

ON THE ORIGIN OF THE MOON BY ROTATIONAL FISSION

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Abstract. Based on simple CIPW norms for the proposed terrestrial upper mantle material, it is shown that if the Moon fissioned from the Earth and gravitationally differentiated, it could have a 72 km thick anorthosite (An₉₇) crust, a calcium poor (3.8% by weight) pyroxenite upper mantle 100 Mg/Mg + Fe = 75 to 80) ending at a depth of 313 km and a dunite (Fo₉₃₋₉₅) lower mantle below a depth of 313 km. Refinements of these simple norm models, based on the cooling history, crystallization sequence and the variations of the 100 Mg/Mg + Fe ratio of the liquid and crystals during the crystallization sequence, indicate that the final form of such a Moon could have the following properties: (1) a primitive, cumulate anorthosite – minor troctolite crust with intrusive and extrusive feldspathic basalts and KREEP rich norites; the thickness of this crust would be 75 km; (2) a zone in the bottom of the crust and the top of the upper mantle which is rich in KREEP, the incompatible elements, silica, and possibly volatiles; this zone would be the source area for the upland feldspathic basalts, KREEP rich norites and KREEP and silica rich fluids; (3) an upper mantle between the depths of 75 km and 350 to 400 km which consists of peridotite containing 80–85% pyroxene (Wo₁₀En₆₈₋₇₂Fs₁₈₋₂₂) and 15–20% olivine (Fo₇₅₋₈₀); the Al₂O₃ content of the upper mantle is ~3%; the peridotite layer would be the source area for mare basalts and; (4) a lower mantle below a depth of 350–400 km which consists of dunite (Fo₉₃₋₉₇).

The cooling history of such a moon indicates that the primitive anorthosite crust would have been completely formed within 10⁸ yr after fission. The extrusion and intrusion of upland basalts and KREEP rich norites and the metamorphism of the crustal rocks via KREEP and silica rich fluids would have ended about 4 × 10⁹ yr ago when cooling well below the solidus reached a depth of 150 km. As cooling continued, the only source of magmas after 4 × 10⁸ yr ago would have been the peridotite upper mantle, i.e. the source area of the mare basalts. Extrusion of mare basalts ended when cooling below the solidus reached the top of the refractory dunite lower mantle 3–3.3 × 10⁹ yr ago.

Thus, it is shown that the chemistry, primary lithology, structure and developmental history of a fissioned Moon readily match those known for the real Moon. As such, the models presented in this paper strongly support the fission origin of the Moon.

1. Introduction

In a paper published in 1963, Wise revived Darwin's fission hypothesis of the origin of the Moon by suggesting that fission occurred as a consequence of the change of the moment of inertia of a rapidly rotating proto-Earth during core formation. Wise (1969) and O'Keefe (1969) later modified this simple version of fission by considering the possible effects of silicate evaporation from an intensely heated Earth on the chemistry and evolution of the Earth-Moon system. In addition to these versions of the core-formation fission hypothesis, Ringwood, (1960, 1970) has developed a precipitation hypothesis, in which he suggests that the Moon condensed from a hot silicate

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atmosphere of the primitive Earth. The most attractive feature of these various theories is that they explain why the bulk composition of the Moon is similar to that of the Earth's mantle and not to the Earth as a whole. Secondly, Ringwood's precipitation hypothesis has the apparent advantage of explaining the small differences between the proposed bulk composition of the Moon and the Earth's upper mantle. This is possible because the Moon would not have formed directly from a piece of detached mantle, but would have passed through an intermediate stage of evolution, via evaporation and precipitation, before assuming its final chemical form. There are dynamical objections to all of these various fission hypothesis, objections which have been reviewed and, at least, qualitatively answered by Wise (1963, 1969) for his version of fission and, to a lesser degree, by Ringwood for his precipitation hypothesis (1960, 1970).

However, with the general exception of Ringwood's highly modified version of fission, the fission hypothesis has been regarded as implausible by most investigators because of the apparent differences between the composition of the lunar rocks and their terrestrial counterparts (Hubbard and Gast, 1971) and, more importantly, their respective source areas, i.e. the lunar interior and the terrestrial upper mantle (Ringwood and Essene, 1970; Green, 1971; Green and Ringwood, 1973). Thus, accretion in a geocentric orbit (Ganapathy *et al.*, 1970; Hubbard and Gast, 1971; Anders *et al.*, 1971; Morgan *et al.*, 1972) or accretion in a heliocentric, but near Earth orbit and later capture (Anderson, 1973) have been regarded as the most plausible modes of lunar origin. While any geocentric orbit-accretion hypothesis is devoid of celestial mechanics problems, it has difficulties in explaining the complete or nearly complete absence of a lunar iron core and the depletion of both the siderophile and the volatile elements. These differences between the Earth and Moon require that the Moon accreted under different physio-chemical conditions and on a different time scale than did the Earth (Ganapathy *et al.*, 1970; Anders *et al.*, 1971; Morgan *et al.*, 1972). While mechanisms have been proposed (Ganapathy *et al.*, 1970; Anders *et al.*, 1971; Morgan *et al.*, 1972) to account for these differences, these mechanisms are satisfactorily only if one assumes that the Moon contains at least several percent of free iron and that the loss of water is not essentially complete: neither of these conditions are met by the Moon (see Section 2). Anderson (1973) has attempted to circumvent these difficulties by assuming that lunar accretion occurred in highly inclined, heliocentric orbit at 1 AU and that the Earth later captured the Moon. This model requires two dynamically unlikely events to explain the origin of the Moon. While perhaps plausible, the Anderson model is statistically very unlikely. Thus, accretional models for the lunar origin have also not been completely successful in accounting for the apparent bulk composition of the Moon.

However, it is shown in this paper that the bulk chemistry of the Moon may be essentially identical to that of the Earth's upper mantle and that, in addition to the depletion of siderophiles; (1) the depletion of the volatiles; (2) the various primary igneous units found on the Moon; and (3) the time of formation of the primary rock units are the expected results of the fission of the Moon from a hot proto-Earth during core formation.

2. General Lunar Properties

Mainly as a result of the Apollo program, we now have a large number of physical, chemical and petrological constraints which any lunar origin theory must meet. First, as has long been known, the Moon's mean density (3.36 g cm^{-3}) indicates that it must consist mainly of ferro-magnesian silicates and cannot contain more than 1% by mass in an iron core without violating both the mass and moment of inertia boundary conditions (Solomon and Toksöz, 1973). In addition to being deficient in free iron, the lunar siderophile elements are in general depleted by factors of 3 to 10 with respect to the Earth (Ringwood and Essene, 1970).

Similarly the volatile elements are strongly depleted by factors of 3 to 100 (Ringwood and Essene, 1970) and the Moon is apparently anhydrous (Engel and Engel, 1970). The lunar rocks show no evidence of having any water associated with their parent magmas as is indicated by the absence of hydroxial bearing minerals (amphiboles or micas) (Mason and Melson, 1970) and deuteric alteration of early formed minerals. Some 'rusty' rocks have been found, but the goethite in these rocks is generally attributed to exogenous water brought to the Moon by cometary impacts (E. Goresy *et al.*, 1973). It is probably true that the lunar magmas were also very deficient in any gaseous constituent (e.f. CO_2), since pumiceous rocks, which would be expected from volatile rich magmas extruded into a vacuum, are absent or generally absent from the lunar samples. Thus, the Moon is not only deficient in volatile elements, but apparently has either lost, or never had any atmospheric forming gases. This accounts for the Moon's extremely low ($10^{-13.5}$ atm) fugacity of oxygen (Ringwood and Essene, 1970) and the lack of ferric iron (Engel and Engel, 1970).

While there are proposed differences between the Earth and Moon with respect to chemical composition, studies of stable isotopes (e.g., $\text{Si}^{30}/\text{Si}^{28}$ and $\text{O}^{18}/\text{O}^{16}$) show that the Earth and Moon are similar (Epstein and Taylor, 1972).

