

Gravitational differentiation due to initial chemical stratification: origin of lunar asymmetry by the creep of dense KREEP?

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Abstract

Models for the evolution of the Moon must explain a number of important characteristics of its magmatic history. Primary among them is the emplacement of mare basalts primarily on one side of the Moon, and in a region of the surface roughly correlating with that containing high subsurface concentrations of KREEP. It is thus important to explore mechanisms which will regionally concentrate a KREEP layer early in lunar evolution and which ~ 400 – 500 Myr later give rise to mare basalt generation beneath the same region. Gravitational instability of a chemically dense, ilmenite-rich cumulate created during the fractionation of an anorthositic crust is one candidate model. If so, gravitational instability must occur at long wavelength to explain the hemispheric asymmetry. We examine the Rayleigh–Taylor instability of a dense layer near the surface of a sphere, representing the lunar mantle, in which the viscosity varies with depth. Our results show that spherical harmonic degree 1 is the fastest growing wavelength of instability if the viscosity of the dense layer is sufficiently low relative to that of the deeper mantle or if the viscosity of the mantle increases with depth. However, the viscosity increase cannot be distributed over a depth that is too large. We explore whether a residual KREEP-like layer near the base of anorthositic crust might give rise to these needed conditions. © 2002 Elsevier Science B.V. All rights reserved.

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1. Introduction

Meaningful models for the evolution of the Moon must explain a number of important mag-

matic characteristics. Volcanic activity subsequent to the formation of its anorthositic crust was dominated by the eruption of mare basalt. (1) The main phase of mare volcanism began ~ 500 Myr after the crystallization of the anorthositic crust and continued for ~ 1 Gyr. (2) The picritic glasses, considered to be representative of primitive mare basalt liquid, were generated by melting, at 400–600 km depth [1,2], of a source containing components that, on the basis of the magma

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ocean hypothesis, should have crystallized at much shallower depth during fractionation of the anorthositic crust. (3) Mare basalts occur primarily in one region of the Moon. Recent topographic data [3] demonstrate that the earlier idea that mare basalt flooded all areas of sufficiently low elevation is not correct. Large areas of very low elevation do not contain mare basalt. The hemispheric asymmetry of mare basalt distribution on the lunar surface must be explained in some other way [4]. (4) A region of the surface roughly correlating with that containing mare basalts also is thought to contain high subsurface concentrations of KREEP which was excavated during the formation of large impact basins. This so-called Procellarum KREEP Terrane (PKT) [5] is responsible for the Imbrium basin-centered thorium anomaly mapped by Lunar Prospector [6]. KREEP, as used here, refers to the very late stage cumulates of the hypothetical magma ocean that would have crystallized near the base of the anorthositic crust. It is thus important to explore mechanisms which will regionally concentrate a KREEP layer early in lunar evolution and which ~ 400 – 500 Myr later will give rise to mare basalt generation beneath the same region.

A model in which dense ilmenite and pyroxene-rich cumulates (IC), that would have crystallized during solidification of the last 5–10% of a magma ocean, sank into the underlying mantle [7,8,9] may explain important aspects of lunar evolution. The IC would carry with it high concentrations of incompatible elements, including heat producing U, Th, and K, that would have become progressively enriched in the residual liquid as the magma ocean crystallized. A previous study [9] suggested that dense IC cumulates may have differentiated to form a layer surrounding an existing metallic core or perhaps the core itself. In fact, a recent inversion of lunar free oscillation data [10] indicates a density structure that may be more compatible with a central concentration of IC rather than a metallic core. As suggested earlier [9], the depth of the deepest lunar seismicity might mark the outer edge of an ≈ 700 km radius IC core that remains hot and ductile due to radiogenic heating.

Thermal expansion due to radiogenic heating

could make IC that had differentiated into the deep interior buoyant relative to overlying mantle, giving rise to upwelling and melting [4]. However, KREEP-rich materials excavated during the formation of the Imbrium basin must have pre-existed mare basalts that fill the basin. The emplacement of KREEP-rich material beneath the PKT prior to mare volcanism in the same region is not likely to be a natural consequence of the upwelling and melting of initially deep IC cumulates, unless some preconditioning of the mantle beneath this region determined the location of later upwelling.

