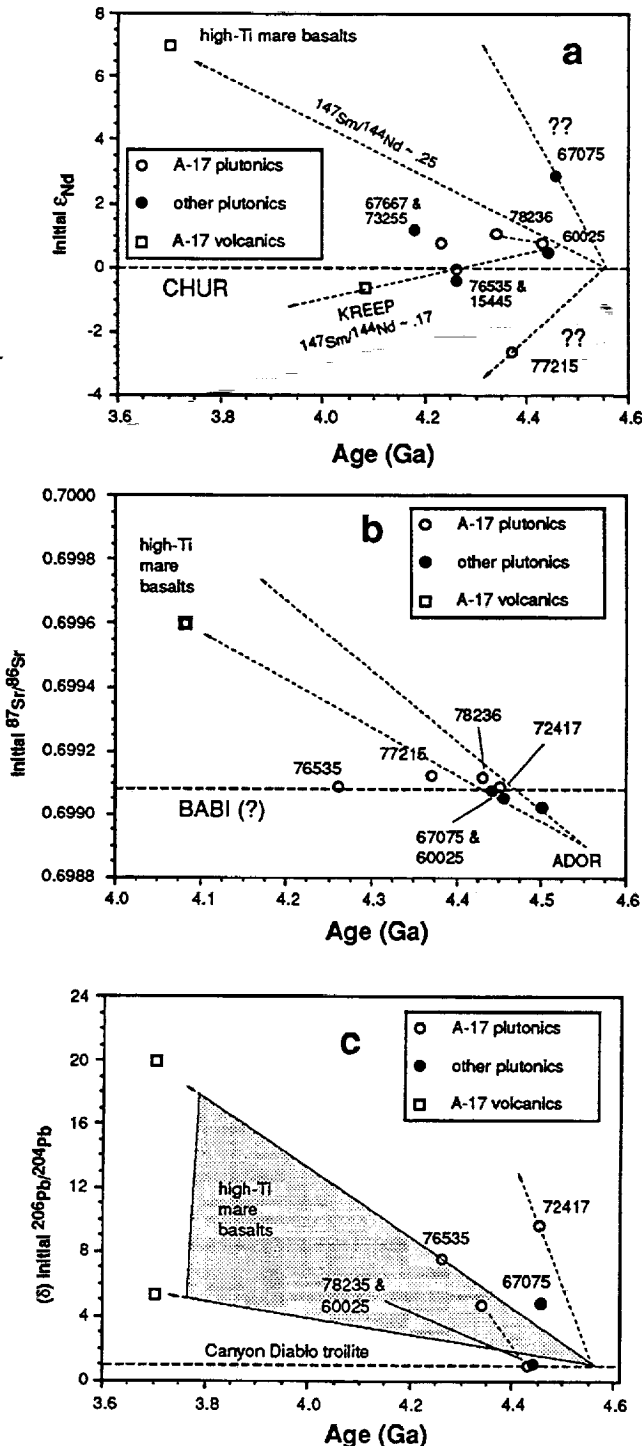


Fig. 2.

In summary, the lack of reliable as well as precise isotope data for Apollo 17 igneous rocks inhibits our ability to make confident statements regarding their ages and origins. However, the present state of isotopic art is progressing in the right direction, such that we should be able to obtain more precise data in the coming years, even with the present set of lunar samples. Nonetheless, it's obvious that we need larger, pristine, monomict highland samples if we are to finally arrive at real answers.

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THE APOLLO 17 SAMPLES: THE MASSIFS AND LANDSLIDE. Graham Ryder, Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston TX 77058-1113, USA.

More than 50 kg of rock and regolith samples, a little less than half the total Apollo 17 sample mass, was collected from the highland stations at Taurus-Littrow. Twice as much material was collected from the North Massif as from the South Massif and its landslide. (The apparent disproportionate collecting at the mare sites is mainly a reflection of the large size of a few individual basalt samples.) Descriptions of the collection, documentation, and nature of the samples are given in [1-3]. A comprehensive catalog is currently being produced (Ryder, in preparation). Many of the samples have been intensely studied over the last 20 years and some of the rocks have become very familiar and depicted in popular works, particularly the dunite clast (72415), the troctolite sample (76535), and the station 6 boulder samples. Most of the boulder samples have been studied in Consortium mode, and many of the rake samples have received a basic petrological/geochemical characterization.

Sample Numbering: Samples from the South Massif are numbered 72xxx (station 2), but the 721xx samples are from LRV stops on the mare plains), 731xx (station 2 and 2a/LRV-4), and 732xx (station 3, on the landslide). Samples from the North Massif are numbered 76xxx, 77xxx, and 78xxx, with the second digit specifying the station, with the exception of a few LRV stop samples. Rock samples have numbers whose last digit is from 5 to 9; unsieved regolith (including cores) have last digits of 0, and sieved fractions end in 1 to 4 according to the size fraction.

Rock Sampling: Sampling of rocks at the Taurus-Littrow massifs was very comprehensive in style. Multiple rock samples were chipped from boulders of varied sizes ranging from less than a meter to the bus-sized station 6 boulder. The multiple samples were taken to evaluate the visible textural and possible chemical variations of the matrices of the boulders necessary to elucidate their origins, as well as to sample clasts that give insight to older lunar events. The station 7 boulder not only had different matrix textures visible during the field study, but also dikelets that cross-cut an extremely large clast as well. Different subsamples of boulders also had different exposure geometries, providing greater input to cosmic and solar radiation models. Individual documented (i.e., photographed *in situ*) and undocumented rock samples, not obviously directly related to boulders,

were also collected. At several stations, samples were collected by raking with a 1-cm-separation rake. Two rake samples were taken at station 2, one at the base of the massif, a few meters from boulder 2, the other on the landslide 50 m away from the base. No rake sample was collected further out on the landslide. On the North Massif, rake samples were taken on the ejecta of a small crater at station 6, and on the rim of a small crater at station 8.

Regolith Sampling: Regolith samples were taken at all massif stations. Samples were taken of general regolith and of material on, under, and adjacent to boulders, mainly by scooping and some by trenching. A double drive tube (total about 70 cm depth) was collected on the light mantle at station 3, and a single drive tube on the North Massif at station 6, to a depth of about 37 cm.

Rock Types: The rock samples collected are different in character as a population from those collected on the Apollo 16 mission or from the Apennine Front on the Apollo 15 mission. In particular, all the larger boulders except boulder 1 at station 2 are very similar in chemistry and crystalline nature to each other, and to a very large proportion of the individual rock and rake samples collected. They have an aluminous basaltic composition with a KREEP incompatible element signature, falling in the general group of low-K Fra Mauro basalts (LKFM) originally defined for samples from the Apollo 14 site. They have crystalline melt matrices ranging from fine grained with olivine microphenocrysts (e.g., 76035) to poikilitic or ophitic (e.g., station 6 boulder). All of them contain mineral and lithic clasts (the latter mainly feldspathic granulites and pristine igneous fragments such as the dunite) and have meteoritic siderophile contamination, with the general similarity of Ir/Au about 1.5 (see summary in [4]). Radiogenic isotope data suggest a common age of around 3.87 Ga for the melting of these samples. The consensus is that these samples represent the impact melt produced by the Serenitatis Basin event. They dominate the sample collection in part because boulders were sampled in accordance with the field plans, and boulders tend to be from the coherent part of the Serenitatis rubble pile, which is the impact melt; older coherent units were broken up by the Serenitatis event.

Boulder 2 at station 2 is different in that it has a friable matrix consisting of a crushed volcanic KREEP basalt and aphanitic melt. This is texturally perhaps (but not chemically) the closest thing to the fragmental breccias at the Apollo 16 site. However, the bulk of boulder 1 consists of aphanitic melt alone, and these melts are different in chemistry from the "Serenitatis" melts, particularly in that they have much lower TiO₂. Some individual rocks collected on the light mantle are very similar to the boulder 1 aphanites, and one of them is a melt "bomb" [5]. As a group they have a clast population different from that of the "Serenitatis" melt (as well as a greater clast content); in particular, they contain more conspicuous granitic (or felsitic) fragments. If these aphanites do not represent an impact distinct from Serenitatis, then they must represent a substantially different phase of it than do the poikilitic rocks (see [4,6] for discussions). The radiogenic age of the aphanitic melts is indistinguishable from that of the poikilitic rocks.

Other impact melts of different composition are present in the Apollo 17 collection, but the range is not as great as that observed among Apollo 16 samples; extremely aluminous melts appear to be absent.

Feldspathic granulites (metamorphosed polymict breccias, most with more than 25% Al₂O₃) are fairly common as small clasts, but a few are individual rake or rock fragments as well. Far greater attention has been given to the pristine plutonic igneous rock fragments that occur both as clasts in the impact melt rocks and as

individual fragments (which may once have been clasts in melt units). Most are in the Serenitatis melts, e.g., the dunite 72415, the large norite 77215 (from a meter-sized clast within the melt), and the gabbro in the station 6 boulder. Others occur as smaller clasts within the aphanitic melts, e.g., the norite clast and the granitic fragments in boulder 1, station 2. Among the individual rocks are the troctolites 76335 and 76535, and the samples of shocked and melted norites (78235) that represent a single meter-sized boulder. These pristine igneous rock fragments have produced radiogenic ages mainly in the range of 4.0 to 4.3 Ga, but some may date back to as much as 4.5 Ga. Conspicuously absent (or perhaps just extremely rare) are the ferroan anorthosites common at the Apollo 16 site and present at the Apollo 15 site.

Regolith breccias are uncommon among the massif samples, although some, including glassy breccias, do occur. The sampling bias toward coherent and boulder samples, and the steep slopes working against the production of lithified regolith, may be responsible for this lack. Furthermore, the South Massif landslide would have diluted the uppermost regolith (the source of regolith breccias) with fresher bedrocks, at least at that one location. Mare basalts are rare among the massif samples, even the rake samples, certainly not surprising in view of the downslope movement, including the landslide, at most of sampling sites. The regolith particles tend to reflect the larger rock types, with feldspathic granulites, poikilitic impact melts, and plutonic fragments recognizable. There is little dilution with mare components, and orange glasses are rare. The regolith samples from the North Massif are contaminated with mare basalt or volcanic glass (TiO_2 3% or 4%), but the purest regoliths from the South Massif have little mare contamination. They are only a little more aluminous than the average LKFM "Serenitatis" melt. However, they have only half the abundance of incompatible elements, demonstrating that the soil contains a much lower proportion of the "Serenitatis" or aphanitic melt than might be expected from the relative abundance among the larger rock, rake, and boulder samples. The component underrepresented in the rocks is not particularly anorthositic, but must be low in incompatible elements.

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IMPACT MELT BRECCIAS AT THE APOLLO 17 LANDING SITE. Graham Ryder, Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston TX 77058-1113, USA.

Impact melt breccias are by far the most common highland rock type collected on the Apollo 17 mission. They tend to be fine grained, with virtually no clast-free impact melt rocks having been identified. All the highland boulders sampled are impact melt breccia, with the possible exception of one South Massif boulder that might have a friable matrix (but nonetheless consists dominantly of impact melt) and a shocked igneous norite boulder from the North Massif. The impact melt breccias were originally described as metaclastic, but their melt origin became apparent as work progressed [1,2]. Chemical compositions appear to allow natural groupings of the impact melt breccias. The present abstract relies in part on some new chemical analyses performed by the author (Fig. 1).

