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Mare volcanism and lunar crustal structure

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Abstract—On the basis of a popular scenario for the early history of the moon, application of simple hydrostatic concepts to the eruption of lunar basalts provides an explanation of several heretofore unrelated features of lunar crustal structure. The scenario includes initial melting of the outer portions of the moon, flotation of cumulate plagioclase from the cooling liquid to form a lunar crust, and a zone of remelted mafic cumulates, the source of mare basalt liquids, that deepened with time. Because the early lunar crust was less dense than most of the mafic or ultramafic liquids that were at one time beneath it, a number of conclusions may logically be drawn: (1) Mare basalt liquids derived from immediately below the crust would not have risen to the lunar surface except at the bottoms of the deepest impact basins. This provides an explanation for the upper limit to the ages of returned mare basalt samples. (2) A minimum depth of origin for the high-titanium (3.7–3.8 AE) mare basalts emplaced at the lunar nearside surface is about 100 km, consistent with estimates from experimental petrology, with thermal history models, and with the absence of mare basalt fill in farside basins if the offset of lunar centers is due to crustal thickness variations. (3) The hydrostatic rise of mare basalt liquids to the lunar surface became less efficient with time, since younger flows of lower density and derived from greater depth do not completely cover old ones. (4) The structure of mascons depends upon crustal thickness. Where the crust was thin prior to basin formation, basalt fill was immediate and constitutes much of the present anomalous mass. Where the crust was thick, basalt fill was delayed or of small volume, and most of the anomalous mass is due to upwelling of the lunar mantle. (5) Because the density of KREEP basalt liquids is less than that of the lunar crust, the uneven distribution of KREEP material on the lunar surface implies that KREEP liquids were not derived from moon-wide molten or partially molten layers in the lunar crust or mantle.

INTRODUCTION

THE CONCEPT OF HYDROSTATIC HEAD, the height to which a fluid at a given pressure will rise in a gravitational field, can be usefully applied to magmatic liquids and to a number of interesting problems in volcanology. The equilibrium heights of terrestrial, and by inference Martian, volcanoes appear to be directly related to the source depths of the basaltic liquids which just rise to the respective volcanic summits (Eaton and Murata, 1960; Vogt, 1974). For the moon, hydrostatic head arguments are an essential ingredient to explanations of mascons (Wise and Yates, 1970; Wood *et al.*, 1970) and of the scarcity of farside maria (Kaula *et al.*, 1973). In this paper, simple hydrostatic concepts are extended to synthesize results from experimental petrology, lunar chronology, thermal history models, seismology, and gravity. Explanations of several heretofore unrelated features of lunar crustal structure and mare volcanism follow logically from these concepts.

We begin with the evidence for a currently popular scenario for early lunar history: initial melting of the outer several hundred kilometers of the moon, flotation of cumulate plagioclase from the cooling magma to form a lunar crust,

and solidification and later remelting of deeper cumulate layers to produce mare basalt liquids. Experimental petrological studies and thermal history models are used to quantify this scenario where possible. The key to understanding the hydrostatics of mare volcanism is the realization that mare basaltic liquids are denser than the early lunar crust. Thus basaltic liquids derived from shallow depths in the lunar mantle would not rise to the lunar surface except at the bottoms of deep impact basins. Rocks solidified from such liquids would be covered with younger liquids generated at greater depth in the mantle. From simple hydrostatics we derive a minimum depth of origin for mare basalts now on the lunar surface, some interesting conclusions about the efficiency of volcanism with time, a simple explanation for variations in the deep structure and possibly the surface heights of mascon maria, and some constraints on the relationship of non-mare basalts to the processes of mare basalt genesis.

A SCENARIO FOR EARLY LUNAR HISTORY

There is now a wide variety of evidence to indicate that at least the outer several hundred kilometers of the moon were heated to near-liquidus temperatures very early in lunar history. The hypothesis of a deep magma "ocean" covering the young moon was offered in different forms by Wood *et al.* (1970) and Smith *et al.* (1970) to explain a thick, feldspar-rich crust in the lunar highlands as the product of crystal–liquid fractionation on a global scale. Alternative explanations for the highland crust involving heterogeneous accretion of the moon can be effectively rebutted on geochemical grounds (Brett, 1973; Taylor, 1973). In addition, the Rb–Sr systematics of lunar rocks and soils require melting and differentiation of the outer few hundred kilometers of the moon at about 4.5 AE (Papanastassiou and Wasserburg, 1971, 1975), a conclusion consistent with U–Pb systematics as well (Tera and Wasserburg, 1974, 1975).

Subsequent to cooling and solidification of most, if not all, of the original magma ocean, later remelting of the deeper mafic cumulate layers produced the basaltic liquids to fill the maria (Hays and Walker, 1974). The titaniferous mare basalts returned from Apollo 11 and 17 sites, with Rb–Sr internal isochron ages of 3.6–3.8 AE (Tera *et al.*, 1974; Papanastassiou and Wasserburg, 1975), were derived from an olivine–pyroxene–ilmenite source at 100–170-km depth (Longhi *et al.*, 1974; Kesson, 1975). This source has been identified as a cumulate on the basis of the shallow source depth, the high-TiO_2 content of the source, and the relative abundances of rare-earth and other trace elements (Duncan *et al.*, 1974; Philpotts *et al.*, 1974; Hays and Walker, 1974). The low-titanium mare basalts returned from Apollo 12 and 15 sites, with Rb–Sr internal isochron ages of 3.1–3.4 AE (Papanastassiou and Wasserburg, 1974), were derived from an olivine–pyroxene source at 200–300-km depth (Green *et al.*, 1971; Longhi *et al.*, 1972; Grove *et al.*, 1973; Kesson, 1975). Whether this source region is also a mafic cumulate produced by crystal–liquid fractionation in the early magma ocean is still an open question from a petrological standpoint (Walker *et al.*, 1975), but is suggested by the Rb–Sr and U–Pb systematics (Papanastassiou and Wasserburg,

1971; Schnetzler and Philpotts, 1971; Tera and Wasserburg, 1975). The apparent deepening with time of the source region for mare basalts, consistent with a thickening of the lunar lithosphere, is a natural feature of most thermal history models for the moon (e.g. Toksöz and Solomon, 1973).

We shall examine the implications of hydrostatics for mare volcanism within the context of this lunar scenario, which has been developed in some detail by Taylor and Jakeš (1974), Smith and Steele (1975), Walker *et al.* (1975), Wood (1975), and others. The scenario is not universally accepted, and certainly minor, if not major, details will have to be altered to fit new data. Ringwood (1975), Green *et al.* (1975), and Ringwood and Green (1975) have challenged the hypothesis that mare basalts are derived by partial melting of cumulates. One of their arguments is that 74275, a relatively magnesian, high-TiO_2 basalt thought likely to match the composition of its parent basaltic liquid, apparently did not equilibrate with a residual Ti-oxide phase in its source region. This is in contrast to the high-pressure melting experiments on 70215, also a good candidate to be a primitive partial melt, by three independent groups (Longhi *et al.*, 1974; Kesson, 1975; Green *et al.*, 1975). It should perhaps be remarked that the cumulate remelting hypothesis outlined above, while the most plausible explanation of most high-pressure melting experiments on mare basalts, is not the only possible explanation. A separate controversy is that the implied correlation between TiO_2 content and age of mare basalts is in conflict with the suggestion from remote sensing and crater counting that on the moon's western hemisphere there are relatively high-TiO_2 mare surfaces, yet unsampled, that are younger than 3.2 AE (Boyce *et al.*, 1974). For purposes of discussion below, the most important aspects of the scenario as briefly outlined above are the deepening of the source region of mare basalts with time and the identification of particular samples of known age, chemistry, and mare site as likely primary melts from a determinable depth. With the exception of precise source depth, these elements of the scenario are probably valid.

Density of Lunar Rocks and Basaltic Liquids

In order to apply hydrostatics to the eruption of mare basalts, it is necessary to know the densities of lunar crustal material and of the basaltic liquids that take part in mare volcanism. For the lunar crust, we may estimate the density from measured and inferred densities of appropriate lunar rocks and from several gravitational considerations. For the basaltic liquids, their density may be calculated directly from their composition (Bottinga and Weill, 1970; Weill *et al.*, 1970).

It is a reasonable assumption that the lunar crust at the time of mare volcanism was grossly similar to the present highland crust. Direct measurements of density for highland rocks give values of 2.0–3.0 g/cm^3, with most values falling near 2.5 g/cm^3 (Talwani *et al.*, 1973). An upper bound to the density of typical highland rocks (Adler *et al.*, 1974), obtained both from combined density and porosity measurements (Talwani *et al.*, 1973) and from densities calculated from normative mineralogies (Solomon and Toksöz, 1968; Turkevich, 1971; Warren *et al.*, 1973), is

between 2.9 and 3.0 g/cm^3. From the negative gravity anomaly over the crater Langrenus in the highlands east of Mare Fecunditatis, Sjogren *et al.* (1974b) inferred a density of 3.05 g/cm^3 for the surrounding material, though this estimate is probably an upper bound since part of the present crater fill is likely to be of lower density than the adjacent bedrock. The material within the top few kilometers of the lunar crust is likely to be considerably less dense than its theoretical, pore-free value. Results from both the traverse gravimeter experiment (Talwani *et al.*, 1973) and the lunar seismic profiling experiment (Kovach and Watkins, 1973) on Apollo 17 indicate that the highland material underlying mare basalt at the Taurus-Littrow landing site is fractured and porous. A density contrast between highland rocks and mare basalt of about 0.8 g/cm^3 is suggested by the steep gradients in Bouguer gravity anomalies near the edges of the mare fill (Talwani *et al.*, 1973). The high porosity of the surface rocks is not completely eliminated at a depth of 20 km in the lunar crust (Toksöz *et al.*, 1973). In all, a value of about 2.8–2.9 g/cm^3 is probably reasonable for the average density of the highland crust, with perhaps the lower value preferred at the time of mare volcanism when the crust was several hundred degrees hotter than at present.