In this study we explore the conditions under which the accumulation of KREEP beneath the PKT may have occurred during differentiation of the IC cumulates. This requires differentiation of IC cumulates at long wavelength (spherical harmonic degree 1). Could this explain preconditioning of the mantle that controlled the location of later upwelling and melting that generated the mare basalts? Alternatively, differentiated IC cumulates may never have reached depths greater than that of the deepest mare basalt source. Could partial melting of these cumulates due to radiogenic heating have generated the mare basalts that were localized in the region of the PKT [11,12] without decompression from greater depth? If so, the rate of differentiation and the rate of radiogenic heating must be of appropriate magnitudes to explain the timing and depth of origin of mare basalts.

While the sinking of dense material is easy to envision, factors that control the rate of differentiation and the scale of mixing of IC into the underlying mantle may be complex. First, the IC layer, if it did not differentiate, would be only 20–40 km thick. Instability of such a thin layer at a wavelength of 10^5 km (the circumference of the Moon) would require that it have a very low viscosity. A still liquid IC layer would have a very low viscosity relative to underlying solidified olivine–pyroxene mantle. Such a model would be similar to that originally proposed by Elsasser for metallic core formation in the Earth [13,14]. But, in contrast to Fe-rich metallic liquid, the density of a liquid IC layer at pressures near the base of the anorthositic crust, based on estimates

like those of Delano [15], is not expected to exceed that of the underlying olivine–pyroxene mantle. Due to lateral pressure gradients at the base of an isostatically compensated crust, a liquid IC layer may flow to areas of thinner crust, but it should not sink into denser underlying mantle.

Since the IC layer is thin, initially formed, small IC plumes or diapirs may sink creating an IC-enriched mantle layer that is denser than underlying mantle. A mixture of IC with previously fractionated olivine–pyroxene cumulate mantle would provide a suitable source for mare basalt. This IC-enriched mantle layer, being thicker than one consisting of pure IC, could sink more rapidly than IC in the small plumes and at a longer wavelength than the initial instability [9,16,17]. We later present a simple finite amplitude Rayleigh–Taylor (R–T) instability model, which illustrates this behavior.

But an initial, short wavelength instability would not wait to develop until the IC had fully solidified. The IC would tend to sink as it solidified; perhaps solidifying IC would feed already established IC plumes. In this case, the initial wavelength of instability should be determined by the solidification rate of the liquid layer [9]. A simple analysis indicates that the depth to which plumes would sink in the time over which the IC crystallizes is about five times the thickness of the IC, giving an IC-enriched layer about 100 km thick. It is interesting that this simple prediction indicates an IC-enriched layer thickness independent of the solidification rate: the horizontal scale of plumes does depend on the solidification rate, but the product of sinking rates of plumes and the time for an IC layer of given thickness to solidify does not [9].

Since the mineralogy of the IC is dominated by clinopyroxene, its solid-state creep viscosity even at high temperatures may be much higher than that of olivine [18]. At temperatures near the melting temperature of IC cumulates ($\sim 1100^\circ\text{C}$) and at low stresses (~ 5 MPa), laboratory flow laws indicate that the creep rate of dry clinopyroxene is more than a factor of 100 lower than that of dry olivine [19]. Such a high viscosity would result in a low R–T growth rate and would not favor instability at long wavelength. However, if instabil-

ity involved still partially molten IC, even 5–10% melt along grain boundaries would lower the viscosity by a factor of 10–100 [20,21]. We leave for future study questions arising in this complex differentiation of compacting, partially molten solid. Here we will assume that small scale sinking of IC into an olivine-rich mantle generates an IC-enriched layer 50–100 km thick with a rheology approximating that of olivine.

In this study we consider only the simpler R–T instability of a dense fluid layer initially surrounding a viscosity stratified sphere. Our objective is to examine the conditions required for long wavelength gravitational instability possibly leading to hemispheric asymmetry in the distribution of IC. In general this requires a dense layer with a low viscosity relative to underlying mantle. Viscosity stratification in the lunar mantle beneath the IC layer could arise due to the temperature dependence of viscosity and heating by radioactive elements concentrated in the IC layer.