Most of the impact melt breccias are assignable to a single chemical group that is commonly accepted as being produced by the Serenitatis Basin event [1-3]. Most of these samples have crystalline groundmasses with fine- to medium-grained poikilitic matrixes. They dominate the rake samples and the boulders. Most of the remaining impact melt breccias are similar in chemistry to the dominant crystalline melts, but have distinctly higher Al_2O_3 and much lower TiO_2 [4-6]. Their compositions are not as uniform as the "Serenitatis" melts, and have some differences both within and among samples. They are also distinct in having aphanitic melt groundmasses and a greater abundance and diversity of lithic and mineral clasts. These aphanitic melts have been found only among samples from the South Massif.

Three other melt compositions are represented only by single samples: (1) 76055, more magnesian than the "Serenitatis" melts, and with lower incompatible element abundances [7,8]; (2) 72735, a high-K impact melt [9,10]; (3) a poikilitic clast in aphanitic melt 72255, similar to the "Serenitatis" melts but with higher alumina and lower incompatible element abundances [11].

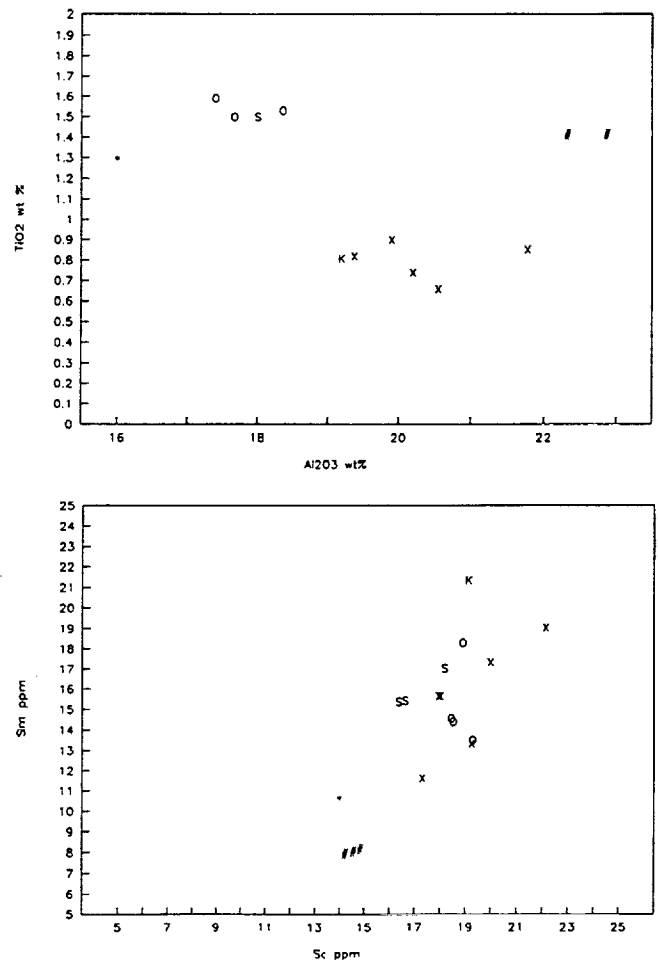


Fig. 1. Chemical analyses (TiO_2 v. Al_2O_3 ; Sm v. Sc) of impact melt rocks acquired in 1992 by the author, using INAA and fused beads (except 76055 = *, data of [8]). O = "Serenitatis" melt rocks, previously analysed. S = Serenitatis melt rocks, previously analysed but reanalyzed by INAA. X = 72255 aphanitic melt phase. # = 72255 poikilitic clast. K = 72735 high-K melt rock.

The only other impact melt samples from the Apollo 17 highlands (apart from glass formed by shallow-level recent impacts) are probably the troctolitic basalts found as clasts in some of the aphanitic rocks [e.g., 4]. A second high-K impact melt breccia proposed by [9] was not confirmed in the present analyses; instead, sample 72558 appears to be a typical "Serenitatis" melt.

A discussion of the origin of the aphanitic melt rocks and the "Serenitatis" poikilitic melt rocks was given by [3]. The poikilitic rocks have a lithic clast population limited almost entirely to plutonic igneous rocks such as norites and feldspathic granulites. The aphanitic rocks were faster-cooled, and contain a greater proportion and variety of lithic clasts; these clasts include surficial types such as impact melt and basalts, and felsite or granite clasts are common. The aphanitic rocks have accretionary and "bomb" characteristics; they are consistent with derivation from a much shallower target than the "Serenitatis" melts. Nonetheless, the targets were similar except that the shallower one was more aluminous and less titanian. The two groups could have been formed in the Serenitatis event, but in that case intermediate compositions might have been expected as well. Neither formed in a significant melt sheet, as they all contain conspicuous clasts and have fine-grained groundmasses. Even in the coarser melts, the clasts have not well equilibrated with the groundmass, even for argon [12]. Indeed, the aphanitic melt rocks do not show argon plateaus; their age is based on degassing of felsite/granitic clasts and is close to 3.86 or 3.87 Ga [13,14]. Clasts in the poikilitic rocks also compromise the melt age, although the best examples again suggest 3.87 Ga, presumably the age of the Serenitatis event [15,16]. Melt rocks with a composition very similar to either the "Serenitatis" melts or the aphanitic melts have not been found among Apollo 15 samples on the opposite side of the Serenitatis basin [17].

Sample 76055 has an older age, with a plateau that seems reliable at 3.91 Ga [18]. The sample is clast-poor (though not clast-free). The other two samples (72255 clast and high-K 72735) have not yet been dated, but are included in a laser Ar-Ar analysis in progress [17]. Samples of the "Serenitatis" and aphanitic melts are also included in the Ar study in an attempt to confirm and refine the age of Serenitatis and the possible relationships among the impact melt groups, as well as characterize the impact history of the Moon.

The characteristics of impact melt rocks are derived from the melt volume produced by the impact, from the clastic material entrained and picked up as the melt moved away from its source, and by the cooling environment. The characteristics can thus provide information on the lunar crust at and around the target site. Differences in composition of melt rocks can be interpreted as vertical or horizontal (or both) variations in crustal composition. Lithic and mineral clasts can be used to define the source rocks. The Apollo 17 impact melts suggest some variation in targets. The crust may be richer in titanium and poorer in alumina at greater depth. The deeper sampled parts seem to consist of pristine igneous rocks, particularly norites and troctolites, and some feldspathic granulites, whereas the shallower part has a greater complement of granitic/felsitic rocks and near-surface lithologies such as basalts, impact melts, and breccias. The lower part does not seem to consist of mixed megaregolith, but some components of the melt are as yet chemically unexplained.

The melt population sampled by the boulders and rakes samples is not representative of the massifs, which contain more alumina and lower abundances of incompatible elements. The soil composition suggests a maximum of 50% of either "Serenitatis" melt or aphanitic melt as a component. There is probably a bias in that coherent material, particularly large boulders, is likely to be a late unit, such as melt produced in Serenitatis; pre-Serenitatis material presumably exists as smaller blocks within the ejecta pile forming the massifs.

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P-4

APOLLO 17: ONE GIANT STEP TOWARD UNDERSTANDING THE TECTONIC EVOLUTION OF THE MOON. Virgil L. Sharpton, Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston TX 77058, USA.

The Moon's landscape is dominated by craters and large multiring impact basins that have obliterated any morphological evidence of the surface and interior processes occurring in the first few hundred million years of lunar history. By ~4.0 Ga, the lunar lithosphere was thick enough to support loads imposed by basin formation and mare infilling without complete isostatic compensation [1,2]. Most of the lunar tectonic features developed since that time are local in scale, and are associated with vertical deformation in and around mare basins. These features include (1) tectonic rilles, or graben, located along the margins of major mare basins, such as Serenitatis; (2) wrinkle ridges of compressional origin located primarily within the mare units; and (3) mare topography [3,4] developed via flexure and in part controlled by basin substructure. The relative ages and spatial arrangements of these features are explained by flexure of the lunar lithosphere around mascon basins in response to volcanic loading and global cooling [1].

Our present understanding of the tectonic history of the Moon has been shaped in large measure by the Apollo program, and particularly the Apollo 17 mission. The landing site, in the Taurus-Littrow valley on the southeastern flank of Mare Serenitatis, allowed extensive examination of the tectonic expression within and around the southern portion of Mare Serenitatis from the CSM orbiting overhead. Serenitatis is particularly well suited for understanding the tectonics of basin loading because its volcanic stratigraphy is conspicuous and reasonably simple [e.g., 3], and numerous tectonic features are developed on and around the mare. The Scientific Instrument Module included the Apollo Lunar Sounder Experiment, which was designed to detect layering in the upper few kilometers of the Moon [5,6]; these data have placed important constraints on the size of the volcanic load and the lithospheric response [1,4-6].

Astronauts on the surface in Taurus-Littrow valley made detailed observations of the Lee-Lincoln scarp and other structures encountered along their extensive traverses. They also made gravity measurements and conducted seismic profiling to constrain the geometry of the subsurface basalts [7]. These data, in conjunction with the samples returned to Earth, permit a far clearer understanding of the origin and evolution of this key valley and its relationship to Mare Serenitatis. In this brief paper, I attempt to summarize some of the interpretations that have emerged since Apollo 17, focusing on some of the problems and uncertainties that remain to stimulate future exploration of the Moon.

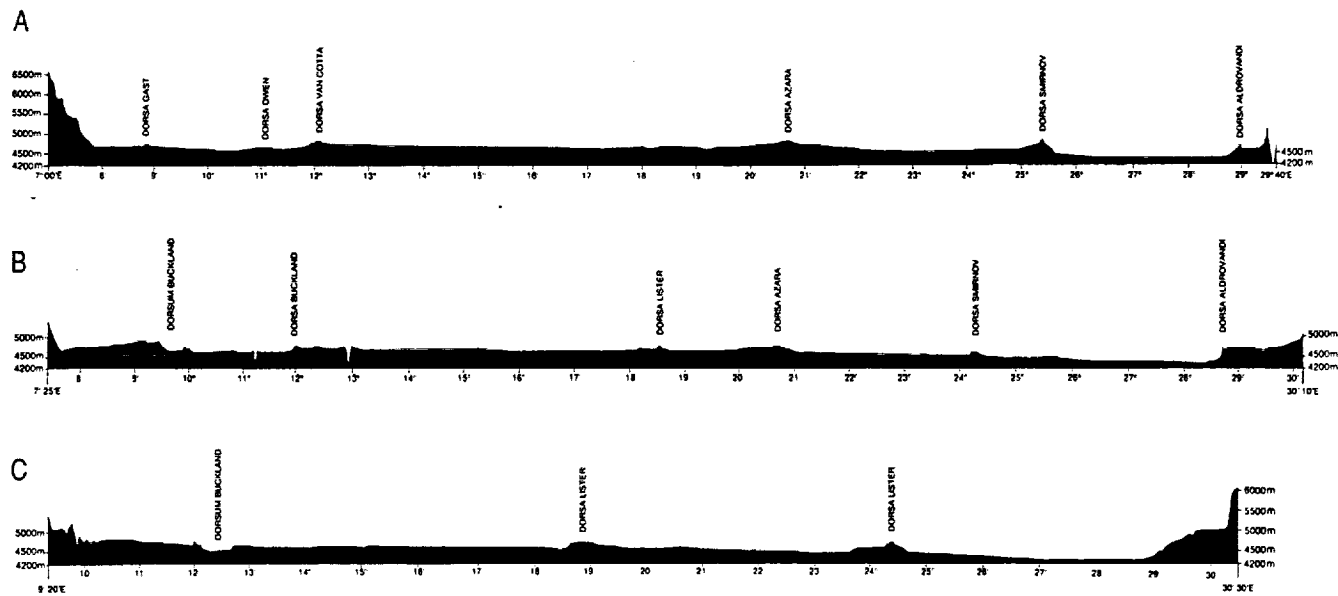


Fig. 2. Topographic profiles across Mare Serenitatis.