Once likely candidates for parent basaltic liquids are identified, the density of the liquids at liquidus temperatures may be estimated using the scheme of Bottinga and Weill (1970). Among the high-titanium mare basalts, 70215, 70017, and 74275 are most likely to have existed on the lunar surface as liquids (Longhi *et al.*, 1974; Green *et al.*, 1975; Shih *et al.*, 1975). Of the low-titanium mare basalts, 12009, 12002, 15016, and 15555 are among the stronger candidates to be primary melts (Green *et al.*, 1971; Longhi *et al.*, 1972; Grove *et al.*, 1973; Kesson, 1974). From the chemical analyses of these samples, densities of the respective basaltic liquids are shown as functions of temperature in Fig. 1.

The early lunar crust, as can be seen in Fig. 1, was lighter than the basaltic liquids that were derived from beneath it. This single fact governs much of the history of mare volcanism.

The densities of solid mare basalt and upper mantle material will also be necessary for the discussion below. The densities of mare basalts at low porosities are 3.3–3.5 g/cm^3 (Talwani *et al.*, 1973; Mizutani and Osako, 1974), with perhaps 3.4 g/cm^3 a representative value. Although the *in situ* density of the uppermost lunar mantle is unconstrained, the mean density and moment of inertia of the moon are consistent with upper mantle densities between 3.4 and 3.5 g/cm^3 (Solomon, 1974).

THICKNESS OF THE LUNAR CRUST

The crustal thickness in highland regions of the moon should have remained nearly constant since the last episode of major basin excavation, prior to substantial filling of the maria. Seismic refraction results have been inverted to obtain crustal structure in one relatively localized region on the lunar nearside (Toksöz *et al.*, 1973), and crustal thickness may be inferred in other regions from considerations of gravity and isostasy (Kaula *et al.*, 1974). Unfortunately, the

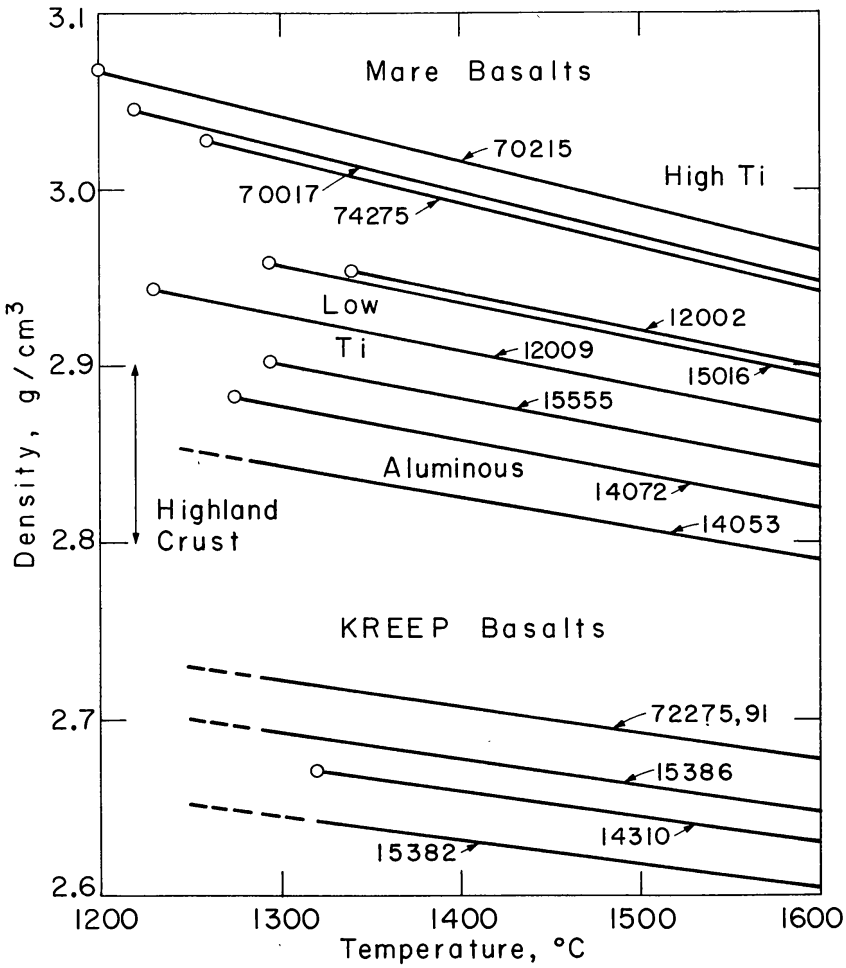

Fig. 1. Density of possible basaltic parent liquids for mare and non-mare basalts. Densities are calculated by the scheme of Bottinga and Weil (1970), using the following chemical analyses: 70215 and 70017 from Rose *et al.* (1974), 74275 from Duncan *et al.* (1974); 12002 from Willis *et al.* (1971), 12009 from Compston *et al.* (1971), 15016 and 15555 from Cuttitta *et al.* (1973), 14053 and 14072 from Reid and Jakeš (1974), 14310 from Rose *et al.* (1972), 15382 from Dowty *et al.* (1973), 15386 from Rhodes and Hubbard (1973), and 72275, 91 from Stoeser *et al.* (1974). The liquidus temperatures at zero pressure, where known (Longhi *et al.*, 1974; Kesson, 1975; Green *et al.*, 1971; Longhi *et al.*, 1972, 1974; Grove *et al.*, 1973; Walker *et al.*, 1972), are shown as circles.

seismic velocity profile cannot be unambiguously interpreted in terms of composition, with the result that highland crustal thickness is far from certain.

For the region including the Fra Mauro area and a portion of eastern Procellarum, the crust is about 60 km thick and is layered (Toksöz *et al.*, 1973). At about 20-km depth, there appears to be a major discontinuity separating materials of differing seismic velocity and velocity gradient. A sharp change in the microcrack density of crustal material at that depth is implied (Toksöz *et al.*, 1973; Wang *et al.*, 1973), though whether the discontinuity represents a superposed chemical change is not resolvable [see the discussion on this point in Toksöz *et al.* (1974) and Mizutani and Osako (1974)].

One interpretation of the 20-km discontinuity is that it marks the base of mare

basalt fill in eastern Procellarum (Toksöz *et al.*, 1973), and that highland material constitutes the lower crust between 20- and 60-km depth. Other interpretations allowed by the seismic velocity profile include both lesser and greater values for the thickness of mare basalt (Wang *et al.*, 1973).

There is spectral, geochemical, and photogeologic evidence in support of the view that the basaltic fill in at least the non-mascon maria is considerably less than 20 km thick. Head (1974b) has argued on stratigraphic grounds that the mare basalt layer overlies the Fra Mauro Formation at the Apollo 12 site and is unlikely to exceed 1 km in thickness. Most of the seismic wave paths on which the crustal profile is based are near the Apollo 12 and 14 sites (Toksöz *et al.*, 1974), and several paths bottom beneath non-mare regions in which the Fra Mauro is the surface formation. McCord *et al.* (1972) observed that the impacts that produced craters Copernicus and Aristarchus (initial crater diameters 72 and 32 km, respectively) apparently excavated highland material from beneath the mare basalt in Procellarum, but that the impact that produced Kepler (diameter 24 km) did not. This is consistent with a mare basalt fill of 2–3 km in central Procellarum, according to the crater depth–diameter relations of Pike (1967). On the basis of enhanced Al/Si, Podwysocki *et al.* (1974) and Andre *et al.* (1975) have identified several small craters in mare material near the edges of Serenitatis that appear to have ejecta containing highland material. A limit of about 1 km to the thickness of mare basalt fill is implied (cf. Talwani *et al.*, 1973), though this limit need have no validity for the central portions of Serenitatis.

We consider below two interpretations of the seismic velocity profile (Toksöz *et al.*, 1973) for the crust in eastern Procellarum. On the basis of the above discussion, the preferred interpretation is that (i) a layer of mare basalt less than 1 km thick overlies 60 km of material similar to rocks at the surface in highland regions. The second interpretation is that (ii) the 20-km discontinuity is a chemical as well as a textural change and that 20 km of mare basalt overlie 40 km of highland material.

