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Two Major Igneous Events in the Evolution of the Moon [and Discussion]

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EXTRATERRESTRIAL MINERALOGY

Two major igneous events in the evolution of the Moon

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An understanding of the origin of the Moon is strongly dependent upon a knowledge of its bulk composition and thermal history. Both aspects require a detailed consideration of the composition and origin of the lunar crust and of the mantle-derived lunar basalts. The evidence for two major igneous events is discussed, the first being a large-scale melting and fractionation into crust and mantle at −4.6 to −4.5 Ga, and the second a partial melting of the uppermost mantle at −3.8 to −3.2 Ga. The distribution of uranium is used to place constraints on the minimum extent of initial melting and on the depth at which the mare basalts were generated, using recent lunar heatflow data for a bulk-Moon uranium content of 30 parts/10^9. The model favours melting of at least 90 % by volume, and a concentration of the high U-contents of the crust and upper mantle by formation of a thick lower mantle of mafic adcumulates 'barren' in heat-producing elements. The 'fertile' mafic orthocumulates from which the mare basalts were generated are restricted by the model to depths of less than 200 km.

A downward revision of the bulk U-content of the Moon results in down-scaling of the other refractory lithophile elements by analogy with the solar–nebula condensation models. This means that the bulk Moon is fairly close in composition to that of the Earth's mantle, including its iron content but excluding the volatile elements which are strongly depleted in the Moon. Low contents of siderophile and chalcophile elements, and high contents of lithophile refractory elements in the lunar basalts are attributable to the large-scale fractionation into a core, mantle and crust.

The hypothesis of an origin for the Moon by fission from a proto-Earth is revived. Earth layering by a heterogeneous accretion sequence would account for non-equilibrium between core and mantle (e.g. nickel distribution) and an outer veneer of volatile-rich condensate that would contribute to subsequent generation of a granitic crust. Early collision with a large body may have caused fission and formation of a proto-Moon from the Earth's iron-poor, proto-mantle, with loss of volatiles. Early melting of most of the proto-Moon led to strong fractionation such that the crust and mantle-derived basalts appear to have more extreme compositions, relative to Earth basalts, than is indicated by the likely bulk composition of the Moon.

1. Differences between lunar and terrestrial basalts

The term 'basalt' is applied to those crystalline, igneous-textured lunar rocks that have closer affinities with terrestrial basalt lavas than with any other terrestrial rock type. They occur chiefly as lava infillings of the large mare basins but more aluminous varieties occur as rare fragments in the breccias that result from impact-reworking of the older feldspathic crust. The coarser-grained anorthositic gabbros and troctolites found in the crustal breccias are not grouped with the basalts.

The lunar basalts are chemically distinct from terrestrial basalts in having low SiO_2 contents associated with high FeO/FeO + MgO ratios, high but variable TiO_2, and low Na_2O and K_2O. The low alkalis give rise to the calcic plagioclases (average An95) that contrast most markedly with any use of plagioclase composition in defining terrestrial basalt. The trace elements also reflect this depletion in volatile elements in the lunar basalts. Taylor (1975) and Ganapathy & Anders (1974), for example, summarize these compositional features and point to the relative depletion in volatile, siderophile and chalcophile elements, and enrichment in refractory lithophile elements in the lunar basalts.

The minerals in these, as in any other rocks, provide a wealth of information in addition to their elemental compositions. The basalts crystallized under conditions of low oxygen fugacity (about 10^{-13} bar compared with 10^{-8} bar on average for terrestrial basalts), as evidenced by phases such as nickel-iron, troilite and ulvöspinel. Low valency states are evidenced by Fe^{2+} (no Fe^{3+}), Eu^{2+} and some Cr^{2+} and Ti^{3+} in certain minerals. Metastable crystallization gave rise to pyroxferroite, subcalcic augites and ferroaugites, cristobalite and tridymite, and the presence of magnesian olivines alongside the silica phases. The Rb–Sr isotopic patterns indicate strongly contrasted crystallization (internal isochron) and model (whole-rock isochron) ages, suggesting that the basalts were not derived by *equilibrium* partial melting of their source material. The natural remanent magnetism of the basalts indicates crystallization in a stronger magnetic field (2000 nT) compared with the present field (38 nT), and hence the possibility of a liquid metallic core to the Moon, that has since solidified.

Hence the lunar basalts provide information related to the nature of the lunar interior, and to the events that gave rise to a Moon differentiated ($I/MR^2 = 0.395$) into a crust, mantle and possible core. In so far as those three units differ in relative volume and composition from such units of the Earth, differences in bulk composition of the two bodies exist that need to be explained in terms of the origin of the satellite Moon relative to the planet Earth.

