OVERTURN OF MAGMA OCEAN ILMENITE CUMULATE LAYER: IMPLICATIONS FOR LUNAR MAGMATIC EVOLUTION AND FORMATION OF A LUNAR CORE.

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We explore a model for the chemical evolution of the lunar interior that explains the origin and evolution of lunar magmatism and possibly the existence of a lunar core. A magma ocean formed during accretion differentiates 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. The dense ilmenite layer sinks to the center of the moon forming the core. The remainder of the gravitationally unstable cumulate pile also overturns. Any remaining primitive lunar mantle rises to its level of neutral buoyancy in the cumulate pile. Perhaps melting of primitive lunar mantle due to this decompression results in early lunar Mg-rich magmatism. Because of its high concentration of incompatible heat producing elements, the ilmenite core heats the overlying orthopyroxene-bearing cumulates. As a conductively thickening thermal boundary layer becomes unstable, the resulting mantle plumes rise, decompress and partially melt to generate the mare basalts. This model explains both the timing and chemical characteristics of lunar magmatism.

Solidification of an 800 km deep magma ocean creates 60 km of anorthosite crust, 40 km of an ilmenite enriched residuum beneath the crust, a 700 km cumulate pile that has relatively dense orthopyroxene-olivine (low Mg*) layers near the top and less dense dunite (high Mg*) layers at the base, overlying about 940 km of primitive moon. The chemical stratification is gravitationally unstable; an estimate of the time needed to develop a Rayleigh-Taylor instablity of the ilmenite layer is given by (1)

$$t = \frac{4\,\mu_2^{2/3}\mu_1^{1/3}}{\Delta\rho g h}$$

where μ_1 and μ_2 are the viscosity of the ilmenite and underlying cumulate, respectively (10^{18}-10^{19} Pa-s for mantle silicates near their melting temperature would be a minimum value), $\Delta\rho$ (0.35×10^3 kg/m^3) is the density difference between them, and h is the layer thickness. Given the uncertainty in a number of the variables, particularly the viscosities, this time ranges from 10-100 Myr .

Part of the ilmenite-rich cumulate layer may mix with the mantle (2), but a large fraction of it sinks through the magma ocean cumulates and primitive moon. A 40 km ilmenite layer formed near the top of a cumulate pile 740 km thick would form a dense 600 km radius lunar core. Provided that the cumulates contain sufficient heat producing elements, the ilmenite core could supply the heat necessary to mobilize and melt the overlying mantle. This mantle, the sources for the picrite mare glasses, is formed from orthophyroxene-olivine cumulate layers (Mg* 75-80) that have small amounts of added ilmenite crystallization products.

Heating of the overlying olivine-orthopyroxene cumulates by the ilmenite core will generate mantle plumes that rise through the cumulates and undergo decompression melting. The ilmenite core will be highly radioactive compared to the overlying cumulates. An estimate of the radiogenic heat production H in the ilmenite core is about 10 times that of the primitive moon. About 4 Byr ago we take heat production in the primitive moon to have been about 4 times the presentday heat production rate in the Earth's mantle (~6x10^{-12} W/kg-s). The conductive thermal boundary layer will thicken with time t as $2\sqrt{\kappa t}$, and the temperature difference across it will increase as Ht/c_p where c_p is the specific heat at constant pressure. When the Rayleigh number based on the boundary layer

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thickness exceeds a critical value Ra_c (~1000), instability will lead to the formation of mantle plumes. The time for plumes to form, corresponding to the time between overturn and the beginning of mare basalt formation, is given by

$$t = \left(\frac{Ra_c \, \mu_1^{1/3} \mu_2^{2/3} c_p}{8\rho_0 \alpha Hg\kappa^{1/2}}\right)^{2/5}$$

where ρ_0 is a reference density (3.3×10^3 kg/m^3), α is the coefficient of thermal expansion (3×10^{-5} °C^{-1}), and κ is the thermal diffusivity (10^{-6} m^2/s). The average temperature difference between the plumes and surrounding mantle, roughly one-half the temperature difference across the thermal boundary layer, is simply $Ht/2c_p$. Here μ_1 is the viscosity in the thermal boundary layer (which we take to be mantle silicate near its melting temperature ~10^{18} Pa-s) and μ_2 is the viscosity in the overlying mantle. After overturn, the base of the olivine-orthopyroxene cumulates will be near their low pressure melting temperature (~1100-1200°C). For a melting temperature which increases 10°K/kbar, the melting temperature of the cumulates at a depth of 1000-1100 km, corresponding to the core-mantle boundary in our model, will be 1400-1500°C. The cumulates must therefore be at least 300°C cooler than their melting temperature. If the viscosity increases by a factor of 6 for each 100°C below the melting temperature, $\mu_2=2\times10^{20}$ Pa-s. With these values and g about 1/3 of the surface value at this 1000 km depth, the time estimated for plume formation is about 100 Myr. After this time the boundary layer thickness is about 100 km and the temperature difference across the thermal boundary layer is 600°C. This would predict temperatures exceeding 1700-1800°C in the ilmenite core, suggesting that it would be largely molten at this time. The viscosites used in this estimate are minimum values. Increasing the viscosities by an order of magnitude would increase the time between overturn and the beginning of mare basalt magmatism to 250 Myr. Decreasing the rate of heat production in the ilmenite core by a factor of 2 would increase this time by about 30%.

High TiO$_2$ picrite mare glasses are near primary melts segregated from mean depths of 400-500 km and generated from an olivine-orthopyroxene source (2,3). If the ages of the picrite glasses overlap those of mare basalts, then picrite volcanism began at least 3.9 Byr ago (4). Thus a time of at least several hundred Myr between overturn and the beginning of mare volcanism is required. The potential temperature within mantle plumes must be in the range of 1450-1550°C, assuming about 10% melting (2). With an initial temperature of the olivine-orthopyroxene cumulates to be 1100-1200°C and an average temperature increase in the thermal boundary layer of 300°C at the time that plumes begin to form, our model predicts temperatures in this range. Thus, both the age of volcanism and the temperatures required for melting are consistent with the model described above. If this model is correct, the heat source for high TiO$_2$ mare basalt volcanism is ilmenite cumulate. But these cumulates exist deep within the moon and may form part or all of a lunar core.

References: (1) Whitehead, J.A., Ann. Rev. Fluid Mech. 20, 61, 1988. (2) Hess, P.C., GRL 18, 2069, 1991. (3) Delano, J.W., Proc. 16th Lunar Planet. Sci. Conf., JGR 91, D201, 1986. (4) Nyquist, L.E., and C.-Y. Shih, Geochm. Cosmochm. Acta 56, 2213, 1992.