The principle of isostasy, modified to include the effect of the mascons, can be used to infer crustal structure in other regions of the moon (Kaula *et al.*, 1974). Let ρ_h, ρ_m, and ρ_M be the densities of highland material, mare basalt, and lunar mantle material, respectively. Let h be the surface elevation with respect to a sphere about the lunar center of mass, z be the depth of isostatic compensation, t be the thickness of mare basalt fill, and T be the thickness of highland crustal material (Fig. 2). Following Kaula *et al.* (1974), we may write

$$\rho_m t_i + \rho_h T_i + \rho_M (z + h_i - t_i - T_i) = C_i, \tag{1}$$

where the subscript $i = 1, 2, 3$, and 4 stands for irregular maria, circular maria, nearside highlands, or farside highlands and where the C_i are constants satisfying $C_1 = C_3 = C_4$ and $C_2 = C_1 + 10^6$ g/cm^2. A variation of gravity with radius between the surface and a depth z has been ignored in writing (1). Under interpretation (i) above, t_1 is negligible. Using the values for h_i determined from laser altimetry

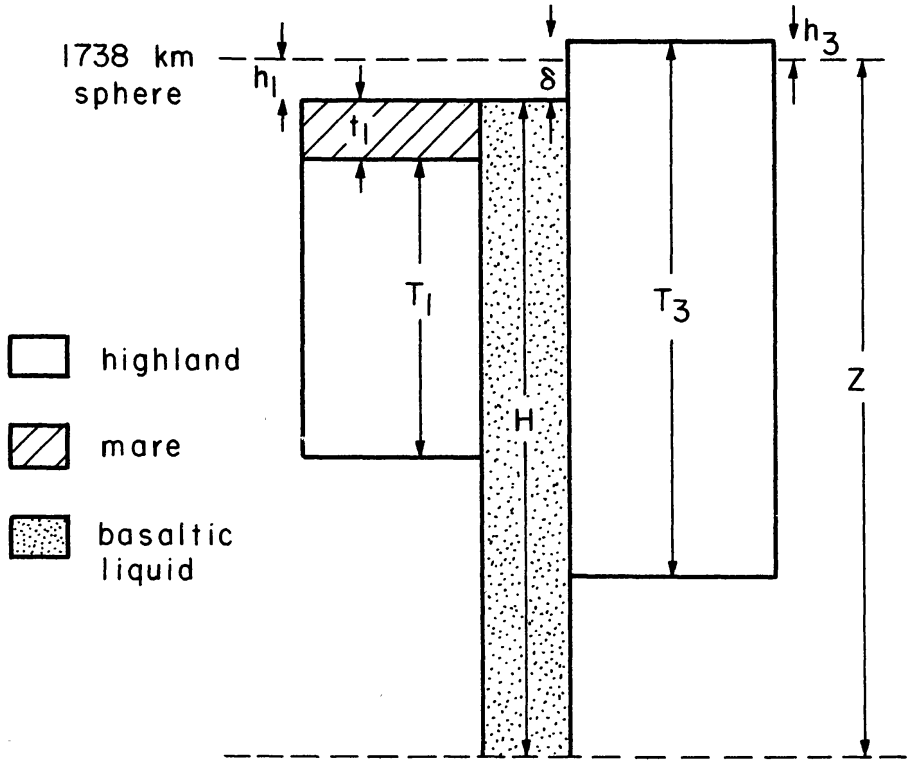

Fig. 2. A schematic view of lunar hydrostatics, illustrating several parameters introduced in the text.

(Kaula *et al.*, 1974), and using $\rho_m = \rho_M = 3.4\ \mathrm{g/cm^3}$, $\rho_h = 2.8\ \mathrm{g/cm^3}$, gives

$$\begin{aligned} T_1 &= 60\ \mathrm{km} \quad \text{(irregular maria, assumed)} \\ T_2 &= 34\ \mathrm{km} \quad \text{(circular maria)} \\ T_3 &= 65\ \mathrm{km} \quad \text{(nearside highlands)} \\ T_4 &= 83\ \mathrm{km} \quad \text{(farside highlands).} \end{aligned} \tag{2}$$

If the base of the crust beneath mascon maria is at the same level (with respect to the center of mass) as the base of the crust beneath non-mascon maria (Kaula *et al.*, 1974), then the mascon maria are filled by mare basalt to a depth of 25 km. If hypothesis (ii) for interpreting the seismic velocity profile is valid, then all thicknesses in (2) would be reduced by 20 km, and mascon maria would, by the above reasoning, have 45 km of mare basalt fill.

An alternative view of the difference in highland crustal thickness between nearside and farside is to regard the offset between the center of figure and the center of mass of the moon as due to a first-degree harmonic term in lunar crustal thickness (Kaula *et al.*, 1972). If this hypothesis holds, then

$$T_4 - T_3 = \frac{2x\bar{\rho}}{\rho_M - \rho_h}, \tag{3}$$

where x is the offset of centers and $\bar{\rho}$ is the mean density of the moon, 3.34 g/cm^3 (Kaula *et al.*, 1972). Using $x = 2.5$ km (Kaula *et al.*, 1974) and other parameters as

above gives

$$T_4 - T_3 = 28 \text{ km}. \tag{4}$$

This value is something of a maximum, in contrast to Eqs. (2) which represent mean values over the respective regions.

SOURCE DEPTH OF MARE BASALT VERSUS TIME

To quantify the discussion of source depths for mare basalts, we next integrate the experimental results on mare basalt petrogenesis with a thermal history model for the moon. The thermal model is calculated in a manner similar to that used by Toksöz and Solomon (1973) and Solomon and Toksöz (1973), except that in the present model, following the scenario outlined above, mare basalt liquids are generated by partial remelting of cumulates rather than partial melting of primitive lunar mantle.

We do not model explicitly the period of lunar history between formation of the large magma "ocean" at the lunar surface and cooling and fractionation of the magma to produce a layered cumulate structure in the outer several hundred kilometers of the moon (see Walker *et al.*, 1975; Wood, 1975). For our purposes it is sufficient that melting of the outer portions of the moon at or shortly after the time of lunar origin is not energetically implausible (Toksöz *et al.*, 1972) and that crystal fractionation and solidification of most, if not all, cumulate layers were essentially complete after 100 or 200 m.y. (Papanastassiou and Wasserburg, 1971, 1975; Tera and Wasserburg, 1974, 1975). Such efficient cooling probably demands that the last liquid to solidify was near the base of, or perhaps within, the lunar crust.

The "initial" temperature distribution is taken at 4.4 AE (Tera and Wasserburg, 1974, 1975), follows a basalt melting curve in the mafic cumulate layers and a conduction solution in the lunar crust, and is close to "primordial" temperatures below the greatest depth of initial melting. The depth to the bottom of the early magma "ocean" is not well constrained by physical or chemical data, and may lie almost anywhere between the source depth for low-Ti mare basalts and the center of the moon (Walker *et al.*, 1975; Smith and Steele, 1975). The thermal history of the upper 200 km or so of the moon, fortunately, is not especially sensitive to the depth of initial melting as long as that depth exceeds about 400 km.

The calculated remelting of cumulates with time to produce basaltic liquids will be sensitive to the adopted "melting curve." The solidus or liquidus of a single basaltic species is not appropriate. As a better approximation we have defined a basalt liquidus for the moon by connecting the points of multiple-phase liquidus saturation determined experimentally on mare basalt parent liquids. This melting curve, which will be used in the thermal history calculation to define melting of a basaltic component in the lunar mantle (Toksöz and Solomon, 1973), is shown in Fig. 3. A finite heat of fusion compensates for the simplification of one melting curve instead of both a solidus and a basalt liquidus.

Two other parameters are treated slightly differently from previous models of

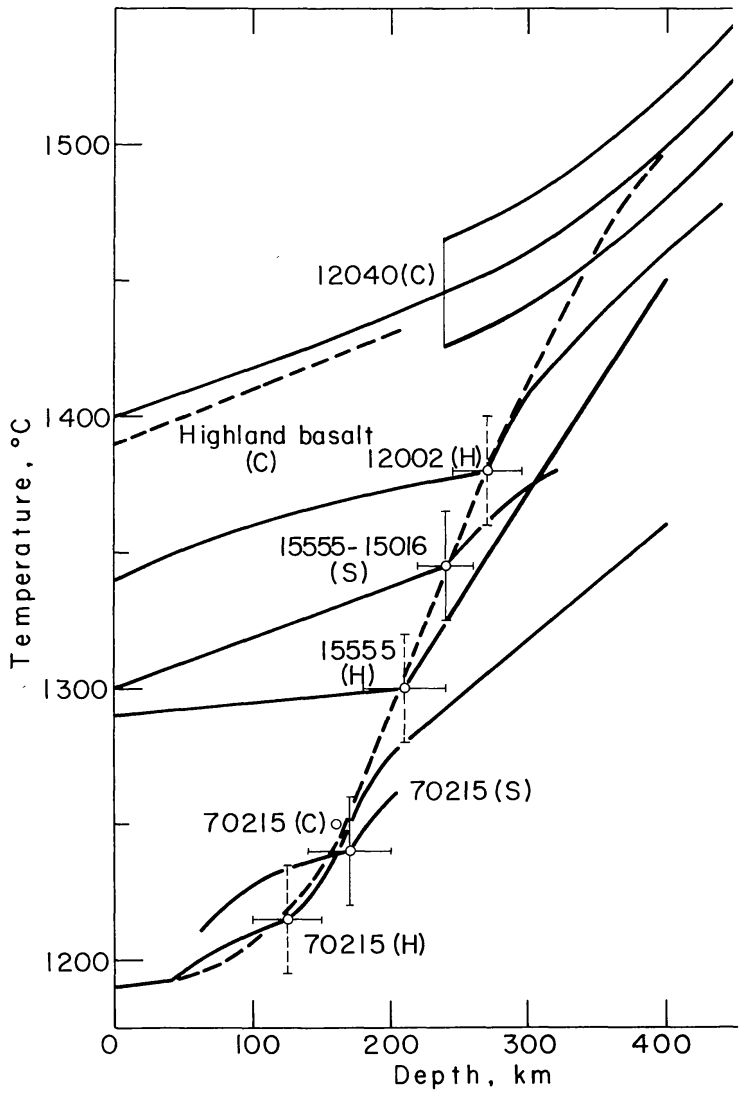

Fig. 3. Experimentally derived liquidus temperature versus depth curves for several possible primary basaltic magmas. The laboratory in which experiments were performed is denoted by C for Canberra (Green *et al.*, 1971), H for Harvard (Longhi *et al.*, 1972, 1974; Grove *et al.*, 1973), and S for Stony Brook (Kesson, 1975). A basalt liquidus for use in thermal history modeling (dashed line) is derived by connecting points of multiple phase liquidus saturation for each liquid (circles, with error bars as given by experiments or dashed if estimated). The liquidus of a "highland basalt" (Råheim and Green, 1974) is also shown.