2. R–T stability analysis

We consider the growth rate of R–T instability of a dense layer of thickness (H) at the surface of a viscosity stratified sphere of radius (R). Simple dimensional analysis shows that the R–T growth rate is proportional to the density difference between fluid in the layer and that beneath it and to the gravitational acceleration. Growth rate is inversely proportional to the magnitude of the viscosity, but also depends on the viscosity stratification beneath the layer. The relative growth rate of a range of wavelengths (or spherical harmonic degrees) is calculated for a range of viscosity stratifications in which viscosity increases exponentially with depth over a depth (D). The depth (D) and the magnitude of the viscosity increase are treated as parameters. Growth rate is calculated using the formulation for gravity-driven deformation in a viscous sphere [22]. We first consider the simple case of a layer of low viscosity surrounding a sphere of higher viscosity, that is $D=0$, for a range of layer thickness and viscosity increase. The R–T growth rate, normalized by its maximum value in each case, is shown as a func-

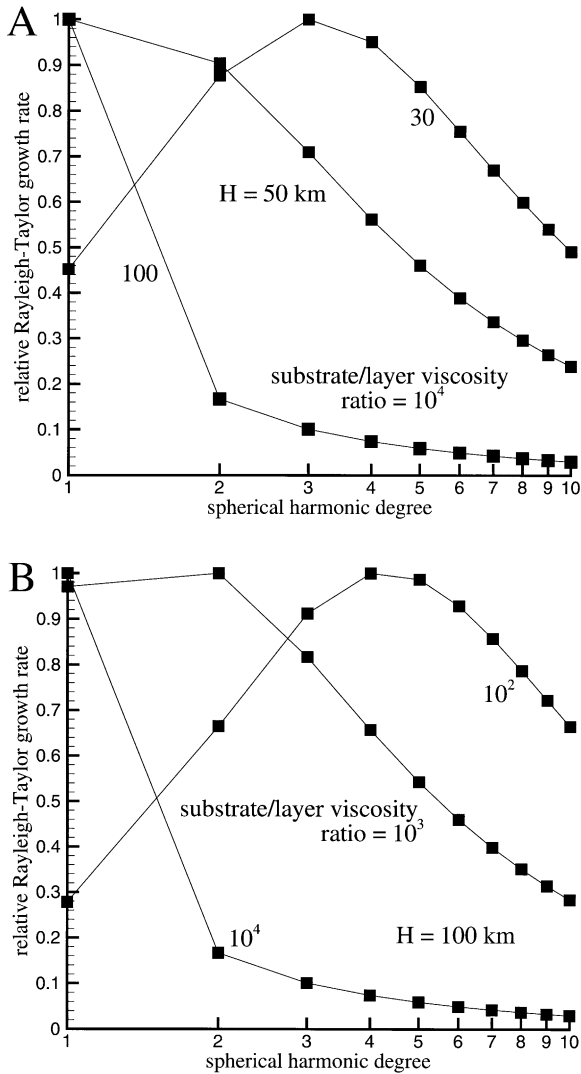


Fig. 1. R–T instability growth rates for a dense layer of thickness (H) at the surface of a viscous sphere with a $R=1700$ km. Growth rate normalized by the maximum growth rate is plotted as a function of spherical harmonic degree (A) for a range of layer thicknesses with a viscosity ratio of 10^4 and (B) for a range of viscosity ratios with layer $H=100$ km.

tion of spherical harmonic degree in Fig. 1. Fig. 1A considers a range of layer thicknesses for a viscosity that increases by a factor of 10^4 . Thicker layers result in a longer dominant wavelength. For this viscosity increase, the degree 1 R–T mode grows the fastest for a layer thickness greater than 50 km. For a layer thickness of 30 km or

less, higher spherical harmonic degrees grow faster than degree 1. Fig. 1B considers a viscosity increase ranging from a factor of 10^2 – 10^4 for a 100 km thick layer. With a viscosity increase greater than about a factor of 10^3 , the degree 1 mode grows faster than modes at shorter wavelength.

A thicker layer and a larger viscosity increase favor a longer wavelength of instability. Earlier studies (cf. [23]) show that the dominant wavelength of a plane layer of thickness H and viscosity (μ_1) overlying a halfspace of viscosity (μ_2) is proportional to $(\mu_2/\mu_1)^{1/3}H$. These results for a Cartesian geometry are expected to be a useful guide in the spherical geometry when the layer is thin compared to its radius. Thus a large substrate/layer viscosity ratio μ_2/μ_1 is needed for instability at a wavelength large compared to H . Alternatively, for a given viscosity ratio, the layer must not be too thin: if a layer were thinner by factor of 2, the viscosity increase would need to be factor of 8 larger for the dominant wavelength to remain the same. This is illustrated in Table 1, which shows the spherical harmonic degree of the maximum R–T growth rate for a range of layer thickness and viscosity ratio.