3. Lunar Lithology and Structure

From the active seismic experiments carried out with the Apollo network, Latham and his coworkers (Latham *et al.*, 1972; Toksöz *et al.*, 1972) have shown that in the area centered on the Apollo 12 and 14 landing sites, the lunar crust consists of an upper mare unit of basalt 25 km thick and a lower crust which ends at a depth of 60–65 km. The depth of the mare unit is consistent with that expected for hydrostatic equilibrium between an anorthositic crust and mare basalt flooding of impact basins (i.e. 25 km (Wood, 1970)) and with the results obtained for the thickness of the mascons (i.e. 20 km (Sjogren *et al.*, 1972)). As is consistent with the seismic velocity data, the crust below the mare fill is most probably a continuation of the anorthositic upland crust (Latham *et al.*, 1972; Toksöz *et al.*, 1972) in which the mare basins have been formed. Based on the 2 km center of figure-center of mass offset of the Moon, Lingenfelter and Schubert (1973) have suggested that the crust on the far side of the Moon is ~ 30 km thicker than on the near side. Thus, the average thickness of the crust may be ~ 75 km.

Excluding for the moment the maria, the ancient crust consists of anorthosites (Kridelbaugh *et al.*, 1973; Prinz *et al.*, 1973) and other plagioclase rich rocks. The anorthosites themselves are highly calcic ($Or_{Trace} Ab_{<5} An_{>95}$) and carry only minor olivine and pyroxene. When original textures are preserved, the anorthosites are clearly seen to be cumulates and hence are thought to be representative of a primitive, cumulate lunar curst (Wood, 1970). Troctolites (Prinz *et al.*, 1973) are also found to be an important crustal rock type. Their feldspars are very calcic ($Or_{Trace}, Ab_{<5}, An_{>95}$), their olivines are very magnesium rich (Fo_{92}), spinel is usually an important minor constituent (e.g. 6%) and like the anorthosites, they have a cumulate texture. Grabbroic anorthosite and anorthositic gabbros are prevalent in the uplands (Hubbard *et al.*, 1973); as a result, the mean plagioclase content of the crust is about 70–75%. In general, the plagioclase of these feldspathic basalts are very calcic, but not as calcic as those of the cumulate anorthosites and troctolites.

A second class of upland crustal rocks require special mention. These are the units rich in K, the rare Earth elements, P, U, Th, etc., i.e. the so-called KREEP basalts or uplands basalts (Hubbard *et al.*, 1972). These rocks generally are orthopyroxene bearing and are more appropriately called norites or feldspathic norites (Meyer *et al.*, 1971) and KREEP is greatly enriched in the 'granitic' or potassium and silica rich glasses found in rock 12013 (Hubbard and Gast, 1970). These KREEP rich norites and basalts also carry feldspars which are somewhat less calcic than the cumulate anortholites; so, the upland feldspathic basalts and KREEP rich feldspathic norites etc. may have formed later than the cumulates.

Thus, the crust consists of cumulate, calcic anorthosites and troctolites, with extrusive feldspathic basalts and norites. The source areas of the magmas of these materials are located at rather shallow depths (<200 km) (Hubbard *et al.*, 1972; Green *et al.*, 1972). The lower age limit of these feldspathic crustal materials, i.e. the age of their last stage of brecciation and metamorphism is $\sim 3.9 \times 10^9$ yr (Papanastassiou and Wasserburg, 1972). Thus, the formation and early, intense modification of the anorthositic crust occurred between 3.9×10^9 and 4.5×10^9 yr ago.

The mare fill, which comprises only about 10% of the volume of the crust, consists of basalts, ilmenite basalts, quartz basalts, olivine basalts, and picrite basalts, all of which have plagioclase contents ranging from 10–70% (LSPET, 1970, LSPET, 1972). Thus, the mare basalts are, in contrast to the remainder of the crust, extremely rich in mafics. Also, in comparison to terrestrial basalts, the lunar mare basalts are quite rich in the incompatible elements, though they do not approach KREEP materials in their content of these elements.

The ages of the mare units sampled range from 3.2×10^9 to 3.8×10^9 yr (Papanastassiou and Wasserburg, 1973; Kirsten *et al.*, 1973). Thus, the mare fill deposition was basically limited to the period after the development of the upland crust. However, it must be noted that these dates refer to the last units deposited at each site and so, the beginning of the filling of the basins most probably predates the 3.8×10^9 yr age of the oldest mare sample. This concept is supported by the observation that rock 12013 contains clasts similar to Apollo 11 breccias and the solidification or, more probably,

the metamorphic age of the rock is 4.0×10^9 yr (Albee *et al.*, 1970). Thus, mare basalts were found on the surface of the Moon at least as long ago as 4.0×10^9 yr.

In series of papers, Ringwood and his associates (Ringwood and Essene, 1970; Green *et al.*, 1971; Green and Ringwood, 1973) have shown that the mare basalt magmas were produced by partial melting of an orthopyroxene-clinopyroxene pyroxenite or olivine poor peridotite source material laying at depths between 200 and 500 km. The 100 Mg/Mg+Fe ratio of the source area must be between 75 and 80, its Al_2O_3 content is on the order of 3–5% and its CaO content is on the order of 3–8%. The identification of a pyroxenite or olivine poor peridotite upper mantle is not inconsistent with the active seismic experiments (Latham *et al.*, 1972; Toksöz *et al.*, 1972).

With the exception of preliminary passive seismic data (Nakamura *et al.*, 1973) there is no evidence which indicates the chemical nature of the materials below depths of 400 to 500 km. However, the available seismic data indicate that the *P* wave velocities are about 8 km s^{-1} and are nearly constant (or decrease by a few percent) with depth. Thus, the entire mantle of the Moon may consist of dunites, peridotites or pyroxenites. Further, the mantle is found to be solid down to a depth of 1000 to 1100 km. Below these depths *S* waves are strongly attenuated, so a small amount of melt may be mixed with the ultramafics in the lower part of the Moon.

4. Differentiated Pyrolite Moon

According to Wise's (1963) fission hypothesis, the separation of the Moon from the Earth occurred as a result of core formation and, hence, after most or all of the free iron and the siderophiles had been removed from the mantle. Thus, as has been long recognized, the newly fissioned Moon would already be deficient in these elements. It is reasonable to assume that during the time fission occurred the proto-Earth was completely molten due to the gravitational energy released by accretion and core formation itself. (Ringwood, 1960; Ringwood, 1970). Thus, the new Moon would also have been molten and probably somewhat above the liquidus. As will be discussed in Section 8, the Moon either remained above the liquidus or may have even heated up somewhat shortly after fission. As a result, the Moon may have been up to several 100's of degrees above the liquidus for a short time after formation and during this time the Moon would most probably have been strongly convecting and could have easily lost its water, CO_2 and the volatile metals discussed in Section 2. Thus, shortly after birth, such a Moon would consist mainly of the refractory elements and have a low fugacity of oxygen as is indicated by the lunar rocks.

However, the fissioned Moon would consist of pyroxenitic dunite, terrestrial upper mantle material which has a 100 Mg/Mg+Fe ratio of 88–89 (Ringwood, 1966; Ringwood and Green, 1967; Green, 1968; Ringwood and Essene, 1970), a value which is considerably higher than that of the olivine pyroxenite source region of the mare basalts (i.e. 75–80) (Ringwood and Essene, 1970; Green *et al.*, 1971; Green and Ringwood, 1973). This is not a fatal problem since the composition of the source region of

the mare basalts does not necessarily represent that of the entire interior of the Moon. At this stage of lunar evolution the Moon would begin to solidify and it is most instructive to see what kinds of, and the amount of materials that a gravitationally differentiated ball of terrestrial upper mantle material 3473 km in diameter could produce via normal fractional crystallization.