evolution: thermal conductivity and heat sources per volume. The thermal conductivity of the lunar crust (Mizutani and Osako, 1974) is taken to be less than that appropriate to the lunar mantle (Schatz and Simmons, 1972); see Fig. 4. Radioactive heat sources will be segregated upward during crystal–liquid fractionation in the cooling magma "ocean." Presumably most of the potassium, uranium and thorium will be concentrated in the last liquid to solidify, but some of the heat sources will also be carried by intercumulus liquids to the mafic cumulate layers. In the particular thermal model presented here, the initial depth of the

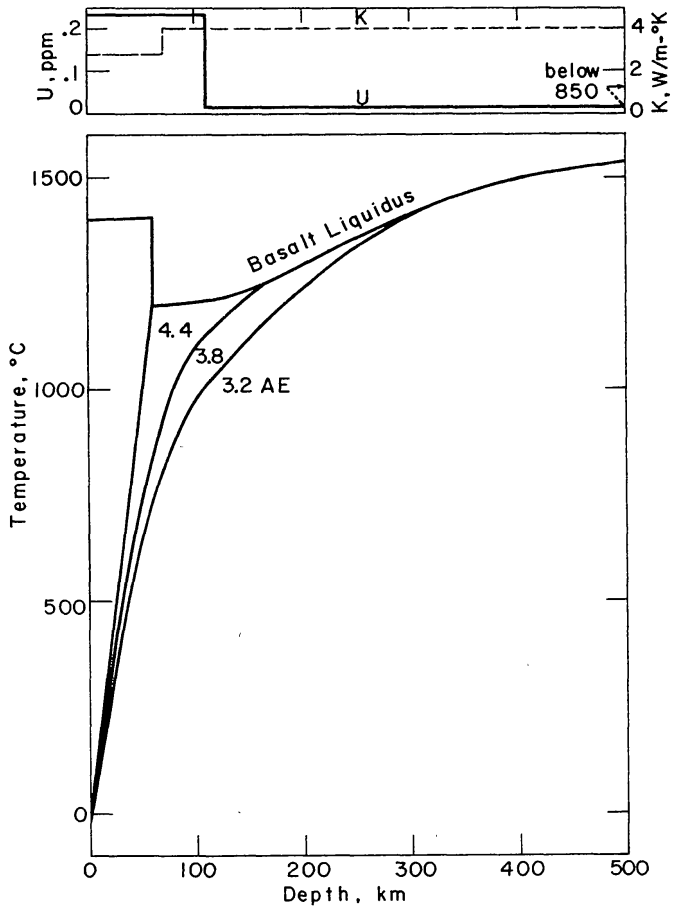

Fig. 4. Temperature versus depth in the moon at several distinct times, according to a thermal history model discussed in the text. The basalt liquidus follows the liquidus of "highland basalt" (Råheim and Green, 1974) in the crust and the mare basalt liquidus from Fig. 3 in the lunar mantle. The thermal conductivity K and the present-day uranium concentration U assumed in the model are shown at top.

magma ocean is taken as 850 km, four-fifths of the total heat sources within the volume above that depth are assumed to have been concentrated uniformly in the upper 100 km of the moon and the remaining one-fifth is distributed uniformly between 100- and 850-km depth. Present-day ratios of Th/U = 3.7 and K/U = 2000 are taken to be uniform, and the primordial (i.e. below 850 km) and bulk U abundances are such as to yield present-day concentrations of .06 ppm (Solomon and Toksöz, 1973, Toksöz *et al.*, 1973); see Fig. 4. The thermal history is qualitatively unchanged by altering the fraction of the heat sources segregated toward the lunar crust. It is only necessary that at least *some* heat sources (.01–.02 ppm U) be retained within the mafic cumulates. Heat transfer by solid-state creep is ignored in these calculations, probably a reasonable assumption for the earliest 1.5 b.y. of lunar history in view of the large spin-up time of convection cells in the lunar mantle (Cassen and Reynolds, 1973, 1974).

Temperature distributions at several times in the thermal model are shown in Fig. 4, and the deepening with time of the zone of shallowest partial melting is shown in Fig. 5, together with the inferred source depths and radiometric ages of

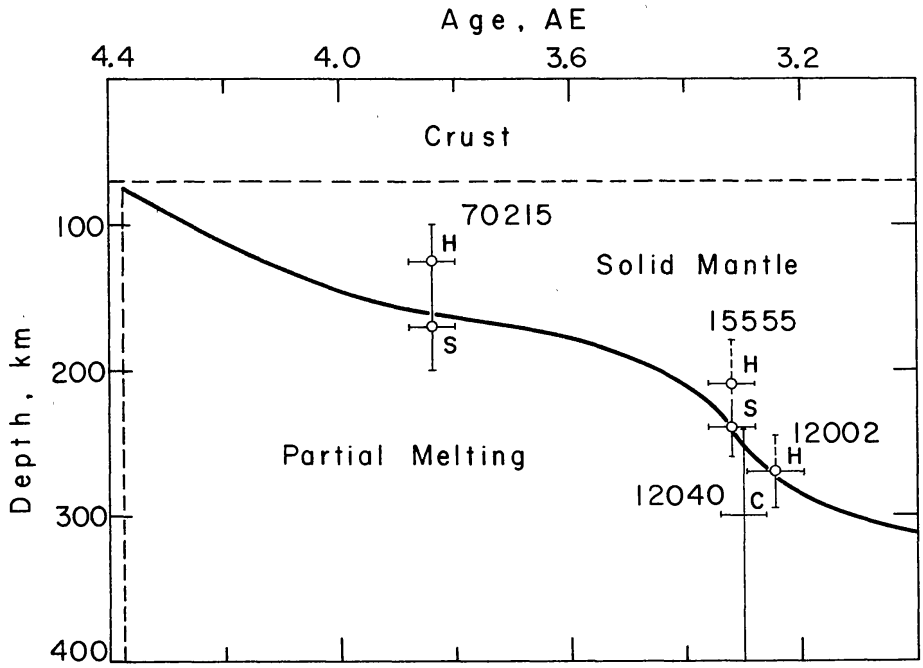

Fig. 5. Extent of the zone of partial melting versus time in the thermal history model of Fig. 4. Also shown are the inferred source depths and radiometric ages of several possible primary basaltic magmas. The information on source depth is as in Fig. 3. The ages are taken from Kirsten and Horn (1974) for 70215, Papanastassiou and Wasserburg (1973) for 15555, Papanastassiou and Wasserburg (1971) for 12040, and Turner (1971) for 12002.

several mare basalts thought likely to represent primary liquids. What is clear from both figures is that the generation of mare basalts by remelting mafic cumulates poses no special difficulties to a thermal model, contrary to statements of Ringwood (1975), Ringwood and Green (1975) and others, and that there is no need to postulate *ad hoc* heating mechanisms such as an extended close approach of the moon to the earth (Turcotte, 1975; Wones and Shaw, 1975). The essential ingredients of the thermal model are that the solidified cumulates are never far from solidus temperatures in the early history of the moon, the small amount of intercumulus heat sources serves to remelt the cumulate layers, and the upward segregation of most heat sources during fractionation of the magma "ocean" and the low thermal conductivity of the crust both serve to maintain upper mantle temperatures at high values. The details of the heat source fractionation and the thermal conductivity structure are somewhat arbitrary, and other values for these parameters can yield an equally good fit to the time-source depth relations for mare basalts portrayed in Fig. 5.

Hydrostatics: High-titanium Mare Basalts

The rise of basaltic liquids from their respective source depths to the lunar surface may be treated by hydrostatics in a manner similar to Eq. (1). By analogy to the earth (Eaton and Murata, 1960; Vogt, 1974), the base of a magma column is

taken to be the top of the partial melt zone. Considering for the moment only nearside highlands, dropping the subscript i in Eq. (1), and equating z to the shallowest depth of partial melting gives

$$\rho_b H = e[\rho_h T + \rho_M(z + h - T)] = eC, \tag{5}$$

where ρ_b is the density of a basaltic liquid, H is the height above the level z to which the liquid will rise, and e is an efficiency factor ($0 < e \leqslant 1$) to account for viscous head losses.

The shallowest depth of first remelting of cumulates in Fig. 5, and presumably in the moon, is immediately beneath the crust. Putting $z + h - T = 0$ in Eq. (5) gives

$$H \leqslant \frac{\rho_h T}{\rho_b}. \tag{6}$$

If these oldest, and most shallowly generated, basaltic liquids are chemically similar to the titaniferous mare basalts, then $\rho_b \simeq 3.1$ g/cm^3 (Fig. 1) and

$$H \leqslant 0.9T.$$

In a nearside crust 60–65 km thick, such early mare basalt liquids would rise to the lunar surface only in lowlands at least 6 km deeper than surrounding terrain, i.e. only in the largest impact basins. If, alternatively, the feldspar-rich portion of the nearside crust is only 40–45 km thick, then early basalt liquids would rise an additional 2 km higher with respect to the highland surface.

With time, the zone of partial melting deepened (Fig. 5), increasing the hydrostatic head of the later mare basalt liquids. Let δ be the local difference in elevation between the adjacent highland surface and the top of the magma column (Fig. 2). Then

$$\begin{aligned}\delta &= z + h - e\left[\frac{\rho_h}{\rho_b}T + \frac{\rho_M}{\rho_b}(z + h - T)\right] \\ &= e\left(\frac{\rho_M - \rho_h}{\rho_b}\right)T - \left(e\frac{\rho_M}{\rho_b} - 1\right)(z + h).\end{aligned} \tag{7}$$

If e is greater than ρ_b/ρ_M (about 0.9), a reasonable assumption for the shallow source depths and the low viscosities of high-Ti mare basalt liquids (Murase and McBirney, 1970), then δ will decrease as the melting depth z increases with time (Fig. 6). The mare basaltic lavas will rise to progressively higher levels in the crust, continuing to fill deep basins and eventually rising high enough to start filling shallower basins and irregular lowlands.