2. Structural and petrological units of the Moon

The seismic data (e.g. Toksöz *et al.* 1973) have provided evidence for a major discontinuity at about 60 km depth, between a crust ($V_p = 7$ km/s) and an upper mantle ($V_p \geqslant 8$ km/s). Subsequent information from deep moonquakes and attenuation of transmitted shear waves has led to the definition of an 'asthenosphere' zone below the thick, rigid lithosphere. The attenuation of the S-waves could be attributed either to a small degree (*ca.* 1 %) of partial melting or (D. C. Tozer, personal communication 1976) to a material with slightly higher volatile contents than the lithosphere. The model presented by Dainty *et al.* (1975) gives an 'asthenosphere' extending for about 1000 km radius from the Moon centre, a 60 km crust, and thus a rigid intervening lithospheric mantle of about 680 km thickness. Nakamura *et al.* (1974) divide the rigid mantle into two zones (upper 250 km, middle 500 km), call the 'asthenosphere' a lower mantle (*ca.* 600 km), and subdivide from the latter a core of 170–360 km radius with low P velocity. Thus within the Moon radius (1738 km) there is clear evidence for a layered structure, broadly analogous to that of Earth except for a smaller-volume core, and a much thicker rigid lithosphere which has supported large positive gravity anomalies (mascons) since at least the period of mare basin filling (-3.8 Ga).

Taylor & Jakeš (1974) proposed a model in which the asthenosphere or lower mantle is undifferentiated, primitive lunar material. Such material would be an attractive possibility because it would have retained its original content of radioactive heat-producing elements and (if they were not lost before accretion) of volatile elements. Hence either effect could account for the present-day attenuation of shear waves in this inner zone, since the first could result in partial melting through 4.6 Ga of radioactive-decay heating. However, this model would require that a major differentiation involving the crust and upper–middle mantle did not extend deeper than about 700–1000 km. In other words, that the Moon was never totally melted. This could cause difficulties in explaining the generation of a core by sinking of an Fe–Ni–S liquid to the central region (discussed later).

The upper, accessible part of the Moon consists of two major units enclosed within the 60 km thick crust. First, the complex assemblage of breccias that constitute the highlands crust and secondly, the suites of mare basalt lavas that have since flooded the large, impact-excavated basins. The breccias consist of rock and glass fragments, of which many are distributed by younger impacts among the soil population at each sampled landing site, consisting chiefly of aluminous material rich in the plagioclase feldspar component. The average composition of such material is of anorthositic gabbro. It is customary, therefore, to refer to the primitive crust as being a 60 km-thick layer of anorthositic gabbro, but this is a risky over-simplification. Other fragments in the breccias include anorthosites, norites and troctolites (referred to as the ANT-suite), as well as rare fragments of dunite, granite or granodiorite, aluminous basalt, and spinel troctolite. Also, there is the type of basalt very rich in K, r.e.e. and P (KREEP basalt) that has been melted and mixed with the other crustal materials during certain major impact events, or possibly only during the Imbrian impact event. This material, rich also in Zr, U and Th, has a problematical history but is almost certainly produced originally by extensive crystal–liquid fractionation that included the evolution of the whole crust and a major part of the mantle. In regard to the troctolites and dunites, and the ferroan anorthosites that contain plutonic-type pyroxene assemblages (Brown & Peckett 1973) it seems likely that the deeper original crust may have been closer to gabbroic than to anorthositic gabbro composition, and that early crystal fractionation occurred in localized magma chambers. Hence the anorthositic character may only persist to shallow depths in the present 60 km crust, and the seismic data could be resolved according to such a model (e.g. Liebermann & Ringwood 1976).

The second major unit of the crust, that of the mare basalts, is less complex but by no means internally uniform. The crystallization ages indicate flooding of the ringed basins between about -3.8 and -3.2 Ga so a period of volcanism spanning at least 600 Ma is involved and would be expected to produce some chemical variation in the products. Secondary variations in texture due to differential cooling rates of flows (including resorbtion of early olivines) has been demonstrated for much of the Apollo 17 basalt suite (Brown *et al.* 1975), while crystal sorting at shallow depths has probably caused some secondary variations in all sampled basalts (O'Hara *et al.* 1975). However, there are more fundamental variables that appear related to the nature of the source material and warrant a classification into several possible 'primary' basalt magma types:

(1) *High-Ti basalts*
  (*a*) Apollo 11 (high K).
  (*b*) Apollo 11 and Apollo 17 (low K).

(2) *Low-Ti basalts*
  (*a*) Apollo 12, Apollo 15, Luna 16 (olivine basalts and quartz-normative basalts).
  (*b*) Apollo 12, Luna 16 (aluminous mare basalts).
  (*c*) KREEP aluminous basalts (pre-mare material, highlands-derived fragments especially from Apollo 14 'Fra Mauro' formation).

