Differentiation of a very thick magma body and implications for the source regions of mare basalts

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Abstract—Mass-balance calculations indicate that the molten layer originally covering the moon may have been several hundred kilometers deep. The solidification of this magma ocean involved some unusual effects resulting from the large-pressure differential in the thick magma layer. The sunken cumulates from this differentiation would be more iron-rich, less refractory, and would contain more incompatible elements than would be expected in simple isobaric fractional crystallization and crystal sinking models of magma differentiation. These lower cumulates can have the mineralogical and chemical properties of the source regions of the low-Ti mare basalts.

INTRODUCTION

Several lines of geochemical and geophysical evidence suggest that wholesale melting of the outer portions of the moon occurred very early in its history, perhaps due to accretional heating. The basaltic samples returned from the lunar missions may be partial melts of source regions which were differentiated during the solidification of that large magma ocean. It is the purpose of this paper to try to estimate the size of this melted zone, to examine its mode of differentiation, and to see if the postulated differentiation products can have the properties of the source regions of lunar basalts.

MAGMA BODY SIZE

Evidence for a global melting event early in lunar history will not be reviewed in detail here but has been deduced from petrologic, trace-element, geophysical, and isotopic considerations. Early identification of this event was made by Wood et al. (1970) and Smith et al. (1970) on the basis of the complementary nature of mare basalts and anorthositic particles found in the soil at the Apollo 11 landing site. The feldspathic particles were presumed to be characteristic of lunar crustal material exposed in the highlands. The complementary nature of the basalts and crustal materials was further indicated by respective negative and positive europium anomalies documented by many including Philpotts and Schnetzler (1970) and Wakita and Schmitt (1970). Seismic, gravity, and rheological considerations indicated that the lunar crust was not of trivial thickness and must have been produced early to support mascons. Rb–Sr systematics of lunar rocks and soils and widespread occurrence of ancient Rb–Sr model ages suggested that the event responsible for this state of affairs occurred very near to the formation of the...
moon at $\sim 4.6 \text{ Gy}$ (Papanastassiou and Wasserburg, 1971). These early deductions have been reinforced by later studies.

The size and initial composition of the magma body are open questions at present. Estimates of the initial composition of the magma body or, alternatively, assuming homogeneous accretion, the bulk composition of the moon have been made by several authors: (1) Anderson (1973), (2) Ganapathy and Anders (1974), (3) Ringwood and Essene (1970), (4) Taylor and Jakeš (1974), and (5) Wänke et al. (1974). None of these suggested compositions appears to be capable of satisfying all geochemical, petrologic, and geophysical constraints but some models appear to be much better candidates than others. There is no obvious way to make model (1) consistent with experimental petrology constraints on the source regions of mare basalts. Model (5) is too low in silica to allow much pyroxene in the source regions of mare basalts. Model (3) may satisfy some gross geophysical constraints on the properties of the moon, but it would be difficult to produce the lunar crust by differentiation of this composition. On the other hand, models (2) and (4) appear to be able to generate rocks of the lunar crust by fractional crystallization and give a plausible constitution for the lunar interior. Hodges and Kushiro (1974) have investigated the phase relations of model (2) as a function of temperature and pressure and reached the conclusion that it could give a generally satisfactory account of lunar petrogenesis and lunar geophysical properties with the following exceptions. The Fe/Mg ratio of lunar crustal rocks was higher than might be expected from model (2) and fractional crystallization of model (2) did not produce troctolite as a rock type. Both these difficulties can be overcome by recalculating slightly more of the total Fe in model (2) as oxidized Fe rather than metal. If model (2) is oxidized it is very similar in major-element composition to model (4) except for its lower silica content. This modified model would precipitate spinel during fractional crystallization at low pressure, leading to Cr depletion of the residual liquid. Since as much as 3% of the crystallization product can be spinel without producing marked Cr depletion (Walker et al., 1975a,b) and since the existence of spinel troctolites requires that there have been some spinel accumulation, a successful model will precipitate only a minor amount of spinel during differentiation. Minor oxidation of model (2) or a slight decrease in silica in model (4) would be an acceptable solution. Therefore we have chosen bulk compositions for the large magma system (or for the moon assuming homogeneous accretion) intermediate between the model of Ganapathy and Anders (1974) and Taylor and Jakeš (1974). (SiO$_2$, 43; Al$_2$O$_3$, 8.2; FeO, 10.5; MgO, 31; CaO, 7.0.)

