Re-examination of the lunar magma ocean cumulate overturn hypothesis: melting or mixing is required

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Abstract

There is a long-standing hypothesis that the last fraction of the lunar magma ocean crystallized into a layer of dense, Ti-rich cumulate minerals at shallow depths ($\sim$100 km) early in the moon's history. Many questions remain about the stability of these high-Ti cumulates. It has been suggested that the cumulates subsequently sank deep into the moon because of gravitational instability, but high-Ti material is required at shallower depths by 3.5 Ga to create the high-Ti mare basalts and picritic glasses. The high-Ti material may have re-erupted from depth, or some or all of it may have remained at shallow depths throughout lunar history. Data on phase stabilities, bulk compositions, densities, and temperatures of melting and crystallizing in addition to results from numerical modeling suggest that the high-Ti cumulates would sink only under highly specific conditions. Five scenarios for sinking high-Ti cumulate materials are examined, and only two are found plausible. In particular, it is found that simple sinking of solidified high-Ti cumulates is unlikely because the temperature at which the cumulates solidify is low, and viscosity under these conditions is very high. It is, however, possible that high-Ti cumulates mixed with a substantial fraction of olivine would have viscosity low enough to allow them to sink as solids. Further, because clinopyroxene and ilmenite melt in a ratio of 2:1, remelted high-Ti cumulates would be negatively buoyant and sink as liquids, percolating downward through the underlying mantle and beginning to recrystallize ilmenite at 200 km depth, making a hybrid, heterogeneous mantle. © 2002 Elsevier Science B.V. All rights reserved.

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1. Introduction

The moon's interior is thought to consist of cumulate zones produced by early ($\sim$4.55 Ga) magma ocean differentiation. Models of lunar magma ocean crystallization [1–4] suggest that the deepest cumulates (300–400 km) are primarily mafic, olivine $+/-$ low-calcium pyroxene that crystallized near or sank to the base of the magma

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ocean, while the uppermost cumulates are primarily plagioclase, which floated upward and formed an anorthositic crust. Highly fractionated, late-stage liquid was trapped between the deep mafic cumulates and the anorthositic crust. Depending on the model composition used, ilmenite begins to crystallize from the late-stage liquid when 89–95% of the magma ocean has solidified (P. Hess and J. Longhi, personal communication). Depending on the thickness of anorthosite crust assumed, ilmenite crystallization would therefore begin at a depth between 150 and 100 km at a temperature between 1180 and 1125°C [2,5]. Crystallization of ilmenite+clinopyroxene ± plagioclase would continue until KREEP, the final liquid differentiate, crystallizes beneath the anorthositic crust. The KREEP model age of 4.42 Ga [6] is thought to mark the completion of magma ocean crystallization.

Once crystallized, the high-Ti cumulates would have a density of 3700–3800 kg/m^3, compared to the underlying olivine+pyroxene mantle density of about 3300 kg/m^3. Ringwood and Kesson [1] proposed that because the solid ilmenite+clinopyroxene cumulate layers were denser than the underlying, less evolved olivine- and pyroxene-bearing cumulates, the ilmenite+clinopyroxene cumulates sank into the underlying cumulate pile after crystallization. If the high-Ti component sank into the deep moon, then the question of how and where the high-Ti picritic glasses and mare basalts obtained their titanium component remains. Ringwood and Kesson [1] and other investigators (e.g. [2,3,7]) hypothesized that the sunken high-Ti cumulates subsequently remelted and contributed to the formation of high-Ti mare basalts and picritic glasses. Other investigators have challenged the hypothesis of deeply foundered, remelted high-Ti cumulates on the basis that high-Ti melts are not buoyant below 200–300 km depth in the moon [8,9] (as shown in Fig. 1); paradoxically, the multiple-saturation depths for the high-Ti picritic glasses are all below this depth limit. Because of the difficulty of eruption from their depths of multiple saturation, we believe the multiple-saturation points are recording different information for the high-Ti glasses than they do for the low-Ti picritic compositions.

