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A. E. Ringwood

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Mare basalt petrogenesis and the composition of the lunar interior

By A. E. Ringwood, F.R.S.
Research School of Earth Sciences, Australian National University

Mare basalts, which are believed to form by partial melting at considerable depths in the lunar interior, are capable of providing a wealth of information concerning the compositions of their source regions. Conversely, any acceptable estimate of the lunar bulk composition must in principle be able to provide source regions capable of yielding mare basalts. A wide range of lunar bulk compositions has been proposed in the recent literature. These differ principally in the proportions of involatile elements, e.g. Ca, Al, to elements of moderate volatility, e.g. Mg, Si, Fe. A detailed experimental investigation has been made of the capacity of the Taylor–Jakes compositional model (8.2 % Al_2O_3) to provide source regions for mare basalts. It is demonstrated that this composition is much too rich in alumina to be acceptable. Other lunar bulk compositions even richer in Al_2O_3 such as those advocated by Ganapathy & Anders, Wänke and co-workers and Anderson can likewise be rejected. In order to produce mare basalts, particularly the least fractionated varieties represented by some Apollo 12 and 15 basalts, lunar bulk compositions containing only about 4 % of Al_2O_3 appear to be required. This is similar to the alumina content of the Earth's mantle. The relative abundances of many other involatile elements, e.g. Ca, U, Ti, r.e.e., Zr, Ba, Sr, may likewise be similar in the Moon and in the Earth's mantle. These relationships point towards a common origin for the Moon and for the Earth's mantle.

1. INTRODUCTION

An embarassingly wide range of bulk lunar compositions has been proposed on various grounds by several authors (table 1). It is seen that they vary principally in the nature of the abundances of highly involatile elements, e.g. Ca, Al, Ti, to the moderately volatile elements Mg, Fe, Si. An acceptable lunar bulk composition must have the capacity to explain the origin of mare basalts, which are believed to have formed by partial melting of the lunar interior at considerable depths. This capacity can be tested by the methods of experimental petrology. A comprehensive investigation of this nature has been carried out upon the Taylor–Jakes (TJ) composition of table 1 (8.2 % Al_2O_3, 6.6 % CaO) and is described below. A considerable amount of experimental data has also been obtained on the other compositions of table 1. The investigation involved a total of 145 experimental runs at pressures up to 30 kbar. Most runs were carried out between the solidus and liquidus and aimed at determining the crystallization sequence of phases as a function of pressure and the compositions of residual liquids. Compositions of coexisting crystals and glasses were determined by electroprobe microanalysis. A detailed description of the experiments has been published elsewhere (Ringwood 1976).

2. SUMMARY OF EXPERIMENTAL RESULTS

Results of experiments upon the TJ composition are shown in figure 1. Noteworthy are the following characteristics. (i) A vast field exists where olivine crystallizes *alone* before being joined by pyroxene. Olivine would continue to be the liquidus phase at the centre of the Moon.

(ii) Note also the narrow field of crystallization of pyroxenes above 10 kbar. The pyroxene field extends only about 50 °C above the solidus. A large degree of partial melting (~50 %) involving mainly the pyroxene component occurs within this small temperature interval. (iii) Plagioclase does not crystallize at pressures greater than 10 kbar (corresponding to a depth in the Moon of 200 km). (iv) The solidus temperature increases sharply with depth below 200 km and

Table 1. Some recent model lunar bulk compositions

	PY	TJ	GA	W	A
SiO_2	46.0	43.4	41.4	39.9	34.5
TiO_2	0.2	0.3	0.6	0.8	1.3
Al_2O_3	3.9	8.1	11.6	17.4	27.2
Cr_2O_3	0.4	0.4	0.2	0.2	—
FeO	8.2	10.4	6.7	8.1	2.0
MgO	38.1	30.6	30.0	19.6	12.9
CaO	3.1	6.6	9.3	13.6	22.1
Na_2O	0.1	0.1	0.1	0.3	—
$\frac{100\ \text{Mg}}{\text{Mg+Fe}}$	90	85	90	83	—

PY, composition derived from Earth's mantle (pyrolite) with adjustments to Na_2O, TiO_2 and Fe_2O_3 contents. (Modified after Binder 1974.)

TJ, composition proposed by Taylor & Jakes (1974) (with CaO increased from 6.0 to 6.6 % to be consistent with chondritic Ca/Al ratio).