These magma bulk compositions are shown on Fig. 1. Oxygen units are used to facilitate volume estimates. Also shown is an estimated average composition for the lunar crust deduced from a compilation of published surveys of lunar feldspathic rocks (Walker et al., 1973). This composition is roughly equivalent to "anorthositic norite." Clearly for the case where the anorthositic-norite crust is a differentiate of the large magma there must be a considerable volume of complementary material stored somewhere in the system to preserve mass balance. The central problem is to deduce the character and volume of this missing material.
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Fig. 1. Estimates of lunar bulk composition are indicated by open symbols. A is from Anderson (1973); GA is from Ganapathy and Anders (1974) calculated with all Fe oxidized; HK is from Hodges and Kushiro (1974) and is based on GA with only part of the Fe oxidized; RE is from Ringwood and Essene (1970); TJ is from Taylor and Jakeš (1974); W is from Wänke et al. (1974). Compositions which satisfy petrological and chemical constraints on the lunar bulk composition should lie nearly on a line between Mf$_2$SiO$_4$ ($Mf = MgO + FeO$) and the peritectic involving spinel, olivine, anorthite, and liquid (Walker et al., 1973, 1975a,b). The successful model compositions are labeled MAGMA. The compositions labeled CRUST correspond to anorthositic norite. MAFIC CUMULATE indicates the nature of the material complementary to the crust formed in differentiation of MAGMA. The diagram is plotted in oxygen units to facilitate estimates of the relative volumes of CRUST and MAFIC CUMULATE as indicated on the lines connecting them since oxygen comprises most of the volume of the materials. Oxygen units are calculated by weighting the component formulae by the number of oxygens they contain.

Also shown in Fig. 1 is a set of low-pressure liquidus phase relations which can serve as a rough guide to crystallization behavior (Walker et al., 1973) although the excess high-calcium pyroxene component (Ca:Al > 0.5) is not considered. Crystallization at the upper surface of the magma where heat loss occurs should be controlled by these liquidus boundaries. Olivine should crystallize first (somewhat above 1400°C) followed by co-precipitation of spinel until plagioclase crystallizes at ~1275°C. Olivine and plagioclase precipitate until ~1225°C when low-calcium pyroxene joins the crystallization. The lunar feldspathic crust is presumably the non-sinking fraction of crystalline material so produced. Differential plagioclase accumulation is responsible for the relative buoyancy of the crust as well as the major component of chemical variation in the crustal rocks. To explain Fe/Fe +
Mg variation in the crustal rocks we need only note that plagioclase fractionation is accompanied by olivine and pyroxene fractionation which can change Fe/Fe + Mg of the evolving liquids. Clearly this ferromagnesian material is not all preserved in the crust and must exist as a complementary cumulate at depth beneath the crust. The volume of this complementary cumulate can be estimated from the bulk composition of the system and the seismically determined volume of the anorthositic-norite crust.

Fig. 1 illustrates how this estimate can be made. It is assumed here that the mafic cumulate is olivine and pyroxene, that it is devoid of plagioclase, and that it has no more than 3% spinel. It is also assumed that the bulk composition is differentiated only into feldspatic crust and complementary mafic cumulate. The residual liquids of the system are lumped with the mafic cumulate for the present. The complementary cumulate must then represent at least \( \frac{2}{3} \) or \( \frac{1}{4} \) of the material of the whole magma system estimated from the volume proportions shown in Fig. 1. If the complementary mafic cumulate contains plagioclase or other aluminous phases then it will represent even more of the total differentiate. Likewise, since there appears to be little excess high-calcium pyroxene in the anorthositic-norite bulk composition for the crust, the excess high-calcium pyroxene in the bulk composition of the magma system must be incorporated in the complementary mafic cumulate. This consideration would also tend to increase the estimated proportion of material represented in the complementary cumulate. Seismic experiments have been interpreted to show that the lunar crust is about 60 km thick on the moon’s nearside (Tóksoz et al., 1974). The amount of shattered feldspatic crustal material versus mare basalt in the upper 20 km is uncertain and undoubtedly highly variable. For \( \sim 60 \) km of anorthositic norite the whole system must extend to a depth of 200–275 km depending on whether the crust is \( \frac{2}{3} \) or \( \frac{1}{4} \) of the system. The 200-km figure agrees with the estimate of Hodges and Kushiro (1974) who also noted that this must be a minimum thickness for the whole system.

Wood (1973) has noted that many aspects of lunar asymmetry may be caused by a greater thickness of crust on the moon’s farside. If we change the crustal thickness estimate to 100 km, the whole system could be as much as \( \sim 500 \) km thick. The presence of plagioclase or other aluminous phases (other than the 3% spinel) and the inclusion of the excess high-calcium pyroxene below the crust will further increase this estimate.

