

SPATIAL AND TEMPORAL SCALES OF LUNAR DIFFERENTIATION: IMPLICATIONS FOR SUBSEQUENT THERMAL AND CHEMICAL EVOLUTION OF THE MOON. E.M. Parmentier and P.C. Hess, Department of Geological Sciences, Brown University, Providence, RI, 02912

The onset of mare volcanism at about 3.9 Gyr (1) from source depths greater than 400-500 km (2,3) provide important constraints on mechanisms of lunar internal evolution. Earlier models of lunar internal evolution do not account for these important characteristics of lunar magmatism. We continue to explore a model for the evolution of the lunar interior that explains the origin of mare basalts (4,5,6) and possibly the existence of a lunar core. We assume that a magma ocean differentiated into the anorthositic crust and chemically stratified cumulate mantle. The cumulate mantle is gravitationally unstable with dense ilmenite cumulate layers overlying olivine-orthopyroxene cumulates with Fe/Mg that decreases with depth. Primitive lunar mantle rises to its level of neutral buoyancy in the cumulates and the dense ilmenite rich layer sinks to the center, forming a core (4,7). After this overturn, radioactivity within the incompatible element-rich ilmenite-cumulate core heats the overlying mantle. Buoyancy due to thermal expansion overcomes the stable chemical stratification in the mantle creating a convectively mixed layer above the core-mantle boundary that is the source region for mare basalts. Predictions of this model for a range of mantle viscosity (10^{19} to 10^{20} Pa-s), core radioactivity values (5 to 10 times bulk earth), and several values of the chemical density gradient above the core-mantle boundary are summarized in Table 1. If the content of radioactive elements in the core and the chemical density gradients above the core are sufficiently high, but within a range of values that might have been present in the moon, deep, high pressure melting at the top of the mixed layer is predicted after a reasonable time interval to explain the 3.9 Gyr onset of mare basalt volcanism.

Temporal and spatial scales for overturn of initial chemical stratification

The time scale for the sinking of the ilmenite-cumulate layer to the center of the Moon can be controlled by the time for diapirs to form by Rayleigh-Taylor instability or by the time for the resulting diapirs to sink through the lunar mantle. This will be determined by the wavelength of instability and the corresponding size of the diapirs. An estimate of the time needed for Rayleigh-Taylor instability of the ilmenite cumulate layer to develop is (e.g. 8)

$$t = \frac{6.5 \mu_2^{2/3} \mu_1^{1/3}}{\Delta \rho g h}$$

with a wavelength of instability

$$\lambda = 2.9 h \left(\frac{\mu_2}{\mu_1} \right)^{1/3}$$

where $\mu_1 \ll \mu_2$. In these expressions μ_1 and μ_2 are the viscosity of the ilmenite-cumulate and underlying olivine-pyroxene cumulates, respectively ($\mu_2 \approx 10^{19}$ Pa-s for mantle silicates near their melting temperature should be a reasonable minimum value), $\Delta \rho$ (0.35×10^3 kg/m³) is the density difference between them, and h is the layer thickness (~ 20 km). The wavelength of instability depends on the viscosity of the ilmenite cumulate layer. If this viscosity were an order of magnitude less than that of underlying cumulates, $\mu_1/\mu_2 \approx 10^{-1}$, the wavelength of instability would be about $6h$ or 120 km. Equating the layer volume in one wavelength of instability, $\lambda^2 h$, to that of the resulting diapir gives a diapir diameter of about 80 km. About 2000 such diapirs would form. The number of diapirs varies inversely with λ^2 and is therefore proportional to $(\mu_1/\mu_2)^{2/3}$. For $\mu_1/\mu_2 \approx 10^{-6}$, only a single diapir would form. This much larger scale of instability would result in less mixing during the differentiation. It is also interesting to

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hypothesize that the lunar center of figure-center of mass offset and corresponding crustal asymmetry (9) might be indicative of this largest scale of instability.

The time scale for the formation of diapirs is 10^{-1} Myr and 2×10^{-3} Myr for $\mu_1/\mu_2 \approx 10^{-1}$ and 10^{-6} , respectively. The time for a diapir once formed to sink to the center of the Moon is estimated using the Stokes settling velocity of a fluid sphere with viscosity μ_1 settling through a fluid with viscosity μ_2

$$V = \frac{1}{3} \frac{\Delta \rho g a^2}{\mu_2} \left(\frac{\mu_2 + \mu_1}{\mu_2 + \frac{3}{2} \mu_1} \right)$$

where a is the diapir radius, $4\pi a^3/3 = \lambda^2 h$. The sinking time over a distance comparable to the radius of the moon (~ 1700 km) for the cases with $\mu_1/\mu_2 \approx 10^{-1}$ and 10^{-6} cited above with $\mu_2 = 10^{19}$ Pa-s are 5×10^{-1} Myr and 5×10^{-3} Myr, respectively. The time for diapirs to form and to sink to the center of the Moon are comparable over a wide range of μ_1/μ_2 . But both the differentiation time and particularly the spatial scale at which differentiation occurs depends strongly on this ratio. For small spatial scales it is important to assess the degree of mixing between the ilmenite-cumulate and the mantle through which it sinks.

Effect of core and mantle melting on convective heat transport

In the models described above, melting first occurs in the ilmenite-cumulate core and then in the mantle just above the core-mantle boundary. Melting just above the core-mantle boundary could increase the rate at which a thermal boundary layer would form by heat conduction alone thus increasing the rate of convective heat transfer into the overlying mantle. The melted mantle would mix with the core making it larger, less rich in radioactive elements, and less dense. If it becomes less dense than the more iron-rich pyroxene-rich cumulates forming the lowermost mantle, core material might rise to form a layer displaced from the center of the moon, at its height of neutral buoyancy in the overlying cumulate mantle. Models that address this possibility are presently being formulated. Core material might also mix with the overlying mantle as it rises to its new equilibrium level.

Table 1. Selected model parameters and results

heating rate	viscosity (Pa-sec)	compositional density gradient (km^{-1})	time to melting (Myr)	melting depth (km)	melting temperature ($^{\circ}\text{C}$)	melting pressure (GPa)
10.	10^{19}	0.3×10^{-7}	996.	336.	1365.	1.84
10.	10^{19}	0.6×10^{-7}	636.	760.	1458.	3.08
10.	10^{19}	0.9×10^{-7}	507.	807.	1491.	3.54
5.	10^{19}	0.3×10^{-7}	1965.	522.	1421.	2.59
10.	10^{20}	0.9×10^{-7}	735.	664.	1459.	3.09

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