From albedo and reflectance spectra, dark titanium-rich mare deposits similar to those of the Taurus-Littrow area have been mapped at a number of widely scattered sites on the lunar surface (Pieters *et al.*, 1973; Charette *et al.*, 1974; Head, 1974a). A plausible inference is that many of these deposits are similar in age to Apollo 17 mare basalts (Head, 1974a), about 3.7–3.8 AE (Tera *et al.*, 1974). Thus by 3.7 AE, titaniferous mare basalt liquids rose to within 1–2-km elevation of the adjacent nearside highlands. From Eq. (7) we may infer a minimum depth of origin for such liquids ($e = 1$). Putting $T = 65$ km, $h = -1.4$ km, $\delta = 2$ km, $\rho_b =$

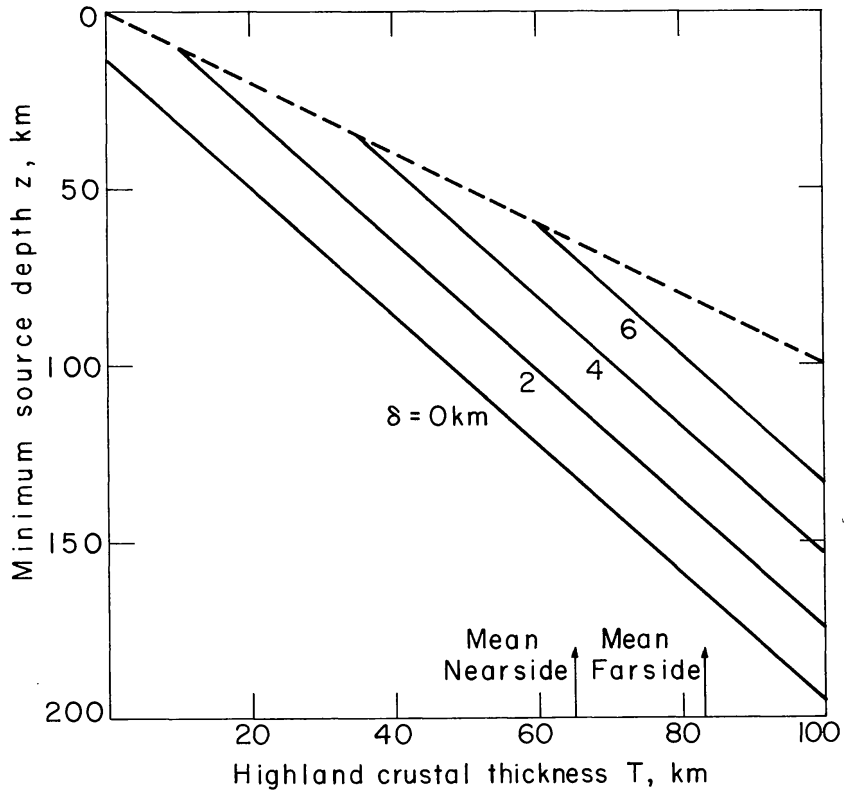

Fig. 6. The minimum source depth of mare basalt versus highland crustal thickness, for various values of the negative hydrostatic head δ of the magma column (see Fig. 2). The quantities ρ_h, ρ_b, and ρ_M are taken to be 2.8, 3.1, and 3.4 g/cm^3, respectively.

3.1 g/cm^3, and other parameters as above gives $z \geqslant 111$ km, or 109 km below the surface of adjacent highlands. This lower bound on z is consistent with the inferred source depth for 70215 (Longhi *et al.*, 1974; Kesson, 1975).

If mare-type basaltic liquid rose to fill deep impact basins on the lunar surface prior to 3.8 AE, the rocks solidified from such liquids were likely covered completely by later flows from magma columns with a greater hydrostatic head. The scarcity of mare basalts with internal isochron ages greater than about 3.8 AE (Papanastassiou and Wasserburg, 1975) is thus not due to any process more profound than the burial of older rocks by younger ones. Further, there is no need to postulate a delay between excavation of a large basin and eruption into the basin of mare basalts (e.g. Baldwin, 1970), but only a difference between the time of basin formation and the time when basaltic liquids were able to rise to levels near the present mare surface.

Both the offset of lunar centers and the paucity of farside maria are consistent with a backside highland crust thicker than on the frontside (Kaula *et al.*, 1972, 1973, 1974). At 3.7–3.8 AE, when titaniferous basalts from at least 100-km depth were rising on the lunar nearside to elevations roughly 2 km less than those of adjacent highland, Eq. (7), with $z = 111$ km, $T = 88$ km [Eqs. (2) and (4)], and $h = 1.8$ km (Kaula *et al.*, 1974) predicts $\delta = 6$ km for the farside highlands. Thus the large farside depression, about 6 km deep with respect to surrounding highlands, observed with the Apollo 15 laser altimeter (Kaula *et al.*, 1972) would

be expected to have negligible mare basalt fill from that episode of mare volcanism.

HYDROSTATICS: LOW-TITANIUM MARE BASALTS

Younger, low-titanium mare basaltic liquids were both less dense (Fig. 1) and derived from greater source depth (Fig. 5) than were the high-titanium mare basalts. On both grounds, Eq. (5) would predict a greater hydrostatic head for the low-Ti basaltic magmas than for the high-Ti magmas. To the contrary, the low-Ti mare basalt flows on the moon do not completely cover the flows of 500 m.y. earlier. These two facts can be reconciled if the efficiency e in Eq. (5) or Eq. (7) decreases with time. For most of the period of eruption of low-Ti mare basaltic magmas,

$$\left(\frac{\partial \delta}{\partial z}\right)_T = e\frac{\rho_M}{\rho_b} - 1 \simeq 0. \tag{8}$$

Taking $\rho_b = 2.95\ \mathrm{g/cm^3}$ (Fig. 1) and $\rho_M = 3.4\ \mathrm{g/cm^3}$ gives $e \simeq .87$.

The decreased efficiency is a measure of greater viscous head loss for the low-Ti mare basalt magmas. If we regard the magma column as a dike, moving at a mean velocity v between two fixed parallel walls separated by a distance h, then

$$e = 1 - \frac{12\eta v}{\rho_b g h^2}, \tag{9}$$

where η is the viscosity of the magma and g is the gravitational acceleration (Landau and Lifshitz, 1959; p. 56). Taking $e = .87$, $\eta = 10$ poise (Weill *et al.*, 1971), $\rho_b = 2.95\ \mathrm{g/cm^3}$ (Fig. 1) and $g = 162\ \mathrm{cm/sec^2}$ gives

$$\frac{v}{h^2} = .5\ \mathrm{cm^{-1}\ sec^{-1}}. \tag{10}$$

If the large lateral flow rates (on the order of 10^3 cm/sec) of lunar lavas, inferred from photogeologic studies of mare basalt flows (Schaber, 1973), are an appropriate measure of v in Eq. (10), then a dike thickness of about 0.5 m can account for the viscous head loss. The concept of a uniform dike extending from 200-km depth to the lunar surface is undoubtedly an oversimplification, but the decrease in e with increasing z is real and is probably due to a narrowing with depth of the channels through which magma can rise.

An apparent decrease in e with time might also be produced by a decrease in the volume of liquid available to feed basalt flows in mare basins or a reduction in the number of vents and pipes through which liquids can reach the lunar surface.

It is worth noting that as the source region of mare basalts deepens with time, the differences in hydrostatic head due to differences in crustal thicknesses are preserved. From Eq. (7),

$$\begin{aligned}\left(\frac{\partial \delta}{\partial T}\right)_z &= e\left(\frac{\rho_M - \rho_h}{\rho_b}\right) - \left(e\frac{\rho_M}{\rho_b} - 1\right)\left(\frac{\partial h}{\partial T}\right)_z \\ &= \left(\frac{\rho_M - \rho_h}{\rho_b}\right)\left(2e - \frac{\rho_b}{\rho_M}\right),\end{aligned} \tag{11}$$

an expression that is independent of z. (There is a second order dependence on z because of variations of gravity, compressibility, and thermal expansion with depth.) Thus basins unfilled at 3.8 AE are not likely to have been filled at 3.2 AE, since the difference in hydrostatic head between low-Ti and high-Ti basaltic magmas is small.

The Structure of Mascons

The above treatment of mare basalt hydrostatics is also relevant to the structure of mascon maria. For purposes of discussion, it is useful to distinguish between the anomalous mass distribution and the excess mass associated with a mascon. The *anomalous mass* is contained within that part of the sub-mascon structure that has a density different from laterally adjacent crust or mantle. Because mascon maria are topographic lows (Kaula *et al.*, 1974), the density contrast associated with the anomalous mass is generally positive. The *excess mass* is obtainable from the mascon gravity anomaly by Gauss's theorem (Kaula, 1969) and constitutes the superisostatic mass of the mascon mare structure.

Two different models have dominated thinking about mascon structure. In the model of Wood *et al.* (1970), most of the anomalous mass is contributed by a thick fill of mare basalt denser than surrounding basin material. In the model of Wise and Yates (1970), most of the anomalous mass is contributed by the upwelling of a dense mantle "plug" beneath the thinned crust of the mare basin. In both models, the excess mass is due to a "superisostatic" rise of basaltic magma to form the last several kilometers of mare basalt fill.