This estimate involves melting of roughly half the volume of the moon. If accretion of cold material is the heating agency to cause the melting, then the magma ocean cannot be very much thicker than this estimate. The total gravitational energy released in accreting the moon is \( \sim 400 \) cal/g, which is about sufficient to heat and melt the whole moon, but the accretion process will release heat in a manner dependent on the square of the radius of the growing planet, so the 400 cal/g is not distributed uniformly. Potential energy converted to heat, even with complete efficiency, will be sufficient to melt (and later superheat) subsequent infall only when the moon has reached \( \frac{3}{2} \) of its present radius. Appreciable kinetic energy of accreting particles or accretion of hot particles could increase
the depth of the melted zone. Rapid accretion is necessary to prevent escape of the released heat. Mizutani et al. (1972) have considered the process of accretion and its implication for lunar thermal history in some detail. They conclude that a satisfactory account of the moon’s thermal properties can be given by an accretion model which assembles cold material in about $10^2$ yr. The depth of the molten layer is about 400 km, which is compatible with possible estimates of the magma thickness based on chemical and mass-balance arguments above. We must now inquire how such a thick magma will solidify and what will be the character of the differentiation products.

WHAT ROCKS ARE AT 200–300 KM IN THE MOON?

Studies of the melting behavior of low-Ti, olivine-rich, mare basalts suggest that they would be in equilibrium with an olivine and low-calcium pyroxene residue at a depth of 200–300 km in the moon (Green et al., 1971; Grove et al., 1973; Kesson, 1974; Longhi et al., 1975; Walker et al., 1975). We have already noted that the complementary mafic cumulate should be composed of those minerals and could easily extend to this depth. An important question in lunar petrogenesis is whether the low-Ti mare basalt source region lies within the complementary mafic cumulates or in undifferentiated material. The former possibility has been considered unlikely by Ringwood (1975) on the grounds that Fe/Fe + Mg and incompatible element abundances attributed to the complementary mafic cumulate are too low to be consistent with the properties of the low-Ti mare basalt source region. For instance, the bulk Fe/Fe + Mg of the model composition we are considering is 0.16; the first olivines to precipitate would have a Fe/Fe + Mg of about 0.06 whereas the source region for the low-Ti basalts must have a Fe/Fe + Mg of 0.24–0.29. The crystallization of the magma at the top of the system where heat loss occurs would precipitate ferromagnesian minerals which sink to form the cumulate. Hodges and Kushiro (1974) quantitatively pursued this approach for a 200-km magma and concluded that about half the system would be composed of dunites and harzburgites. These materials would indeed make poor-mare basalt source regions because of their low Fe/Fe + Mg, low-incompatible element abundance, and refractory nature. This is especially true for a perfect fractionation or totally efficient differentiation model. The very size of the system dictates relatively slow heat loss and hence would promote efficient differentiation by crystal settling. However, for a magma of 400-km depth the very large size also introduces effects which reduce the efficiency of the differentiation process. We will now consider these effects.

EFFECTS OF MAGMA SIZE ON DIFFERENTIATION IN THE GENERAL CASE

The principal agent which introduces unusual effects into the differentiation of a very large magma body is the large pressure difference between top and bottom (≈20 kbar for 400 km on the moon). This pressure differential will change the
crystallization behavior of the magma with depth so that different minerals may crystallize. It will also increase the melting temperature with depth at a rate such that the melting temperature gradient is steeper than an adiabatic temperature gradient. Figure 2 shows a compilation of liquidus curves of various lunar materials which have olivine as the low-pressure liquidus phase. Materials with more normative olivine have higher liquidus temperatures and the change from liquidus olivine to liquidus pyroxene occurs at higher pressure. The temperature

Fig. 2. Liquidus curves as a function of temperature and pressure are plotted for lunar materials having olivine on the liquidus at low pressure (from Longhi et al., 1975; Stolper, 1974; Walker et al., 1972, 1975a,b). The materials with higher temperature liquidus curves have more normative olivine. Note that there is a regular increase of the temperature of pyroxene-olivine co-saturation with increasing pressure. Olivine-pyroxene co-saturated liquids are also increasingly olivine-normative with increasing pressure. Average saturation gradients are indicated. 14072, the most noticeably deviant composition, is distinctly more aluminous than the other compositions. These regularities can be used to roughly extrapolate the liquidus of a composition having olivine as the low-pressure liquidus phase if the low-pressure liquidus temperature is known.
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gradient of a magma in liquidus equilibrium with olivine would be \( \sim 0.15^\circ C/km \); with pyroxene it would be \( \sim 0.4^\circ C/km \). If a magma were saturated with both olivine and pyroxene throughout its vertical extent the temperature gradient would be \( \sim 0.8^\circ C/km \). Note that this situation would be accompanied by an appreciable compositional gradient in the liquid since the olivine + pyroxene liquidus boundary curve shifts to more olivine-normative compositions with increasing pressure.