In an earlier series of papers, Ringwood and his associates (Ringwood, 1966; Ringwood and Green, 1967; Green, 1968) discussed the chemical composition of the terrestrial upper mantle and have called this material pyrolite. The major oxide components of pyrolite III are given in Table I. Of the various potential pyrolite compositions investigated, material of this composition seems best to represent the terrestrial upper mantle material and as such was used as a basis for this study.

TABLE I
Composition of pyrolite III

SiO ₂	45.16%	Na ₂ O	0.57%
MgO	37.47%	K ₂ O	0.13%
FeO	8.04%	Cr ₂ O ₃	0.43%
Fe ₂ O ₃	0.46%	TiO ₂	0.71%
Al ₂ O ₃	3.54%	P ₂ O ₅	0.06%
CaO	3.08%		

CIPW norms were calculated for pyrolite III which was modified slightly to take into account the lack of ferric iron in the lunar samples (Engels and Engels, 1970) (the Fe₂O₃ was recomputed as FeO, so that the total FeO is 8.45%) and for volatilization losses of Na₂O and K₂O (norms were computed for concentrations of Na₂O and K₂O of 0.3, 0.1 and 0.03 of those given for pyrolite). Also the norms were computed with the constraint that the pyroxenes would have 100 Mg/Mg+Fe of 75 and 80 to match that of the source area of the mare basalts. The magnesium to iron ratio of the olivine was left as a free parameter. The results of these norm calculations are given in Table II, from which it is clear that a pyrolite Moon is capable of having a calcic anorthosite crust and a calcium poor clinopyroxene-orthopyroxene upper mantle with 100 Mg/Mg+Fe between 75 and 80 as required by the petrology of the lunar mare rocks. The latter requirement is met simply if there is a magnesium rich dunite lower mantle, so that the bulk composition of the total mantle has 100 Mg/Mg+Fe = 88-89.

When allowances are made for the different densities of the minerals, these norms can be used to give the first order internal structure for pyrolite Moons. In these computations the small amounts of apatite, ilmenite and chromite are assumed to be trapped in the pyroxenes. The result of these calculations are given in Table III, from which it is clear that these models are not very sensitive to the amount of Na₂O and K₂O retained. Hence, only the model with 10% Na₂O and K₂O retention will be used for discussion.

Thus from Tables II and III the norm for a differentiated pyrolite Moon yields an anorthosite (An₉₇) crust 72 km thick, a pyroxenite (Wo₇En₇₀Fs₂₃-Wo₇En₇₄F₁₉)

TABLE II
CIPW norms for pyrolite moons with 30%, 10% and 3%
retention of original K₂O and Na₂O

Retention model	30%	10%	3%
Mineral	Weight percent		
Plagioclase	10.4 (Or ₁ Ab ₈ An ₉₁)	9.9 (Or _T Ab ₃ An ₉₇)	9.8 (Or _T Ab ₁ An ₉₉)
Pyroxene ^a	30.1 (Wo ₈ En ₆₉ Fs ₂₉) (Wo ₈ En ₇₄ Fs ₁₈)	32.0 (Wo ₇ En ₇₀ Fs ₂₃) (Wo ₇ En ₇₄ Fs ₁₉)	33.0 (Wo ₆ En ₇₁ Fs ₂₃) (Wo ₆ En ₇₅ Fs ₁₉)
Olivine ^a	56.5 (Fo ₉₅) (Fo ₉₃)	55.1 (Fo ₉₅) (Fo ₉₃)	54.2 (Fo ₉₅) (Fo ₉₃)
Apatite	0.1	0.1	0.1
Ilmenite	1.4	1.4	1.4
Chromite	0.6	0.6	0.6

^a Upper and lower values represent the mineral compositions for the models having pyroxenes with 100 Mg/Mg + Fe of 75 and 80, respectively.

TABLE III
Internal structure for pyrolite model moons

Retention model	Layer-mineralogy	Volume (%)	Radius to interface (R/R ₀)	Radius to interface (km)	Depth to interface (km)	Thickness of layer (km)
30%	Crust-anorthite	12.62	0.956	1661	76	76
	upper mantle-pyroxene;	30.83	0.827	1436	301	225
	lower mantle-olivine	56.55				1436
10%	Crust-anorthite	11.99	0.958	1665	72	72
	upper mantle-pyroxene;	32.92	0.820	1424	313	241
	lower mantle-olivine	55.09				1424
3%	Crust-anorthite	11.66	0.959	1667	70	70
	upper mantle-pyroxene;	33.47	0.819	1422	315	245
	lower mantle-olivine	54.87				1422

upper mantle which lies between the depths of 72 km and 313 km and a dunite (Fo₉₃₋₉₅) lower mantle below a depth of 313 km, a model which compares remarkably well with the known properties of the Moon (see Sections 2 and 3).

However, before proceeding with the discussion of this pyrolite model, there are two points to be made regarding variations of pyrolite models. The first is, that by trapping the ilmenite and chromite in the pyroxene layer, the 100 Mg/Mg + Fe ratio is reduced by about 3 units for the layer due to enrichment of FeO. Correcting for this 'enrichment' by leaving the ilmenite and chromite out of the system does not significantly change the depths of the interfaces and hence need not be considered further at this point. Potentially more important though, are the uncertainties in the composition

of pyrolite. If pyrolite I (Ringwood and Green, 1967) were the more likely candidate for the Earth's upper mantle, then (for the 10% retention model) the anorthositic crust would be 82 km thick and the depth to the bottom of the pyroxene layer would be only 218 km. The latter value is near the upper limit for the source area of the mare basalt magmas, but would be adequate as such. However, as will be clear from the discussions in Section 5 and as is acceptable (Ringwood and Essene, 1970; Green *et al.*, 1971; Green and Ringwood, 1973), the source area of the mare basalt magmas must contain olivine and, hence, the upper mantle for a pyrolite I Moon would considerably deeper than the 218 km derived from the norm calculations.

Returning to the more likely pyrolite III model, there are some other considerations which must be addressed before leaving this simple norm model. First, as discussed in Section 3, the lunar crust contains about 10% mare basalt and the remaining 90% is only about 70–75% anorthite. Since the available anorthite is adequate to account for the observed thickness of the crust, some 30 to 35% of the available CaO and Al₂O₃ from the anorthite must be located in the mantle. With a starting norm composition (by weight) of 3.8% CaO and no Al₂O₃ for the pyroxenes, the addition of this CaO and Al₂O₃ to the pyroxenes would increase their CaO content to 5.7% by weight and their Al₂O₃ content would become about 3.7% by weight. From Ringwood *et al.* (see Section 3), the Al₂O₃ content of the pyroxenite layer is 3–5%. Thus, the Al₂O₃ is nicely taken care of. Similarly the CaO content of the pyroxenites is 3–8%, so the 5.7% value is also quite satisfactory and results in the pyroxenes having a composition of about Wo₁₀En_{68–72}Fs_{18–22}.

5. Crystallization Sequence of a Pyrolite Moon

As suggested in the early part of Section 4, the newly fissioned Moon may have been above the liquidus by perhaps a few 100 °C. The details and time scale of the cooling of the Moon are discussed in Section 8, but the current discussion begins at the point, when the surface has reached the liquidus. Because the cooling of the molten Moon occurred only by radiation from its surface, it is reasonable to assume that crystallization would have also occurred only at the surface and that the crystals either sank or floated depending on their density. Thus, it is probably a very good approximation to use normal 1 atmosphere pressure studies to derive the lunar crystallization sequence. Since a pyrolite Moon is very magnesium and calcium rich, the equilibrium system anorthite-forsterite-silica (Figure 1) as described by Anderson (1915) can be used as a good approximation to the lunar system. Also because a pyrolite Moon is rich in normative olivine, we can use the forsterite-fayalite system (Figure 2) to determine approximately the changes in the 100 Mg/Mg + Fe ratio of the liquid and crystals for the early part of the crystallization history of such a Moon.