GA, composition proposed by Ganapathy & Anders (1974).

W, composition proposed by Wänke *et al.* (1974).

A, composition proposed by Anderson (1973).

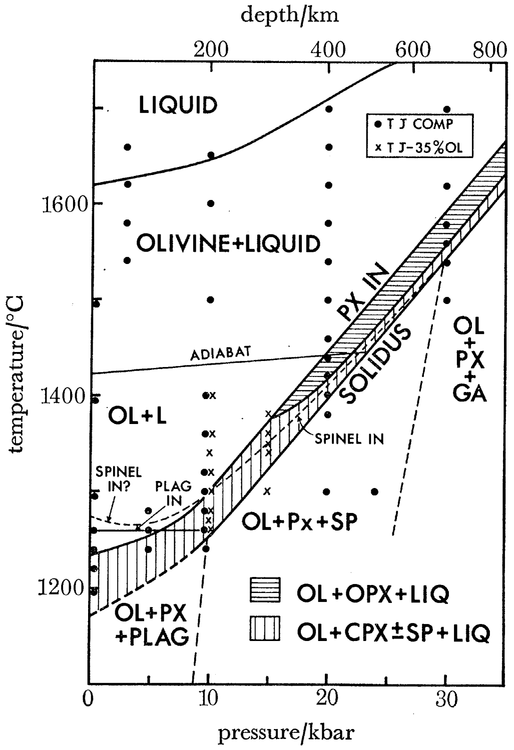

Figure 1. Melting relationships displayed by TJ and TJ minus 35 % olivine model lunar compositions over the range 0–30 kbar. Each dot or cross represents an experimental run.

its gradient is much greater than the adiabatic gradient for the residual liquid. (v) The temperature of plagioclase entry does not increase strongly with pressure between 0 and 8 kbar (0–160 km).

Compositions of phases crystallizing from the TJ composition at 10, 15 and 20 kbar are given in tables 2 and 3. The clinopyroxenes crystallizing between 10 and 20 kbar are seen to be notably aluminous, containing 7–10 % Al_2O_3. The residual liquid after 50–70 % of crystallization is also highly aluminous, containing over 17 % Al_2O_3.

Table 2. Compositions of phases crystallizing from TJ (minus 35 % olivine)† composition at 10 and 15 kbar

	10 kbar. 1300 °C				15 kbar. 1350 °C		
	Ol	Cpx	Sp	glass	Ol	Cpx	glass
SiO_2	40.3	49.0	0.3	47.4	40.3	50.8	46.4
TiO_2	—	0.5	0.1	0.6	—	0.2	0.6
Al_2O_3	—	9.5	50.0	17.0	—	8.4	16.1
Cr_2O_3	0.35	1.5	18.4	0.4	0.4	1.5	0.6
FeO	13.1	5.4	11.3	10.0	12.2	6.1	10.2
MgO	45.8	17.6	19.3	10.8	46.5	21.3	12.8
CaO	0.5	16.5	0.5	13.4	0.5	11.8	13.0
Na_2O	—	—	—	0.4	—	—	0.3
$\frac{100\,Mg}{Mg+Fe}$	86	86	76	66	87	86	69

† TJ-35 % Ol represents starting composition obtained by subtracting 35 % of olivine (Fo_{84}) from the TJ composition. For some purposes, this composition is more convenient to study experimentally than the TJ composition. All experimental runs were saturated with olivine so that results on TJ-35 % Ol apply directly to TJ composition.

Table 3. Compositions of phases crystallizing from TJ composition at 20 kbar

	1420 °C			1400 °C†		
	Ol	Opx	Gl	Ol	Cpx	Sp
SiO_2	40.6	52.5	46.1	40.3	51.7	1.2
TiO_2	—	—	0.7	—	0.15	0.1
Al_2O_3	—	7.0	17.6	—	7.3	61.9
Cr_2O_3	0.2	0.6	0.5	0.1	0.7	4.9
FeO	12.1	7.7	12.4	13.5	6.4	9.8
MgO	46.7	29.2	8.2	45.8	22.5	21.7
CaO	0.4	2.9	14.1	0.4	11.2	0.2
Na_2O	—	—	0.4	—	—	—
$\frac{100\,Mg}{Mg+Fe}$	87	87	55	86	86	80

† Glass present but not analysable.