Fig. 2 considers a viscosity increase distributed over a depth $D > 0$ beneath a 100 km thick layer ($H=100$ km). Fig. 2A considers a viscosity increase ranging from a factor of 10^2 – 10^4 over a depth $D=100$ km. The fastest growing wavelength is short (spherical harmonic degree > 50),

Table 1

Wavelength (spherical harmonic degree on a 1700 km radius sphere) with maximum R–T growth rate as a function layer thickness and layer/substrate viscosity ratio

Viscosity increase	Layer thickness (km)	Spherical harmonic degree
1	30	45
1	50	26
1	100	13
10^2	30	16
10^2	50	9
10^2	100	4
10^4	30	3
10^4	50	1
10^4	100	1

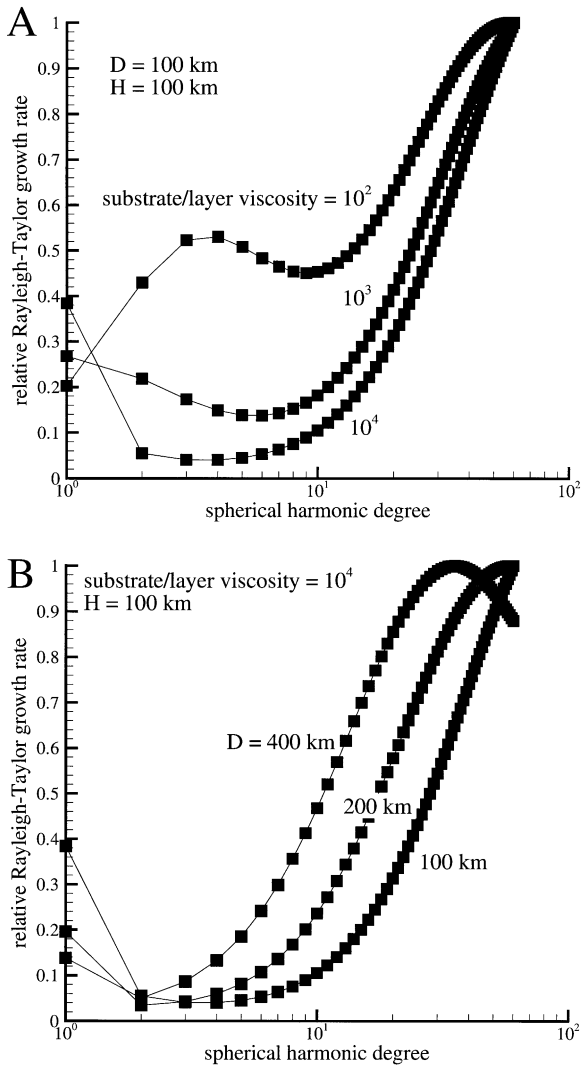


Fig. 2. R–T instability growth rates for a dense layer of $H = 100$ km at the surface of a viscosity stratified sphere with a $R = 1700$ km. Growth rate normalized by the maximum growth rate is plotted as a function of spherical harmonic degree for different viscosity stratifications (A) with $D = 100$ km, a range of viscosity increase and (B) with a viscosity increase by a factor of 10^4 occurs over a range of depth (D).

indicating that short wavelength structure develops quickly, but the resulting small plumes or diapirs are not able to penetrate to a large depth before a growing long wavelength structure becomes predominant. This is confirmed by finite amplitude numerical calculations described below.

For a viscosity increase greater than about a factor of 10^3 , degree 1 grows faster than a range of immediately longer wavelengths. Similar growth rate spectra are also obtained if the viscosity increase occurs over larger depths. Fig. 2B shows results for D ranging from 100 to 400 km with viscosity increasing by a factor of 10^4 . The growth rate of degree 1 relative to immediately longer wavelengths decreases with increasing D but remains larger than immediately longer wavelengths for this whole range of D values. For $\Delta\rho = 60$ kg/m³, $g = 1.63$ m/s², and a 100 km thick layer with viscosity 10^{20} Pa s, absolute degree 1 growth times (reciprocal of the growth rate) predicted by this analysis are given in Table 2.