There is considerable chemical variation between and within these groups. Average compositions (e.g. Rose *et al.* 1975) indicate, between groups, Al_2O_3 variation from 8.6 to 21.4 %, TiO_2 from 1.2 to 13.0 %, and K_2O (as a trace) from 0.06 to 0.48 %, for example. Within group 1*b*, the Apollo 17 'high-Ti' suite shows TiO_2 varying from 9.6 to 13.3 % which could be attributable to near-surface crystal sorting but indicates the problems in selecting a 'primary' magma type (Walker *et al.* 1976). However, a division of the low-alumina mare basalts into high-Ti and low-Ti groups appears legitimate and, as will be discussed below, is supported by a grouping according to l.i.l.-element abundances and crystallization ages.

The classification of the high-alumina basalts (cf. < 9 % Al_2O_3 for other mare basalts) is beset by difficulties in nomenclature and interpretation (see Taylor 1975). The aluminous mare basalts are rare, with 12–14 % Al_2O_3, 3–4 % TiO_2, and 0.1–0.2 % K_2O. All other high-alumina basalts are pre-mare formation, and occur as highlands breccia fragments. They have suffered melting, contamination and recrystallization during impacts but some have a consistently unique composition that warrants consideration in terms of early crust–mantle generation. Taylor (1975) divides them into low-K, medium-K and high-K Fra Mauro basalts. They are not to be confused with the dominant, feldspathic highlands rocks that constitute most of the breccias, which are chemically very high-alumina basalts (> 23 % Al_2O_3) but are viewed as ANT feldspathic cumulates in origin (named anorthositic gabbros, gabbroic anorthosites, anorthositic norites, troctolites, etc.). The low-K Fra Mauro basalts (19 % Al_2O_3) are probably misnamed and belong to the less feldspathic of the cumulative (ANT) suite. The medium-K and high-K Fra Mauro basalts contain high but variable amounts of K, r.e.e. and P and are better viewed collectively as KREEP basalts (18–19 % Al_2O_3, > 1 % K_2O).

For the purposes of this paper, only the KREEP basalts, of the aluminous suite, will be considered in relation to the evolution of the younger mare basalts. The most significant chemical grouping is related to the abundances of the refractory, large-ion lithophile (l.i.l.) elements. Especially, r.e.e. contents relative to chondrites (Taylor 1975) are enriched by factors of about 20 (low-Ti, quartz and olivine normative basalts), 50 (high-Ti basalts) and 200 (KREEP basalts).

3. CHEMICAL AFFINITIES BETWEEN LUNAR CRUST AND MANTLE

The highly aluminous, refractory nature of the lunar crust, its enrichment in heat-producing elements such as U and Th, and its 'primitive' age compared with the mare basalts, led to early views that it was accreted from material chemically distinct from that of the sub-crustal part of the Moon. Element fractionation in the solar nebula would give a condensation sequence, related to temperature and pressure distribution, such that the early condensates (higher bond-energy condensates; G. Arrhenius, personal communication, 1976) would consist chiefly of Ca, Al, Ti and other refractory lithophile elements (see, for example, Grossman & Larimer 1974). Although such condensates have accreted within the solar nebula (as evidenced by the white inclusions in the Allende carbonaceous chondrite meteorite), they would not be expected to accrete during the latest stages of formation of a planetary body. Also, whereas the lunar crust is rich in Ca and Al, the enrichment in Ti resides in the mantle-derived mare basalts, and the high r.e.e., U, Th and other l.i.l. elements in the KREEP basalts (believed to originate at the crust/mantle interface). For these reasons alone, the crust cannot be viewed as having formed by heterogeneous accretion of the proto-Moon. The hypothesis advanced early by Wood *et al.* (1970) and Smith *et al.* (1970), that the aluminous crust is chiefly a result of plagioclase fractionation from a molten or partly molten, homogeneously accreted Moon, is the one that has since received extensive petrological and geochemical support. Since the mare basalts originated by later partial melting of a sub-crustal zone, their composition must be associated with the chemical differentiation into primitive crust and upper mantle. Chemical differences and affinities between crustal rocks and mare basalts are hence a reflexion of much of that primitive differentiation event.

Taylor (1975) had made an extensive summary of the evidence for chemical affinities between the early highlands crust and the mantle-derived mare basalts. The most significant factors are:

(1) Plots of volatile against non-volatile elements show similar ratios in both highlands and mare rocks (e.g. K/Zr, K/U, K/La, Rb/Ba, Tl/U) except for the rare cases of volatile-element contamination from meteoritic components in highlands rocks.

(2) The 'europium anomaly' shows a complementary relation between highlands and mare rocks. The entry of Eu^{2+} preferentially into plagioclase gives mostly a positive anomaly for the anorthosites, anorthositic gabbros, troctolites and norites. The fact that the mare basalts show a strong negative anomaly means that they are derived from a source from which the plagoclase-rich rocks were earlier subtracted. The KREEP basalts show the strongest negative Eu anomaly and are thus related to a sub-crustal source, probably at the crust/mantle interface.