Interpretations of the gravity anomalies associated with frontside mascons, particularly Serenitatis and Crisium, have shown that the anomalous mass is relatively shallow and is confined in horizontal extent to the more deeply filled portions of the maria (Phillips *et al.*, 1972, 1974). This conclusion favors the model of Wood *et al.*, (1970) as the dominant contribution to the mascon structure. From a photogeologic study of the Orientale Basin, however, Head (1974b) concluded that only a small thickness of mare basalt had flooded that basin and therefore the Wise and Yates (1970) model is the preferred explanation of the Orientale mascon.

These two disparate interpretations can be understood in terms of mare basalt hydrostatics and a laterally variable crustal thickness. If the offset of lunar centers is due to a sinusoidal variation of highland crustal thickness with lunar longitude (Kaula *et al.*, 1972), then the thinnest highland crust should be near longitude $\lambda = 24°E$ (Kaula *et al.*, 1974), about the longitude of Serenitatis. The highland crust, with this assumption and Eq. (4), progressively thickens as $T(\lambda = 24°) + 14[1 - \cos(\lambda - 24°)]$ km. At the longitude of Orientale, the highland crust is then about 21 km thicker than in the Serenitatis area. From Fig. 6, the behavior in time and space of mare volcanism is strongly dependent on the local thickness of the highland crust and its controlling effect on the hydrostatic head of the basaltic magma column.

We propose that the deep structure of a mascon mare is dependent on the thickness of the original crust in which the mare basin was formed. Where the crust was relatively thin immediately prior to basin excavation, filling of the basin

with mantle-derived basalt began essentially at once and was a more rapid process than isostatic upwelling of the mantle. The anomalous mass associated with the present mascon in such maria (e.g. Serenitatis, Crisium, Imbrium) has a large contribution from the mare basalt fill. Where the crust was 20–30 km thicker just prior to mare basin excavation, subsequent mare basalt filling was either negligible or delayed until the basalt magma column acquired sufficient head to reach the bottom of the basin. In such basins, upwelling of the crust and mantle would be more efficient than mare volcanism at reducing the nonhydrostatic load. The anomalous mass associated with the mascon in such maria (e.g. Orientale) is primarily due to the mantle "plug." The mantle plug of itself does not produce a mascon. The excess mass must be mainly due to mare basalt fill, for several reasons. On the lunar farside, ringed basins unfilled by mare basalt appear to be characterized by either negative gravity anomalies or relative gravity lows (Ferrari, 1975). Further, the complete Orientale structure is characterized by a negative excess mass, with the positive (mascon) anomaly located only over the central portion of the basin now covered by mare basalt (Sjogren *et al.*, 1974a).

Probably a necessary, but not sufficient, condition for a mascon to form in a mare is that the basalt at the mare surface be relatively young, perhaps younger than 3.4 AE. Only after that time did the moon have a lithosphere sufficiently thick (Fig. 5) to support most of the mascon load for over 3 b.y.

From the Apollo laser altimeter and lunar sounder experiments, it has been noted that the surfaces of the mascon maria are topographically lower than those of the non-mascon maria and that the mascon mare surfaces are not at the same distance from the lunar center of mass (Kaula *et al.*, 1972, 1973, 1974; Brown *et al.*, 1974; Sjogren and Wollenhaupt, 1975). The generally lower elevation of mascon maria compared to other mare surfaces can be ascribed to some combination of post-volcanic solidification and thermal contraction of mare basalt, a process necessary for generating a mascon completely by basalt fill (Runcorn, 1974), and sinking of the mascon maria in response to the excess load (Arkani-Hamed, 1973). The variable departure of mascon mare surfaces from a 1736-km sphere about the lunar center of mass (Sjogren and Wollenhaupt, 1975) may also be related to one or both processes, though other mechanisms involving changes in the shape of equipotential surfaces with time have been proposed. If post-volcanic sinking of the mascon maria is responsible for the elevation differences among maria, then the excess masses of greatest lateral extent should have sunk the furthest, contrary to observation. If the amount of contraction of the mare basalt subsequent to the last volcanic flow is responsible, then some of the smallest basins (e.g. Smythii) would be inferred to be the deepest, an unlikely situation.

In view of the increased inefficiency with time of the youngest stage of mare volcanism noted above, it is not unlikely that the last volcanic fill in each mare, the main contributor to the mascon in the circular maria, did *not* rise to the same level in every location. If this should be the case, then the departure of the mascon surface from a sphere about the lunar center of mass would vary inversely with

the excess surface density attributed to the mascon. Such a variation is observed for Grimaldi, Serenitatis, Crisium, and Smythii (Sjogren *et al.*, 1974a,c; Muller *et al.*, 1974; Sjogren and Wollenhaupt, 1974), though the differences in elevation are larger than predicted by the differences in surface load by a factor of about three. Even in view of the uncertainties in estimating surface load from the shape and magnitude of the gravity anomaly, a variation in the extent of superisostatic fill among the mascon maria can be at most only a partial answer to their elevation differences, and other processes must also contribute.

Hydrostatics: Other Basalts

In addition to the basaltic liquids considered above, it is of interest to extend the concepts of hydrostatics to the parent liquids of other basaltic rocks returned from the lunar surface.

Reid and Jakeš (1974) identified a class of mare-like basalts higher in Al_2O_3 and lower in FeO than most mare basalt samples. These aluminous "mare" basalts have been recovered from several Apollo sites and have been dated at 3.3–4.0 AE (Reid and Jakeš, 1974; Ridley, 1975). It has been speculated that at least the oldest aluminous mare basalts may have been derived by partial melting in a plagioclase-bearing region intermediate in depth between the base of the crust and the source region for high-Ti mare basalts (Reid and Jakeš, 1974; Taylor and Jakeš, 1974). The hydrostatics of aluminous mare basalt liquids are marginally consistent with this idea. Liquids identical in composition to 14053 and 14072 are comparable to or somewhat greater in density than the highland crust (Fig. 1). Thus aluminous mare basalts might have partially filled many deep impact basins, only to be subsequently covered by younger high-Ti and low-Ti mare basalts. If this hypothesis is correct, then a large fraction of the volume of basaltic fill in the deeper maria may be aluminous mare basalt (Reid and Jakeš, 1974; Ridley, 1975).

The relationship of KREEP basalts to melting events in the lunar crust or mantle is not well established (Walker *et al.*, 1973). There are grounds for believing that Apollo 15 and 17 KREEP basalts may have been originated as volcanic magmas produced by partial melting of the lunar interior (Irving, 1975), though at least the Apollo 14 KREEP basalts are probably impact melts (Green *et al.*, 1972; Irving, 1975). Because the densities of KREEP basaltic liquids are substantially less than that of the lunar highland crust (Fig. 1), however, the likelihood of a volcanic origin for KREEP is lessened. If KREEP basaltic liquids were once concentrated more or less uniformly beneath or within the lunar crust either as an end-product of the initial episode of melting and fractionation or as an early partial melt (Taylor and Jakeš, 1974), then it is difficult to explain why such liquids did not rise to high levels throughout the lunar crust, producing a more spatially even distribution of KREEP on the present lunar surface than is observed (Metzger *et al.*, 1973). Local melting events, either impact-generated or volcanic, are more likely the mechanism for producing KREEP basalts.

CONCLUSIONS

The application of simple hydrostatics to mare volcanism on the moon, together with information from experimental petrology, thermal history, seismology, and gravity, can provide a rational basis for studying the behavior in time and in space of mare basaltic liquids and rocks. Age relationships among mare samples, source depths of basaltic liquids, and mare and highland crustal structure can all be quantitatively interrelated by these concepts. Some new ideas have emerged from this synthesis, including a demonstration that the efficiency of mare volcanism decreased with time, a simple explanation of structural differences among mascon maria, and an additional constraint on the origin of KREEP basalt liquids.

A philosophical thread throughout this discussion has been the implicit assumption that wherever possible lunar events should be interpreted in terms of a spherically symmetric moon. One obvious exception to this rule is the presence of discrete impact basins and associated ejecta. Another is the hypothesis of a pronounced lateral variation in the thickness of the early feldspar-rich lunar crust, plausibly attributed to convective processes in the cooling magma "ocean" (Lingenfelter and Schubert, 1973). Are there not additional lateral heterogeneities in chemistry or temperature of the lunar interior due to these or other effects? No doubt there are. The scenario and quantitative models adopted in this paper were intended to account for the characteristics of a large but still limited data set from a variety of lunar studies. These models can provide a basis for further testing of the lunar scenario.

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REFERENCES

Adler I., Trombka J., Gerard J., Lowman P., Schmadebeck R., Blodget H., Eller E., Yin L., Lamothe R., Osswald G., Gorenstein P., Bjorkholm P., Gursky H., and Harris B. (1972) Apollo 16 geochemical x-ray fluorescence experiment: Preliminary report. *Science* **177**, 256–259.

Andre C. G., Hallam M. E., Clark P. E., Podwysocki M. H., Weidner J. R., and Adler I. (1975) Synthesis of remote sensing data from the Tranquilitatis/Serenitatis area (abstract). In *Lunar Science VI*, p. 15–17. The Lunar Science Institute, Houston.

Arkani-Hamed J. (1973) Viscosity of the moon, 1, After mare formation. *The Moon* **6**, 100–111.

Baldwin R. B. (1970) Summary of arguments for a hot moon. *Science* **170**, 1297–1300.

Bottinga Y. and Weill D. F. (1970) Densities of liquid silicate systems calculated from partial molar volumes of oxide components. *Amer. J. Sci.* **269**, 169–182.

Boyce J. M., Dial A. L., and Soderblom L. A. (1974) Ages of the lunar nearside light plains and maria. *Proc. Lunar Sci. Conf. 5th*, p. 11–23.