The multiple-saturation depths and saturation phases of picritic glasses and mare basalts contain clues about the location of high-Ti material in the moon. Eight experimental multiple-saturation points are shown in Fig. 2. Experimentally determined multiple-saturation points indicate the approximate depth and temperature of melt segregation from its source, and the phases at multiple saturation are thought to indicate the residual source mineralogy. The mare basalts thus far investigated all have multiple-saturation points shallower than 300 km depth, while the picritic glasses all have origination depths greater than 250 km. The green glass multiple-saturation points indicate a source mineralogy of olivine+orthopyroxene, and so, surprisingly, do the high-Ti glasses: none of the high-Ti glasses yet studied has ilmenite on its liquidus at any pressure. If the high-Ti glasses were the result of deep remelting of sunken high-Ti cumulates (e.g. [1]), then each should show high-Ti cumulate mineralogy on its liquidus. The absence of ilmenite on the liquidus of any of the high-Ti compositions means either that the melt fraction was high enough to exhaust the high-Ti phase in the source, or that the Ti content was added as a later solid assimilant, which would not be reflected in experimental phase equilibrium results.

The green glass compositions also provide evidence for the spatial distribution of Ti in the moon. The compositional trends of the Apollo 14B green glasses show increasing titanium from assimilation simultaneous with fractional crystallization of olivine. Therefore the Apollo 14B data require high-Ti material for assimilation shallower than their depth of multiple saturation (480 km)[10] at between 3 and 4 Ga.

Questions remain, however, about the processes of high-Ti cumulate redistribution; most importantly, whether it is possible for solid or liquid cumulates to sink. Zhong et al. [7] suggested that the high-Ti cumulates sank to the core–mantle boundary, and then rose, melted, and erupted at the time of mare basalt genesis. This mechanism would probably leave high-Ti material at shallow mantle depths to accommodate the picritic glass compositional trends, and could nicely explain the predominance of mare basalts on the
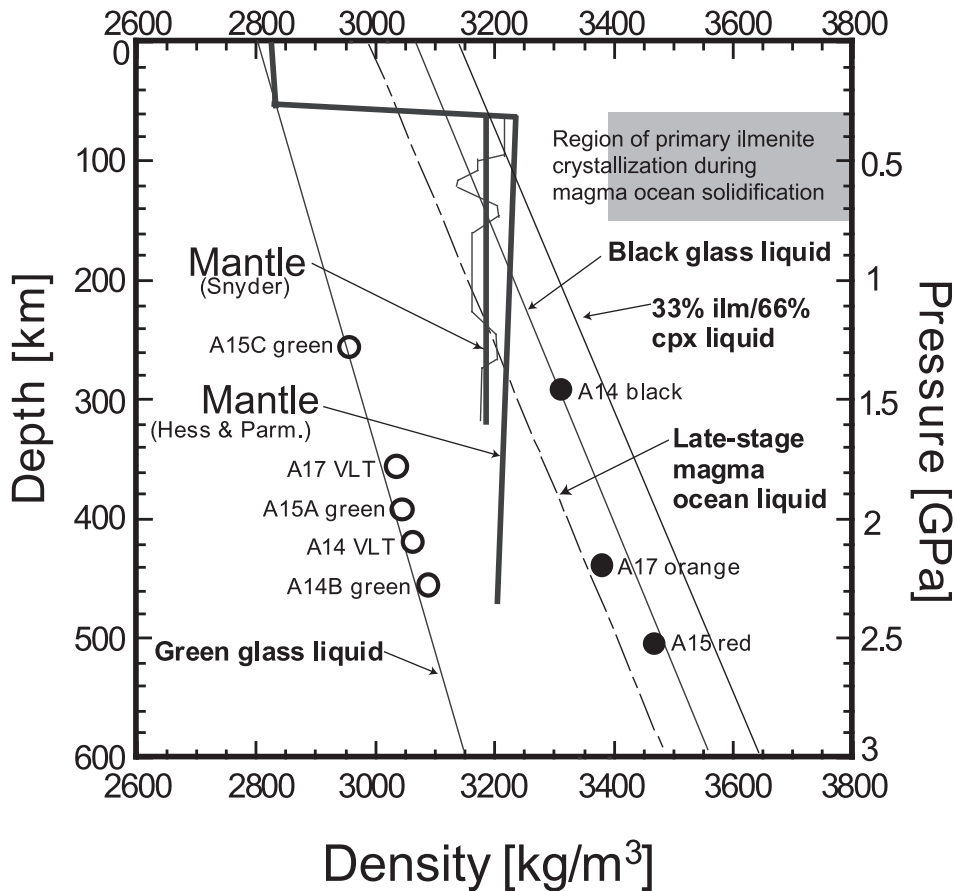
Fig. 1. Lunar depth versus density curves for liquid compositions and two lunar mantle models [2,3]. The vertical scale shows pressure in GPa and corresponding depth in the moon. High-density spikes above 100 km representing solidified high-Ti cumulates have been removed from the mantle models. Multiple-saturation points are plotted at their experimental pressure and calculated density (A15C Green = Elkins Tanton and Grove, in preparation; A17 VLT = [30]; A15A Green = Elkins Tanton and Grove, in preparation; A14 VLT = [31]; A14B Green = [10]; A14 Black = [9]; A17 Orange = [32]; A15 Red = [18]). The black glass density curve is from [8]. The green glass density curve is from [15], and our calculations. Other liquid curves calculated as described in text.