The mare basalts thus provide evidence for a chemical relation with the highlands crust. They also possess chemical characteristics that relate to their upper mantle source. Certain chemical properties common to all the mare basalts are to be noted in so far as they may relate to overall lunar chemistry. Compared with terrestrial primary basalts and solar-system abundances, the lunar basalts are enriched in the refractory lithophile elements, and depleted in the chalcophile, siderophile and volatile elements (e.g. Taylor 1975; Ganapathy & Anders 1974). Enrichment in the l.i.l. elements is characteristic also of the highlands crust, and hence is likely to be characteristic of the bulk Moon. Volatile/involatile element ratios, similar in both highlands and mare rocks, indicate also a bulk-Moon depletion in the volatile elements (rather than local surface loss from the low gravity field). Depletion in chalcophile and siderophile elements (neglecting local meteoritic contamination of highlands rocks) could be attributed to concentration in an Fe–Ni–S core but terrestrial basalts do not show this depletion (see §6).

The mare basalts are derived from shallow depths in the mantle, probably not deeper than about 250 km (see, for example, Walker *et al.* 1976), and from mafic material that, according to the Eu-anomaly and other chemical data, has participated in an early chemical fractionation process involving the feldspathic crust. Hence the chemistry of the mare basalts alone cannot reflect derivation from a 'bulk-Moon' source. However, grouping the crust and mare basalts together, and assuming an underlying refractory residuum of upper mantle material, rich in olivine and clinopyroxene (from the experimental data), such an outer 250 km zone of melted Moon ought to reflect the significant elemental enrichments and depletions of the bulk Moon.

4. DATED EVENTS IN LUNAR EVOLUTION

The isotopic age data indicate that there were two major igneous events in the Moon's history. The first took place shortly after accretion (–4.6 to –4.5 Ga) and the second began about 800 Ma later and lasted for about 600 Ma (–3.8 to –3.2 Ga). A summary based on

Rb–Sr internal isochrons is given by Papanastassiou & Wasserburg (1976, Fig. 2) and a range of age data is tabulated by Taylor (1975, p. 181).

The crystallization ages (T_c) of the mare basalts (internal isochrons) fall into two groups. The high-Ti basalts (Apollo 11 and 17) crystallized at about -3.8 to -3.6 Ga, and the low-Ti basalts (Apollo 12 and 15, and Luna 16) at about -3.5 to -3.2 Ga. This represents the second major igneous episode.

Model ages (T_{BABI}) based on whole-rock isochrons (e.g. soil analyses) show, in contrast, a majority with ages of about -4.6 to -4.5 Ga. The only exceptions are the Apollo 11, high-K basalts with model ages of -3.9 to -3.8 Ga. The high model ages represent the first major igneous episode. The data suggest that at about -4.6 Ga, when the Earth and Moon accreted (and other solar system planetary bodies such as the meteorites), the Moon then melted and crystallized, at least throughout the outer 200 km or so, over a very short period of time (10^6 years has been suggested). The major Rb–Sr fractionation occurred then, associated with differentiation into a feldspathic crust and a mafic upper mantle.

The heat-producing elements were concentrated in the crust and KREEP layer, but the underlying mafic-layer source for the mare basalts contained enough for radioactive decay, over 800 Ma, to generate partial melting and the first eruption of the high-Ti mare basalts at -3.8 Ga. Deeper mafic layers were more depleted in heat-producing elements and generation, there, of low-Ti mare basalts did not begin until -3.5 Ga. Even deeper layers were probably so depleted that no partial melting could occur, and mare basalt generation apparently ceased at -3.2 Ga. This model differs in one vital respect from models of basalt generation from the Earth's mantle. The partial melting of the lunar mantle must have been non-equilibrium melting, such that the liquid did not equilibrate with a Sr-bearing residual phase. Hence the model ages (*ca.* 4.6 Ga) continue to reflect the early Rb–Sr fractionation and no subsequent Rb–Sr fractionation was superimposed except for the Apollo 11 high-K basalts which were presumably generated in equilibrium with plagioclase near the base of the crust. Albee & Gancarz (1974), who explain the Rb–Sr systematics in the context of the model ages, suggest that lack of water on the Moon may have inhibited equilibrium (i.e. diffusion between partial melt and residual crystals).