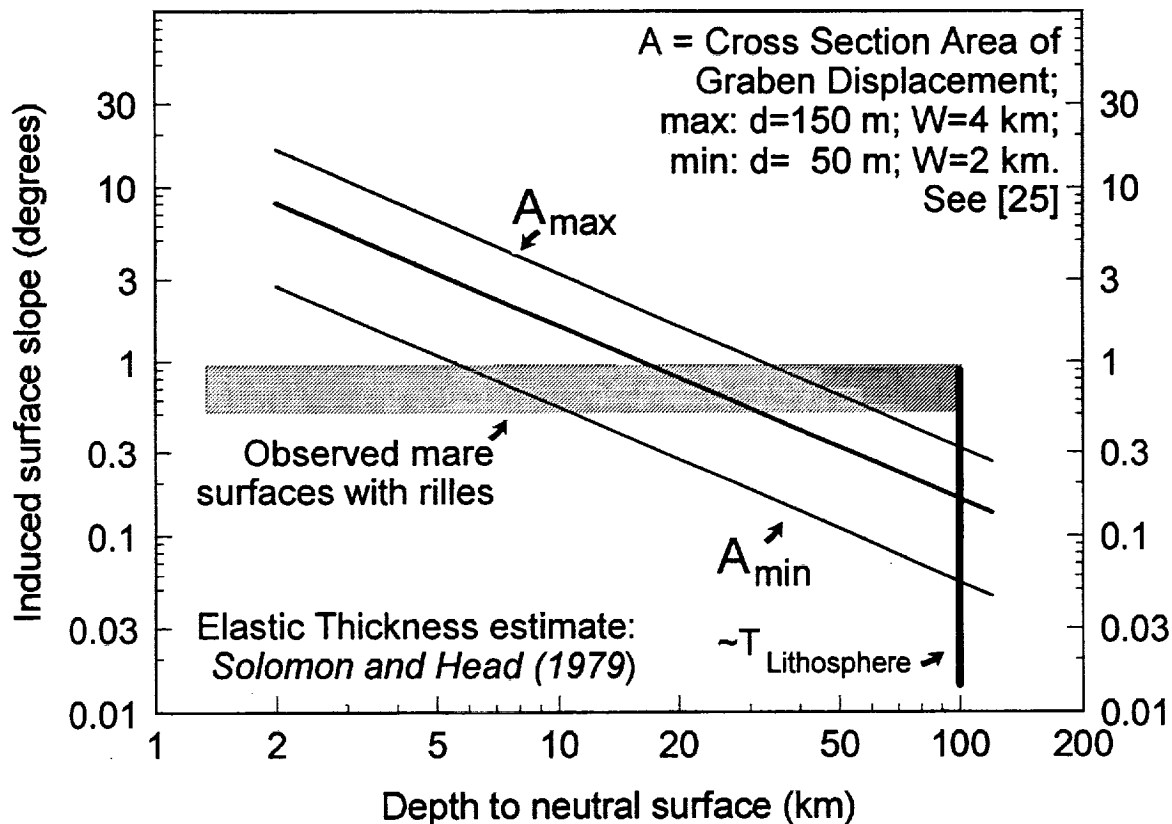


Fig. 3. Plot of the surface slope angle generated from rille formation as a function of depth to the neutral surface.

floor, and as the scarp cuts the valley wall of North Massif, its trend cuts sharply to the west. The slope of North Massif approaches 20° , therefore if the scarp is the expression of a westwardly dipping reverse fault, the dip must exceed 20° . The problem, however, is that the morphology of wrinkle ridge systems is highly variable and permits a spectrum of tectonic styles to be involved. Within Serenitatis alone, there are complex systems of ridges (Fig. 1), some of which have distinct vertical offsets, while others do not (Fig. 2). Two models have been proposed to account for the vertical offsets: thrust faulting [25] and nearly vertical faulting [24]. There does not appear to be any correspondence between the vertical offset across a ridge element and its topographic relief. Furthermore, apparent offsets across some ridges are produced because the ridge is developed on a sloping mare surface; when the regional trend is removed, so is the offset in several cases. This suggests that wrinkle ridges include a variety of compressional tectonic structures, perhaps ranging from simple thrust faults to nearly vertical reverse faults to complex zones of buckling [22]. There does not, however, appear to be any indication of substantial overthrusting.

Nature and Timing of Tectonic Rille Formation: Concentrically oriented tectonic rilles deform the flanks of many of the large mare basins including Serenitatis. These structures have been attributed to the extensional deformation associated with mascon loading [1]. Golombek [25] proposed that these graben were produced through simple extension, that the bounding faults intersected at a depth of a few kilometers, and that the faulted layer corresponded to the "megareolith." In that analysis, graben formation due to bending was dismissed because under the assumed conditions, the mare surface would slope away from the graben by up to 10° . However, Golombek apparently did not consider the effects of increasing the thickness of the faulted layer. Figure 3 shows the surface slope that would result from layer bending to produce the size of lunar rilles observed (50–150 m deep, 2–4 km wide; [25]). If the depth to the neutral surface were half the thickness of the elastic lithosphere (~ 100 km; [1]), then the slopes induced by bending would be $\leq 1^\circ$, in excellent agreement with measurements of the slopes on mare surfaces containing linear rilles [4]. This analysis indicates that graben around lunar basins can be accounted for solely by bending of the elastic lithosphere.

The stratigraphic relationships between tectonic rilles and the major volcanic units exposed in mare basins provide clues to the timing of basin deformation. It has long been recognized that rille formation ended prior to 3.4 Ga [26]. In addition, assessment of southeastern Serenitatis shows that the oldest volcanic unit (unit I; Plinius basalts equivalent; [3]) was emplaced prior to the onset of rille deformation. There are no cases where unit I clearly floods or embays any rilles nor is there any indication that rilles are truncated at the boundary between this unit and the highlands. In contrast, rilles that intersect the younger unit II surface are consistently truncated and embayed by these lavas. Elongate collapse features, indicative of buried rilles, are observed on unit II in the southeastern portion of the basin, but no such features are evident on exposed unit I surfaces. Consequently, it seems that rille formation in southeastern Serenitatis began after unit I emplacement and culminated before unit II.

Samples returned from Apollo 17 indicate that the unit I basalts were deposited over a range of ~ 150 Ma from ~ 3.84 Ga to ~ 3.69 [27]. Thus the post-unit-I onset of rille formation appears to signal a relatively slow response of the lithosphere to the increasing volcanic load in the Serenitatis Basin. The most likely explanation for this involves the isostatic state of the Serenitatis Basin during early mare emplacement. It is conceivable that appreciable quantities of the early volcanics would be required to offset the mass deficiency created

during the impact basin formation. If this is the case, volcanic infilling of the southern portion of Mare Serenitatis did not reach superisostatic levels until the majority of unit I was emplaced.

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p. 3

MELTING OF COGENETIC DEPLETED AND ENRICHED RESERVOIRS AND THE PRODUCTION OF HIGH-TI MARE BASALTS. Gregory A. Snyder¹, Lawrence A. Taylor¹, and Alex N. Halliday², ¹Department of Geological Sciences, University of Tennessee, Knoxville TN 37996, USA, ²Department of Geological Sciences, University of Michigan, Ann Arbor MI 48109, USA.

Implicit in current understanding of the location of terrestrial enriched and depleted reservoirs is the notion that they are spatially separated. The depleted reservoir on Earth is situated in the upper mantle, and the complementary enriched reservoir is located in the crust. However, Earth reservoirs are continually being modified by recycling driven by mantle convection. The Moon is demonstrably different from Earth in that its evolution was arrested relatively early—effectively within 1.5 Ga of its formation [1]. It is possible that crystallized trapped liquids (from the late stages of a magma ocean) have been preserved as LILE-enriched portions of the lunar mantle. This would lead to depleted (cumulate) and enriched (magma ocean residual liquid) reservoirs in the lunar upper mantle. There is no evidence for significant recycling from the highland crust back into the mantle. Therefore, reservoirs created at the Moon's inception may have remained intact for over 4.0 Ga.

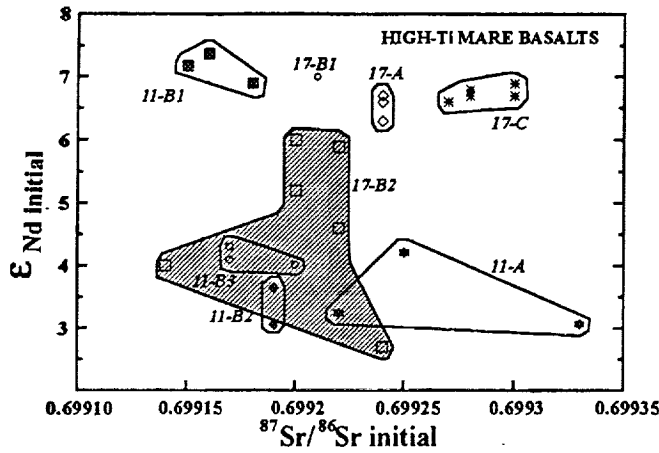


Fig. 1.

Radiogenic Isotopes in High-Ti Mare Basalts: Evidence of Heterogeneity? Data from all high-Ti basalts (Apollo 11 and Apollo 17 landing sites, as well as intermediate-Ti ilmenite basalts from Apollo 12) display a broad scatter in initial Nd and Sr isotopic composition as shown in Fig. 1 [2-5]. Basalts from Apollo 11 generally plot with less radiogenic Nd compared with those from Apollo 17. Although Apollo 11 group B1 basalts have ϵ_{Nd} values that are similar to Apollo 17 A, B1, and C basalts, these groups are distinct in Sr isotopic composition. These variations in initial Nd and Sr isotopic compositions have led previous workers to postulate that the basalts were derived from separate sources and have been used as evidence of heterogeneity in the lunar upper mantle [4]. Though this interpretation seems quite adequate, it is difficult to reconcile with the view that the lunar upper mantle was a simple system that crystallized early and has not been subject to later mixing and recycling processes. Furthermore, this interpretation, which includes myriad different sources, does not explain the homogeneity in mineralogy and major-element chemistry for these rocks.

Formation of Cogenetic Depleted and Enriched Reservoirs: It is generally believed that for less than the first 0.2 Ga of its history, the Moon differentiated into a small Fe-rich core, an olivine-rich

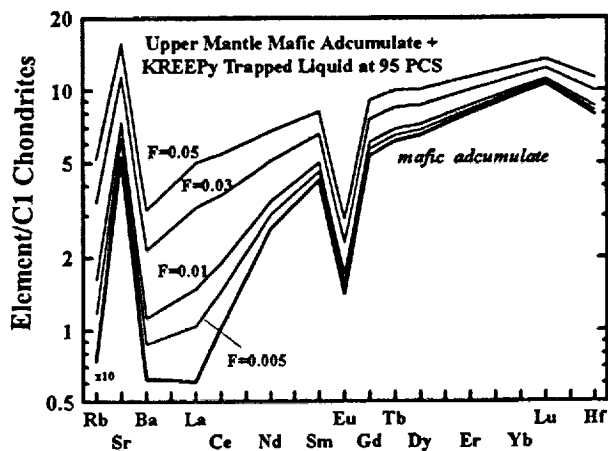


Fig. 2.