Treating all the FeO as MgO, all the Na₂O and K₂O as CaO and ignoring the small amounts of apatite, chromite and ilmenite and a 1.3% excess of CaO, a pyrolite Moon consists of 10.95% anorthite, 14.60% quartz and 74.45% forsterite and lies on the equilibrium diagram (Figure 1) at point A. As the melt cooled, forsterite would begin

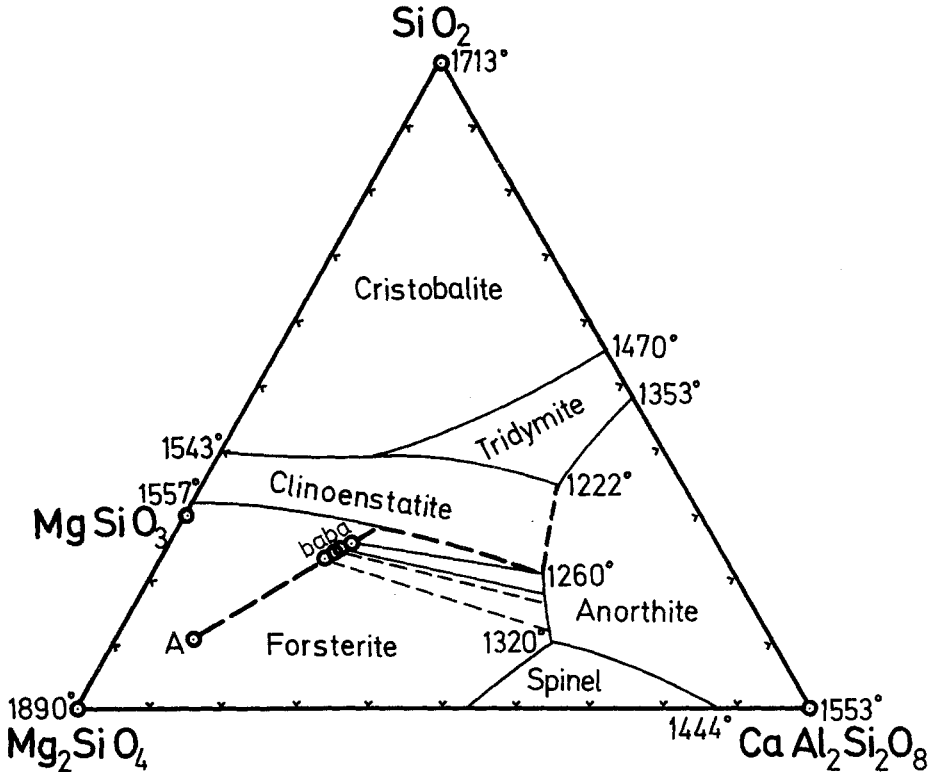


Fig. 1. The anorthite-forsterite-silica system as described by Anderson (1915). Point *A* is a good approximation to the lunar composition and the crystallization sequence for the Moon follows the heavy dashed line. Points *a* and *b* are the points, for different models, in the crystallization sequence where olivine stopped being removed from the melt and equilibrium crystallization began.

to crystallize out and fall, quite rapidly (see Section 7) towards the center of the Moon. In doing so, the early formed olivine would be removed from the system and not react with it. This process would continue and the composition of the liquid would move along the dashed line in Figure 1 until the boundary between forsterite and clinoenstatite was reached, at this point about 63% of the Moon would have solidified as olivine. Before proceeding further with the crystallization sequence, the variations in the magnesium-iron ratio of the crystals and residual liquid, and the isolation of the lower, dunite mantle must be addressed.

For a starting liquid with $100 \text{ Mg/Mg} + \text{Fe} = 88$, the first olivine crystals to form would have had a composition of $\text{Fo}_{97.5}$ (Figure 2). As crystallization proceeded, the composition of the newly formed crystals and the residual liquid would have become more iron rich. In order to yield the present Moon, this process must have continued until the residual melt had $100 \text{ Mg/Mg} + \text{Fe} = 75-80$, at which time the isolation of early formed olivine crystals in the lower mantle would have had to stop and the remaining liquid and the crystals formed after this time must have remained mixed and thus able to react – yielding the observed $100 \text{ Mg/Mg} + \text{Fe}$ ratio of $75-80$ throughout

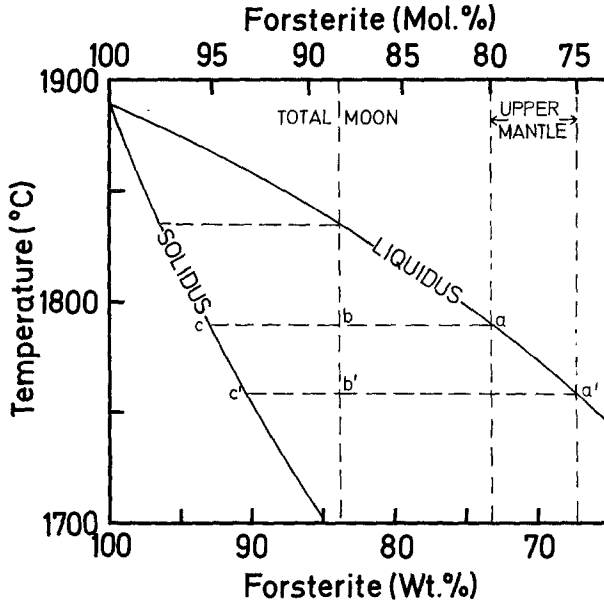


Fig. 2. The forsterite-fayalite system.

the upper mantle. At the point where olivine must have stopped being removed from the crystallizing system, the last olivine would have a composition of $Fo_{95.2}$ for a 100 Mg/Mg+Fe=80 upper mantle and $Fo_{93.0}$ for a 100 Mg/Mg+Fe=75 upper mantle (see Figure 2).

Now from Figure 2, the ratios $a-b/a-c$ and $a'-b'/a'-c'$ indicate that for equilibrium crystallization, 54% and 73% of the available ferromagnesium-silica liquid would have crystallized when the residual liquid had reached 100 Mg/Mg+Fe=80 and 75 respectively. However, for the case under consideration, i.e. disequilibrium crystallization due to removal of olivine from the residual melt, these percentages of olivine crystals would be reduced to about 49% and 62%, respectively. From the norm calculations, some 85% of the Moon consists of $(Mg, Fe) SiO_3$ and $(Mg, Fe) SiO_4$. So from these figures, and from Figure 3, which shows the evolution of olivine in the lower mantle during disequilibrium crystallization, it is apparent that about 42% of the Moon would have a mean composition of $Fo_{96.7}$ for the 100 Mg/Mg+Fe=80 upper mantle model, or about 53% would have a mean composition of $Fo_{96.4}$ for the 100 Mg/Mg+Fe=75 upper mantle model. Now, the norm calculations show that the Moon must contain 55% olivine in order for the SiO_2 to be properly used up. Thus, 2 to 13% of the olivine must have 100 Mg/Mg+Fe=75 to 80 and be in the upper mantle in order to satisfy the olivine to pyroxene ratios required by the chemistry. This simple correction yields mean olivine compositions of $Fo_{92.8}$ and $Fo_{95.5}$ for the two models, the values of which are in excellent agreement with the norm calculations.

Alternatively, the problem can be treated using the curves in Figure 2 and Figure 3. Considering the model with the upper mantle having 100 Mg/Mg+Fe=75, the norm

calculations require that the olivine has a mean composition of $Fo_{95.2}$. From Figure 3, the mean composition of the olivine precipitated in a lower mantle containing 55.1% of the lunar mass is $Fo_{96.3}$. Thus to a first order, the mixture of $Fo_{96.3}$ and $Fo_{75.0}$ need to give an average composition of $Fo_{95.2}$ is found to be 96% $Fo_{96.3}$ and 4% $Fo_{75.0}$. Thus, of the 55.1% olivine in the Moon, 52.8% must have a mean composition of $Fo_{96.3}$ (range $Fo_{97.4}$ to $Fo_{93.0}$) and 2.3% of the olivine must be mixed with the pyroxene and equilibrated at $Fo_{75.0}$ in the upper mantle.