3. MARE BASALT PETROGENESIS

To be viable, any model lunar bulk composition must be capable of explaining the formation of mare basalt magmas. Three hypotheses of mare basalt formation should be considered. (*a*) Direct partial melting of the bulk composition, (*b*) partial melting of cumulates formed in an earlier differentiation event, (*c*) hybridization between magmas from the primordial interior and early-formed cumulates.

The experimental results discussed above show that the liquids formed by up to 40% of partial melting of the TJ composition contain over 17% of Al_2O_3 and have the character of high-alumina basalts. They do not resemble primitive mare basalts which characteristically contain only about 8.5% of Al_2O_3. The discrepancy shows that, according to the direct melting hypothesis, the source region should contain *much less* than the 8.2% Al_2O_3 of the TJ model.

Differentiation of a completely or partly molten Moon

It is widely believed that the lunar anorthositic crust was formed in a major magmatic differentiation event some 4.4 to 4.6 Ga ago which involved an outer shell at least a few hundred kilometres thick (e.g. Wood 1972). A corollary following on general petrologic grounds is that beneath the plagioclase-enriched crust, there must exist a zone of complementary ferromagnesian cumulates and/or, refractory residua, extending at least to this depth. Some workers have argued that this early major differentiation event affected the entire lunar interior (e.g. Smith *et al.* 1970). A currently popular hypothesis interprets mare basalts as being formed by the remelting of these cumulates (e.g. Schnetzler & Philpotts 1971).

The preceding experimental data permit this hypothesis to be tested rigorously for the TJ bulk composition. The following possible cases are considered (*a*) melting and differentiation of the entire Moon, (*b*) melting and differentiation of an outer shell five hundred kilometres thick. In each case, two important boundary conditions apply. First, because the melting point gradient exceeds the adiabatic gradient below 200 km, crystallization necessarily commences at the base of the molten zone and progresses upwards to about 200 km. Above 200 km, the adiabatic and melting point gradients are similar; moreover, the residual liquid has a basaltic composition (~17% Al_2O_3) and plagioclase is able to crystallize thereby making it possible to form a crust. Secondly, the presence of significant quantities (e.g. > 5%) of trapped intercumulus liquid in *thick* (> 100 km), superadiabatic layers of cumulates causes extreme convective instability (proportional to fourth power of thickness). This is augmented by the increase of Fe/Mg ratio as fractional crystallization proceeds so that the intrinsic density increases with height. Thus, convective motions within cumulates will cause segregation of intercumulus liquids into overlying regions. Intercumulus liquids accordingly cannot play an important role in providing heat sources for subsequent remelting of early cumulates as suggested by Solomon (1975).

The internal structures of differentiated lunar models of TJ composition are shown in figures 2 and 3. In the case of a totally molten Moon (figure 2), crystallization of olivine (Fo_{94}) begins at the centre and proceeds outwards. An olivine ‘core’ thus develops, gradually becoming more iron-rich, reaching Fo_{78} by a depth of 380 km. Above this depth olivine is joined by orthopyroxene which crystallizes over a limited interval before being replaced by clinopyroxene. Between 380 and 200 km, by far the major phase to crystallize is clinopyroxene averaging 8% Al_2O_3 and 11–17% CaO. At a depth of 200 km (10 kbar) the Al_2O_3 and CaO contents of the residual liquid have risen to 17 and 13.4% respectively. Plagioclase first appears at 200 km, accompanied by clinopyroxene ± olivine. Because of the similarity between melting point and adiabatic gradients between 0–200 km, it may be possible for some plagioclase to be elutriated and to accumulate near the surface while olivines and pyroxenes continue to sink. At this stage, a frozen crust consisting of a suspension of plagioclase (~40%) in intercumulus liquids (60%)

may develop (e.g. Walker *et al.* 1975*a*). This composition provides a good match to the composition of the crust (Taylor *et al.* 1973). It should be noted that even if the required amount of plagioclase is introduced into the crust to match its observed near-surface composition, the region between depths of 60 and 200 km still contains over 14 % Al_2O_3, equivalent to more than 25 % of *modal* plagioclase.