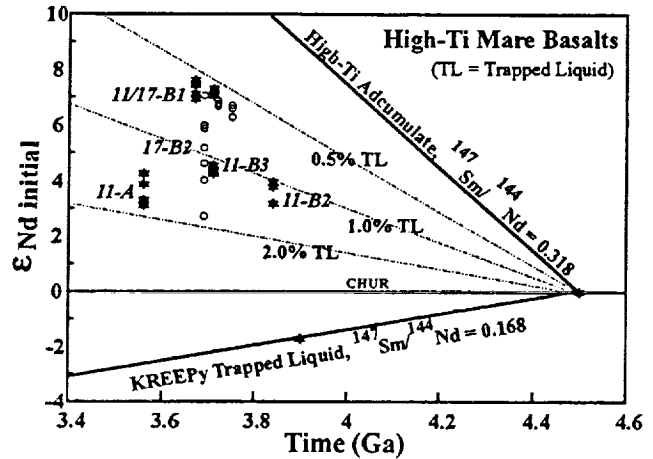


Fig. 3.

lower mantle, and a differentiated magma ocean [6]. This outer magma ocean, or magmasphere, progressively crystallized to form the upper mantle of the Moon and, once plagioclase became a liquidus phase, its anorthositic crust. The crystallization of the magmasphere probably led to layering of the upper mantle and included phases, such as ilmenite, late in its differentiation. The residual liquid magmasphere became more evolved with time, leading to enrichment in the LILE. This LILE-enriched liquid could have been trapped in variable, yet small, proportions and effectively "metasomatized" the relatively LILE-depleted crystallizing mafic cumulate. In this way, adjacent regions or layers of the mantle could maintain mineralogical and major-element homogeneity while exhibiting heterogeneities in their trace elements.

The LILE-enriched, trapped liquid end member of the source is represented by residual liquid after 95% crystallization of the LMO (as per Snyder et al. [7]). The isotopic and trace-element composition of this KREEPy liquid component may be represented by KREEP basalt sample 15382 (Rb = 16 ppm, Sr = 195 ppm, $^{87}Sr/^{86}Sr(3.84 \text{ Ga}) = 0.70115$; Sm = 31 ppm, Nd = 112 ppm, $\epsilon_{Nd}(3.84) = -3$; [8,9]). The cumulate portion of the source is modeled as a cpx-pigeonite-ilmenite-olivine perfect adcumulate (Rb = 0.01 ppm, Sr = 4.16 ppm,

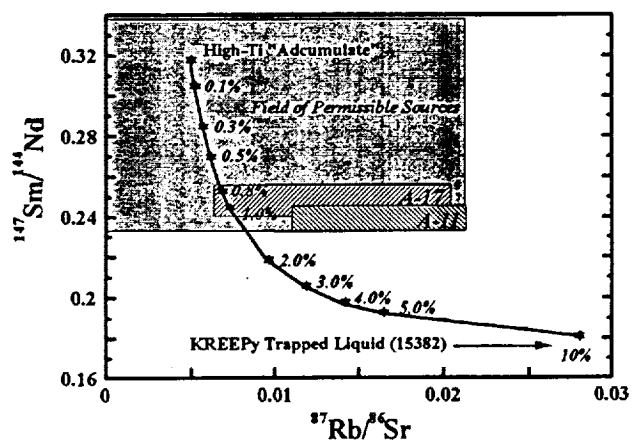


Fig. 4.

Sm = 0.631 ppm, Nd = 1.2 ppm, again as per [7]; Figs. 2 and 3) that has an extremely elevated ϵ_{Nd} (+8 at 3.84 Ga) and the lowest Sr initial ratio (0.69910) at 3.84 Ga. However, it is not likely that these two components remained distinct over a period of 500 Ma, when the interior of the Moon was still hot [10]. Recrystallization of the cumulate-trapped liquid pile could have occurred, yielding a source that was heterogeneous in trace elements on the scale of meters. Due to the low Sm/Nd ratio of the trapped liquid relative to the cumulate, those portions of the mantle that contained a larger proportion of this component would evolve with more enriched isotopic signatures.

Melting of the Source to Achieve High-Ti Mare Basalts: After an extended period of evolution (e.g., >0.5 Ga), earliest melting of the trapped liquid-cumulate pair would probably affect regions that were relatively enriched in the LILE (containing heat-producing elements U, Th, and K). Therefore, those regions that trapped the largest proportion of residual LMO liquid would melt first. Melts of these regions would exhibit relatively enriched isotopic signatures. Later melting would tap regions with less trapped liquid and would yield more isotopically depleted melts. Obviously, the degrees of enrichment and depletion of the melts are highly dependent upon the proportion of trapped liquid and the extent of melting of the cumulate + trapped-liquid pile. However, the trapped-liquid component is LILE-enriched (generally by at least an order of magnitude over the mafic cumulate) and would have originally consisted of low-temperature melting phases that would readily remelt. Therefore, even a small proportion (e.g., 1%) in the cumulate pile will greatly affect the isotopic signature of initial derived melts. However, because of its small proportion, the trapped liquid would have a lesser effect (inversely proportional to the degree of melting) on the major-element composition of the melt.

The low $^{147}\text{Sm}/^{144}\text{Nd}$ of this KREEPy trapped liquid, in concert with the relatively high abundances of Sm and Nd, obviates a large proportion of trapped liquid in the source (Fig. 2). This is illustrated in Fig. 2, where small proportions ($\leq 5\%$) of trapped liquid have been added to a model high-Ti adcumulate. This addition of trapped liquid has the effect of lowering the Sm/Nd ratio, yet increasing the abundances of Sm and Nd, thereby leading to a viable cumulate source region. Again, only 2–3% of trapped liquid is required in the source for modal melting, and less if the trapped liquid is an early melting component (if the source was not recrystallized).

This can be further appreciated by looking at the parent/daughter ratios of the high-Ti basalts (Fig. 4, diagonally hatched areas). Simple two-component mixing of a high-Ti "adcumulate" source with varying percentages of KREEPy trapped liquid (where sample 15382 represents the KREEPy liquid) yields a curve of chemical mixtures. The compositions of the high-Ti basalts from both Apollo 11 and Apollo 17 lie along this curve. Any source that contains residual mafic minerals, such as pigeonite, clinopyroxene, and olivine, would be more depleted (lower in $^{87}\text{Rb}/^{86}\text{Sr}$ and higher in $^{147}\text{Sm}/^{144}\text{Nd}$) than the basalts from which it was generated. Therefore, the field of permissible sources (shaded) indicates <1.5% trapped liquid. Although this trapped KREEPy liquid is minor in volume, it controls the radiogenic isotope signature of the derived melt.

Similar calculations to discern the proportion of KREEP in these basalts were performed by Hughes et al. [11] and Paces et al. [4]. Both groups concluded that small percentages (generally <1%) of a Rb-, Sr-, and REE-enriched component, with high Rb/Sr and low Sm/Nd ratios, are required to explain the compositions of parental magmas for the high-Ti basalts. However, both groups envisioned this component as distal to the cumulate source and added to the source prior to its fusion, but not cogenetic with its inception. Paces et al. [4] pointed out the problems inherent in such a scenario, but neither group

explored the possibility of an ancient KREEPy reservoir that was spatially associated with the cumulate source—as trapped intercumulus liquid from the late stage of evolution of the LMO—since its inception over 4.4 Ga.

Summary: The interpretation of cogenetic depleted and enriched reservoirs in the Moon is the consequence of events unique to the Moon. First, the late-stage LREE- and Rb-enriched residual liquid from a crystallizing LMO was trapped in variable and small proportions in the depleted upper mantle cumulates. A lack of recycling in the lunar environment would allow these reservoirs to diverge along separate isotopic evolutionary paths. This portion of the mantle would remain undisturbed for ≥ 0.5 Ga, prior to being melted to form the oldest high-Ti mare basalts. The isotopic character of the melts would be controlled by the degree of melting, as the least radiogenic reservoir would be melted first, i.e., that portion of the cumulate containing the greatest proportion of trapped liquid would melt first. The range in Sr and Nd isotopic ratios seen in basalts from Mare Tranquillitatis (Apollo 11) is due to melting of a clinopyroxene-pigeonite-ilmenite-olivine cumulate layer with variable proportions of trapped intercumulus liquid. Types B2 and B3 basalts were melted from a portion of the cumulate layer with intermediate amounts of trapped KREEPy liquid. Type B1 basalts from both Apollo 11 and 17 are melted from a "near-perfect" adcumulate portion of this layer. Apollo 12 ilmenite basalts represent the final known melting of this cumulate source, after it had been nearly exhausted of its ilmenite and trapped-liquid components. Type A basalts were probably extruded from a vent or vents near the Apollo 17 landing site [12] and could, therefore, represent melting of a similar source, albeit with the added complexity of neoKREEP assimilation.

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92-09-18813/7 f-3
BASALTIC IMPACT MELTS IN THE APOLLO COLLECTIONS: HOW MANY IMPACTS AND WHICH EVENTS ARE RECORDED? Paul D. Spudis, Lunar and Planetary Institute, Houston TX 77058, USA.

Many of the rocks in the Apollo collections from the lunar highlands are impact melt breccias of basaltic bulk composition [1–3]. They are known by a variety of names, including "low-K Fra Mauro basalt" [1], "VHA basalt" [2], and "basaltic impact melts" [3]. These rocks have been studied to understand the compositional nature of the lunar crust [1,4], to decipher the processes of large body impact [4], and to comprehend the record of impact bombardment of the Moon [5].

Study of terrestrial craters has led to a model for impact melt generation (e.g., [6]) whereby diverse target lithologies are totally, not partially, melted during impact. The impact melt makes up a few percent of the total volume of crater material; superheated silicate liquids of the impact melt have extremely low viscosities and mix intimately. This mixing thoroughly homogenizes the melt chemically during the excavation of the crater. Colder, unmelted debris is

overridden by the melt sheet as the crater cavity grows. Incorporation of these cold clasts rapidly chills the melt, with zones of greater and lesser amounts of clasts being primarily responsible for modestly differing thermal regimes [6]. The net effect of this process is the production of a suite of rocks that have extreme chemical homogeneity, but wide petrographic diversity (see [7]).

Strict application of this model to the petrogenesis of basaltic impact melts from the Moon has some fairly significant consequences for how we interpret early lunar history. First, total amounts of impact melt are small, usually a few percent of the volume of ejecta (although this fraction may increase as a function of increasing crater size [8]), and such small total melt volumes facilitate rapid cooling. Thus, coarse-grained impact melts must come from the central parts of the melt sheets of relatively large (tens of kilometers) diameter craters [9]. Second, because the chemical composition of melt sheets is extremely homogeneous, the supposed wide chemical diversity of lunar melt compositions reflects the sampling of multiple melt sheets [3,10]. These melt sheets formed in a variety of craters, most of which occur close to or beneath the Apollo highland landing sites [3,9,11]. Third, impact melts are the only products of impact suitable for radiometric dating [5,12]; thus, because (1) only the ages of melt rocks should be considered in reconstructing the cratering rates and (2) the Apollo impact melts are from multiple events that formed large craters, the data from the Apollo samples are telling us that Moon underwent a cataclysmic bombardment about 3.8–3.9 Ga ago, at which time nearly all the Moon's craters and basins were formed [5].

These conclusions are significant to how we perceive the evolution of the Moon as a planetary object, yet few stop to consider that the paradigm of impact melt petrogenesis upon which this depends is itself an incomplete model based on the geology of some poorly preserved terrestrial craters and a few inferences about the physics of

large impacts. In particular, the application of this model to the problem of the generation of basaltic impact melts on the Moon creates some difficulties in understanding all the lunar data. Are there really a large number of impact events represented by these melt rocks? What is the role of the largest impact structures (basins) in the genesis of basaltic impact melts?