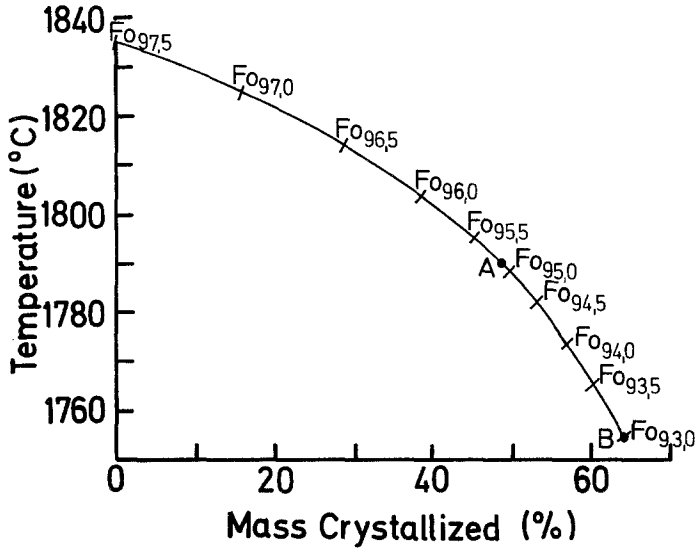


Fig. 3. Evolution of olivine during the solidification of the lower mantle. Disequilibrium crystallization of olivine stopped at point *A* for the models with an upper mantle 100 Mg/Mg + Fe ratio of 80 and at point *B* for the models with an upper mantle 100 Mg/Mg + Fe ratio of 75.

Table IV gives the results obtained for similar calculations done for both 100 Mg/Mg + Fe = 75 and 80 upper mantle models under consideration, those where the ilmenite and chromite have been left out of the system (see Section 4), and the results obtained first by considering the mass balance requirements within the MgO-FeO-SiO₂ system. Also included in Table IV is the ratio of olivine to olivine + pyroxene in the upper mantle as a result of the addition of the excess and equilibrated Fo_{75} or Fo_{80} to the upper mantle and the depth to the bottom of the resulting upper mantle. (It is noted here that similar consideration of the mantle composition would increase the depth of the bottom of a pyrolite I upper mantle, see p. 60). Thus from Table IV a pyrolite Moon would have a dunite (Fo_{96-97}) lower mantle below a depth of 330–440 km and a pyroxenite (with 6% olivine) to peridotite (with 29% olivine) upper mantle with the required 100 Mg/Mg + Fe ratio of 75–80. The addition of olivine to the upper mantle would dilute its Al₂O₃ and CaO content to 2.6–3.5% and 4.0–4.5%, respectively. Thus, if the amount of olivine in the upper mantle were as high a 30% its Al₂O₃ and CaO content might be a little low. Given the computed range of values

TABLE IV
Upper and lower mantle compositions

Basis for calculations	$\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}}$; upper mantle	Total olivine		Lower Mantle		Olivine (%) in upper mantle	100 OI Ol + Py in upper mantle	Depth to bottom of upper mantle (km)
		Mean comp.	Vol. (%)	Mean comp.	Vol. (%)			
MgO-FeO-SiO ₂ balance	75	FO _{96.5}	55.1	FO _{96.4}	53	2	6	330
	80	FO _{92.8}	55.1	FO _{96.7}	42	13	28	440
Mg to Fe ratios, simple norm models	75	FO _{95.2} ^a	55.1	FO _{96.3}	52.8	2.3	7	330
	80	FO _{93.9} ^a	55.1	FO _{96.8}	47.9	7.2	18	380
Mg to Fe ratios, norm models without ilm. and chr.	75	FO _{93.5}	56.9	FO _{96.5}	48.8	8.1	19	370
	80	FO _{91.8}	56.9	FO _{96.9}	43.5	13.4	29	420

^a Norm calculations

and considering the restrictions on the Al_2O_3 and CaO content and olivine content of the source area of the mare basalt magmas, the upper mantle is probably a peridotite consisting of 80–85% $\text{Wo}_{10}\text{En}_{68-72}\text{Fs}_{18-22}$ with 3% Al_2O_3 and 15–20% Fo_{75-80} , reaching a depth of 350–400 km.

Returning now to the crystallization sequence: from the above discussion the isolation of the early formed olivine from the melt must have ceased when 42–53% of the melt had crystallized and the remaining system of liquid and crystals would have continued evolve – but in equilibrium. The points where the isolation of the magnesium olivine stops for the 100 Mg/Mg+Fe = 75 and 80 upper mantles for both the simple norm models (a) and the norm models without ilmenite and chromite (b) are shown in Figure 1. Thus, continuing from the 42–53% solidification points, olivine would continue to crystallize until the boundary line between forsterite and clinoenstatite was intersected, at this point 63% of the original liquid would have solidified as olivine. Pyroxene would then begin to crystallize and that olivine which was not isolated, i.e. the latter 10–21% to crystallize, would begin to react with the liquid and the composition of the melt would begin to move along the boundary line towards the forsterite-anorthite-clinoenstatite peritectic. During this process and the peritectic crystallization, the norm calculations require that 8% olivine would be converted to pyroxene, leaving the required 55% total olivine with 2 to 13% of it being converted from Fo_{95} to Fo_{75-80} and being part of the upper mantle as discussed above.

In Figure 1, the lines (continuous for the 75 and dashed for the 80 upper mantle models) drawn radially from the MgSiO_3 point and through the points *a* or *b* (where the isolation of olivine ceased) intersect the anorthite-forsterite boundary line and not the forsterite-clinoenstatite boundary line; thus, the crystallization path passes through the forsterite-anorthite-clinoenstatite peritectic. This peritectic is reached when about 80% of the original liquid has solidified. From that point on, pyroxene and anorthite would crystallize out together. The anorthite would float up and form the primitive anorthositic crust and the pyroxene and previously, but late, formed olivine would sink to form the 100 Mg/Mg+Fe = 75–80 upper mantle. At this point, we now have the three layered Moon predicted from the norm models and which matches most of the known parameters of the real Moon.

Further consideration of the melt available after the time when the peritectic point is reached, yields some additional important points. First, the composition of the fluid near the peritectic point (see Table V) is very similar to that of rock 14310 and KREEP rich norite (Green *et al.*, 1972); thus, the experimental crystallization studies done on these materials (see for example Figures 1 and 2 of Green *et al.*, 1972) can be used to predict potential rock types derived by the fluids available when the anorthositic crust of the Moon was forming. From the diagrams, it is clear that in addition to the anorthosites, troctolites, norites, gabbroic anorthosites and anorthositic gabbros could also easily form from this fluid as late differentiation products. So, the final solidification of the Moon could produce a cumulate anorthosite (Wood, 1970) and troctolite crust (Prinz *et al.*, 1973) containing intruded and extruded norites, gabbroic anorthosites and anorthositic gabbros.

TABLE V
Composition of the remaining melt near the peritectic point,
Rock 14310 and KREEP

	Melt	14310	KREEP
SiO ₂	48.0	47.6	48.5
MgO	8.5	7.6	8.3
FeO	8.5	8.2	11.1
Al ₂ O ₃	19.2	20.7	16.7
CaO	11.8	12.5	11.5
Na ₂ O	0.3	0.7	0.6
K ₂ O	0.1	0.5	1.2
Cr ₂ O ₃	0.5	0.1	—
TiO ₂	1.0	1.2	2.0
P ₂ O ₅	0.1	0.4	0.6

Second, after final crystallization began and the anorthositic crust began to accumulate, the final portion of the melt would be trapped between the downward growing crust and the upward growing upper mantle. Thus the last bit of fluid, which would be rich in KREEP, the incompatible elements, residual silica and whatever volatiles the Moon had, would be concentrated between the upper mantle and the lower crust. As such, they would be enriched in the final feldspathic basalts injected into and extruded onto the crust and would be readily available to be injected as KREEP and 'granite' fluids (like those in 12013) into the earlier formed crustal rocks.