Between the first and second of the igneous events, lies 800 Ma of indistinct and highly complex lunar history. Intense meteoritic bombardment of the feldspathic crust occurred, and culminated in the 'cataclysmic' production of the huge ringed basins such as Imbrium (-3.9 Ga). The KREEP basalts, that formed a sub-crustal layer during the first igneous event (from model ages), were excavated and recrystallized by the Imbrian impact and possibly in part by earlier impacts around -4.0 Ga (e.g. Nectaris). Working back in time from then, anorthositic fragments are revealing ^{39}Ar/^{40}Ar ages of about -4.2 Ga (Eberhardt *et al.* 1976). The most dramatic age results of all, however, relate to crystallization (T_c) ages of -4.5 to -4.6 Ga for a troctolite and a dunite fragment, with the most primitive (BABI) of solar-system initial ^{87}Sr/^{86}Sr ratios. Papanastassiou & Wasserburg (1976) give these data and the internal isochron for the troctolite, a remarkably fresh 'terrestrial-type' troctolite (Apollo 17) which crystallized virtually when the planets formed. Clearly the first igneous event must have been more complex than we had envisaged. Plutonic troctolites, norites and dunites, as well as ferriferous anorthosites with plutonic exsolved pyroxenes (Brown & Peckett 1973) indicate the existence of early crustal, fractionating magma chambers. If we consider, however, that the early crust formed from a floating plagioclase-rich crystal layer, solidifying downwards over about 60 km, then stoping was probably a common event. That is, the crustal layers would tend to fracture and permit block displacements, particularly under the influence of magma convection. Magma chambers could thus develop within the hot crust and the primitive mafic magma could have fractionated within such crustal enclaves.

5. Source of mare basalts

If the mare basalts were derived simply by partial melting of primitive lunar mantle material, it would be necessary to place their source deeper than the thick, outer shell that gave rise to the plagioclase-rich crust and, inevitably, to an underlying complementary layer of mafic cumulates. It would also be necessary to provide an alternative explanation for the basaltic negative Eu anomaly that is complementary to the positive Eu anomaly of the highlands rocks. In the absence of any compelling reason for seeking a source in an undifferentiated layer of the mantle, the most likely source would be a mafic cumulate layer beneath the feldspathic crust. A third, but still debatable reason for that choice hinges on whether plagioclase is a liquidus phase of the mare basalts. Ringwood & Essene (1970) first popularized the view that it is not a liquidus phase, in which case the basalts could not be derived from a layer which had been modified from its initial composition simply by plagioclase fractionation. The europium anomaly would be explained by that mechanism, but plagioclase would remain a liquidus phase in the residual material. However, if mafic minerals (e.g. olivine and pyroxene) were concentrated in a cumulate layer complementary to the plagioclase-rich layer, later partial melting of that mafic cumulate would, if the partial melt were enriched in the mafic cumulus-phase components, give rise to a more mafic liquid which need not crystallize plagioclase on the liquidus. That mechanism (proposed by the writer during the Apollo 11 Lunar Science Conference discussions) was not accepted until Taylor & Jakeš (1974) developed a refined model for two-stage melting involving the mafic cumulates.

The Taylor–Jakeš model (summarized in Taylor 1975) has many attractive features. It requires melting of the outer 1000 km of the Moon during the -4.6 Ga event, the inner ‘asthenosphere’ being possibly undifferentiated primary material, discussed earlier (§1). It then provides for the deposition of a series of mafic cumulus minerals, later joined by plagioclase which floated to form the crust. The upper mafic cumulus layers would thus be depleted in Eu^{2+} and be the source of the mare basalts. Such a mafic cumulate sequence would, by analogy with terrestrial layered intrusions (Wager & Brown 1968), show upward enrichment in the l.i.l. refractory elements. Hence the mare basalts richest in those elements (the high-Ti basalts) would be derived by partial melting of mafic cumulate layers overlying the source of the low-Ti basalts. As noted in §4, this also correlates with the eruption ages which are presumably related to the relative concentration of heat-producing elements in the mafic cumulates, being highest in the source of the earliest-erupted mare basalts.

The Taylor–Jakeš model was not related to a layered-intrusion model in detail, because of neglect of the rôle of intercumulus material and other more subtle aspects of igneous cumulate formation. The writer has discussed some of those features (Brown 1976) although a more rigorous approach is still needed. Igneous cumulates consist of crystals concentrated under the influence of gravity (cumulus phases), and the products of the intercumulus liquid which accounts for about 40–50 % by volume under simple packing conditions. The intercumulus liquid may be trapped (to crystallize later as lower-temperature phases), usually because of rapid crystal accumulation, the final rock being termed an 'orthocumulate'. Slower deposition rates can, however, allow element diffusion between intercumulus and overlying (main bulk) liquid such that the pore spaces are filled by growth of the high-temperature cumulus crystals. The resultant rock, from which the contemporary liquid products have therefore been expelled, is termed an 'adcumulate' (Wager *et al.* 1960). Adcumulate layers within, say, the lunar mantle would thus be refractory and would be responsible for concentrating the l.i.l. trace elements in an overlying zone more effectively than would layers of orthocumulates (analogous to the 'zone refining' process of Hubbard & Minear 1976). Conversely, orthocumulates would, by partial melting of their less refractory pore material, yield liquids enriched in those trace elements. To account for the high Ti contents of certain mare basalts, for example, melting of pore material enriched in Ti would therefore be more feasible than melting of an assemblage of Ti-rich cumulus minerals. The latter would require near-total melting of the layer, and such a layer would occur at a higher level in the fractionated layered sequence than the Ti-rich pore material. The next point is that to obtain, say, Ti-rich pore material would require, from a melted Moon material of moderate Ti content, extensive earlier removal of Ti-poor phases. This could be best effected if those phases were cemented not by trapped Ti-bearing pore material (contemporary liquid) but by adcumulus-growth material. Clearly, therefore, any consideration of large-scale melting of the Moon, followed by crystal–liquid fractionation, leads to the recognition of variables that cannot easily be quantified, but which could be used to explain certain anomalies.