One of the sites on the Moon where it is most appropriate to question the ruling paradigm for impact melting is the Apollo 17 Taurus-Littrow highlands [10,13]. The highland rocks from this site mostly consist of a variety of impact melt breccias, which have been broadly subdivided into two groups based on petrographic texture: the aphanitic and poikilitic melt rocks [6,11–17]. The aphanites appear to be a relatively heterogeneous group [11] and differ from the poikilitic melts in bulk composition [11,13], clast populations [11,14], and ages [15–17]. The group of melts collectively named “poikilitic” [16] are actually diverse texturally, having a variety of igneous textures, but showing remarkable chemical homogeneity [11]. In terms of chemical composition, the aphanites display considerable variation (Fig. 1), especially in comparison with the well studied Manicouagan impact melt (black in Fig. 1); the Apollo 17 poikilitic melts show chemical diversity comparable to that of Manicouagan Crater. These observations led Spudis and Ryder [11] to suggest that the two classes of melts formed in different impact events, with only the poikilitic melts being direct products of the Serenitatis impact. However, other workers preferred to interpret the aphanites as being a part of the Serenitatis Basin melt complex [13,14,17], the differences between the poikilitic and the aphanitic melts being attributable either to derivation of the latter from the margins of the melt sheet or early ejection [18] or the differences being considered insignificant [13,14].

Spudis and Ryder [11] noted an alternative interpretation of the Apollo 17 data: Our understanding of impact melt petrogenesis is incomplete and the terrestrial analogue should be applied to the Moon only with caution. Since that paper was written, a large amount of data has been collected for basaltic impact melts on the Moon: their compositions, their ages of formation, and their regional distribution and geological setting (summarized in [19]). In addition to data on Apollo 17 melt rocks, Fig. 1 also shows the principal melt groups found at other Apollo landing sites (these groups also appear well defined in plots other than Ti-Sc; see [19,20]). Note that with the exception of the Apollo 17 aphanites (and “group” A of Apollo 15, a three-member collection), the melt compositions appear to form diversity envelopes of size roughly comparable to each other and to the terrestrial Manicouagan impact melt sheet (Fig. 1). However, the groups also cluster *by site*, with the Apollo 16 melts making up a diffuse group with low Ti and Sc (groups 1–3, Fig. 1), the Apollo 17 melts having moderate Ti and high Sc (Poikilitic and Aphanitic, Fig. 1), and the Apollo 15 melts having high Ti and Sc (groups A–E, Fig. 1). Finally, note that if the melt groups are considered collectively by site, the resultant envelopes show diversity *no greater* than that displayed solely by the Apollo 17 aphanitic impact melts (Fig. 1).

In addition to these compositional data, we now understand several more things about basaltic impact melts on the Moon than we did 10 years ago. First, these impact melts are distinct in chemical composition from typical upper crust, as determined by remote sensing; they are both richer in KREEP and transition metals and are more mafic (less Al and more Mg) than the anorthositic composition of the upper crust [21,22]. Second, all these melts formed in a very short interval, between about 3.95 and 3.82 Ga ago [5,12–17]. Finally, each of these three Apollo sites is located within, on, or near the rims of three of the largest, youngest [23] basins on the lunar nearside: Apollo 17 occurs within the Serenitatis Basin [11,13], Apollo 15 is on

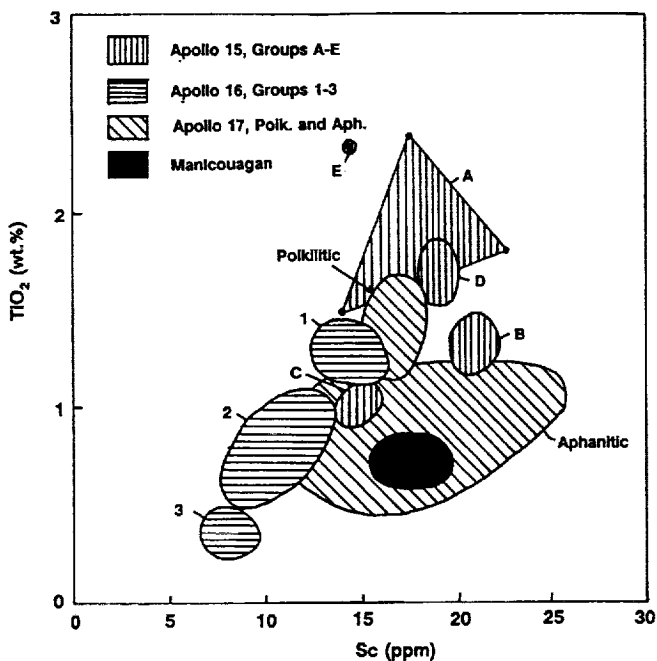


Fig. 1. Variation in Ti and Sc for basaltic impact melts of the Apollo 15, 16, and 17 landing sites (data from the literature). Impact melt sheet of the terrestrial Manicouagan Crater (black) shown for comparison. From [19].

the main rim of the Imbrium Basin [20,22,23], and Apollo 16 is on the backslope of the rim of the Nectaris Basin [19,23,24]. Each of these Apollo sites is in proximity to recognizable deposits of each basin; indeed, such deposits were high-priority sampling targets during these missions [23].

Taking the compositional data (typified by Fig. 1) and the above considerations at face value, I suggest that most of the basaltic impact melts in the Apollo collections represent impact melt from the Nectaris (Apollo 16 groups), Serenitatis (Apollo 17 groups), and Imbrium (Apollo 15 groups) Basins. (From this assignment as basin melt, I exclude Apollo 16 group 3 melts and Apollo 15 group E melt, none of which are "basaltic" in the sense that term is used here (see above) and which are probably from local impacts [20,24].) I believe that the terrestrial Manicouagan Crater, while giving us important insight into certain processes during melt generation, is an incomplete guide to understanding the origin of basaltic impact melts in the Apollo collection. The paradigm of Manicouagan (and other terrestrial craters) has been taken too literally and has been applied incautiously and uncritically to the Moon. Basin formation is an impact event at scales that greatly exceed our experience [19]. There is no independent reason to believe that sheets of basin impact melt are as thoroughly homogenized as is the melt of the terrestrial Manicouagan Crater. Recent study of the impact melt rocks from the suspected K-T boundary crater, Chicxulub, indicates that significant variation in the chemical composition of impact melt may occur in basins on the Earth [25]. Moreover, both the great size of basin-forming impacts and the thermal conditions within the early Moon suggest great quantities of impact melt are generated, not only making complete chemical homogenization less likely, but possibly providing a heat source for a variety of geological effects, including thermal metamorphism of breccias (granulites).

If this scenario is correct, the implications for the geological evolution of the Moon are significant. First, we must revise our model of impact melt genesis and subsequent evolution; such revision, in slightly different contexts, has been proposed for some terrestrial craters [25,26] and impact process in general [27]. Second, the principal evidence for a lunar cataclysm [5] is weakened, although such a cratering history is not excluded in this reinterpretation. If most of the melt samples from these highland landing sites are in fact melt from the three basins listed above, the absence of old impact melts in the Apollo collection reflects dominance of those collections by melt samples from these latest basins (of the over 40 basins on the Moon, Nectaris, Serenitatis, and Imbrium are among the youngest dozen; [23]). However, the argument of Ryder [5] that old impact melts should have been sampled as *clastic debris* from the ejecta blankets of these basins is still valid and their absence remains a puzzling and troublesome fact in this interpretation (although no basin ejecta blanket is well characterized). Finally, the several small to moderately sized "local craters" that have long been invoked to explain the geology of Apollo sites (e.g., [11]) are much less important than often has been assumed [3,9]. Most of the basaltic melts from these sites are from basins, not local craters, a fact evident by virtue of their bulk composition, which cannot be made by small or moderately sized impacts into the local substrate [24]. The only alternative to a basin origin for these rocks is derivation by crater impact into targets far removed (tens of kilometers) from the Apollo sites; the rocks would then have to be ballistically transported to these sites by other impacts [24].

While differing significantly from conventional wisdom, this interpretation of the basaltic impact melts in the Apollo collections is consistent with what we know about the Moon and what we think we understand about the impact process, a field that continues to evolve

with new knowledge, insights, and appreciation for the complexity of geological processes. Although this view of the Moon is not proven, I believe it to be a viable alternative that should be considered as we continue our study of the Moon and its complex and fascinating history.

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FUTURE SCIENTIFIC EXPLORATION OF TAURUS-LITROW. G. Jeffrey Taylor, Planetary Geosciences, Department of Geology and Geophysics, SOEST, University of Hawaii, Honolulu HI 96822, USA.

The Apollo 17 site was surveyed with great skill and the collected samples have been studied thoroughly (but not completely) in the 20 years since. Ironically, the success of the field and sample studies makes the site an excellent candidate for a return mission. Rather than solving all the problems, the Apollo 17 mission provided a set of sophisticated questions that can be answered only by returning to the site and exploring further. This paper addresses the major unsolved problems in lunar science and points out the units at the Apollo 17 site that are most suitable for addressing each problem. It then discusses how crucial data can be obtained by robotic rovers and human field work. I conclude that, in general, the most important information can be obtained only by human exploration. The paper ends with some guesses about what we could have learned at the Apollo 17 site from a fairly sophisticated rover capable of *in situ* analyses, instead of sending people. This is an important question because the planned first return to the Moon's surface is a series of rover missions. As discussed below, it seems clear that we would not have learned as much as we did with expert human exploration, but we would not have come away empty handed.

Unsolved Problems: Moonwide and at Taurus-Littrow: Primary differentiation. It is widely supposed that the Moon was surrounded by an ocean of magma soon after it formed. Ferroan anorthosites formed from this system, accounting for the high Al content of the bulk upper crust. Because the magma ocean was global, accumulations of ferroan anorthosites ought to be global as well. If so,

why are there virtually none at the Apollo 17 site? Perhaps there was no magma ocean. Perhaps ferroan anorthosites were stripped away by impacts, or assimilated by the abundant Mg-suite magmas that are represented by clasts in Apollo 17 breccias. Or perhaps we simply did not sample ferroan anorthosites at the site because they are buried beneath thick deposits of Serenitatis and Imbrium ejecta. A return to Apollo 17 can only test the last idea by searching specifically for anorthosites. Possible locations are deep inside the massifs, which might be revealed by craters on their flanks, or, less likely, far into the Sculptured Hills, which were not well sampled by Apollo 17. If we were lucky enough to find large blocks of Mg-suite rocks we might find evidence for assimilation reactions between them and anorthosites; this is a long shot.

Highland magmatism. A wide diversity of magmas intruded the original crust or extruded on it between 4.35 and 3.9 Gy ago. These include troctolites, norites, gabbro-norites, KREEP basalts and their differentiates, alkali anorthosites, and granites. Troctolites and norites themselves represent many separate intrusions. Other lithologies are inferred from breccia compositions to have been present, and others may have existed but have not been sampled. During Apollo 17 field work, rocks like these were collected as clasts in breccias that had rolled downhill from the massifs. It is likely that some large clasts could be sampled from apparent outcrops of breccia on the massifs, but blocks approaching outcrop size are probably not present. Nevertheless, a thorough search of lenses of impact melt rocks would be valuable, but probably requires field work and sample returns.

Bombardment history and dynamics of basin formation. The Moon is an ideal laboratory in which to study the details of the formation of immense impact craters. Apollo 17 is located just outside the second basin ring of Serenitatis, so the site was greatly affected by the formation of the Serenitatis Basin. In addition, deposits from other large craters and basins that predate Serenitatis, such as Crisium, must underlie the Serenitatis deposits, and ejecta from Imbrium overlies it. Thus, the Apollo 17 site is an excellent place to both study basin formation dynamics (quantity and morphology of ejecta from each source) and to determine the ages of at least two basins, Imbrium and Serenitatis. Impact deposits occur in the massifs, beneath the mare floor, and in the Sculptured Hills. Previous results from field observations and sample analysis indicate that the massifs are composed of breccias, including impact melt rocks, probably of basin origin and possibly representing the composition of the lower lunar crust. Field observations suggest that outcrops of breccias occur on top of South Massif and on the slopes and top of North Massif. These units contain important information about the nature of basin ejecta, including ejected melt. They demand thorough study. The Sculptured Hills are lower than the massifs and are gradational from knobby to smooth facies. Some authors have interpreted the formation as Serenitatis Basin ejecta, analogous to knobby material in the Orientale Basin. Others argue that it is younger than this basin, perhaps related to Imbrium. We still know very little about this interesting deposit. Samples obtained from this site on the lower flanks of the Sculptured Hills (station 8) contain a significant component of mare basalts from the valley floor and there were no boulders at the locality, so the lithologic make-up of the Hills is unknown. The Sculptured Hills clearly need to be studied during a return to the Apollo 17 site.