6. Trace Elements

While no quantitative studies of the partitioning of trace elements were made in this study, it is clear by inspection of the liquid/crystal partitioning coefficients (Hubbard and Gast, 1971) that the europium, barium and strontium will be enriched in the anorthositic crust, leaving the source areas of the mare basalts and feldspathic upland basalts and norites depleted in these trace elements.

Also, Ganapathy *et al.*, (1970) have discussed the depletion of Au, Re, Os, Co, Ir, Ni and Pd and concluded that the depletion pattern of these siderophile elements is 'unearthly' and hence that the Moon could not have lost its siderophile element while part of the Earth. They conclude therefore that the Moon could not have fissioned and suggest that the very low fugacity of oxygen of the Moon and lack of water may have resulted in a depletion pattern between these elements which is different from that found on the Earth. However, in the sequence of events discussed in Sections 4 and 5, the pyrolyte, fissioned Moon would have lost the bulk of its siderophiles before fission. Afterwards, during the initial phase of fractional crystallization, the remaining siderophiles would have undergone a second stage of evolution as they would have been carried to the center of the Moon along with what little iron (if any) was left in the terrestrial mantle before fission or with any iron formed from Fe₂O₃ and FeO as the Moon devolatilized. This second differentiation of the siderophiles would have occurred

ed under nonterrestrial conditions and hence could have produced the 'unearthly' pattern observed.

7. Crystal Settling

Since crystal settling is an important part of the hypothesis discussed above, it is necessary to determine if the settling times for crystals is compatible with the requirement that the early formed crystals are removed from the melt efficiently and on a time scale compatible with the cooling time scale discussed in Section 8. As a starting point we can assume that Stokes Law is a good approximation for computing the velocities of crystals moving through a viscous melt (Jaeger, 1968). We have then

$$V = \frac{2gr^2\Delta\rho}{9\eta}, \quad (1)$$

where V is the velocity, g the acceleration of gravity, r the radius of the crystal (spherical), $\Delta\rho$ the density difference between the liquid and the crystal and η the viscosity of the liquid. Now, for the Moon,

$$g_{(R)} \cong 162 \text{ cm s}^{-2} \frac{R}{R_0}, \quad (2)$$

where R is the distance of the crystal from the center of the Moon and R_0 the radius of the Moon; $\Delta\rho \cong 0.3 \text{ g cm}^{-3}$ for olivine crystals; and η can be roughly estimated from the work of Weill *et al.* (1971) on mare basalts. However, since the overall composition of the Moon is more basic than the mare basalts, these values for η are probably upper limits. From Weill *et al.*, η is a strong function of the temperature (T) of the melt and is given by

$$\log \eta = \frac{-6 + 1.1 \times 10^4}{T(^{\circ}\text{C}) + 273^{\circ}\text{C}}, \quad (3)$$

where the coefficients are averages of the values listed (Weill *et al.*, Table II). The temperature as a function of lunar radius can be obtained for the pyrolite solidus from Green and Ringwood (1967). Based on the liquidus-solidus relationships of pyroxenite (Ringwood and Essene, 1970) it is assumed that the pyrolite liquidus is about 300°C higher than the solidus.

Using all the above data, we can derive $V(R)$ for olivine crystals and hence fall times as a function of crystal radius for liquidus and near solidus temperatures. Remarkably, the fall time from the surface to the center of the Moon is only about 105 days for a 1 cm radius crystal at liquidus temperatures. Due to the low viscosity of the lunar magmas, crystals grow rapidly and hence 1 cm radius crystals are more than reasonable. Using this case as representative, Figure 5 gives the fall times to different depths in the Moon for a 1 cm radius crystal at liquidus temperatures. From Figures 4 and 5, it is apparent that early formed crystals could have been very

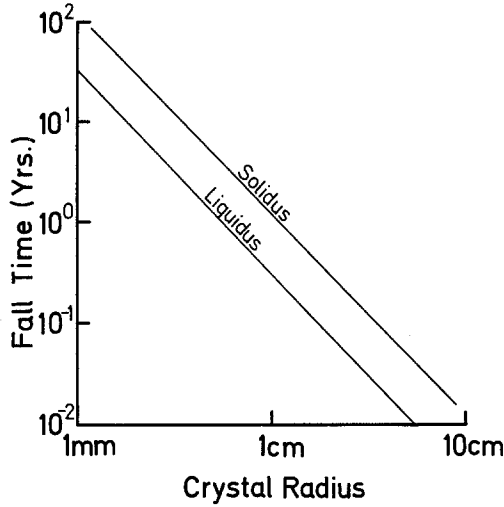


Fig. 4. Fall times to the center of the Moon for crystals, as a function of crystal radius, for crystals moving through a melt at liquidus and near solidus temperatures.

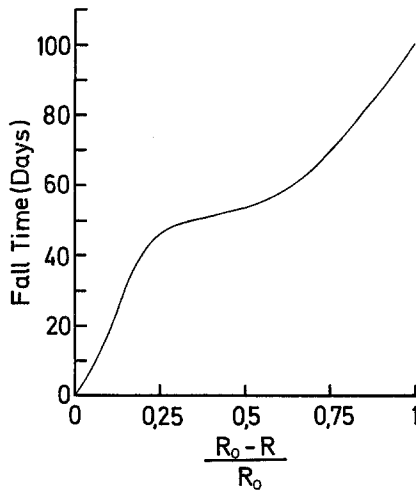


Fig. 5. Fall times to different depths in the Moon for 1 cm radius in a melt at liquidus temperatures.

rapidly removed from the melt by gravitational settling alone. Also, this process may have been even faster due to convective motion within the cooling melt.

8. Thermal History of a Fissioned Moon

As suggested in Section 4, the newly fissioned Moon probably was at or slightly above the liquidus just after fission. It was also suggested that the Moon may have also heated up immediately after fission so the early temperature of the Moon may have been 200°C to 500°C above the liquidus. There are at least two sources of heat available

to the Moon at this time. The first is that of ohmic heating. Sonett (personal communications) has shown that ohmic heating due to an early, strong solar magnetic field was capable of causing even small bodies, as far from the Sun as asteroids, to have partially melted. Now, just after fission, the Moon would have most probably been embedded in a very strong terrestrial magnetic field and as a result would have undergone ohmic heating. Based on Sonett's analysis it is not unreasonable to assume that the temperature of the Moon increased by at least a few hundred degrees due to this effect.

Second, at the moment of fission the Moon may have been rotating with a period of 2.65 hr; but by Wise's hypothesis the Moon would have been released from the small end of the Poincaré-shaped proto-Earth at a distance of ~ 3 Earth radii (Re) and thus had a period of revolution of ~ 7.3 hr. Tidal interactions with the rapidly rotating Earth (2.65 hr) would have immediately caused the distance between the Earth and Moon to increase and the period of rotation of both the Earth and Moon to decrease. From tidal theory (Mac Donald, 1966), the distance of the Moon from the Earth (R) as a function of time (t) can be written as

$$R = 60.2 Re \left(\frac{t + 15.7 \text{ yr}}{4.6 \times 10^9 \text{ yr}} \right)^{2/13}, \quad (4)$$

where $t=0$ and $R=3 Re$ at fission. Also, since the period (P) varies with $R^{3/2}$ we have

$$P = 656.5 \text{ hr} \left(\frac{t + 15.7 \text{ yr}}{4.6 \times 10^9 \text{ yr}} \right)^{3/13}. \quad (5)$$

From this equation, the period of revolution of the Moon would be 9 hr about 20 yr after fission. Now if the Moon achieved synchronous rotation within a few years after fission and maintained it from then on, the Moon would have lost almost all of its initial rotational energy within a few 10's of years after fission and this energy would have been dissipated in the interior of the Moon. The rotational energy of a Moon initially rotating in 2.65 hr is about 1.5×10^{35} ergs and this is enough to raise the temperature of the Moon by 200°C (for a mass of 7.36×10^{25} g and a specific heat of 10^7 ergs g^{-1}C). Thus, the new Moon was potentially heated up by a total of a few hundred degrees in the first few 10's of years after fission when it was near enough to the Earth to interact with the geomagnetic field and to have strong tidal interactions with the Earth.