The large-magma system may be expected to crust at the top where rapid heat loss occurs. Growth of the crust by heat loss through it will precipitate olivine, plagioclase, and pyroxene. We have noted above that much of the ferromagnesian material was not retained in the crust but sank due to its high density. Olivine which sinks from the bottom of the solid crust (at \( \sim 1200^\circ C \)) will undergo compression while sinking. An adiabatic temperature rise of about 20\(^\circ C\) would be experienced in sinking 400 km. However, the actual temperature rise in the magma must be considerably more than this just for it to be in the liquid state. Furthermore, as we will discuss later, the accretion-produced thermal profile would have a considerable body of superheated magma at depth. Clearly the sinking olivine will not be in thermal equilibrium with the liquid. The olivines will dissolve until the superheated magma is cooled to its liquidus temperature by supplying the heat of fusion. Note also that this process introduces a compositional gradient in the liquid. The magma will have a higher liquidus temperature and become more olivine normative with depth as olivine dissolves. When magma at one depth is cooled to its liquidus temperature, olivine will no longer dissolve but can sink to saturate a deeper level by dissolution there. However, in sinking through the saturated level the cooler olivine must induce extra crystallization of the magma as it approaches thermal equilibrium. Crystal sinking in a large magma may be considered a negative mode of heat transfer which induces crystallization at depth even though heat is lost at the top of the system.

Continued crystal sinking will bring the magma at depth to co-saturation with olivine and pyroxene with a corresponding compositional gradient with depth. The temperature gradient for this situation is seen in Fig. 2 to be \( \sim 0.8^\circ C/km \). The ability of the system to reach this gradient, rather than "choking" and solidifying at some intermediate depth, depends on the original amount of superheat with depth, the rate of heat loss at the top, and the rate of crystal settling. Rapid crystal addition and small superheat will prevent this gradient from being established over large ranges of depth. To give a feeling of the possible magnitude of the effect for the case where such a situation is reached, we may calculate the amount of heat necessary to bring an olivine sinking from the crust into equilibrium with the magma as it sinks to the bottom of the system. This heat must be supplied by crystallization of the saturated magma through which the cold crystal sinks. In sinking 400 km the olivine experiences a temperature rise of 20\(^\circ C\) due to adiabatic compression. At 400-km depth the magma is 300\(^\circ C\) (400 \(\times\) 0.8 – 20) hotter than the olivine. Using representative heat capacity data from Robie and Waldbau (1968) for olivine, about 14 kcal/gfw is necessary to raise the olivine to the magma temperature. For a heat of fusion of olivine of \( \sim 22 \) kcal/gfw about half as much
olivine as sank into the magma would be required to crystallize in order to keep the olivine in thermal equilibrium with the magma. (These heat capacity values are uncorrected for pressure but serve to give a rough idea of the order of magnitude of the process.) This extra crystallization would, of course, not all take place at 400 km but would happen continuously during sinking.

The pressure gradient also may change the nature of the phases crystallizing with depth. For the case we have just discussed, the composition gradient with depth probably will keep olivine as the liquidus phase with depth. However, if there is appreciable mechanical stirring (e.g. from impacts) or convective stirring (to be considered later) and the compositional gradient is suppressed, the sinking olivine may in some cases become unstable with respect to pyroxene. Should the sinking olivine armor itself with a reaction product of pyroxene, the lesser heat of fusion of pyroxene (15–18 kcal/gfw) would require that a larger number of moles of pyroxene precipitate than was the case of precipitation of olivine to heat the cold sinking olivine. The number of moles of precipitated pyroxene must be comparable to the number of moles of sinking olivine rather than the previous estimate of about half as much for the simple olivine case if olivine precipitation alone is to supply the heat. At depth the reaction relation between olivine, pyroxene, and the liquid becomes a co-precipitation relationship if the liquid composition is allowed to adjust itself according to the pressure constraints. Therefore a combination of olivine and pyroxene precipitation would supply the necessary heat to warm the sinking olivine. In this case the sinking crystal would induce extra crystallization of the liquid by an amount from about a half to a whole of itself, depending on the proportions of olivine and pyroxene precipitated. This is to say that \( \frac{1}{2} \) to \( \frac{3}{2} \) of the crystal accumulate is material precipitated at depth even though all the heat loss leading to this crystallization occurred at the top of the system.