TABLE 1. URANIUM DISTRIBUTION ESTIMATED FOR LAYERS OF THE MOON

(Less significant values rounded-off for a generalized model)

unit (subtraction stages in parenthesis)	shell thickness (radial km)	volume (% of Moon)	U content (parts/10^9)	U contribution (parts/10^9) to total volume
1. total Moon (see text)	1738	100	30	30
2. undifferentiated part	738	8	30	2
3. differentiated part (1–2)	1000	92	30	28
4. l. mantle adcumulates (3–5)	800	62	0	0
5. crust + u. mantle orthocumulates	200	30§	94	28
6. average feldspathic crust	60	10	250†	25
7. u. mantle orthocumulates including low-volume KREEP layer (5–6)	140	20	13‡	3

† From γ-ray orbital values (Trombka *et al.* 1974) of 450 parts/10^9 for average crust, less about 6% surface-distributed KREEP material (3500 parts/10^9) that distorts upward the γ-ray spectrometer estimate of average crustal value. Values for various crustal rock-types summarized by Brown (1976).

‡ A maximum value. Any increase would reduce the crustal contribution which is based on a probable minimum (†) of about 250 parts/10^9. As this uranium (13 parts/10^9) would reside in the orthocumulate pore material, 5% partial melting could generate basalts with 260 parts/10^9 uranium, as observed in the average low-K mare basalts (240 parts/10^9).

§ Volume (and hence shell thickness) calculated to best-fit the need to generate the crustal U-content (6). A lower volume would contribute too little, and a higher volume too much U for the required 28 parts/10^9 of the differentiated part (3). A lower volume for the undifferentiated part would have only a slight effect on these U figures and an appreciably higher volume (i.e. less initial melting) could not be tolerated if the values in (6) are to be explained.

The distribution of an element that is rejected by the major crystallizing phases can be used to illustrate likely constraints. *Uranium* has been chosen because there are independent means for calculating its abundance in the bulk Moon, and plenty of data on its abundance in the observed rocks and its distribution across the lunar surface from γ-ray spectrometer data

(Trombka *et al.* 1974). The calculations are summarized in table 1. Initially, a mean U-value for the Moon was taken to be 50 parts/10^9 (Brown 1976) but a drastic revision downward of the heatflow values (Langseth *et al.* 1976) now indicates that a value closer to about 30 parts/10^9 is more likely.

In order to obtain average crustal abundances of about 450 parts/10^9 uranium, a very large volume of the Moon must be 'reworked'. If only the outer 300 km had melted, this would correspond to 40 % melted and differentiated, and 60 % undifferentiated (by volume). It seems impossible for such a thin, 300 km outer shell (30 parts/10^9) to differentiate to give 450 parts/10^9 in the average crust. The latter (60 km thick) constitutes about 10 % of the lunar volume and so would contribute, if U is distributed evenly throughout the 60 km thickness, about 112 parts/10^9 to the 40 % volume differentiated zone. In fact, if the value for average crust were accepted for the whole crust, then 10 % by volume would give a contribution of 45 parts/10^9 to a Moon assumed to have only 30 parts/10^9 uranium! In order to make any type of meaningful calculation, it is therefore proposed that because the KREEP-rich materials are erratically distributed on the lunar surface and would contribute to high average, near-surface, U-contents out of proportion to their volume in the whole crust, an average crustal U-value of about 250 parts/10^9 is preferable by subtracting about 6 % of KREEP-rich material (with 3500 parts/10^9).

As shown in table 1, even with a 1000 km zone of melting and an assumed low U-value for the average crust, much of the early melt must have crystallized as deep-level adcumulates (lower mantle) so that elements such as uranium could be concentrated in an outer zone of mafic orthocumulates (upper mantle, basalt source) and feldspathic cumulates (crust). Total melting would not affect the balance significantly, since the postulated asthenosphere is only 10 % of the lunar volume. As a result of this calculation, the upper mantle + crust is only 30 % of the lunar volume. Hence mare basalt generation would be confined to a zone extending no deeper than about 200 km. Deeper zones would be ultramafic adcumulates and thus devoid of heat-producing elements and also of basaltic components.