However, at the same time, the Moon would have been radiating energy and assuming that the Moon was a perfect black body, the rate of heat loss was, using the Stefan-Boltzmann Law, 1.8×10^{26} ergs s^{-1} for a radiation temperature of 1700 K, i.e. at the liquidus, and 3.5×10^{26} ergs at 2000 K, i.e. 300 K above the liquidus. For a specific heat of 10^7 ergs gm^{-1}C , the Moon would have cooled from 2000 K to the liquidus in about 8.5×10^8 s or 27 yr. Thus, without a more detailed analysis of the thermal history of the first few 10's of years of Moon evolution it is difficult to know whether the Moon succeeded in heating up by a total of several hundreds of degrees over the liquidus or simply remained above the liquidus for some 30 or so years after

formation. Though the former case is more favorable, either case would probably have allowed a strongly convecting Moon to devolatilize.

Once the Moon stopped being heated by conversion of rotational energy into heat and by ohmic heating, the Moon's surface would have quickly reached the liquidus and olivine and later pyroxene would begin to crystallize. Assuming that no significant chill crust developed and the convection (aided by gravitational settling of crystals) efficiently kept the melt stirred, the Moon would have been radiating at 1.8×10^{26} ergs s^{-1} initially and at 8×10^{25} ergs s^{-1} later at near solidus temperatures. Now for a Δt of $\sim 300^\circ\text{C}$ between the liquidus and solidus, a specific heat of 10^7 ergs $g^{-1}^\circ\text{C}$ and a latent heat of fusion of 4.2×10^9 ergs g^{-1} , 80% of the Moon would have solidified in 100 yr. That is, the point in time, at which peritectic crystallization began and the anorthositic crust started to form was about 100 yr after crystallization began. It is to be noted here that the fall of time of crystals (Section 7) is compatible with such a short crystallization time and hence the crystallization sequence discussed in Section 5 is acceptable.

Once the anorthositic crust began to form, the surface temperature and hence radiation temperature would drop drastically and more importantly, conduction through the new crust would begin to dominate the transfer of heat from the still liquid layer, just below the crust, to the surface. When peritectic crystallization began, about 1.5×10^{25} g or 3.9×10^7 $g\text{ cm}^{-2}$ (of surface area) remained to be solidified. Thus, about 5×10^9 ergs g^{-1} would have been released by solidification and lowering the temperature to the solidus. Using a time average crystal thickness of 35 km and conductivity of $\sim 3 \times 10^5$ ergs $\text{cm}^{-1}\text{ s}^{-1}^\circ\text{C}$, the average rate of transfer of heat through the crust would have been about 1.0×10^2 ergs s^{-1} and about 6×10^7 yr would have been required to solidify the final 20% of the melt. Thus, the entire process of cooling to the solidus would have taken less than 10^8 yr.

Once the Moon was everywhere at the solidus – i.e. $\sim 10^8$ yr after fission or 4.5×10^9 yr ago – the upper part of the Moon would have begun to drop well below the solidus. It is pointed out here that as long as any part of the crust or mantle was at the solidus, magma could have been produced by partial melting if the confining pressure on the solid were reduced by a small amount e.g. by impact removal of part of the lithostatic load and by the accompanying fracturing of the lower crust and upper mantle. Thus, the cooling history of the Moon after 4.5×10^9 yr ago can be used to predict the depths from which magmas would originate as a function of time. Fortunately, Toksöz and Solomon (1973) have calculated the thermal history of a Moon which was at the solidus 4.6×10^9 yr ago, a model which is applicable for the remaining part of the thermal history of the fissioned Moon.

From Toksöz's and Solomon's (1973) Figure 7, cooling well below the solidus would have reached a depth of 150 km after 5×10^8 yr or 4×10^9 yr ago. From the discussion in Section 5, the zone between the bottom of the crust and a depth of 120–150 km is rich in trapped KREEP, incompatible elements, etc. and is the source area of the feldspathic upland basalts and norites (Hubbard *et al.*, 1972; Green *et al.*, 1972). Thus, because of the fall of the temperature well below the solidus by 4×10^9 yr ago, the

metamorphism of upland rocks by KREEP etc. fluids and the intrusion and extrusion of KREEP and upland basalts and norites would have stopped by that time. Whatever magmas reached the surface after 4×10^9 yr ago would have come from depths greater than 150 km, i.e. from the source area of the mare basalt magmas. Cooling to the bottom of the peridotite upper mantle, i.e. to a depth of 350–400 km would have taken another 7 to 10×10^8 yr, and so mare basalts could have been extruded only between $3.0\text{--}3.3 \times 10^9$ and 4.0×10^9 yr ago. Thus, the cooling history predicted for the pyrolite Moon would produce the exact age sequence seen in the lunar rocks, without resorting to an early phase of superficial heating, followed by a pulse of radioactive heating (Wood, 1972). Also, the shut-off on lunar volcanism some $3.0\text{--}3.3 \times 10^9$ yr ago was caused when the cooling reached the top of the refractory dunite, lower mantle, whose solidus would be 250–300°C higher than that of the adjacent peridotite upper mantle.

There is one final suggestion to be made before leaving the cooling history of the Moon. As reviewed in Section 3, seismic data (Nakamura *et al.*, 1973) indicate that below a depth of 1000 to 1100 km there is a zone which may contain a small amount of melt. Because of the low gravity field within that zone (g varies between 0 and about 65 cm s^{-2}), gravitational settling and filter pressing by the olivine crystals may not have been 100% effect. If so, then the small amount of liquid which may be contained within the lower part of the lower mantle would simply be the part of the original melt of the Moon.

9. Physical Properties of a Pyrolite Moon

Detailed calculation, taking into account temperature and pressure effects on the density, have not yet been made for the differentiated pyrolite Moon model. However, based on the first-order approximation that pressure and temperature effects cancel each other for the lunar case, a simple calculation shows that $I/MR^2 = 0.3981$, which is in good agreement with the value of 0.3985 which Kaula reported at the 4th Lunar Science Conference.

As reviewed in Section 3, the P wave velocities ($\sim 8 \text{ km s}^{-1}$) of the upper mantle are consistent with it being composed of peridotite and the P wave velocities are nearly constant with increasing depth in the mantle. According to Akimoto (1972) the P wave velocities of Fo_{95} is about 8.5 km s^{-1} , i.e. 0.5 km s^{-1} higher than observed for the lunar interior. However, Nakamura and Latham (1969) have shown that P wave velocities would be reduced by about 0.5 km s^{-1} below depths of 300 or 400 km in the Moon due to thermal effects. Thus, the available seismic velocity data do not preclude a magnesium rich dunite lower mantle.

10. Cratering History

From the cooling history discussed in Section 8, it is clear that the lunar crust would have been thick enough to begin recording impacts 1 or 2×10^7 yr after the Moon fissioned from the Earth. Wise (1963) has suggested that a large number of small bodies would have been formed from part of the material in the 'neck' of the Poincaré

figure when the Moon fissioned, and that these small bodies would have accounted for part or most of the earlier craters and mare basins on the Moon. Recent work on the time dependency of the flux during the period between 4.5×10^9 and 3.3×10^9 yr ago by Neukum and Arkani-Hamed (1973) has indicated that the craters and basins formed within this time interval can be accounted for by a single family of objects whose flux decreased with a time constant of $\sim 2 \times 10^8$ yr. This time constant is the same as that for objects in near Earth, heliocentric orbits (Öpik, 1961). Thus, if most of the swarm of small bodies produced during fission were quickly ejected into near Earth, heliocentric orbits, they would have been available on the appropriate time scale to cause the upland cratering and formation of the mare basins. Based on Neukum and Arkani-Hamed's work this bombardment would have effectively ceased about 3.8 or 4×10^9 yr ago and Mare Imbrium and Mare Orientale are simply the last two basins to have formed. Thus, there is no need to call upon a catastrophic bombardment phase centered on 4×10^9 yr ago in order to account for the young ages of these two basins and the metamorphism and brecciation of the primitive lunar crust (Tera *et al.*, 1973). These two processes, i.e. impact brecciation and base surge blanket-'KREEP-granitic' fluids alteration of upland rocks would have both acted together to erase the age records between $\sim 4 \times 10^9$ and 4.5×10^9 yr ago and fortuitously, would have both ceased 3.8×10^9 to 4.0×10^9 yr ago.