near side of the moon, but questions remain about re-erupting dense high-Ti liquids, and, more importantly, about the initial fall of the cumulates.

In this study we examine several processes that may have allowed the shallow, late-crystallizing high-Ti material to sink into the deep moon. Using finite element numerical fluid dynamic models of the moon's interior and data on phase stability, bulk compositions, densities, and temperatures of melting and crystallizing, we show that the conditions that would allow high-Ti cumulates to sink are restrictive. Five scenarios are examined,

and only two are found likely to allow significant volumes of high-Ti material to relocate deeper into the moon.

2. Models

Addressing the possibility of sinking solids or downward-percolating liquids requires models of the host mantle density and viscosity, the high-Ti cumulate mineralogy, density, liquidus and solidus temperatures, and viscosity, as well as expressions for density as a function of depth and com-
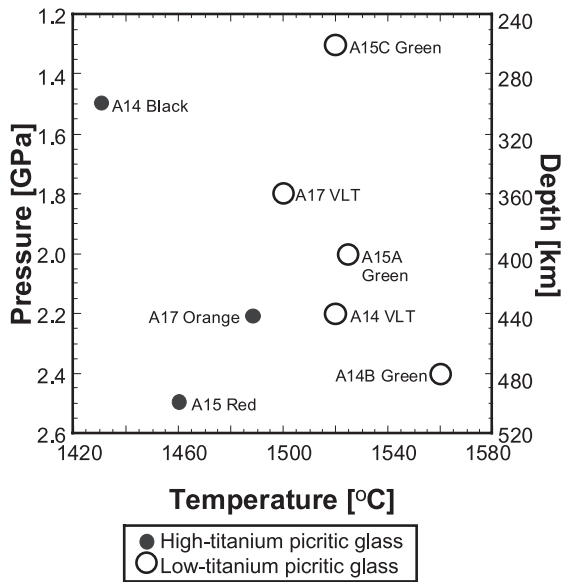
Fig. 2. Pressures and temperatures of multiple saturation for the eight picritic glass compositions listed in Fig. 1. High-Ti compositions are shown with filled circles, and low-Ti compositions with empty circles.

position for liquids. These parameters allow us to answer simple questions of comparisons, such as, at 90 km depth, will a late magma ocean liquid be negatively buoyant with respect to the lunar mantle? To look further at the complex dynamic problem of flowing solid materials, a numerical code is required, such as the finite element fluid dynamical code used in this paper.

The density of the lunar mantle as a function of depth is calculated using the cumulate pile models of Hess and Parmentier [2] and Snyder et al. [3]. For the mantle cumulates we calculated the weighted average density of the phases thought to be present at each depth by finding the phase composition in equilibrium with the modeled Mg# for each depth, and then estimating the density of each phase. This yielded a model curve of mantle density with depth just after crystallization of the magma ocean. We removed the spike of dense high-Ti cumulate from its original position under the light anorthosite crust to make the data interpretation clearer (it is indisputable that the solid high-Ti cumulate compositions modeled by Hess and Parmentier [2] and Snyder et al. [3] are

denser than any possible underlying mafic mantle cumulates). No correction for pressure is included, as the densities of the solid phases under consideration increase by only a few percent over the pressure range of interest [11].