At such a late stage of major fractionation, whereby the highlands crust and the mare basalt source material have evolved in the proportion, to early adcumulates, of 1:2, major elements such as Fe and Ti, as well as the incompatible trace elements, would have risen appreciably over lunar bulk abundances. Further fractionation of plagioclase and mafic minerals to form the crust and upper mantle cumulates would extend this fractionation to give material such as the KREEP-basalt layer, highly enriched in r.e.e. and other l.i.l. elements and probably located as the final unit of a 'sandwich' between feldspathic crust and underlying mafic cumulates (cf. Hubbard & Minear 1976). Partial melting of the mafic cumulates would concentrate elements such as Fe, Ti, r.e.e., etc. in the mare basalts, beyond the levels reached by the early, major fractionation process. Before the 200 km thick 'sandwich' had become stabilized, opportunities would arise for magma to invade the proto-crust (to form fractionated plutons), at a stage when the crustal skin was only a few kilometres thick and hence the 'upper mantle' magma was little advanced and could precipitate fairly magnesian olivines and pyroxenes.

Ringwood (1975) has pointed out several problems in the Taylor–Jakeš model, chief of which is the need for the high-Ti mare basalts to have a higher Fe/Mg ratio than the low-Ti basalts, if they are tapped from a higher level of the upper mantle cumulates. This ratio is anomalous compared with the trace-element evidence in support of the cumulate model.

The KREEP basalts also are relatively magnesian, suggesting contamination with crustal rocks during impact reworking. Hence the high-Ti basalts, if derived from not far below the original KREEP horizon, may also have suffered some contamination, or a more complex type of hybrid history (cf. Kesson & Ringwood 1976).

6. Bulk composition and origin of the Moon

The aluminous nature of the lunar crust led to an early conclusion that the Moon has a refractory composition compared with the Earth and solar-system elemental abundances. This view was substantiated by comparison with carbonaceous chondrites in terms of the lunar basalt contents of U, Th, r.e.e. and other refractory lithophile elements. Since the Moon is clearly a strongly fractionated body, however, the crustal and upper mantle derived basalt compositions cannot give more than a qualitative estimate of bulk composition.

Consideration of condensation processes in cooling solar nebula (e.g. Grossman 1972) led to the assumption that the inner planets formed by the processes responsible for chondrite formation. Ganapathy & Anders (1974) present a model based on these processes for the formation of the Moon and the Earth. The equilibrium condensation sequence of a solar gas (Larimer 1967) would give a sequence of dust condensates in the sequence (*a*) early refractory condensate, (*b*) metallic nickel-iron, and (*c*) magnesium silicates, there being subsequent reactions with volatiles to produce, for example, FeS, and FeO for the silicate phases.

The refractory component of a planet (i.e. proportion of early condensate) is estimated from the bulk uranium content, taken to be 18 parts/10^9 for the Earth (Larimer 1971) and 59 parts/10^9 for the Moon, from heatflow data (Ganapathy & Anders 1974). Other established ratios (e.g. Th/U, K/U, Ca/Al) are then used to calculate abundances for a planet relative to cosmic (solar system) abundances. Unfortunately this model has always produced anomalies in regard to the postulated Ca and Al contents versus the petrological evidence for the composition of the Earth and Moon mantles. Hutchison (1976, Appendix) points to the Earth problem, where Ca and Al are over-estimated by 25 % (Al) and 40 % (Ca). Ringwood (e.g. 1975) has been concerned for some time that the estimated Ca and Al contents for the Moon's mantle are far too high in relation to the basalt petrology, as evidenced by the pyroxene compositions.

The recent discovery (Langseth *et al.* 1976) that the Moon's heatflow is much less than estimated from earlier measurements, and based on more reliable determinations over $3\frac{1}{2}$ years (Apollo 15 station) and 2 years (Apollo 17 station), requires a reappraisal of the compositional estimate. The revised heatflow values indicate a bulk U-content closer to about 30 parts/10^9 than the 59 parts/10^9 used in the Ganapathy & Anders model (1974). In that case, all values for the refractory lithophile elements need to be scaled downward by a factor of 2. They become $1\frac{1}{2}\times$ rather than $3\times$ cosmic abundances. This brings them much closer to abundances for the Earth (*ca.* $0.7\times$ cosmic), although the latter may be slightly on the high side (Hutchison 1976).

This less refractory composition for the Moon is particularly significant in terms of its thermal history. As shown earlier (table 1), large-scale melting is now even more necessary in order to concentrate the U into the high crustal values, from the lower initial bulk content now indicated. It also means that the models for bulk-Moon composition that show high Al and Ca contents, and which were unacceptable in details from the experimental petrology data (e.g. Kushiro & Hodges 1974; Delano 1976), are no longer necessary. Instead, the model composition proposed by Binder (1975) is more attractive. This composition approaches closely the composition of the Earth's mantle, either in terms of a pyrolite composition (Ringwood 1966) or by subtraction of the Fe + FeS core from the solar–nebula condensation model for the Earth (Ganapathy & Anders 1974).