We have also considered the finite amplitude evolution of R–T instability using finite element analysis like that reported previously [4]. Fig. 3 shows two times in the evolution of an example with a dense layer 100 km thick beneath which the viscosity increases by a factor of 5×10^3 over a depth of 400 km. Consistent with the stability analysis in Fig. 2, instability begins at short wavelength but at a longer wavelength spherical harmonic degree 1 flow then develops. The initial short wavelength instability creates an effectively thicker dense layer that is then unstable at long wavelength.

3. Discussion

The wavelength and rate of R–T instability are controlled by the rheology of IC and underlying mantle deforming by solid-state thermally activated creep. Creep rate, or effective viscosity, depends on temperature, composition, and textural

Table 2

Degree 1 R–T instability growth time for a 100 km thick dense layer with $\Delta\rho = 60$ kg/m³ and a viscosity 10^{20} Pa s

D (km)	Viscosity increase	Growth time (Myr)
100	10^2	76.6
100	10^3	95.9
100	10^4	105.0
200	10^4	86.7
400	10^4	64.3

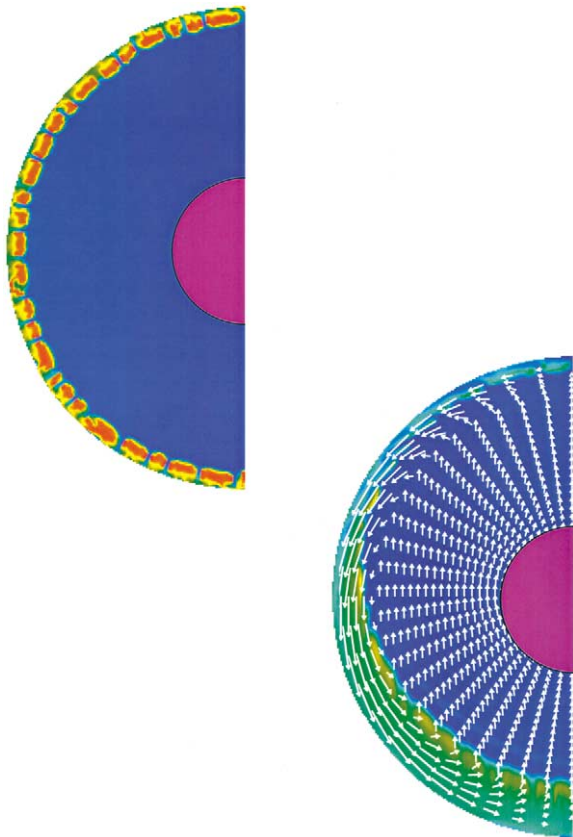


Fig. 3. Finite amplitude R–T instability of an initially 100 km thick layer overlying an interior region in which the viscosity increases by a factor of 5000 over a depth interval of 400 km. Early time is on the left; later time is on the right. Velocity vectors show the spherical harmonic degree 1 flow pattern present at the later time. Central region corresponds to a dense liquid metallic core. Consistent with expectations from linearized stability analysis, instability begins at a wavelength comparable to the layer thickness; but at a much longer wavelength spherical harmonic degree 1 flow subsequently grows more rapidly because of its longer wavelength. The early short wavelength instability creates an effectively thicker dense layer that is then unstable at long wavelength.

variables such as grain size. With a creep activation energy typical of that in olivine (~ 500 kJ/mol; cf. [18]), the temperature dependence of viscosity is particularly strong: a temperature increase of only 100 K at a temperature of ≈ 1500 K results in a factor of 10 lower viscosity. The activation energy for creep in dry clinopyroxene is reported to be even higher, more than 700 kJ/mol [18].

IC is expected to be rich in clinopyroxene while the underlying mantle may contain significant amounts of olivine. Since clinopyroxene deforms more slowly than olivine at the same temperature, pressure and deviatoric stress [18], an IC-enriched layer may be more viscous than an underlying, more olivine-rich mantle at the same temperature [24]. The mantle beneath an IC-enriched layer may be compositionally stratified with lower density, high Mg/Fe mantle at the top [9]. Higher Mg/Fe compositions will have a higher melting temperature and, at the same temperature, a lower homologous temperature. Because creep rate generally decreases with decreasing homologous temperature, a higher Mg/Fe mantle should have a higher viscosity. Both of these compositional effects will act to increase the viscosity of an IC-enriched layer and decrease any possible increase in mantle viscosity with depth beneath this layer. However, both effects are likely to be dominated by the very strong temperature dependence of thermally activated creep.