The result of this process of heat transfer by crystal sinking is not only to accelerate crystallization at the bottom of the system but also to affect the constitution of the cumulate produced. Simple efficient differentiation models not considering this effect give sunken cumulates of dunite then harzburgite which have very high Mg/Mg + Fe at the bottom. The model presented here favors harzburgites over dunites although dunites are still possible. The cumulate will be less refractory as a result of the introduction of pyroxene. Furthermore, the Fe/Fe + Mg of the cumulates will be higher at the bottom of the system. This effect is illustrated schematically in Fig. 3. The upper part of the figure indicates schematic liquid-solid "loops" for phases entertaining Fe–Mg solid solution. At lower pressure \( P_0 \) the solidification curves are at lower temperature \( T \) than at high pressure \( P_\infty \). The lower part of the figure indicates the variation of Fe–Mg in cumulates produced in two different situations. The solid curve is for fractional crystallization at a constant pressure \( P_0 \). The initial phase crystallizing and sinking to the bottom to form the basal layer of the cumulate has a composition appropriate to the liquidus phase for the bulk composition (open symbol) at low pressure. Subsequent fractional crystallization produces extreme iron enrichment. The dashed curve indicates the variation in the resulting cumulate if there is
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Fig. 3. Schematic illustration of the effect of pressure on the distribution of Fe and Mg in the cumulates produced during differentiation of a large-magma system. See text for discussion.

a substantial pressure gradient between the top and bottom of the melt. Crystals sinking at the top of the system have the liquidus composition as in the first case. However, at deeper levels the shift of the "loop" indicates that the sinking phase becomes more iron-rich as a consequence of the greater degree of crystallization. This is a schematic representation, since the sinking will not be isothermal, but it serves to illustrate that the extra crystallization induced during sinking drives up the Fe/Fe + Mg of the crystalline fraction of sinking material. The composition of the basal cumulate layer will be intermediate between the two vertical dashed lines. The exact composition depends on the degree of Fe exchange with the melt during sinking, the heat capacity, and heat of fusion of the phase. This basal layer will always be more iron-rich than in the first case. As the magma fills up with its settling products the pressure gradient and settling distance decrease, so the "loops" at later stages in the differentiation merge with the $P_0$ "loop." Extreme
iron-enrichment occurs at the end of the sequence but a smaller volume of iron-rich material is produced since more iron was incorporated in the early cumulate. The solid and dashed curves are shown intersecting at an Fe/Fe + Mg value corresponding to the right vertical dashed line. This situation is possible but need not be the case in general. The composition and depth of this intersection is not uniquely specified by the position of the “loop.”

It should be noted this increase in Fe/Fe + Mg of the basal cumulate occurs whether the material transfer is by convection (adiabatic) or crystal sinking (not adiabatic). Either way the increased degree of crystallization induced during sinking increases the Fe/Fe + Mg of the crystalline material deposited at the bottom of the system. The FeO component of the residual liquid is more uniformly distributed in the system than during efficient differentiation.

It is doubtful that the mode of solidification discussed above would ever operate fully for reasons to be discussed below. However, the discussion does show that simple crystal settling models of differentiation are quite inadequate to describe a 400-km magma body and may give incorrect insights into the constitution of the differentiation products. A further factor to be considered is that the increase in the liquidus temperature of the system with pressure is considerably greater (factor of 3–8) than the adiabatic temperature gradient (about 0.05°C/km in the lunar case). The adiabatic gradient should approximate the temperature distribution in a convecting liquid. If our magma system convects then it should solidify chiefly from the bottom up since the adiabat will intersect the steeper liquidus at the bottom of the system first (Fig. 4). This mode of solidification will also subvert the efficiency of differentiation since the opportunity for crystal–liquid separation by crystal settling is diminished when the crystals form at the bottom of the system. The basal members of the solidification product would probably be harzburgite or lherzolite (with garnet or spinel). A complication to be noted is that the model moon composition has an appreciable melting interval so that crystals may precipitate over quite a distance in the column and settle before the bottom is solidified fully. Another complication, to be developed later, involves likely initial temperature distributions which may be gravitationally stable over large parts of the magma. However, neglecting these special initial situations, the magma, when cooled at the top, becomes exceedingly unstable with respect to convection. In the presence of a superadiabatic temperature gradient convection will occur in any magma body deeper than about 10 m for reasonable values of viscosity, thermal expansion, and diffusivity. The Rayleigh number for a 400-km magma with an adiabatic gradient is about 10^{22}. The size outweighs all the other factors. It is clear then that convection should be an important part of any analysis of magma differentiation on this scale. If solidification from the bottom upwards were the only process operating, however, the existence of a thick lunar crust containing material of low Fe/Fe + Mg would be unaccounted for.