The bulk density of the Moon (3.34 g/cm^3) is close to that of uncompressed, low-temperature Earth mantle material. In terms of the refractory lithophile element abundances, it now seems likely that the bulk Moon composition is not far removed from that of the Earth's mantle. The bulk Earth contains about 35 % Fe (Mason 1966) whereas the Moon contains about 9 % Fe (± 4 %) based on magnetometer data (Parkin *et al.* 1973). However, the Earth's mantle probably contains about 7 % Fe (Mason 1966), similar again to that of the bulk Moon.

The bulk Moon is depleted in chalcophile, siderophile and volatile elements relative to cosmic abundances (Ganapathy & Anders 1974, Fig. 3). Whereas the latter depletion must be a bulk characteristic, the other two depleted groups could be due to concentration of those elements in a metallic core. Their abundances are related to Fe abundances and Fe–S reactions, and to depletion in the lunar basalts. If large-scale melting of the Moon has occurred (§5) then the basalt source would be depleted in siderophile and chalcophile elements relative to the core. (If the central 'asthenosphere' of the Moon is hot undifferentiated material, as discussed in §2, the metal-rich phase could have concentrated at the interface between this and the molten shell and later diffused to a core region.)

It is noteworthy that the Earth's mantle is not strongly depleted in siderophile elements such as nickel, suggesting that the mantle was not at any time in chemical equilibrium with the iron-rich core. In contrast, the Moon's mantle is so strongly depleted in nickel that the element could not be accommodated in the core (restricted in volume by bulk density and moment of inertia) if the Moon had accreted a cosmic abundance of nickel. Nickel could be so accommodated, however, if the Moon had begun with a partly-depleted, Earth-mantle nickel abundance, and large-scale melting had 'scavenged' the siderophile elements to the core, more effectively than on Earth.

This raises the question of a possible difference in the evolution of the Earth and Moon. If chemical equilibrium were not attained by the Earth's core and mantle, heterogeneous accretion could have been responsible for a layering in the sequence metal → silicates → volatile-rich veneer, and large-scale initial melting would not be necessary. The volatile-rich outer layer would contribute to the generation of a granitic crust. Shortly after accretion, collision with another body (Cameron 1976, suggests a 'Mars-sized object') could have caused fission and the production of a proto-Moon from the proto-mantle of the Earth. The proto-Moon would need to have been heated extensively at this stage, such that the volatile-rich material was lost and the bulk of the body melted. Loss of volatiles would impart to the proto-Moon its slightly more refractory composition relative to that of the Earth's proto-mantle. Large-scale fractionation then concentrated the siderophile and chalcophile elements towards the core region, and the lithophile refractory elements towards the crust region.

The hypothesis of a fission-origin for the Moon has not been popular for dynamic reasons, but on chemical grounds there is no doubt that the Moon shows much in common with the Earth's mantle. The oxygen isotope data (Clayton *et al.* 1976) are particularly compelling in that respect. Dynamical problems need to be resolved for the Earth–Moon system, as do the problems concerning the heat-source for large-scale melting of the proto-Moon. In so far as we do not yet have a satisfactory hypothesis for the origin of the Moon, however, there is a case for reconsidering the fission hypothesis.

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Discussion

M. J. O'Hara (*Geology Department, Edinburgh University*). The model of structure and mineral composition of the upper mantle of the moon rests upon two important assumptions, namely that the outer several hundred kilometres melted and fractionally crystallized at *ca.* 4.6–4.3 Ga ago, and that the hand specimens of basalts represent the compositions of unmodified primary partial melts of that recrystallized mantle. In addition to aspects of these assumptions which have already been questioned (Ford *et al.* 1977; O'Hara & Humphries 1977) there is at least one consequence of accepting the second assumption which is, apparently, difficult to reconcile with the first.

Granted that selected hand specimens represent primary liquids, experimental petrology studies can identify the pressure, temperature, and nature of the crystalline residuum from which they came. The array of data points defines, for the accepted lunar mantle mineralogy, a series of points on a putative selenotherm (or selenotherms, because the lavas range in age from 3.8 to 3.2 Ga). The implied selenotherms are exceptionally steep, indicating thermal gradients at depth which are about 16 times the adiabatic gradient. What mechanisms exist for freezing such a superadiabatic thermal gradient into the layered crystallization products of a cooling, convecting magma or, alternatively, how can such a gradient be imposed subsequently? Why, if such a superadiabatic gradient existed for at least 600 Ma, is there so little topographic evidence in the surface rocks for extensive convection of the lunar mantle?

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