Brett R. (1973) The lunar crust: A product of heterogeneous accretion or differentiation of a homogeneous moon? *Geochim. Cosmochim. Acta* **37**, 2697–2703.

Brown W. E., Jr., Adams G. F., Eggleton R. E., Jackson P., Jordan R., Kobrick M., Peebles W. J., Phillips R. J., Porcello L. J., Schaber G., Sill W. R., Thompson T. W., Ward S. H., and Zelenka J. S. (1974) Elevation profiles of the moon. *Proc. Lunar Sci. Conf. 5th*, p. 3037–3048.

Cassen P. and Reynolds R. T. (1973) Role of convection in the moon. *J. Geophys. Res.* **78**, 3203–3215.

Cassen P. and Reynolds R. T. (1974) Convection in the moon: effect of variable viscosity. *J. Geophys. Res.* **79**, 2937–2944.

Charette M. P., McCord T. B., Pieters C., and Adams J. B. (1974) Application of remote spectral reflectance measurements to lunar geology classification and determination of titanium content of lunar soils. *J. Geophys. Res.* **79**, 1605–1613.

Compston W., Berry H., Vernon M. J., Chappell B. W., and Kaye M. J. (1971) Rubidium–strontium chronology and chemistry of lunar material from the Ocean of Storms. *Proc. Lunar Sci. Conf. 2nd*, p. 1471–1485.

Cuttitta F., Rose H. J., Jr., Annell C. S., Carron M. K., Christian R. P., Ligon D. T., Jr., Dwornik E. J., Wright T. L., and Greenland L. P. (1973) Chemistry of twenty-one igneous rocks and soils returned by the Apollo 15 mission. *Proc. Lunar Sci. Conf. 4th*, p. 1081–1096.

Dowty E., Conrad G. H., Green J. A., Hlava P. F., Keil K., Moore R. B., Nehru C. E., and Prinz M. (1973) Catalogue of Apollo 15 rake samples from stations 2 (St. George), 7 (Spur crater) and 9a (Hadley rille). Spec. Pub. 8, University of New Mexico Institute of Meteoritics.

Duncan A. R., Erlank A. J., Willis J. P., Sher M. K., and Ahrens L. H. (1974) Trace element evidence for a two-stage origin of some titaniferous mare basalts. *Proc. Lunar Sci. Conf. 5th*, p. 1147–1157.

Eaton J. P. and Murata K. J. (1960) How volcanoes grow. *Science* **132**, 925–938.

Ferrari A. J. (1975) A comparison of near and farside lunar gravity (abstract). In *Lunar Science VI*, p. 260–262. The Lunar Science Institute, Houston.

Green D. H., Ringwood A. E., Ware N. G., Hibberson W. O., Major A., and Kiss E. (1971) Experimental petrology and petrogenesis of Apollo 12 basalts. *Proc. Lunar Sci. Conf. 2nd*, p. 601–615.

Green D. H., Ringwood A. E., Ware N. G., and Hibberson W. O. (1972) Experimental petrology and petrogenesis of Apollo 14 basalts. *Proc. Lunar Sci. Conf. 3rd*, p. 197–200.

Green D. H., Ringwood A. E., Ware N. G., and Hibberson W. O. (1975) Experimental petrology and petrogenesis of Apollo 17 mare basalts (abstract). In *Lunar Science VI*, p. 311–313. The Lunar Science Institute, Houston.

Grove T., Walker D., Longhi J., Stolper E., and Hays J. F. (1973) Petrology of rock 12002 and origin of picritic basalts at Oceanus Procellarum. *Proc. Lunar Sci. Conf. 4th*, p. 995–1011.

Hays J. F. and Walker D. (1974) Lunar igneous rocks and the nature of the lunar interior (preprint). *Proc. Soviet–American Conf. Cosmochemistry of the Moon and Planets* (Moscow). The Lunar Science Institute, Houston.

Head J. W. (1974a) Lunar dark-mantle deposits: Possible clues to the distribution of early mare deposits. *Proc. Lunar Sci. Conf. 5th*, p. 207–222.

Head J. W. (1974b) Some geologic observations concerning lunar geophysical models (preprint). *Proc. Soviet–American Conf. Cosmochemistry of the Moon and Planets* (*Moscow*). The Lunar Science Institute, Houston.

Irving A. J. (1975) Chemical and petrographic characteristics of non-mare melt rocks: Implications for lunar impact and volcanic processes (abstract). In *Lunar Science VI*, p. 426–428. The Lunar Science Institute, Houston.

Kaula W. M. (1969) Interpretation of lunar mass concentrations. *Phys. Earth Planet. Interiors* **2**, 123–137.

Kaula W. M., Schubert G., Lingenfelter R. E., Sjogren W. L., and Wollenhaupt W. R. (1972) Analysis and interpretation of lunar laser altimetry. *Proc. Lunar Sci. Conf. 3rd*, p. 2189–2204.

Kaula W. M., Schubert G., Lingenfelter R. E., Sjogren W. L., and Wollenhaupt W. R. (1973) Lunar topography from Apollo 15 and 16 laser altimetry. *Proc. Lunar Sci. Conf. 4th*, p. 2811–2819.

Kaula W. M., Schubert G., Lingenfelter R. E., Sjogren W. L., and Wollenhaupt W. R. (1974) Apollo laser altimetry and inferences as to lunar structure. *Proc. Lunar Sci. Conf. 5th*, p. 3049–3058.

Kesson S. (1974) Petrogenesis of high-Ti mare basalts: Melting of 70215 in vacuum and at high pressures (abstract). *EOS* (*Trans. Amer. Geophys. Union*) **55**, 1200.

Kesson S. E. (1975) Melting experiments on synthetic mare basalts and their petrogenetic implications (abstract). In *Lunar Science VI*, p. 475–477. The Lunar Science Institute, Houston.

Kirsten T. and Horn P. (1974) Chronology of the Taurus-Littrow region III: Ages of mare basalts and highland breccias and some remarks about the interpretation of lunar highland rock ages. *Proc. Lunar Sci. Conf. 5th*, p. 1451–1475.

Kovach R. L. and Watkins J. S. (1973) Apollo 17 seismic profiling: Probing the lunar crust. *Science* **180**, 1063–1064.

Landau L. D. and Lifshitz E. M. (1959) *Fluid Mechanics*. Pergamon Press, London.

Lingenfelter R. E. and Schubert G. (1973) Evidence for convection in planetary interiors from first-order topography. *The Moon* **7**, 172–180.

Longhi J., Walker D., Stolper E. N., Grove T. L., and Hays J. F. (1972) Petrology of mare/rille basalts 15555 and 15065. In *The Apollo 15 Lunar Samples* (editors J. Chamberlain and C. Watkins) p. 131–134. The Lunar Science Institute, Houston.

Longhi J., Walker D., Grove T. L., Stolper E. M., and Hays J. F. (1974) The petrology of the Apollo 17 mare basalts. *Proc. Lunar Sci. Conf. 5th*, p. 447–469.

McCord T. B., Charette M. P., Johnson T. V., Lebofsky L. A., Pieters C., and Adams J. B. (1972) Lunar spectral types. *J. Geophys. Res.* **77**, 1349–1359.

Metzger A. E., Trombka J. I., Peterson L. E., Reedy R. C., and Arnold J. R. (1973) Lunar surface radioactivity: Preliminary results of the Apollo 15 and Apollo 16 gamma-ray spectrometer experiments. *Science* **179**, 800–803.

Mizutani H. and Osako M. (1974) Elastic-wave velocities and thermal diffusivities of Apollo 17 rocks and their geophysical implication. *Proc. Lunar Sci. Conf. 5th*, p. 2891–2901.

Muller P. M., Sjogren W. L., and Wollenhaupt W. R. (1974) Lunar gravity: Apollo 15 doppler radio tracking. *The Moon* **10**, 195–205.

Papanastassiou D. A. and Wasserburg G. J. (1971) Lunar chronology and evolution from Rb–Sr studies of Apollo 11 and 12 samples. *Earth Planet. Sci. Lett.* **11**, 37–62.

Papanastassiou D. A. and Wasserburg G. J. (1973) Rb–Sr ages and initial strontium in basalts from Apollo 15. *Earth Planet. Sci. Lett.* **17**, 324–337.

Papanastassiou D. A. and Wasserburg G. J. (1975) A Rb–Sr study of Apollo 17 boulder 3: Dunite clast, microclasts, and matrix (abstract). In *Lunar Science VI*, p. 631–633. The Lunar Science Institute, Houston.

Phillips R. J., Conel J. E., Abbott E. A., Sjogren W. L., and Morton J. B. (1972) Mascons: Progress toward a unique solution for mass distribution. *J. Geophys. Res.* **77**, 7106–7114.

Phillips R. J., Conel J. E., and Sjogren W. L. (1974) The nature of circular maria based on gravity studies. *Space Research* **XIV**, 635–645.

Philpotts J. A., Schuhmann S., Kouns C. W., Lum R. K. L., and Winzer S. (1974) Origin of Apollo 17 REE Patterns. *Proc. Lunar Sci. Conf. 5th*, p. 1255–1267.

Pieters C., McCord T. B., Zisk S., and Adams J. B. (1973) Lunar black spots and nature of the Apollo 17 landing area. *J. Geophys. Res.* **78**, 5867–5875.

Pike, R. J. (1967) Schroeter's rule and the modification of lunar crater impact morphology. *J. Geophys. Res.* **72**, 2099–2106.

Podwysocki M. H., Weidner J. R., Andre C. G., Bickel A. L., Lum R. S., Adler I., and Trombka J. I. (1974) The application of trend surface analysis to a portion of the Apollo 15 x-ray fluorescence data. *Proc. Lunar Sci. Conf. 5th*, p. 3017–3024.