The densities of model lunar magmas, magma ocean liquids, and cumulate remelts are calculated as a function of composition, temperature, and pressure. The density at 0.1 MPa was calculated using the method of Lange and Carmichael [12]. For each of the liquids considered, the density as a function of pressure was then calculated using a third-order Birch–Murnaghan equation of state following Delano [13].

Circone and Agee [8] determined the density versus pressure relationship experimentally for a high-Ti lunar glass, and found Ti content to be a determining variable in the density relationship. Our calculations closely match their experimental results. For the very low Ti Apollo green glasses, the equation of state determined by Miller et al. [14] for a komatiite liquid, with $K = 25$ GPa and $K' = 5$, matches our Birch–Murnaghan calculation as well as the green glass density curve determined by Longhi [15].

Fig. 1 shows the multiple-saturation points, liquid density curves, and mantle cumulate curves plotted on axes of density versus pressure (and corresponding depth in the moon). This presentation of the data aids in determining the gravitational stability of a given material in comparison to the hypothesized lunar cumulate mantle. All the liquid compositions are buoyant compared to the mantle at shallow depths, and all the liquid compositions cross the mantle curves at some depth and become negatively buoyant. The green glasses are buoyant at their depths of multiple saturation, but each high-Ti glass is more dense than the lunar mantle at its depth of multiple saturation. The high-Ti magmas could never have risen and erupted if formed in entirety at their depths of multiple saturation.

To model the fluid dynamics of the lunar interior, more specifically, the onset of Rayleigh–Taylor instability (formation of falling or rising diapirs because of a density contrast between layers), we use a version of the finite element code Con-Man [16] that includes compositionally and ther-

mally driven convection, and a stress-dependent viscosity.

Stress-dependent rheology is particularly important when modeling the formation of instabilities: nascent instabilities quickly build up stress while deforming, thereby lowering their viscosity and enabling flow. This feedback process speeds the fall of instabilities significantly when compared to simple temperature-dependent rheology. Though the extrapolation of the stress dependence of viscosity from laboratory to planetary scales is an inexact science, addition of stress dependence to the viscosity law only increases the likelihood of the model predicting that the high-Ti cumulate fell into the deeper moon, and therefore the inclusion of stress-dependence is doubly important for modeling this process. The viscosity law used depends on stress, temperature, and depth:

$$\eta = \eta_0(\sigma_0/\sigma)^2(\exp[((E+vz)/(T+T_0))-$$

$$((E+vz_0)/(1+T_0))]) \tag{1}$$

where η_0, σ_0, T_0, and z_0 are reference values of viscosity, stress, temperature and depth, respectively, E is the activation energy for solid-state creep, and v scales viscosity to depth. E is set to create an order of magnitude change in viscosity for each 100° change in temperature, and v is set to create a factor of five total viscosity change across the depth of the model. Because an accurate viscosity law for ilmenite+pyroxene is not known, in these models it is assumed to follow the same law as olivine. This is a generous assumption: pyroxene is stiffer than olivine [17–19], and a higher viscosity caused by pyroxene

in the cumulate would inhibit the formation and sinking of cumulate diapirs.

Thermal convection is governed by the Rayleigh number:

$$\mathrm{Ra} = (\rho g \alpha \Delta T h^3)/(\eta_0 \kappa) = 38000 \tag{2}$$

where α is thermal expansivity, κ is thermal diffusion, ρ is a characteristic density, ΔT is the temperature change over the model box, and h is a characteristic length. Convection due to the negative buoyancy of the high-Ti cumulate is controlled by the compositional Rayleigh number:

$$\mathrm{Ra_c} = (\Delta\rho g h^3)/(\eta_0 \kappa) = \mathrm{Ra}[\Delta\rho/(\rho\alpha\Delta T)] =$$

$$-100000 \tag{3}$$

where $\Delta\rho$ is the difference in densities between the compositions. The numerical model, which is 260×260 km, is illustrated in Fig. 3, and model parameters are given in Table 1. The boundary conditions for each model are the same: free-slip in the horizontal along the top boundary, free-slip in the vertical along the side boundaries, and a flow-through bottom boundary. The bottom of the KREEP layer is also constrained to be a free-slip boundary with no flow-through.