Mare basalt volcanism. Mare volcanic deposits contain information about the Moon's thermal history, mantle composition, mechanisms of transport to the surface, and processes in magma chambers and conduit systems. We have a good idea about the number of magma types represented among mare basalts returned by Apollo 17, the extent to which fractional crystallization operated, basalt ages, and something about their source areas in the lunar mantle. However, we

have a much less secure knowledge about basalt stratigraphy, unit thicknesses, modes of emplacement, where the fractional crystallization took place, processes that operated inside thick flows, and whether regolith layers developed between eruptive events. These problems can be addressed at Apollo 17 by a systematic study of blocks ejected from craters and those exposed in crater walls. Any clear ledges in crater walls could represent flow units like those exposed in Hadley Rille at the Apollo 15 site. Getting to such outcrops may be extremely difficult with robotic devices, and perhaps even with people. The Apollo 17 site also contains old mare basalts as clasts in breccias. Studies of the clasts populations in impact melt rocks could reveal additional types of mare basalts.

Pyroclastic volcanism. Pyroclastic deposits are important because they are composed of relatively unfractionated basaltic magmas (hence contain vital information about the lunar interior), contain completely unexplained enrichments of volatile elements, and might constitute important resources to use on the Moon. Apollo 17 is the type section. The deposit discovered at Shorty Crater was sampled with great skill, but time did not permit a more complete examination. A return to the site needs to address many questions: How thick are the deposits? Does the thickness vary around the site? Is more than one event represented? That is, are there alternating orange and black glass bands? Do the concentrations of volatile elements or the bulk chemical composition vary laterally or vertically?

The regolith and the Sun's history. The most important aspect of the lunar regolith is that it contains a record of the Sun, or at least of the solar wind. The regolith has proven to be disturbingly complicated and although isotopic variations in solar N and He have been observed, correlating these with time has not been possible. At the Apollo 17 site it might be possible to sample older regolith at craters on the maria near the massifs, where the basalt cover is thin. This would excavate highland materials exposed for 100 m.y. prior to basalt extrusion 3.8 Gy ago. We could also obtain a more detailed understanding of regolith stratigraphy by sampling ejecta from craters in the 1–100-m size range, and by digging trenches and drilling numerous deep cores. With luck, a few time markers, like the coarse-grained layer in the Apollo 17 drill core, could provide an absolute chronology of the stratigraphy revealed.

Tectonics. The Taurus-Littrow valley is almost certainly a fault valley associated with the Serenitatis impact event. However, we do not know its detailed structure, the extent to which it was modified by the Imbrium event, its initial depth, or the amount of massif material that has subsequently filled it and whether the fill includes material deposited with the Sculptured Hills. The Apollo 17 site is also decorated with a pronounced postmare fault scarp (Lee-Lincoln scarp) and an impressive landslide. The mechanisms and timing of these features could shed light on lunar tectonic processes. Although the landslide was apparently caused by secondaries from Tycho, the details of the avalanche mechanisms are not understood.

Origin of the Moon. No one site can address such a major, global question, and no return mission can be planned to observe just the right thing, return just the right sample, or make just the right measurement to determine how the Moon formed. However, the Apollo 17 site holds some key pieces of the puzzle: (1) abundance of anorthosite, (2) nature of highland magmatism, (3) nature of the lower lunar crust (basin-derived impact melts), (4) composition of the mantle (mare basalts and volcanics), and (5) total volatile inventory of the Moon and storage areas of volatiles inside the Moon (pyroclastic deposits).

Return to Taurus-Littrow: Robots and People: When we return to the Moon we will send both robotic roving vehicles and people. Suppose we return to the Apollo 17 site. What could we do

with a rover, such as those being planned by the Artemis team? What problems require people to solve? This can be evaluated by considering specific areas in and around the Taurus-Littrow valley that need to be studied to address the problems outlined above. To do this, I assume that the rover has a range of many tens of kilometers, cannot return samples to Earth, and carries an imaging system, a device to obtain mineralogical information such as an imaging spectrometer, and an instrument to make accurate analyses of major and selected minor elements. The chemical analyzer needs to be able to either sample rocks easily with a reliable drill or make analyses from a small distance (for example by laser emission spectroscopy). Other instruments could also be useful, such as gadgets to determine regolith maturity or determine the contents of solar wind gases, but I will assume that such contraptions will not be carried on the first lander. To compare to human exploration, I assume that geologist-astronauts will be able to travel 25 km from an outpost, have sufficient time to study rocks in the field, can make it to the top of North and South Massifs, and will return samples to Earth. The field sites are listed in priority order.

Sculptured Hills. We know so little about these deposits that significant gains can be made with a rover. By traveling far into the Hills and making analyses of soils and rock samples along the way, a solid idea of the mineralogical and chemical composition of the Sculptured Hills will be obtained. We could also determine the compositions of clasts in boulders, though determining whether they were coarse or fine grained may be difficult. However, it is not clear that we will be able to determine the amounts of impact melts and fragmental breccias, and we certainly could not determine ages, thus leaving open the question of when the Sculpture Hills formed. Nevertheless, a rover mission would add substantially to our knowledge of these basin deposits. Human explorers would be able to obtain samples for detailed study (including ages and isotopes) and could examine boulders, crater ejecta, crater walls, and other possible outcrops. Their observations would be far superior to the rovers because of better vision and agility.

Outcrops on massifs. We learned a lot from field and laboratory study of the boulders that rolled down the massifs, but we will learn much more by examining the outcrops the boulders came from. These are probably direct deposits of basin fragmental and melt ejecta. A rover (assuming it could ascend the slopes) might be able to send back images of sufficient quality to allow types of breccias to be distinguished and to observe their structural relationships to each other. Possibly the rock types present in the clast population could be recognized. However, distinguishing poikilitic impact melts from aphanitic impact melts may be impossible in the field (even for an astronaut). The chemical distinction is routine for returned samples, but *in situ* analysis would require an instrument capable of distinguishing rocks with >1.5 wt% TiO_2 from those with <1.3 wt%; this is a tall order. On the other hand, analytical devices on a rover could determine that many fine-grained materials have LKFM composition (18 wt% Al_2O_3) and detect the presence of other types of LKFM (high alumina, 22 wt% Al_2O_3 ; ferroan, mg# of 60 rather than the conventional 70). Overall, though, an astronaut could make better field observations (principally because of better eyesight and agility) and analyses of returned samples would allow us to make significant though subtle distinctions among mapped units and, most important, determine ages of impact melts, hence of basins.

Pyroclastic deposits. A rover might have discovered the orange soil, and even grabbed a scoop full of it, but it could not have determined the geologic context. The emphasis during a return excursion should be on physical volcanology, as outlined above. Little of the data we need could be obtained by a rover, including detailed

study of deposits in the walls of Shorty Crater, although some observations could be done and we might learn something useful. We need detailed field observations and careful sampling, including core samples. The field observations should not be confined to Shorty Crater, but ought to include smaller ones nearby that show hints of orange ejecta and numerous craters throughout the landing site to determine the extent of the deposit.

Mare basalts. Apollo 17 basalts are coarse grained, implying thick flows. It would be interesting to sample individual flows in detail to see how crystal size varies and if late-stage liquids segregate and migrate throughout the flow. It is also possible that the flows were inflated during emplacement, a process akin to intrusion, causing them to thicken and allowing slow cooling of the interior. Careful field work is clearly called for. Furthermore, the key outcrops are in crater walls, probably inaccessible to simple rovers. Finally, many interesting processes that operate inside lava flows are revealed by trace-element analysis, which can be done best on Earth.

Regolith. To determine secular variations in solar wind isotopic composition, samples of known or determinable ages are essential. This job is impossible without sample returns. However, other interesting properties of the regolith and the contents of solar wind gases could be determined by a properly equipped rover. Such a payload could be included on a resource assessment mission, rather than one designed strictly for science.

Suppose All We Had Originally Was a Rover: A return to Taurus-Littrow requires people to be present to make substantive progress in understanding the geology of the site and the Moon. Rovers will not add significantly to our knowledge, except for exploration of the Sculptured Hills. However, suppose we had never been to the Taurus-Littrow and sent a rover mission to the site (or a similar one). What would we learn? Here's a guess: (1) We would determine that the valley floor contains high-Ti mare basalts, but probably not determine that there are four groups of basalts and definitely not measure their ages. (2) Unless we were lucky, we would probably not discover the orange soil; even if we did we would probably not be able to demonstrate that it was a pyroclastic deposit. (3) We could deduce that the boulders at the base of the massifs are impact breccias and have the characteristic LKFM basaltic composition, though we would not know their levels of REE or Sc. (4) We could determine much about the nature of the Sculptured Hills. This is less than we learned by sending skilled people, but still a solid contribution to our knowledge of one place on the Moon. What rovers lack when compared to humans they make up in much longer time spent exploring and in enhanced abilities while in the field (chemical analysis, multispectral imaging). Of course, astronauts could carry such devices as well.

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THE SUDBURY-SERENITATIS ANALOGY AND "SO-CALLED" PRISTINE NONMARE ROCKS. Paul H. Warren, Institute of Geophysics and Planetary Physics, University of California, Los Angeles CA 90024, USA.

The Serenitatis Basin is the one lunar basin from which we confidently identify a suite of samples as pieces of the impact melt sheet: the distinctive Apollo 17 noritic breccias (at least the typical poikilitic variety; the aphanitic breccias might not be from the same impact [1]). Recent studies of the Sudbury Complex (e.g., [2]) indicate that its "irruptive" is almost entirely of impact-melt origin, making it the closest terrestrial analogue to the Serenitatis melt sheet. Any attempt to model the evolution of the Moon's crust should be

compatible with the relatively well-understood Sudbury Complex. The textures of Sudbury Complex rocks are mostly fine-grained (~1 mm), with relatively elongate plagioclase [3], compared to typical terrestrial, lunar, and meteoritic cumulates. However, many of the Sudbury rocks are nonetheless cumulates, and the Sudbury magma clearly underwent extensive fractional crystallization. For example, the *mg* ratio in augites shows cryptic variation from 0.79 to 0.25. The upper granophyre layer (roughly 1/2 of the total "irruptive") is enriched in K and incompatible elements by a factor of ~2.8, and depleted in CaO by a factor of ~3.6, compared to the other main layer (the norite). Sudbury produced a crater of apparent diameter D_a ~ 200 km, and the diameter of the transient crater D_t was roughly ~110 km [4]. For Serenitatis, D_t has been estimated at roughly 370 km [5]. A superficial analysis would suggest that the volume of the impact melt was greater, and thus its rate of cooling was slower, at Serenitatis; thus, the Serenitatis melt sheet underwent a comparable, if not more extensive, fractional crystallization. Besides raising questions about the origin of the A-17 breccias, this Sudbury-Moon analogy has led Grievé et al. [2] to conjecture that "some misinterpretation of the origin of . . . so-called pristine lunar highland samples has been made and some are primordial impact melt rocks from large impact events."