It should be noted that once a material becomes partly crystalline it is rather effectively “buffered” against normal convective cycling on an adiabat. Should a parcel of magma with suspended crystals begin to rise, melting of the crystals begins which cools the parcel as the heat of fusion is supplied. The temperature is
buffered on the melting curve which makes the parcel cooler than its surroundings since the melting curve is steeper than the adiabat. Parcels of magma which begin to sink experience extra crystallization. Heat of fusion is liberated, warming the parcel. In general the density contrast produced by melting or solidification is much greater than the density contrast produced by thermal expansion. Therefore sinking parcels continue to sink and rising parcels continue to rise even though they may be respectively warmer and colder than their surroundings. This is the reverse of the temperature relations for normal convective cycling in a homogeneous liquid which has an adiabatic temperature distribution.

We have now considered two "end-member" processes which introduce unusual effects into the solidification of large magmas. Both processes will operate in the solidification of a magma to spread the heat loss sustained through the top crust to the lower parts of the magma and accelerate solidification there. The net result is a reduction in the efficiency of differentiation by fractional crystallization and crystal settling because crystallization will take place throughout the magma. We now consider a possible outcome of the solidification of the large lunar magma body for plausible initial conditions when these effects are taken into account.

Fig. 4. The effect of having melting curves steeper than adiabats is to induce crystallization at the bottom of the system and during sinking. This effect has been discussed in some detail by Irvine (1970). Ascending convective magma cools at the top of the system and descends, intersecting the melting curve at depth. In actual magma differentiation this effect is complicated by the presence of a melting interval with generally more than one crystallization curve.
DIFFERENTIATION OF THE LUNAR MAGMA OCEAN

To outline the solidification process of such a large-magma body we must know the phase relations of the material as well as the initial temperature distribution. The phase relations as a function of pressure are deduced in the following way. The solidus is taken from Hodges and Kushiro (1974). The solidus temperatures turn out to be rather insensitive to bulk composition since this solidus is similar to the one proposed by Ringwood and Essene (1970). The low-pressure liquidus temperature (olivine) is taken from putting the bulk FeO and MgO into Roeder and Emslie’s (1970) olivine-saturation model. The high-pressure liquidus and olivine + pyroxene saturation curve can then be deduced by extrapolation of the trend shown in Fig. 2. The result is shown in Fig. 5. The principal heat source for the initial melting event is presumed to be accretion. The accretionary temperature profile shown in Fig. 5 is essentially that of Mizutani et al. (1972) but has been adjusted for the larger melting interval in the present model composition.

Fig. 5. Possible initial conditions for the primordial lunar magma ocean are illustrated as discussed in the text. Important features include melting to a substantial depth, a large amount of superheated magma, and a gravitationally stable zone deeper than the temperature maximum.
Inspection of the initial conditions postulated in Fig. 5 reveals several points of interest. The material is molten to a depth of \(\sim 400\) km. (Accretion of hot material would make the molten layer thicker.) Subsequent material infalling after the melting temperature is reached will be superheated. A decline in the accretion rate at the end of the accretional process allows cooling of the surface layer. A maximum in temperature is induced in the system. The temperature profile at depths below the maximum is gravitationally stable and there would be no tendency for convective instabilities to develop. At depths shallower than the temperature maximum, the profile is exceedingly unstable with respect to convection indicating that this initial state of affairs must be very transitory. In actual fact as the rate of infall declines, the subsequent portion of accreting material will probably be falling into a convecting bath. This state of affairs or that which occurs very soon after the decay of the convectively unstable profile is illustrated by profile 1 in Fig. 6. A final feature to be noted in this differentiation

Fig. 6. Illustration of possible mode of solidification of the magma. Initial conditions in Fig. 5 are given by curve 0. Curves and cartoons 1, 2, 3, and 4 are for subsequent stages in the solidification. Temperature profiles are schematic. The gravitationally stable zone which traps residual liquids in the lower cumulates amplifies the normal tendency to trap residual liquids discussed in the text.
model is the presence of a crust at the top of the system. This is imposed as a boundary condition even though there is some tendency for the crust to be disrupted by the convecting layer underneath. This boundary condition is geologically reasonable since large terrestrial lava lakes develop frozen crusts in the presence of convection. Furthermore, the presence of high-Mg/Mg + Fe material in the present lunar crust is most easily explained in this manner (Taylor and Jakeš, 1974). The problem of crust formation is discussed further in a later section.