We ran a large suite of models, and report four here. For each of the models, the top 15 elements (60 km) are anorthosite crust, which is modeled with an initial viscosity 100 times that of the mantle. The next layer is three elements (12 km) of KREEP. This is probably a significant overestimate of the thickness of the KREEP layer, but numerical considerations require that at least three elements be used in each material. A thicker

Table 1
Numerical model parameters

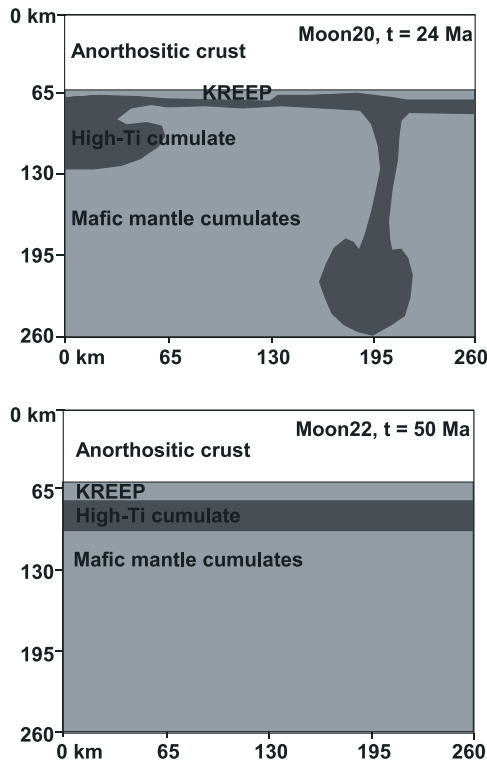|		
h	height and width of model box	260 km
ρ	reference mantle density	3250 kg/m³
$\Delta\rho$	difference between cumulate and mantle densities	−400 kg/m³
ΔT	temperature across the model box	1400°C
η_0	reference viscosity	1.00E+20 Pa s
α	thermal expansivity	3E−05 /°
κ	thermal diffusivity	1E−06 m²/s
e	number of elements in model box, in each dimension	67

Fig. 3. Results of numerical models of solid high-Ti cumulate instabilities. Top panel shows the only model in which instabilities formed. Overturn progressed only under entirely unrealistic conditions, where no surface cooling occurred and the temperature of the cumulates was above their solidus. Bottom panel shows the results of more realistic starting conditions, where as a starting condition the model was conductively cooled until the high-Ti cumulate layer was at its solidus.

KREEP layer will only facilitate descent of the high-Ti layer. The initial viscosity of the KREEP layer is 100 times less than the mantle. Below the KREEP layer are five elements (20 km) of high-Ti cumulate. The initial viscosity for the high-Ti layer and the initial temperature conditions are varied among the models. The remainder of the model box is made up of 43 elements (170 km) of silicate mantle.

Three numerical model runs describe the parameter space of high-Ti cumulate descent. In the simplest model, Moon20, the entire model box begins at 1400°C, and the high-Ti cumulate and the silicate mantle begin with the same initial viscosity. This is an extreme and unrealistic model: there is no conductive cooling of the surface, and no allowance for the relative stiffness of pyroxene in comparison to olivine. The high-Ti cumulate forms diapirs that sink to 260 km depth by about 25 Ma (shown in Fig. 3a). This is the sinking behavior found in previous models of cumulate overturn, but we found this behavior only in this extreme model. Note that for the hot initial condition assumed, a real high-Ti cumulate layer would be fully liquid and positively buoyant, not negatively buoyant, as in the calculations.

A second model, Moon21, starts with this same 1400°C initial condition (no conductive cooling), but with the initial viscosity of the high-Ti layer 10 times that of the silicate mantle, to approximate the stiffer rheology of pyroxene in the cumulate layer as compared to the olivine-dominated mantle. In this model the high-Ti cumulate layer never forms instabilities and does not fall. This model was designed to test the effects of a higher viscosity pyroxene cumulate, and we found that if the mineralogy of the cumulate layer raises its viscosity an order of magnitude above that of olivine under the same conditions, then cumulate sinking is halted.