The Sudbury-Moon analogy might be a misleading oversimplification, if applied too rigidly. The A-17 poikilitic impact breccias that appear to be from the Serenitatis melt sheet have major-element compositions very similar to many pristine noritic cumulates and monomict-brecciated cumulates (e.g., 78235, which has grains up to 10 mm across). Yet the textural contrast between the least-brecciated pristine norites and the poikilitic impact breccias is obvious. The textures of the A-17 poikilitic breccias could hardly be mistaken for endogenous igneous rocks, as the Sudbury "irruptive" rocks once were [3].

What caused the Serenitatis impact melt to evolve so differently from the Sudbury impact melt? The total volume of melt V_m was far greater at Serenitatis. Equation (6) of [4] estimates V_m as a function of crater diameter. This method is of course imprecise, but the accuracy of the implied *slope* for V_m vs. D_a is supported by comparison to various terrestrial craters. Assuming that D_a for South Serenitatis is roughly $6 \times D_a$ for Sudbury, and correcting ($\times 0.23$) for the Moon's lower *g* (and thus, lower ratio of melted/displaced material: equation 7.10.2 of [6]), V_m should be roughly 240 \times greater for Serenitatis than for Sudbury. Adjusted for the roughly 36 \times greater area of the Serenitatis melt sheet (assuming analogous melt sheet shapes), the melt sheet thickness at Serenitatis should have been roughly 7 \times that at Sudbury, *assuming* similar aggregation efficiencies for the melts.

Besides cooling rate, the efficiency with which a melt body will fractionally crystallize and generate cumulates is probably sensitive to the ability of convection or other fluid motions to continually supply "fresh" melt to crystal/melt interfaces. The tendency to convect is governed by the Rayleigh number *Ra*, which is proportional to thickness³ and *g*¹. Thus, a lunar melt sheet 7 \times thicker than an otherwise similar terrestrial one would have a 57 \times higher *Ra*. The melt viscosity μ would also be a key factor (*Ra* is proportional to μ^{-1}). The ~1.24 wt% water in the Sudbury Complex [3] would be offset by its high average SiO₂ (~63 wt%), and at a likely differentiation *T* of 1000°C, μ calculated à la [7] would be 1.3×10^5 poise; even assuming 2.48 wt% H₂O and *T* = 1200°C, μ would be 1.6×10^3 poise. The 1200°C μ for a melt of A-17 noritic breccia composition [8] is far lower: 86 poise.

I suggest that the key factor that stifled differentiation of the Serenitatis impact melt was an adverse density relationship. The 1000°C, 1-kbar density of the Sudbury Complex composition [3], calculated à la [9], is 2.43 g cm⁻³. Even at 10 kbar, it is only 2.46 g

cm⁻³. This is 0.30 g cm⁻³ lower than the average density of the country rock [10], 0.22 g cm⁻³ lower than the density of the least-dense liquidus phase (feldspar), and 0.40 g cm⁻³ lower than the aggregate density of the cumulates of the lower half of the complex. Thus, the Sudbury impact melt must have efficiently segregated up and away from the country rocks with which it was initially interspersed, and from the crystals it grew as it cooled. In contrast, the 1200°C density calculated for the average A-17 noritic breccia composition [8] is 2.759 g cm⁻³. A typical estimate for the average zero-porosity density of the lunar crust would be 2.9 g cm⁻³. In the uppermost few kilometers, this density is reduced by breccia porosity. The porosity is roughly 15–20% in the uppermost 2–3 km. It diminishes with depth, probably in a stepwise fashion, but seismic data suggest that it is not entirely squeezed out until a depth of ~20 km. Assuming 5% porosity is representative of the region where most of the impact melt first forms, and that this region is compositionally "average," the implied country rock density is 2.76 g cm⁻³—identical to that of the melt. Assuming the Serenitatis and Sudbury melts were originally dispersed amidst country rock to similar degrees, the Serenitatis melt sheet was probably far less efficiently aggregated into a single large mass. Instead, pockets of impact melt that originally formed deep in the Serenitatis crust may have typically remained almost stationary, or rose only to a level where the porosity of the surrounding country rock translated into neutral buoyancy. These dispersed small masses would have undergone relatively rapid thermal equilibration with the country rocks (and much of the country rock would have been baked into granulitic breccias, which are common among A-17 rocks).

During crystallization of whatever Serenitatis melt managed to aggregate into a large, nearly clast-free sheet, the density of the melt (~2.76 g cm⁻³) would have been only 0.05 g cm⁻³ greater than the 1200°C density of a major liquidus phase (Ca-rich feldspar), and ~0.20 g cm⁻³ less than the 1200°C density of the aggregate liquidus assemblage (feldspar + Mg-rich low-Ca pyroxene). Under these conditions (including lunar *g*), unless the magma was very thick, it would tend to become choked with feldspar, turning off convective motions, and thus also fractional crystallization.

Also, the original dispersal of the impact melt was greater beneath Serenitatis. During an impact, most of the melting tends to occur at depths greater than 1–2 projectile radii [6]. Assuming for Sudbury $D_a = 200$ km and impact velocity $v_i = 20$ km/s, equation (7.8.4) of [6] (the intermediate of three scaling laws discussed) implies an estimated Sudbury projectile diameter $D_p = 28$ km. Assuming that for Serenitatis $D_a = 600$ –1000 km, its $D_p = 73$ –140 km. Lower v_i , as commonly invoked for the early Moon, imply even larger projectiles; e.g., reducing the Serenitatis velocity to 10 km/s implies $D_p = 107$ –204 km. Of course, besides this depth effect, the Serenitatis melt would also have been more widely dispersed horizontally. Note that these calculations imply that the Serenitatis melt sheet may have included a component of mantle-derived melt (but only if the deepest Serenitatis melt managed to migrate all the way up to the near-surface melt sheet). A minor mantle-derived component might help to explain why the average *mg* ratio of the A-17 noritic breccias (0.706) is almost as high as a typical estimated bulk-Moon *mg* ratio.

The same considerations apply to lunar vs. terrestrial large-scale cratering events in general. The dichotomy between apparently nonpristine and apparently pristine lunar rocks is remarkably sharp. A lunar crust exposed to steadily declining bombardment by basin-scale impacts might be assumed to acquire a less distinct dichotomy. The accretion rate did not necessarily decline steadily. But even supposing it did, if the above interpretation of the role of density in the movement and crystallization of lunar impact melts is correct, then once the magma ocean produced a thick ferroan anorthositic crust

with near-surface (porous) density $\sim 2.76 \text{ g cm}^{-3}$, impact melts probably almost never managed to pool together well enough, and thus cool slowly enough, to produce coarse-grained, pristine/cumulate-seeming rocks.

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TROCTOLITIC ANORTHOSITE FROM 77115: A MAGNESIAN MEMBER OF THE ALKALIC SUITE. Paul H. Warren and Gregory W. Kallemeyn, Institute of Geophysics and Planetary Physics, University of California, Los Angeles CA 90024, USA.

Alkalic suite pristine nonmare rocks are distinctly enriched in plagiophile elements such as Na and K, as well as generally incompatible elements, despite modes and textures more characteristic of typical crustal cumulates (most commonly anorthosites) than of the basaltic KREEP rocks that appear to account for the bulk of the lunar crust's total complement of incompatible elements. Most of the ~ 17 previously reported alkalic suite samples have come from Apollo 14 or 12 (only 180 km to the west of A-14), except for clasts from one A-15 breccia (15405) and one A-16 breccia (67975). Our studies indicate that the 77115 troctolitic clast of Winzer et al. [1] is actually a troctolitic anorthosite (or anorthositic troctolite), probably best classified as a member of the alkalic suite. Winzer et al. [1] analyzed a 30-mg chip and found a high normative olivine content (60%, plus 40% plag. and 1% apatite) and bulk-rock $mg = 87.3 \text{ mol}\%$, despite high contents of rare-earth elements (e.g., Sm = 42 $\mu\text{g/g}$, or 0.88 \times average high-K KREEP). Norman and Ryder [2] classified this sample as KREEP, but the pattern of incompatible elements of the Winzer et al. [1] analysis was far from KREEP-like (e.g., Ba/Ce = 0.23 \times the KREEP ratio, Ce/Lu = 1.6 \times the KREEP ratio). Chao et al. [3] reported that two thin sections were made from this clast, but "only plagioclase of the clast was sectioned."

We managed to obtain a thin section with pyroxene and olivine, and analyzed a 13.4-mg chip by INAA. This chip, like all the thin sections, is highly anorthositic, with only 0.87 wt% FeO. It has an even higher LREE/HREE ratio than the Winzer sample (e.g., La/Lu = 2.2 \times the KREEP ratio), and extraordinarily high contents of plagiophile elements (e.g., Ga = 6.3 $\mu\text{g/g}$, Eu = 4.0 $\mu\text{g/g}$, Sr = 340 $\mu\text{g/g}$), in typical alkalic suite fashion. However, Winzer et al. [1] only found Sr = 134 $\mu\text{g/g}$. Extraordinary, by alkalic suite standards, is the magnesian nature of the mafic silicates: olivine averages $\text{Fo}_{99.3}$ (range among 14 analyses 97.5–89.1), low-Ca pyroxene clusters very tightly near $\text{En}_{87.9}\text{Wo}_{1.7}$ (average $mg = 0.894$). An uncommonly magnesian Cr-spinel is also present, containing 17.75 wt% Al_2O_3 , 16.31 wt% FeO, 12.64 wt% MgO, and 2.40 wt% TiO_2 . The plagioclase averages $\text{An}_{95.1}$ (range among 35 analyses: 94.3–95.8), which is extraordinarily Na-poor by alkalic suite standards.

Nonetheless, the alkalic affinity indicated by the Ga, Sr, and REE (especially Eu) data, and the strangely P-rich composition determined by Winzer et al. [1] (0.53 wt% P_2O_5), all point toward a complex

petrogenesis, probably involving either assimilation of KREEP into a Mg-suite magma, or metasomatism of an Mg-suite troctolitic anorthosite by an extremely evolved fluid or melt. In the past, we were unable to resolve between these two models for alkalic anorthosites from Apollo 14 [Warren et al., 1983]. However, the mass balance for mixing KREEP into a hypothetical 77115c Mg-suite parent magma is difficult, unless the KREEP component is remarkably REE-rich and the Mg-suite component is remarkably magnesian. Thus, 77115c tends to strengthen the case for metasomatic alteration in alkalic suite genesis. However, this sort of metasomatic activity (which probably requires a volatile-rich fluid) surely only affected a tiny fraction of the Moon's crust, and tentative acceptance of a metasomatic model for one alkalic suite rock need not imply that this model is preferable over the physical mixing/assimilation model for alkalic suite rocks in general.

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LAST CHANCE AT TAURUS-LITTROW. D. E. Wilhelms, U.S. Geological Survey, retired, 2027 Hyde St. Apt. 6, San Francisco CA 94109, USA.

The Problems: By the fall of 1971 we knew that only two more Apollos would land on the Moon. Most geoscientists agreed that both should concentrate on the previously neglected terrae (highlands). In June 1971 the Apollo Site Selection Board (ASSB) had chosen Descartes as the site of the Apollo 16 terra landing, scheduled for April 1972. Therefore we had to assess how many pre-Apollo objectives the first four landings had met, how many Apollo 16 was likely to meet, and how to meet the remaining ones with Apollo 17.