11. Summary and Conclusions

The analysis and discussions presented show that, allowing for volatile losses, the bulk composition of the Moon may be identical with that of the upper mantle of the Earth. Also, if the Moon fissioned from the Earth at the end of its accretion and as a result of core formation, the expected differentiation and thermal evolution of the Moon would result in a three layered body whose lithological and physical characteristics and developmental history match all those presently known for the Moon. As such, the evidence presented strongly supports Wise's fission hypothesis for the origin of the Moon.

The internal structure of the Moon, based on the differentiation of a pyrolite Moon is given in Figure 6 and the evolutionary history of the Moon is as follows:

(1) Origin of the proto-Earth via accretion 4.6×10^9 yr ago; simultaneous melting of the proto-Earth as a result of accretional energy, ohmic heating, etc; formation of the core and the resulting depletion of free iron and siderophiles in the mantle of proto-Earth.

(2) Fission of the Moon near or at the end of core formation 4.6×10^9 yr ago, the Moon is completely molten and depleted in free iron and siderophiles.

(3) For the first few 10^7 's of years after fission, the Moon is either heated to a few hundred degrees above the liquidus by ohmic heating and as rotational energy is converted to heat via tidal interactions with the Earth or the new Moon is held at the liquidus with radiative losses keeping up with the heating. During this short phase, the Moon loses H_2O and other volatiles and the volatile metals (Na, K, Pb, etc.).

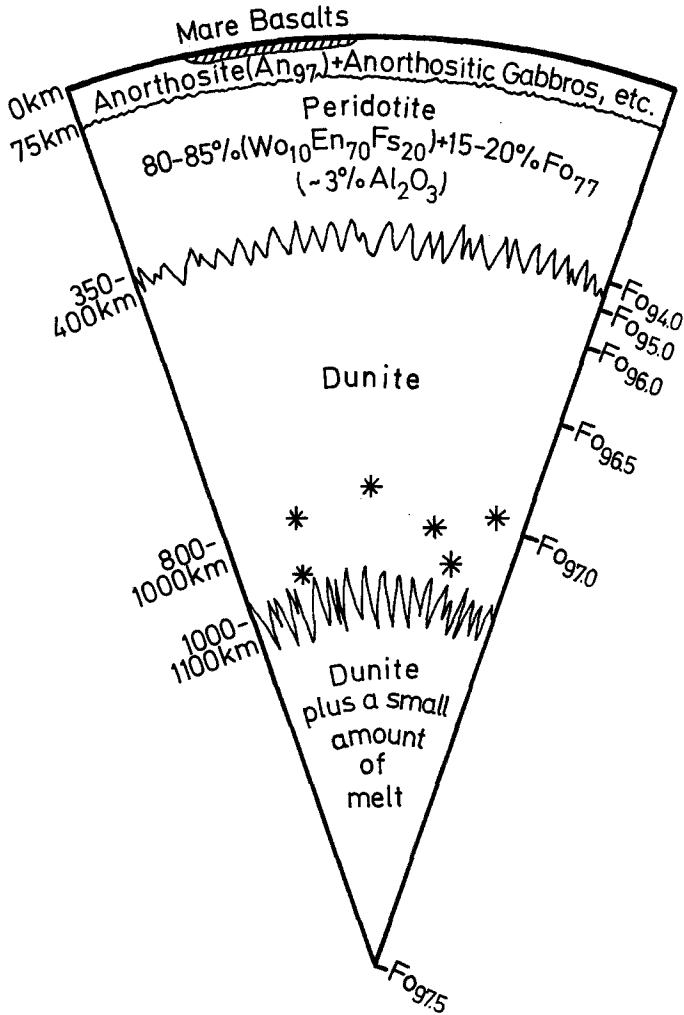


Fig. 6. Proposed lunar structure based on the differentiated pyrolite Moon models. The compositions of the peridotite upper mantle and dunite lower mantle are those for a model with upper mantle 100 Mg/Mg + Fe of 77, i.e. the mean of the range of 75 to 80 as proposed for the upper mantle (Ringwood and Essene, 1970; Green *et al.*, 1971; Green and Ringwood, 1973). Included in the diagram, but not discussed in the text, is the Moon quake zone (800–100 km depth) just above the zone of partial melting (Nakamura *et al.*, 1973).

As the Moon retreats from the Earth via tidal interactions, the heating phase ceases and the Moon cools to the liquidus at the surface via radiation.

(4) Refractory olivine Fo_{93-97} crystallizes at the surface and rapidly sinks (~ 105 days) to the center of the Moon to form a dunite lower mantle. The 100 Mg/Mg + Fe ratio of the residual liquid decreases from 88 to 80 or 75 at which time 42% to 53% of the melt has solidified and at which time disequilibrium crystallization ceases. Because of incomplete filter pressing near the center of the Moon, a small amount of

original melt may be trapped between the olivine crystals, thus accounting for the zone of *S* wave attenuation below a depth of 1000 to 1100 km.

(5) Crystallization of olivine continues, but in equilibrium with the melt, until 63% of the original melt is solid. At this point pyroxene begins to crystallize; some of the olivine which crystallized out (in equilibrium) is converted to pyroxene, the remaining is equilibrated with the pyroxene and has a $100 \text{ Mg/Mg} + \text{Fe} = 75$ to 80. The total amount of olivine remaining is 55%. The crystallization of pyroxene and equilibration and reabsorption of the 10–21% excess, early olivine continues until 80% of the original melt is solid about 100 yr after crystallization began.

(6) When 80% of the melt has solidified, crystallization of pyroxene and anorthite (An_{97}) begins. The anorthite floats to the surface and starts to form the primitive anorthosite crust. As soon as the cumulative anorthosite and minor troctolite crust begins to develop, the rate of cooling of the Moon is drastically reduced, so the cooling of the final 20% of the melt requires 6×10^7 yr.

(7) Towards the end of phase 6, feldspathic basalts begin to be produced locally and are extruded onto and intruded into the crust; so the final crust is only about 70% anorthite. Also, as a result of the final eutectic crystallization, KREEP, the incompatible elements, 'granite' fluids and volatiles are concentrated in a zone within the bottom of the crust and at the top of the peridotite upper mantle. These materials are available to form KREEP rich norites and to intrude earlier formed rocks. The emplacement of these feldspathic, upland and KREEP rich norites and the metamorphism of the crust via KREEP and 'granitic' fluids terminates (4×10^9 yr ago), when cooling below the solidus reached a depth of 150 km.

(8) Mare basalts are extruded into the basins on the shallow front side crust of the Moon. The magmas are produced by partial melting of the peridotite [$80\text{--}85\%$ ($\text{Wo}_{10}\text{En}_{68\text{--}72}\text{Fs}_{18\text{--}22}$) with 3% Al_2O_3 and 15–20% ($\text{Fo}_{75\text{--}80}$)] upper mantle between depths of 150 to 350 or 400 km.

(9) The cooling below the solidus reaches the refractory dunite ($\text{Fo}_{93\text{--}97}$) lower mantle, and lunar volcanism ceases (between 3 and 3.3×10^9 yr ago).

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