Råheim A. and Green D. H. (1974) Experimental petrology of lunar highland basalt composition and applications to models for the lunar interior. *J. Geology* **82**, 607–622.

Reid A. M. and Jakeš P. (1974) Luna 16 revisited: The case for aluminous mare basalts (abstract). In *Lunar Science V*, p. 627–629. The Lunar Science Institute, Houston.

Rhodes J. M. and Hubbard N. J. (1973) Chemistry, classification, and petrogenesis of Apollo 15 mare basalts. *Proc. Lunar Sci. Conf. 4th*, p. 1127–1148.

Ridley W. I. (1975) Petrology of aluminous mare basalts in breccia 14063 (abstract). In *Lunar Science VI*, p. 668–670. The Lunar Science Institute, Houston.

Ringwood A. E. (1975) Some aspects of the minor element chemistry of lunar mare basalts. *The Moon* **12**, 127–157.

Ringwood A. E. and Green D. H. (1975) Mare basalt petrogenesis (abstract). In *Lunar Science VI*, p. 677–679. The Lunar Science Institute, Houston.

Rose H. J., Jr., Cuttitta F., Annell C. S., Carron M. K., Christian R. P., Dwornik E. J., Greenland L. P., and Ligon D. T., Jr. (1972) Compositional data for twenty-one Fra Mauro lunar materials. *Proc. Lunar Sci. Conf. 3rd*, p. 1215–1229.

Rose, H. J., Jr., Cuttitta F., Berman S., Brown F. W., Carron M. K., Christian R. P., Dwornik E. J., and Greenland L. P. (1974) Chemical composition of rocks and soils at Taurus-Littrow. *Proc. Lunar Sci. Conf. 5th*, p. 1119–1133.

Runcorn S. K. (1974) On the origin of mascons and moonquakes. *Proc. Lunar Sci. Conf. 5th*, p. 3115–3126.

Schaber G. G. (1973) Lava flows in mare Imbrium: Geologic evaluation from Apollo orbital photography. *Proc. Lunar Sci. Conf. 4th*, p. 73–92.

Schatz J. F. and Simmons G. (1972) Thermal conductivity of earth materials at high temperatures. *J. Geophys. Res.* **77**, 6966–6983.

Schnetzler C. C. and Philpotts J. A. (1971) Alkali, alkaline earth, and rare-earth element concentrations in some Apollo 12 soils, rocks, and separated phases. *Proc. Lunar Sci. Conf. 2nd*, p. 1101–1122.

Shih C. Y., Wiesmann H. W., and Haskin L. A. (1975) On the origin of high-Ti mare basalts (abstract). In *Lunar Science VI*, p. 735–737. The Lunar Science Institute, Houston.

Sjogren W. L. and Wollenhaupt W. R. (1975) Lunar maria figure (abstract). In *Lunar Science VI*, p. 750–752. The Lunar Science Institute, Houston.

Sjogren W. L., Wimberly R. N., and Wollenhaupt W. R. (1974a) Lunar gravity via the Apollo 15 and 16 subsatellites. *The Moon* **9**, 115–128.

Sjogren W. L., Wimberly R. N., and Wollenhaupt W. R. (1974b) Lunar gravity: Apollo 16. *The Moon* **11**, 35–40.

Sjogren W. L., Wimberly R. N., and Wollenhaupt W. R. (1974c) Lunar gravity: Apollo 17. *The Moon* **11**, 41–52.

Smith J. V. and Steele I. M. (1975) Comprehensive petrologic model based on total melting of low-volatile ultra-basic moon (abstract). In *Lunar Science VI*, p. 753–755. The Lunar Science Institute, Houston.

Smith J. V., Anderson A. T., Newton R. C., Olsen E. J., and Wyllie P. J. (1970) A petrologic model for the moon based on petrogenesis, experimental petrology, and physical properties. *J. Geology* **78**, 381–405.

Solomon S. C. (1974) Density within the moon and implications for lunar composition. *The Moon* **9**, 147–166.

Solomon S. C. and Toksöz M. N. (1968) On the density distribution in the moon. *Phys. Earth Planet. Interiors* **1**, 475–484.

Solomon S. C. and Toksöz M. N. (1973) Internal constitution and evolution of the moon. *Phys. Earth Planet. Interiors* **7**, 15–38.

Stoeser D. B., Marvin U. B., Wood J. A., Wolfe R. W., and Bower J. F. (1974) Petrology of a stratified boulder from South Massif, Taurus-Littrow. *Proc. Lunar Sci. Conf. 5th*, p. 355–377.

Talwani M., Thompson G., Dent B., Kahle H., and Buck S. (1973) Traverse gravimeter experiment. In *Apollo 17 Preliminary Sci. Report*, NASA publication SP-330, pp. 13-1 to 13-13.

Taylor S. R. (1973) Chemical evidence for lunar melting and differentiation. *Nature* **245**, 203–205.

Taylor S. R. and Jakeš P. (1974) The geochemical evolution of the moon. *Proc. Lunar Sci. Conf. 5th*, p. 1287–1305.

Tera F. and Wasserburg G. J. (1974) U–Th–Pb systematics on lunar rocks and inferences about lunar evolution and the age of the moon. *Proc. Lunar Sci. Conf. 5th*, p. 1571–1599.

Tera F. and Wasserburg G. J. (1975) The evolution and history of mare basalts as inferred from U–Th–Pb systematics (abstract). In *Lunar Science VI*, p. 807–809. The Lunar Science Institute, Houston.

Tera F., Papanastassiou D. A., and Wasserburg G. J. (1974) Isotopic evidence for a terminal lunar cataclysm. *Earth Planet. Sci. Lett.* **22**, 1–21.

Toksöz M. N. and Solomon S. C. (1973) Thermal history and evolution of the moon. *The Moon* **7**, 251–278.

Toksöz M. N., Solomon S. C., Minear J. W., and Johnston D. H. (1972) Thermal evolution of the Moon. *The Moon* **4**, 190–213, and *The Moon* **5**, 231–232.

Toksöz M. N., Dainty A. M., Solomon S. C., and Anderson K. R. (1973) Velocity structure and evolution of the moon. *Proc. Lunar Sci. Conf. 4th*, p. 2529–2547.

Toksöz M. N., Dainty A. M., Solomon S. C., and Anderson K. R. (1974) Structure of the moon. *Rev. Geophys. Space Phys.* **12**, 539–567.

Turcotte D. L. (1975) Could an earth–moon close approach be responsible for mare formation? (abstract). In *Lunar Science VI*, p. 824–825. The Lunar Science Institute, Houston.

Turkevich A. L. (1971) Comparison of the analytical results from the Surveyor, Apollo and Luna missions. *Proc. Lunar Sci. Conf. 2nd*, p. 1209–1215.

Turner G. (1971) ^{40}Ar–^{39}Ar ages from the lunar maria. *Earth Planet. Sci. Lett.* **11**, 169–191.

Vogt P. R. (1974) Volcano height and plate thickness. *Earth Planet. Sci. Lett.* **23**, 337–348.

Walker D., Longhi J., and Hays J. F. (1972) Experimental petrology and origin of Fra Mauro rocks and soil. *Proc. Lunar Sci. Conf. 3rd*, p. 797–817.

Walker D., Grove T. L., Longhi J., Stolper E. M., and Hays J. F. (1973) Origin of lunar feldspathic rocks. *Earth Planet. Sci. Lett.* **20**, 325–336.

Walker D., Longhi J., and Hays J. F. (1975) Differentiation of a very thick magma body (abstract). In *Lunar Science VI*, p. 838–840. The Lunar Science Institute, Houston.

Wang H., Todd T., Richter D., and Simmons G. (1973) Elastic properties of plagioclase aggregates and seismic velocities in the moon. *Proc. Lunar Sci. Conf. 4th*, p. 2663–2671.

Warren N., Trice R., Soga N., and Anderson O. L. (1973) Rock physics properties of some lunar samples. *Proc. Lunar Sci. Conf. 4th*, p. 2611–2629.

Weill D. F., McCallum I. S., Bottinga Y., Drake M. J., and McKay G. A. (1970) Mineralogy and petrology of some Apollo 11 igneous rocks. *Proc. Apollo 11 Lunar Sci. Conf.*, p. 937–955.

Weill D. F., Grieve R. A., McCallum I. S., and Bottinga Y. (1971) Mineralogy–petrology of lunar samples. Microprobe studies of samples 12021 and 12022; viscosity of melts of selected lunar compositions. *Proc. Lunar Sci. Conf. 2nd*, p. 413–430.

Willis J. P., Ahrens L. H., Danchin R. V., Erlank A. J., Gurney J. J., Hofmeyr P. K., McCarthy T. S., and Orren M. J. (1971) Some interelement relationships between lunar rocks and fines, and stony meteorites. *Proc. Lunar Sci. Conf. 2nd*, p. 1123–1138.

Wise D. U. and Yates M. T. (1970) Mascons as structural relief on a lunar 'Moho'. *J. Geophys. Res.* **75**, 261–268.

Wones D. R. and Shaw H. R. (1975) Tidal dissipation: a possible heat source for mare basalt magmas (abstract). In *Lunar Science VI*, p. 878–880. The Lunar Science Institute, Houston.

Wood J. A. (1975) Lunar petrogenesis in a well-stirred magma ocean (abstract). In *Lunar Science VI*, p. 881–883. The Lunar Science Institute, Houston.

Wood J. A., Dickey J. S. Jr., Marvin U. B., and Powell B. N. (1970) Lunar anorthosites and a geophysical model of the moon. *Proc. Apollo 11 Lunar Sci. Conf.*, p. 965–988.