

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. Photogeologic 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 ± 00.05 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 μ m 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