Material cooled and partially crystallized at the top of the system becomes dense enough to sink due both to thermal contraction and to the presence of suspended solids. It should be noted here that plagioclase (ρ = 2.7) should not sink and would probably float in the fractionated liquids from which plagioclase crystallizes (calculated ρ = 3.0 by the scheme of Bottinga and Weill, 1970). A further argument that plagioclase preferentially remains at the top to be incorporated in the crust is that if more than 15% of the sunken material is plagioclase then the mass-balance calculations presented earlier would require a body of magma larger than the whole moon be melted to extract the crust. The separation of plagioclase from the fractionated liquid and crystals which sink depletes Eu in this descending material.

As the cooled liquid and its mafic minerals begin to descend, further crystallization is induced since the melting curves (e.g. solidus) increase in temperature faster than the liquid adiabat. Even if there is no heat exchange with the surroundings, the sinking material undergoes extra crystallization during sinking. Therefore the actual temperature gradient followed by a descending convective body will be steeper than an adiabat due to release of heat of fusion. The gradient, however, will not exceed the crystallization curve the body follows. It should be noted that there was a “floor” for convective updrafts introduced by the top of the gravitationally stable regime (note dotted line in Fig. 6). This is not a “floor” for downdrafts since they are colder than the surroundings and contain suspended crystals. Sinking material will descend to the bottom of the melted zone. The extra crystallization induced during descent will make the solid material of higher Fe/Fe + Mg than at the beginning of the descent. When the material stops descending there is no incentive for any residual liquid remaining in the batch to rise again for another cycle of convective overturn and further differentiation since the material is now in the gravitationally stable regime. Any further separation of residual liquid occurs only by upward physical displacement of the liquid by denser settled crystals.

The thermal effects of such a cycle can be seen in the sequence of curves 1 through 4 on Fig. 6. Rising of material and cooling at the top of convective curve 1 will lower the maximum temperature of the system from 1 to 2. Sinking of the cool material displaces the remainder of the system upwards. The “floor” of convective updrafts moves up from 1 to 2. To the extent that the sinking material exchanges heat with its surroundings, the temperature slope of the gravitationally stable regime in 2 is less than in 1. The ceiling of the convective zone is lowered as the crust at the top grows. Heat loss out the cold bottom of the system induces
final solidification of sunken material and its trapped juices from the bottom up. Further stages of this process are illustrated by curves 3 and 4 in Fig. 6.

The geochemical novelties introduced in this differentiation scheme are that Fe/Fe + Mg of the sunken crystalline cumulate does not show as wide a variation on a large scale as might be expected during efficient fractionation and that residual liquids on the bottom are not convectively recycled for further enrichments but are trapped with the cumulates. (This material is of course depleted in Eu as a result of plagioclase separation which occurs at the top of the system before this material descends.) The first effect results from the extra crystallization introduced during sinking and operates independently of the second effect. The second effect is a result of the accretionary thermal profile. Inasmuch as this scheme of differentiation overcomes most of the geochemical objections (Ringwood, 1975) to regarding the mafic cumulates as the source material of low-Ti olivine-rich mare basalts, we may conclude that it is still possible to regard the cumulates as a reasonable source region on geochemical grounds.

DISCUSSION

The solidification process outlined in the previous section indicates that the bulk of lower cumulate layers produced in a large-magma body are relatively undifferentiated. If the process operates fully the lower cumulates will differ from the bulk composition of the system only in having lost plagioclase (and Eu, etc.) to the crust. The Fe/Fe + Mg will be roughly the same as the bulk composition and the full complement of Eu-depleted residual liquid is successively incorporated in accumulating layers. This product contrasts with the very magnesian dunite which would be expected in a simple fractional crystallization and crystal sinking model where half the volume of the system crystallizes as magnesian olivine and sinks before plagioclase precipitation initiates crust formation. Instead of a barren, magnesian dunite, a harzburgite forms with some layers enriched in Eu-depleted residual liquid component providing a more attractive source region for low-Ti mare basalts during later remelting.

A more quantitative treatment of the process does not seem possible at this time. Details of the initial temperature distribution, the rates of crust formation and heat loss at the surface, the flow rates and patterns in the magma, and the degree to which chemical equilibration is achieved remain unknown. Further complications which require exact treatment for proper understanding are the effects of compositional gradients introduced by precipitating and dissolving material at intermediate depths. Two features of the model, however, seem to require further discussion. First, how are the trace-element enriched residual liquids produced and where are they to be found at the end of the solidification process? Second, what is the process of crust formation atop a convecting, superheated, magma body?