A strongly temperature dependent viscosity may play several fundamental roles. Conductive cooling to the surface would increase the viscosity of the IC-enriched layer, thus inhibiting longer wavelengths of instability and increasing the time scale on which instability would occur. However, sufficiently concentrated radioactive heating in the IC-enriched layer would prevent significant cooling on the time scales over which R–T instability may develop. Simple models predict that radioactive heating may more than offset conductive heat loss [12].

The IC cumulates are thought to have formed by crystallization of the last 10% of a magma ocean that generated the anorthositic crust [25]. For initially chondritic U and Th concentrations and approximating these heat producing elements as perfectly incompatible, the rate of radiogenic heating in the IC would be $10\times$ chondritic. In the early Solar System, chondritic abundances of U and Th give a heating rate of ≈ 50 K in 100 Myr, so that, in the absence of conductive heat loss, $10\times$ chondritic heating would result in a temperature increase of 500 K over 100 Myr. Heat conduction to the surface must certainly reduce this temperature increase, but even in the presence of

conductive cooling, the IC layer beneath a 60 km thick crust with a porous, impact fractured regolith of low thermal conductivity can remain hot for a long time [12].

Heat generated in the IC layer will also be conducted downward into the underlying mantle. In time (t), this heating would extend to a depth $2\sqrt{\kappa t}$ where thermal diffusivity $\kappa \approx 30 \text{ km}^2/\text{Myr}$. If R–T instability developed on a several hundred Myr time scale, temperature gradients would extend downward over a depth of several hundred kilometers. Based on our results above, a several hundred K temperature decrease over this depth interval should be sufficient to make degree 1 the fastest growing wavelength.

Fractional crystallization of a lunar magma ocean initially 800 km thick would generate a 20 km thick layer of ilmenite-bearing cumulates with a density about 300 kg/m^3 greater than the underlying cumulates [9]. If an initial short wavelength instability created an IC-enriched layer 100 km thick, mass conservation (constant product of the density difference and layer thickness) would result in an effective $\Delta\rho = 60 \text{ kg/m}^3$. The R–T growth times listed in Table 2 then provide an estimate of the time for this layer with a viscosity of 10^{20} Pa s to sink into the underlying mantle at a wavelength corresponding to spherical harmonic degree 1. This time is directly proportional to the layer viscosity and depends on the viscosity structure as shown. The high- and low-Ti mare basalts filling the major near-side basins were emplaced within 500 Myr of the crystallization of the anorthosite crust. An IC-enriched layer viscosity as large as 10^{21} Pa s would be too large to allow IC to sink to depths of 400–600 km in this time. A viscosity of 10^{20} Pa s would allow the layer to thicken beneath the PKT and sink into the mantle on the appropriate 100 Myr time scales.

As it sinks, both the temperature and melting temperature of IC-enriched mantle will increase due to radioactive heating and the increase of melting temperature with pressure, respectively. With a melting temperature increase of $\approx 0.5 \text{ K/km}$, the melting temperature will rise about 200 K at 400 km depth. If IC-enriched mantle starts out near its melting temperature, a temperature rise at least this large is required to cause melting at a

depth of 400 km. Except for a small adiabatic contribution, this temperature increase must be derived from radioactive heating. The heating in a 20 km thick IC layer containing $10\times$ chondritic heating (the last 10% of a chondritic magma ocean) will be reduced by mixing of IC with underlying mantle. U and Th abundances in an IC-enriched layer 100 km thick of $2\times$ chondritic would give a heating rate $\approx 100 \text{ K}$ in 100 Myr. Thus IC-enriched mantle initially near its melting temperature would begin to melt at $>400 \text{ km}$ depth in times of a few hundred Myr. Large lateral scale of the degree 1 downwelling will minimize the effect of conductive cooling.

Assessing the possibility of long wavelength gravitational instability of a dense ilmenite-rich layer as a mechanism for the generation of mare basalts and the asymmetry of lunar structure depends strongly on the scale and rate of instability. These in turn depend on the depth to which dense IC mixes into the mantle as this layer progressively solidifies. If an IC-enriched mantle layer that develops by mixing is not too thin and if the viscosity of this layer is sufficiently low, this may be a viable mechanism for explaining the degree 1 structure of the Moon. Heating due to U and Th concentrated in the IC combined with the strong temperature dependence of thermally activated creep may create a viscosity structure favorable for this mechanism.

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