In a third model, Moon22, the most realistic starting conditions are used, and high-Ti cumulate sinking did not occur. To crystallize, the ilmenite+clinopyroxene cumulate must be below 1150°C [5,20], and so with the uncooled initial conditions used in models Moon20 and Moon21 the high-Ti cumulate would still be liquid, and this liquid would be positively buoyant. In this model, Moon22, the model box is cooled conductively until the bottom of the high-Ti layer reaches 1170°C (a value probably still above the solidus of the material, but chosen to facilitate flow by providing the minimum viscosity). With a cooled start, using the olivine viscosity law for the high-Ti cumulates, no instabilities form. The stationary cumulate layers are shown in Fig. 3b. With the conductively cooled starting condition, instabilities do not form until the background viscosity (the starting viscosity of the mantle and high-Ti cumulate) is lowered to 10^{16} Pa s at the start of the model run.

In a variation of this most realistic model, Moon22, the lower boundary of the high-Ti layer

was perturbed to form a nascent instability. This is an important test, since Conrad and Molnar [21] demonstrated that, especially with stress-dependent viscosity, the time of development of instabilities is strongly controlled by initial perturbations. For this test we placed a hanging bulge of cumulate in the center of the model box 24 km wide and 20 km at its deepest point was added to the existing 20-km thick high-Ti cumulate layer. Though this central bulge created a very sluggish material flow downward beneath its center, encouraging growth of the instability, the material was stiff enough that over 500 Ma no instability formed and none of the high-Ti cumulate fell.

The existence of a lubricating, liquid KREEP layer above the high-Ti cumulate helps the Rayleigh–Taylor instability to develop [22], but as KREEP has a solidus only slightly cooler than the high-Ti solidus, and the KREEP will be closer to the cold surface of the moon, it is unlikely that the KREEP layer will stay liquid long enough to facilitate the instability. To maximize the possibility of instability, the KREEP layer in this model is given a viscosity 100 times less than background. Still, no instabilities are formed, and the cumulate layer is frozen at its depth of formation.

3. Discussion

There are several possible mechanisms that could allow redistribution of high-Ti liquids or solids into underlying cumulates. The most commonly suggested process for moving high-Ti material into the deep moon is the sinking of solid high-Ti ilmenite+pyroxene cumulates as instabilities from their depth of crystallization [e.g. [1,2,7]]. We find this process unlikely due to viscosity constraints. Attempting to find a process that does allow high-Ti material to move more deeply into the moon, we then investigate models that attempt to lower the viscosity of the solid cumulate (models 2 and 5), and models in which a heavy liquid may be negatively buoyant with respect to the lunar mantle, and sink through percolative flow (models 3 and 4).

3.1. Could the solid ilmenite+pyroxene cumulate layer sink via a Rayleigh–Taylor instability?

No. Numerical models demonstrate that the process becomes a race to form Rayleigh–Taylor instabilities in the high-Ti cumulate before it freezes in place. We find that under realistic initial conditions, taking into account the temperature at which high-Ti cumulates become negatively buoyant, and even with a perturbed lower boundary, the viscosity is too high to allow high-Ti cumulates to fall.

To allow instabilities to form, the high-Ti material must be hot enough that its viscosity allows it to flow. Using the viscosity law shown above, we found that instabilities would not form until the shallow moon was ∼1300°C, hot enough that the cumulate would be well above its solidus, and therefore no longer gravitationally unstable at shallow depths. This is the falling behavior described in previous models of cumulate overturn, but we found this behavior only in this high-temperature model without solidus and density constraints. Previous investigators have simplified this problem by modeling without conductive cooling and temperature-dependent viscosity constraints, and that is why some papers have shown instabilities forming and falling on appropriate time scales. It is common to simplify a modeling problem by using constant viscosity, but in this case, with a near-surface material having a low solidus temperature, more realistic modeling is necessary.