Geologists convened at Caltech in November 1971 by Lee Silver and geology-team leader William Muehlberger formulated the following list of major lunar problems (edited here): (1) ancient crustal and interior materials; (2) early impact history; (3) major basins and mascons, a broad category that included the basins' ages, the petrology of their ejecta, the nature of the deep rock they excavated, the origin of their rings and radial sculpture, and the cause of the positive gravity anomalies (mascons) detected over their mare fillings; (4) large craters and their products—their ages, the subcrater rock brought up in their central peaks, their superposed pools and flows (generally assumed to be volcanic), and even the hoary question of their origin still doubted by caldera advocates; (5) highland igneous evolution, then widely believed to be an important process affecting terra morphology; (6) maria—the variability of their compositions and ages; (7) postmare internal history, mostly meaning the dark pyroclastic blankets thought to postdate the already-sampled mare basalts; (8) present physical and chemical state of the interior; (9) lunar heterogeneity, both vertical and lateral; and (10) regolith evolution and radiation record.

From this list only one major impact structure (Imbrium Basin), the maria, and the regolith were thought to have been well explored through the time of Apollo 15 (August 1971). Apollos 14 and 15 had sampled the Imbrium ejecta. Apollos 11, 12, and 15 had abundantly sampled three points on the maria. Crews of all four successful Apollo landings had collected regolith cores, and Apollo 16 could be expected to obtain comparison cores in the heart of the highlands. Before it flew, most people still thought that Apollo 16 would elucidate the types of volcanism and magmatic evolution endemic to the terrae.

That left large chunks of the list for Apollo 17 to tackle. Discovery of lunar anorthosite and formulation of the magma ocean hypothesis had suggested what the early crust may have been like, but no petrologist or geochemist was satisfied with the sample record then in hand, and Apollo 16 as then fancied did not promise to add more. The crucial dating of Imbrium at about 3.84 or 3.85 aeons (post-1977 decay constants) had shown that most basins and large craters had formed in the Moon's first 700 m.y., but further specification of the impact rates depended on dating some pre-Imbrian basins. The only possible samples from a large post-Imbrium crater were those from Apollo 12 thought, not universally, to have come from Copernicus ray. Since all the returned mare samples were extruded between 3.84 and 3.16 aeons, nothing was known about later thermal history. Geophysical probing had produced only tentative conclusions about the interior by late 1971. In other words, the main objectives remaining for Apollo 17 were at the extremes of lunar history: primitive non-Imbrium terra at the old end, and the state of the interior and the postmare volcanics at the young end.

The Site: An Ad Hoc Site Evaluation Committee chaired by Noel Hinners of Bellcomm had recommended Descartes as the landing site of Apollo 16, and in January 1972 it received recommendations for Apollo 17 from the Caltech meeting and other interested parties [1,2]. Several old favorites were rejected once and for all. The possible Apollo 12 dating of Copernicus had downplayed the importance of that otherwise scientifically desirable, though operationally difficult, target; anyway, Copernicus is in the Imbrium region, and its supposed volcanic features were thought "well understood." Marius Hills might satisfy the young-volcanics objective but would not yield any terra material; also, it was barely accessible by the winter launch being planned for Apollo 17. Apollo 16 photographs would not be available in time to plan a mission to Rima Davy, a chain of small craters then widely counted on as a source of xenoliths because it looks like a string of maars localized by a deep fault. Alphonsus, a perennial contender for all missions and favored for Apollo 17 by the ASSB in June 1971, was considered once again, but it was thought probably contaminated by Imbrium ejecta because it is crossed by Imbrium sculpture and seemed softened by a mantling blanket. MSC decisively vetoed the scientifically very desirable Tycho because it looked too rough and too far south. Jack Schmitt had proposed a landing at Tsiolkovskiy on the farside, but no funds were available for the necessary communication relay satellite.

There is plenty of non-Imbrium, pre-Imbrian terra on the Moon, but Apollo 17 was restricted to those parts of it that were covered by good Lunar Orbiter or Apollo photos and that satisfied the many restrictions imposed by propellant capacity, launch reliability, solar lighting, communications, Earth splashdown point, and so forth [3]. Only two general zones survived preliminary screening. One was Gassendi Crater, which offered excellent non-Imbrium, pre-Imbrian terra and a good geophysical station, though only dubious volcanic units other than more mare. Gassendi was also Apollo's last chance to explore a large crater and moreover one with a central peak and a geophysically interesting uplifted floor. Orbital overflights could have continued over the very attractive target of the Orientale Basin on the west limb. MSC engineers, however, thought the astronauts would be blocked by rilles and a ring trough from reaching Gassendi's main target—the central peak—and the Apollo program managers did not accept orbital science as a valid consideration in landing-site selection.

Enlargements of Apollo 15 pan photos drew all eyes to the second region, the highlands east of Mare Serenitatis, west of Mare Crisium, and north of Mare Tranquillitatis. As usual on the Moon, most parts of these highlands were nondescript and too lacking in mappable

geologic units to provide a context for the point samples. MSC considered a scientifically suitable site near Proclus to be too far east for adequate tracking and communication with Earth during approach. A region southwest of Mare Crisium was rejected because it was accessible to the Soviet sample returners and thus might be sampled redundantly; in fact, Luna 20 did sample the Crisium Basin rim in February 1972. That left the western reaches of the highlands, near Mare Serenitatis. There were disturbing signs of Imbrium influence in the form of radial striations and blanketing deposits, but the ancient crustal rock seemed likely to be exposed in relatively sharp-looking massifs of the pre-Imbrian Serenitatis Basin rim that are part of Montes Taurus.

The other half of the site's name, derived from the nearby 1-km Littrow Crater, was originally applied to a supposedly young dark-mantled site at the margin of Mare Serenitatis that had been intended as the Apollo 14 landing site before the Apollo 13 accident in April 1970. The dark surface extended eastward into a valley lying amidst the Serenitatis massifs. A landing on this Taurus-Littrow valley floor therefore seemed likely to provide access to a young pyroclastic deposit. This interpretation was bolstered by the beautiful Apollo 15 orbital photos and by visual observations by Apollo 15 command module pilot Al Worden of dark-halo craters that looked like cinder cones scattered all over the region's brighter surfaces. Shorty Crater was one of these. The dark mantle also showed up clearly as streaks on the massifs, supporting its interpretation as a pyroclastic deposit that had been forcefully fountained from numerous vents. It might furnish two coveted items that had not turned up earlier: volatiles and xenoliths.

A young "bright mantle" derived from South Massif promised to place samples of the massif, therefore of the ancient rock, within easy reach of the astronauts. With luck, the massif samples would also shed light on basin-forming processes, as would a distinct unit of tightly packed domical knobs called Sculptured Hills that resembles knobby ejecta units of the Orientale and Imbrium Basins called Montes Rook and Alpes formations respectively (though to some the Hills looked like volcanic domes). The plains beneath the dark mantle of the valley floor ("smooth plains" or "subfloor material") constituted yet another distinct geologic unit. So Taurus-Littrow offered a diverse geologic banquet [4].

It also seemed good for geophysics because it lies in a contact zone between a mare and its containing basin. Although this setting is similar to that of Apennine-Hadley, most of the surface instruments differed from those of Apollo 15. Photo-loving geologists were bothered because an orbital track tied to Taurus-Littrow would largely duplicate that of Apollo 15, but geochemists and geophysicists were less worried because they would have different instruments on board. On 11 February 1972, its last meeting, the ASSB unanimously approved Taurus-Littrow for Apollo 17.

Today: Gene Cernan and Jack Schmitt returned a fine collection from the massifs, bright mantle, Sculptured Hills, subfloor basalt, and dark mantle of Taurus-Littrow [5-7]. They answered many of 1971's questions, showed others to have been wrongly asked, but left others for us to ponder still today.

1. The lunar crust consists not only of anorthositic and KREEPY rock, as might have been thought if the Apollo program had ended after Apollo 16, but also includes large amounts of a magnesian suite unrelated to the magma ocean [8]. The question remains, why does the Mg-suite dominate this one of the sampled localities?

2. Early lunar impact history is still not well known because the apparent absolute ages of the massif and bright-mantle samples, 3.86 or 3.87 aeons [compiled in 6,9], are not old enough. This is true no matter what basin they date—Serenitatis itself [6,9,10], Imbrium,

Crisium, or more than one basin or crater [11]. If the collected samples are from the Serenitatis ejecta, if Serenitatis is as stratigraphically old as its many superposed craters and degraded appearance suggest, and if 20 or 30 m.y. can really be resolved analytically, then the small differences between the Apollo 17 absolute ages and those from the Apollo 14 and 15 Imbrium samples would support the hypothesis that all large basins formed in a cataclysm. The age differences have less bearing on the cataclysm hypothesis, however, if Serenitatis is late pre-Imbrian (late Nectarian) and looks old only because it is degraded by deposits and secondary craters of Imbrium [9].

3. We found out that major basins make a lot of impact melt and create highly heterogeneous ejecta [12,13], important findings that were not clear from Apollos 14, 15, and 16. Theoretical massaging of Apollo 15 and 17 orbital data, in particular, has pretty well cleared up the problem of the mascons by showing that they are caused both by incompletely sunken slabs of mare basalt and by mantle uplifts [14]. However, the formational mechanism of massifs is still not agreed on, nor is the source of the Sculptured Hills. Cernan and Schmitt remarked on their distinctiveness; they are not volcanic and are probably a discrete deposit of high-trajectory basin ejecta like the Alpes and Montes Rook Formations [6]. But which basin ejected them? The superposition relations and distribution of similar though less distinctive hills on adjacent terrain, including the massifs, suggest that they are an outlier of the Alpes Formation cut off from the main exposure by Mare Serenitatis. If this is their origin, Apollo 17 may have failed to escape Imbrium's dominion.

4. Large craters would have been better investigated at Cassendi; we still have only Copernicus ejecta, if that. However, continued experimental, photogeologic, and geophysical research, combined with negative evidence from all Apollos and Lunas, has shown to most people's satisfaction that volcanism has played no role in the formation of large craters or even of their superposed pools and flows. The trend of a ray from Tycho and the clustered secondary craters visible on South Massif indicate that the bright mantle is either a landslide triggered by the impact of Tycho ejecta on the massif or a spray of ejecta from the secondaries; in either case, dating of the bright mantle and of the Central Cluster added Tycho to the list of dated craters, at 109 m.y. [15].

5. Apollos 16 and 17 have shown that impact and not volcanism has created the many diverse landforms of the terrae [9], with the possible exception of some plains that remain unsampled. Highland igneous evolution therefore probably completely or nearly ended in pre-Imbrian time.

6. Apollo 17 brought back abundant additional mare basalt from the valley floor, though this added little to existing knowledge of the variability of the visible maria except to demonstrate that mare flows can pour out more voluminously and quickly than they did at the

Apollo 11, 12, and (probably) 15 sites [16,17]. More novel was the return of numerous clasts from disrupted prebasin maria, showing that mare volcanism was active in pre-Imbrian time [18].

7. The dark mantling deposit consists of pyroclastic glasses [19] formed way back during the main epoch of mare formation in the Imbrian Period (an antiquity also perceptible from photogeologic relations); therefore "postmare" internal history was not as active as thought, although independent photogeologic work has identified small Copernican mare and dark-mantle units in several places on the Moon.

8. These geologic findings when added to the sum of findings about the interior from the Apollo 15 and 17 heat-flow experiments and the seismic experiments of all missions have shown that the Moon is and has long been cool or lukewarm and much more quiescent than had been widely believed in the 1960s, but the thickness of the crust is known at only a few places, and the existence of a core has not been established [17].

9. The diversity of both mare and terra samples reveals a heterogeneous Moon, though more samples and orbital surveys are required to show the degree and scale of the heterogeneity.

10. The regolith is understood to a first order but still contains a rich record waiting for future explorers.

We have come a long way since 1971 and the hot-cold controversy about the origin of lunar surface features. Now let us look again at the rich trove of data we have for answers to the remaining questions.

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