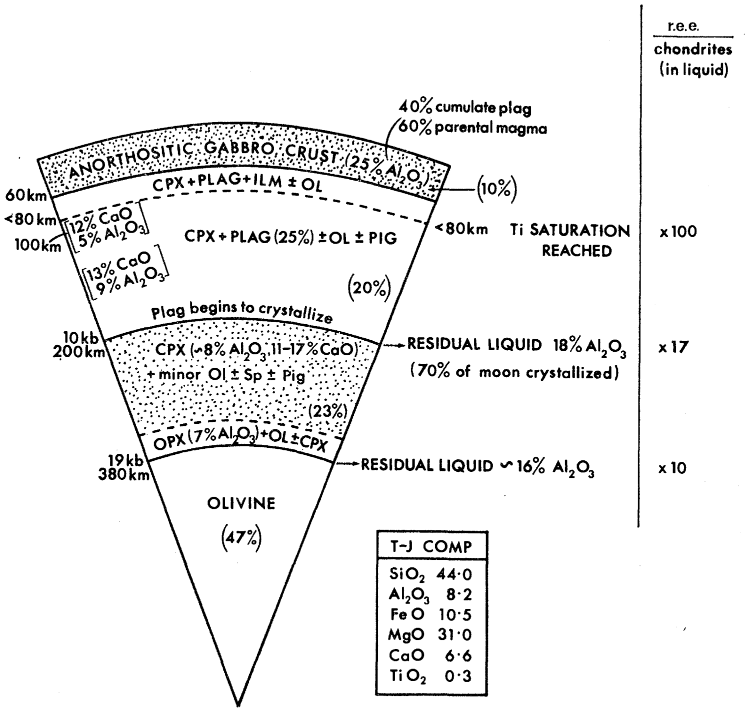

FIGURE 2. Diagram showing the mineralogical zoning which would occur if a moon of TJ bulk composition were completely melted throughout, followed by fractional crystallization. (Not to scale.)

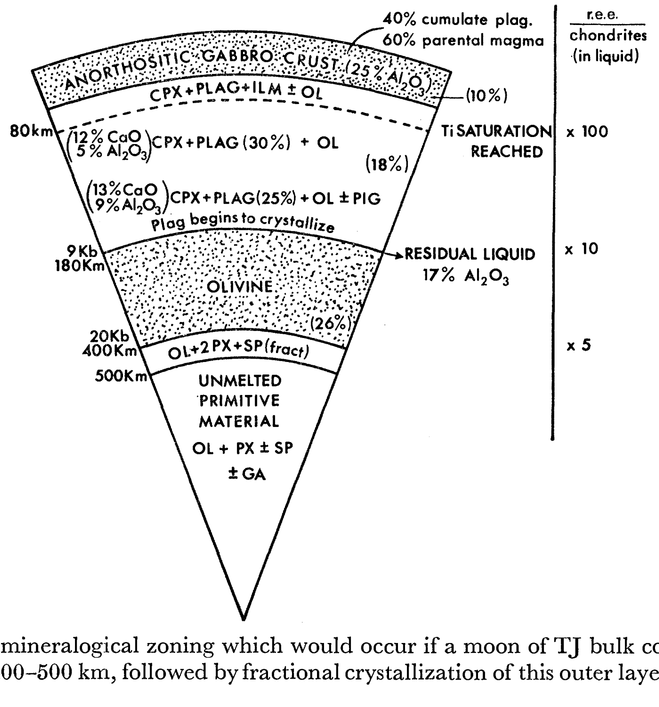

FIGURE 3. Diagram showing the mineralogical zoning which would occur if a moon of TJ bulk composition were completely melted to a depth of 400–500 km, followed by fractional crystallization of this outer layer. (Not to scale.)

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The TJ composition contains 5 times the chondritic abundances of incompatible elements including titanium. By the time crystallization has reached 200 km, the residual liquid has 17 times the chondritic abundances of incompatible elements and the TiO_2 content of the residual liquid has reached 1.5% (allowing for some entry of Ti into pyroxenes). Experimental observations imply that in order for ilmenite to be precipitated as a liquidus phase from residual ferrobasaltic compositions at lunar redox conditions, the TiO_2 content of the liquid must exceed 5 %. This requires fractional crystallization of more than 70 % of the residual liquid between 60 and 200 km. It is clear that precipitation of ilmenite cumulates is possible only at a very late stage from a residual liquid which will accumulate at the top of the earlier cumulates and immediately beneath the crust – i.e. *at a depth shallower than* 100 *km*.