3.2. Could the ilmenite+clinopyroxene cumulate sink as a solid if it were heated radiogenically?

No. Radiogenic heating of the high-Ti layer would not lower its viscosity and allow it to fall; the numerical models examined above assume the high-Ti layer is just below its solidus, and so any heating will simply melt it. Radiogenic heating in the KREEP layer would keep it liquid longer, but if the top of the high-Ti cumulate were at 1200°C (and the cumulate were somehow still solid) then viscosities would be such that not even a liquid KREEP layer would lubricate sufficiently to allow the cumulate to fall.

3.3. Could a late-stage liquid from magma ocean crystallization become negatively buoyant?

No. At the point when ilmenite appears on the liquidus and begins to crystallize, the late-stage liquid has only about 6% TiO_2, and would not be negatively buoyant at any depth above 250 km (see Fig. 2). It is not possible to sink a late-stage magma ocean liquid.

3.4. If the ilmenite+clinopyroxene cumulate could be remelted, would that secondary liquid sink?

Yes. Radiogenic heating or shock heating from giant impacts could raise the ilmenite+clinopyroxene cumulate above its solidus and produce a dense liquid. Several experimental studies (e.g. [5,23]) have demonstrated that clinopyroxene and ilmenite melt in a ratio of ∼2:1, resulting in a liquid with about 33% ilmenite at maximum. Such a liquid would be negatively buoyant with respect to mafic mantle cumulates at depths below 20–80 km, depending on the mantle composition used, making it a good candidate for carrying a high-Ti component deeper into the moon (Fig. 1). The bulk of the ilmenite+pyroxene cumulate probably crystallized between 60 and 100 km, and thus it is likely that any radiogenic remelting (for example, from the Procellarum KREEP terrane [24]) would occur in a region of negative buoyancy and any resulting high-Ti liquid would be denser than the surrounding lunar mantle. The giant impacts of the late heavy bombardment then could have shock-melted any remaining shallow portions of high-Ti cumulate.

The ilmenite+clinopyroxene cumulate is likely to have also contained plagioclase. The composition studied by Van Orman and Grove [5] contained 7% plagioclase by mass. When fully molten, the density of their high-Ti cumulate is about 3040 kg/m^3 at 1 atm. This composition will be negatively buoyant with respect to the cumulate models of Hess and Parmentier [2] and Snyder et al. [3] at between 140 and 200 km. These depths are probably too great to be helpful in high-Ti migration; the bulk of the ilmenite probably crystallized at shallower levels, and remelting mechanisms such as impact or radiogenic heating prob-ably also occurred more shallowly. However, if the melting extent is small enough that clinopyroxene or, better yet, ilmenite is not exhausted, then the liquid would still be negatively buoyant at its likely source depths. On the other hand, if the source contained more than 7% plagioclase, then obtaining a negatively buoyant liquid by re-melting is increasingly unlikely.

Downward liquid flow-through percolation would likely be fast enough to create a hybrid mantle for later remelting, in part because of the very low viscosity of clinopyroxene–ilmenite melt. Using percolation theory from Wark and Watson [25], we find that a matrix grain diameter of 0.1 cm and porosity of 1% results in a melt velocity of 0.4 cm/yr. If the porosity is 5%, the melt velocity rises to 9 cm/yr. Because melt velocity scales with the square of the grain size, larger grains in the matrix will significantly speed the downward liquid flow.

The steep liquidus of the ilmenite+pyroxene liquid would cause ilmenite to begin crystallizing at ∼1200°C and less than 200 km depth [5,20]. At shallow depths, the liquid will also most likely crystallize olivine and dissolve pyroxene. At greater depths, the liquid will be in its pyroxene primary phase volume, and will dissolve olivine and crystallize pyroxene. This makes possible the formation of a heterogeneous, hybrid mantle from downward flow of high-Ti liquids. If the liquid did percolate downward at 9 cm/yr, then it would have reached 200 km depth in about 2 Myr. This timing is adequate to account for the onset of high-Ti mare basalt eruptions, which began near the end of the late heavy bombardment [6].