There is important circumstantial evidence that the trapping of residual liquid in the cumulate has not been complete and uniform. Titaniferous lunar basalts which have strong enrichments in incompatible elements (and Ti) appear to be
derived from source regions in the lunar interior at depths of ~100–150 km, perhaps from a series of ilmenite cumulates derived from late residual liquids (Walker et al., 1975a,b). If this is the correct explanation for the source region of high-Ti basalts then there may have been some recycling of residual liquid during differentiation to achieve the observed trace-element enrichments. In the framework of the proposed differentiation model this could happen in several ways. Crystal sinking may to a certain extent physically displace residual liquid upward from the lower cumulates. Alternatively the solidification by heat loss through the bottom can have an effect analogous to zone refining. As the solidified zone advances upward through the lower cumulates, the incompatible elements may be driven upward if there is good intercumulus plumbing. Either way the \( \text{Fe}/(\text{Fe} + \text{Mg}) \) of the residual liquid is "buffered" by the crystalline phases of the cumulate while the degree of incompatible element enrichment increases. Finally it should be noted that as substantial degrees of solidification of the system are obtained, the pressure differential in the smaller volume of residual magma is less, so that the effects described here become less important. To the extent that the unusual size effects do not totally control the differentiation, the remaining magma fraction will differ from the original bulk composition and later cumulates may increase in \( \text{Fe}/(\text{Fe} + \text{Mg}) \). Specifically, curve 4 in Fig. 6 probably does not correspond to a real situation because changes in the liquid will have been introduced which modify the phase relations and temperature profiles. The thick magma effects will be most effective in the early stages of differentiation.

The formation of a crust on a large body of superheated magma also requires further inspection. Existence of a crust was assumed as a boundary condition in the discussion above because the moon has a crust and its relatively high \( \text{Mg}/\text{Fe} \) ratio is most easily explained if it is produced by freezing the top of the magma rather than by later differentiation from material beneath. For crustal formation to occur in this way it is necessary that ascending magma parcels lose enough heat at the top of the system to crystallize at least to the point of plagioclase precipitation. If the feldspar is free to physically separate from the cooled and partly crystallized material which then sinks, this feldspathic segregate becomes the crust. To the extent that feldspar can separate, the crustal material will be gravitationally stable against sinking (\( \rho \) for anorthositic norite \( \approx 2.9 \) whereas \( \rho \) for the liquid \( \approx 3.0 \)). It remains to be shown that enough cooling of the superheated magma can occur at the top of the convective zone to allow partial crystallization and feldspar separation before the magma has substantially solidified from the bottom upward.

Details of the initiation of crust formation are not clear but may resemble the following. Initially the heat flux cooling the top of the superheated magma is radiation into space and has a magnitude \( \sim 7 \text{cal/cm}^2 \cdot \text{sec} \) for assumed "blackbody" behavior. This flux is sufficient to cool and crystallize the magma at the rate of \( 10 \text{m/day} \). This rate of crustal growth cannot be maintained after solidification of the top layer limits the heat transfer to the conductive mode. Furthermore this rate of crustal growth is too rapid to allow concentration of plagioclase in the crust by differential buoyancy. Therefore the rapidly frozen crust will sink if disrupted,
as it most certainly would be by continued accretionary bombardment from above
and turbulence in the magma below. Crustal fragments would tend to dissolve
rather quickly during sinking into the superheated magma beneath. Since the crust
at this stage is undifferentiated "chill margin," this dissolution does not introduce
compositional gradients but does serve to reduce the temperature of the magma at
shallow depths beneath the rapidly reforming crust. As the rate of crust growth
decreases during the change to conductive (rather than radiative) heat loss at the
top and as the thermal gradient decreases at shallow depths in the magma, it
becomes possible to separate plagioclase and olivine. Upward plagioclase con-
centration leads to the establishment of a crust which no longer sinks when
interrupted. This insures that the slow conductive heat loss, conducive to plagi-
oclase separation and further crustal growth, continues. The important feature of
this proposed mode of crust formation is that the cooling power of radiative heat
loss can be distributed to shallow depths by crustal resorption. A relatively
slow-cooling regime which is then established directly beneath the "chill margin"
allows olivine–plagioclase separation and formation of a non-sinking crust. This
non-sinking crust is then stable against perturbations and grows by differential
plagioclase retention from magma crystallizing slowly at its base by conductive
heat loss. Ruptures in this crust should heal by a process like that leading to
crustal formation when hot magma is again exposed to radiative cooling at the
surface.

For the proposed process to be successful, the stable crust must be established
before large quantities of the partially cooled layer beneath the crust are
subducted by convection. It is possible that the proposed process may succeed.
Crystal separation rates leading to crust formation (subtraction of olivine by
settling) are driven by liquid–solid density contrasts and may be more rapid than
the convection rates driven by thermal contrasts in the liquid alone which act to
reduce the opportunity for establishing a zone conducive to plagioclase–olivine
separation.

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