The structure of the lunar interior for the case in which an outer shell some 500 km thick is melted and differentiated is shown in figure 3. An unmelted, unfractionated primitive 'core' would extend from the centre to 500 km followed by a fractionated layer of olivine, orthopyroxene, clinopyroxene and spinel extending to 400 km. Above that level, between 400 km and the surface, an adiabat passes through a zone of olivine plus liquid. Olivine would immediately settle to form a thick dunite layer extending from 400 km to above 250 km. At this point, the residual liquid is still in the primary field of crystallization of olivine. Convection occurs, accompanied by radiative surface cooling, and a further crop of olivine crystals continues to settle until the pyroxene crystallization surface is finally reached between 200 and 180 km. The zone between 400 and 180 km thus consists overwhelmingly of olivine. By a depth of 180 km, the Al_2O_3 content of the residual liquid (now of 'basaltic' composition) reaches 17 % and plagioclase and clinopyroxene begin to coprecipitate, together with minor olivine. Further crystallization above 180 km follows a similar course to the complete melting case.

The cumulate re-melting hypothesis

Walker *et al.* (1975*b*) and others have argued that low-titanium basalts were produced by partial melting of olivine + pyroxene cumulates at considerable depths in the Moon (e.g. 200–400 km) whereas high titanium basalts were produced at shallower depths (around 120 km) by partial melting of late stage cumulates consisting of olivine + pyroxene + ilmenite. We will evaluate this hypothesis in terms of cumulates produced in a differentiated Moon of TJ composition as shown in figures 2 and 3. Before proceeding, however, it should be noted that more recent experimental petrology studies (Green *et al.* 1975; Kesson 1975) do not support this hypothesis since they imply that high-titanium basalts were derived from depths greater than 170 km, not significantly different from the depths of source regions of low-titanium basalts.

Attention is now drawn to some specific difficulties encountered in the cumulate remelting hypothesis for a *completely melted and differentiated* moon of TJ bulk composition.

(*a*) The dunite 'core' below 380 km is not a possible source of mare basalts. The proportion of intercumulus liquid trapped between olivine crystals is likely to be small as discussed earlier. This region thus contains insufficient Ca, Al and incompatible trace elements to provide a second generation of mare basalts by partial remelting, and moreover, does not possess a negative Eu anomaly.

(*b*) Arguments based on experimental petrology show that the most likely source region for many mare basalts is between depths of 200 and 400 km (e.g. Green *et al.* 1971, 1975; Kesson 1975). By far the predominant phase is clinopyroxene averaging 8 % Al_2O_3. In contrast, the pyroxenes on the liquidi of primitive mare basalts in this depth interval contain only 2–4 %

Al_2O_3 (Green *et al.* 1971, 1975; Kesson 1975). Thus, the clinopyroxenes in the potential source region between 200 and 400 km are too high in Al_2O_3, after acceptable degrees of partial melting, to have been in equilibrium with primary mare basalt magmas.

(*c*) Plagioclase does not begin to separate from the residual liquid until depths shallower than 200 km. Thus, clinopyroxene cumulates between 200 and 400 km have never 'seen' plagioclase. The assumed negative Eu anomaly in this region cannot have been caused by preferential partition of Eu in plagioclase which was subsequently removed by flotation as postulated by Taylor & Jakes (1974), Walker *et al.* (1975*a*) and many others.

(*d*) Crystallization of plagioclase commences at a depth of about 200 km (figure 3). We allow, according to the TJ model, sufficient plagioclase to be elutriated upwards to form a crust of mean composition corresponding to the average highlands composition. This requires the removal of 20 % of the residual liquid between 200 and 60 km as plagioclase. Using the experimentally determined plagioclase-liquid distribution coefficient for Eu of 1.0 (Ringwood 1970) the mean negative Eu anomaly between 60 and 200 km is only 20 %, which is an order of magnitude smaller than the Eu anomalies observed in Apollo 11, 12 and 17 basalts. We conclude again, that plagioclase removal into the highlands does *not* provide an explanation of the negative Eu anomalies of mare basalts, on the basis of this model.