3.5. Could the cumulate composition reasonably be modeled in another way, that would allow it to fall as a solid?

Maybe. Recognizing the viscosity limitations of the ilmenite+pyroxene cumulate, Van Orman and Grove [5] suggested that mixing olivine-rich cumulate with the high-Ti cumulate could lower its viscosity enough to allow the formation of instabilities. If the late-stage magma ocean liquid entrained and assimilated 10–20% olivine from

deeper cumulates, then the change in bulk composition would alter the liquidus relations in such a way that ilmenite would no longer crystallize; instead, spinel would be the titaniferous phase. The solidus temperature of this material would be higher, allowing a negatively buoyant cumulate layer to form while the layer remained hot, and had lower viscosity. MELTS [26] predicts that the solidus of this new cumulate would be about 115°C higher (1240°C at 100 km depth) than the original ilmenite+clinopyroxene cumulate, leading to a reduction in viscosity by about a factor of 30. The density contrast between the new cumulate and the underlying mantle would be about half of what it was with the ilmenite+clinopyroxene cumulate, but the decrease in density is small compared to the sharp viscosity reduction. (It is interesting to note that Wagner and Grove [9] found orthopyroxene+olivine+spinel, not ilmenite, on the liquidus of the Apollo 14 black glass.) According to our numerical models, if the viscosity law for the spinel+clinopyroxene+olivine cumulate is similar to an olivine law, and the cumulate layer is at or above 1240°C, then it is tantalizingly close to conditions necessary to form a sinking instability. If the MELTS calculation is incorrect and ilmenite crystallizes rather than spinel, the resulting solid will be even denser and more likely to fall. Solomatov and Stevenson [27] have suggested that, while crystallizing, a magma ocean would simultaneously be convecting. Even sluggish, local convective currents may be enough to mix olivine into the ilmenite+pyroxene cumulate in quantities sufficient to allow the dense layer to flow and fall.

Longhi (personal communication) suggested that entrainment of plagioclase was as likely as entrainment of olivine into the late-stage liquid. Using the results of Philpotts et al. [28], an interesting scenario can be constructed from this suggestion. Philpotts et al. [28] found that when a basaltic magma was only 25% crystallized, structural chains of plagioclase and pyroxene had formed a solid three-dimensional network, through which the remaining liquid was strained. The pyroxene made the chains negatively buoyant in the liquid, and this structure actually caused the plagioclase to fall, not rise to the surface. If this happened at any point in the final crystallization of the lunar magma ocean, there would be several important results: (1) plagioclase would be trapped at depth in the moon, and the crustal highlands component may not represent the vast majority of the aluminous lunar component, as has previously been supposed; (2) the final magma ocean liquids would be further depleted in pyroxene component, and therefore enriched in ilmenite; (3) this ilmenite-enriched final liquid would be equally unlikely to fall through gravitational instability when solidified, because it will be resting on a very stiff network of plagioclase and pyroxene; and (4) the final solids will produce even denser liquids if remelted by a later event, making percolative downward flow of a high-Ti liquid even more likely.

4. Conclusions

We suggest that mass movement of a solid high-Ti cumulate from the shallow into the deep moon is unlikely, due to viscosity and temperature constraints.

A remelted ilmenite+clinopyroxene liquid (see Section 3.4) is negatively buoyant even at shallow depths and will sink. Falling high-Ti liquids from remelted cumulates would hybridize the lunar mantle as deeply as 300 km. Remelting the cumulate layer by giant meteorite bombardment may be a more viable mechanism than remelting by radioactive heating. Their rate of radioactive heat production is highest early in the moon's history, so while radioactive heat may have kept the high-Ti cumulates molten for longer, it is a poor candidate for later remelting. The giant impacts of the late heavy bombardment, however, were sufficient to remix and remelt high-Ti cumulates [29].

The most likely cumulate to sink as a solid is the result of late-stage magma ocean liquids assimilating 10–20% olivine before crystallizing (see Section 3.5). This material would form sinking diapirs, leaving a fraction of the high-Ti cumulate at shallow depths. A lunar mantle with varying Ti content at depths shallower than 560 km is consistent with the timing and composition of lunar

glasses, and is permitted by rheological constraints.

These two mechanisms, a remelted high-Ti cumulate percolating downward as a liquid, and a solid high-Ti cumulate mixed with olivine foundering as a solid, may both have occurred. Some early cumulate may have sunk as a solid, and later cumulate may have been remelted in the late heavy bombardment and sunk as a liquid.

Acknowledgements

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