(*e*) After elutriating the required amount of plagioclase into the highlands, the region between 60 and 200 km still contains 25 % (av) of *modal plagioclase at the solidus*. Thus, basalts formed by less than 50 % partial melting in this region would necessarily be plagioclase-saturated, contrary to observations on primitive mare basalts (Ringwood 1970; Green *et al.* 1971, 1975; Walker *et al.* 1975*b*; Kesson 1975).

(*f*) Residual liquids do not become ilmenite-saturated until depths smaller than 80–100 km. Between 400 and 80 km, more than 90 % of the liquid has crystallized (after allowance for crust formation). Using measured distribution coefficients (Ringwood 1976) we find that the residual liquid will have been depleted in chromium by more than a factor of 5 while Fe/Mg ratios will be notably increased (Ringwood 1975).

According to the cumulate melting hypothesis, low-Ti basalts are formed by partial melting of early ilmenite-free cumulates, whilst high-Ti basalts formed by partial melting of late ilmenite-bearing cumulates – the product of advanced fractional crystallization. Primitive high-titanium basalts should therefore have much lower Cr contents and higher Fe/Mg ratios than primitive low-Ti basalts. This is not observed. The above characteristics are generally similar both for high-Ti and low-Ti basalts (Ringwood 1975).

(*g*) Experimental petrology investigations (Green *et al.* 1975; Kesson 1975) show that *if* some high-Ti basalts formed by partial melting of olivine + pyroxene + ilmenite cumulates, then the source must lie at depths *exceeding 170 km*. However, in a fractionally crystallized moon of TJ composition, ilmenite does not begin to crystallize until depths less than 80–100 km.

(*h*) The mascons presumably formed before or during the flooding of the mare basins. Their continued existence for up to 3.8 Ga implies the existence of a strong, cool and thick (> 150 km) lithosphere at the time of mare volcanism (e.g. Kaula 1969). This is consistent with thermal history considerations which show that a relatively cool outer shell > 150 km thick would develop in a timescale exceeding 10^9 years (e.g. Solomon & Toksoz 1973). The derivation of high-Ti mare basalt from a source region immediately underlying the crust (60–100 km) requires high temperatures (~ 1200 °C) and partial melting in this region. This is extremely difficult to reconcile with thermal history considerations and moreover, implies a very weak and thin lithosphere which could be incapable of supporting the loads represented by mascons.

Turning now to the model in which melting occurs only in the outer 400 km of the Moon (figure 3), we note that above 500 km, the significant difference between this model and the totally molten Moon model is that the aluminous clinopyroxenite layer of figure 2 (complete melting) is replaced by the dunite layer of figure 3 (molten outer shell). For reasons given previously, the objections to the dunite layer of figure 3 as a subsequent source of mare basalts are even more compelling than those applying to the clinopyroxenite layer of figure 2. The remaining problems encountered in trying to derive mare basalts by partial melting of cumulates in the region above 180 km are similar to the complete melting case.

Discussion

In the light of evidence discussed above, it is difficult to escape the conclusion that a Moon of TJ bulk composition, if totally melted, or if melted only in the outer regions, would differentiate in such a manner as to form a series of cumulates possessing compositions which are unlikely to be parental to mare basalts in subsequent partial melting episodes. The primary weaknesses of the TJ composition from which most of the difficulties described earlier stem are the high abundances of Al_2O_3 and CaO. In order to minimize these difficulties, the abundances of Ca, Al and related involatile elements such as U, Ti, etc., relative to moderately volatile major elements like Mg, Si, Fe should be reduced considerably, perhaps by a factor of two compared to the TJ model. The lunar bulk composition would then approach that of the Earth's mantle, except that the Moon is relatively depleted in voltatile elements.

Several authors have postulated bulk lunar compositions still higher in Ca and Al than the TJ model discussed previously, e.g. Ganapathy & Anders (1974), Wänke *et al.* (1974) and Anderson (1973). These compositions are listed in table 1. Experimental investigations of these models have been carried out (Ringwood 1976) and it is found that they encounter corresponding difficulties to the TJ model except that they are more severe because of their higher Ca and Al contents.

A further difficulty with the differentiated TJ and all the other models mentioned above concerns the composition of the pyroxene component in the lunar crust. The bulk pyroxenes from highland anorthositic gabbros and gabbroic anorthosites tend to be relatively poor in CaO. The mean normative pyroxenes (diopside + hypersthene) in the Wänke *et al.* (1975) and Taylor *et al.* (1973) model compositions for the highlands contain only 6 and 3 mol % of CaO respectively.

For the TJ composition, the pyroxene(s) crystallizing at the stage of plagioclase saturation contain a bulk average of at least 11 % CaO and this increases with further fractionation. It is not readily apparent how models such as that of Taylor & Jakes (1974) which postulate extended fractional crystallization before and during plagioclase precipitation, can account for the low mean CaO content of pyroxenes in the lunar highlands. This problem becomes increasingly acute for other bulk composition models even richer in calcium, such as those of Ganapathy & Anders (1974), Wänke *et al.* (1974) and Anderson (1973).

Hybrid models of mare basalt petrogenesis

A third model of mare basalt petrogenesis seeks to explain many of their unique chemical and isotopic characteristics by a process of hybridization (Hubbard & Minear 1975; Kesson 1975). The outer few hundred kilometres of the Moon is believed to have melted during or soon after accretion, forming a wide range of differentiates including the crust and also late stage cumulates strongly enriched in titanium and incompatible elements. The deeper primitive 'core' which did not participate in the early melting and differentiation was heated over a long time-scale by its original complement of U, Th and K, reaching near-solidus temperatures between 3 and 4 Ga ago. A complex process of assimilative interaction occurred between late stage differentiates and primordial material. According to the model of Kesson & Ringwood (1976) this occurred via the sinking of dense iron-rich pods of late stage differentiates through the differentiated zone to the primordial interior, thereby triggering off extensive partial melting and hybridization processes.

Regardless of detailed mechanisms it is clear that for hybrid models, it will be the mare basalts which have experienced the least contamination by early (ca. 4.5 Ga) differentiates which possess the most direct 'memory' of their primordial source region. The low abundances of incompatible elements in some Apollo 15 basalts (e.g. 15555) and Green glass, combined with small europium anomalies, show that these magmas have been contaminated only slightly by the early cumulates and were produced mainly by direct partial melting from the primordial interior. These magmas contain only 8% of Al_2O_3 whereas we noted earlier that up to 40% partial melting of the TJ composition would yield magmas containing over 17% Al_2O_3. It seems clear that the source region of Apollo 15 basalts and Green Glass contained no more than about 4–5% of Al_2O_3, i.e. about twice chondritic levels. Likewise, the low abundances of involatile incompatible elements (e.g. r.e.e., U) in these samples (X4–12 chondritic) in relation to the degree of partial melting and permissible contamination required to explain their composition imply a source region containing only about twice the ordinary chondritic abundances of r.e.e., U, Al, Ca, etc.

4. CONCLUSIONS

This investigation has shown that the TJ bulk lunar composition containing 8.2% Al_2O_3 as well as other proposed bulk compositions containing more than 8% Al_2O_3 are unable to provide acceptable source regions for mare basalts. This conclusion holds whether mare basalts are produced by direct partial melting, by the remelting of cumulates produced in an earlier differentiation event, or by hybridization processes. Moreover preliminary investigations on a composition containing 6% of Al_2O_3 (Kesson & Ringwood, in prep.) show a similar series of difficulties. It seems clear that, in order to provide low-alumina source regions capable of ultimately producing mare basalts, the bulk Moon should contain only about 4% of Al_2O_3. This is similar to the Al_2O_3 content of the Earth's mantle, or twice the (ordinary) chondritic abundance of Al_2O_3. It appears likely that other involatile elements such as Ca, Ti, Zr, r.e.e., Ba, Sr, U, Th, etc., are also present in the Moon at about twice the (ordinary) chondritic abundances and are similar in this respect to their abundances in the Earth's mantle. The composition of the Moon appears generally to resemble that of the Earth's mantle, the principal difference being that the Moon is relatively depleted in volatile elements. This resemblance is enhanced by the identity of oxygen isotope compositions in the Moon and Earth (Clayton & Mayeda 1975) and by the discovery of xenon of terrestrial isotopic composition as a tightly bound component of lunar breccias (Lightner & Marti 1974). These similarities strongly imply a genetic relationship between Moon and Earth and, in addition, support the hypothesis that the Moon was ultimately derived from the Earth's mantle subsequent to the formation of the Earth's core (Darwin 1908; Ringwood 1960, 1970).

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