MAGMA OCEAN CUMULATE OVERTURN: GENERATION OF AN EARLY MAGNETIC FIELD. L.T. Elkins-Tanton, S. E. Zaranek, and E.M. Parmentier, Department of Geological Sciences, Brown University, Providence, RI 02912 (Linda_Elkins_Tanton@brown.edu, Sarah_Zaranek@brown.edu, EM_Parmentier@brown.edu).

Introduction: Dynamical models of Martian differentiation and early evolution need to be consistent with several major attributes of Mars believed to have developed before 4.0 Ga: differentiation of mantle source regions into isotopically distinct reservoirs; development of an early, brief, strong magnetic field; and the formation of an early crust to record that field. Significant and perhaps complete melting of the large terrestrial planets is expected due to the conversion of kinetic energy to heat during accretion of planetesimals, and to the potential energy release of core formation [e.g., 1-5]. Previous results of Martian magma ocean investigations indicate that magma ocean crystallization and subsequent overturn on Mars could be fast and complete [6], and is consistent with magma source region differentiation and the development of an alumina-poor Martian mantle [7]. The further results presented here demonstrate that magma ocean crystallization and overturn can produce a magnetic field of between 10 and 50 million years duration.

Magma ocean model: The fundamental, relevant processes of crystallizing the magma ocean are well summarized in Solomatov [8]. Vigorous convection in the low viscosity liquid magma ocean is assumed to result in an adiabatic variation of temperature with depth. Because an adiabat is steeper than the solidus a solid region develops at the bottom of the mantle as the ocean cools.

The simplified Martian mantle mineralogy used follows from the bulk mantle composition of Bertka and Fei [9] and the phase relations of Longhi *et al* [10]. All phases are fractionally crystallized in increments of one-half percent from the evolving liquids of the magma ocean. One percent of interstitial liquid is retained in the solids throughout the magma ocean, acting as a reservoir for radiogenic incompatible elements, which are partitioned from the solid phases using coefficients from [11-13].

As crystallization progresses the solids remain at their solidus temperatures, but the temperature of the solidus decreases as liquid composition evolves. The final cumulate stratigraphy, lying at its solidus, ranges in temperature from 2,100°C at the core-mantle boundary to about 700°C near the surface.

Overturn and subsequent mantle evolution: The resulting cumulate stratigraphy is unstable to gravitational overturn mainly due to the effects of iron enrichment as fractional solidification proceeds. Because the time scale for Rayleigh-Taylor overturn is inversely dependent upon the thickness of the layer, overturn is not likely to initiate until the magma ocean is largely crystalline, whereupon the cumulates flow as solids into an equilibrium density profile [7].

All the models in this suite produced a final cumulate overturn stratigraphy where the coolest, shallowest cumulates fell to the core-mantle boundary. (Results from [7] showed the garnet layer falling to the core-mantle boundary, but the more evolved liquid compositions allowed in these more sophisticated models prevent complete fall of the garnet layer in every case.)

The density profile after overturn is highly stable, as shown in Fig. 2 in the companion abstract in this volume and in Elkins-Tanton *et al.* [7]. This stratification will strongly inhibit thermal convection. In the solid state, a temperature at the bottom of the mantle significantly higher than the melting temperature would be required to cause solid-state thermal convection. Many earlier studies attribute Tharsis volcanism to decompression melting in a solid-state mantle plume, but this hypothesis will need to be revised if strongly stable compositional stratifications are present after overturn.

Core heat flux model: Because cumulate overturn is expected to occur on a time scale much shorter than that of heat conduction over significant distances, the core-mantle boundary will effectively retain its pre-overturn temperature during overturn. The density profile of the solidified magma ocean predicts the stratigraphy of the overturned cumulates, from which the temperature profile directly follows since the solids move adiabatically during the rapid reshuffling. Moving cold cumulates from near the surface to the core-mantle boundary during overturn can produce a brief and intense heat flow out of the hot core and into the cold cumulates.

A finite difference computer program in spherical coordinates has been written to predict heat flux from the core by solving the conductive heat equation while incorporating conductive cooling at the surface, cumulate heating from radiogenic elements (whose distribution is produced by the initial fractional crystallization of the magma ocean), and the initial temperature distribution in the overturned cumulates. No convection is allowed in this model; the stable density profile created by cumulate overturn should prevent initiation of convection on these time scales.

Stevenson [14] estimates that a superadiabatic core cooling rate on the order of 80K per Ga (~0.022

J/m^{2}sec) is required to initiate a Martian core dynamo. Our calculations indicate that the superadiabatic heat flux required to initiate a dynamo may be as high as 0.10 J/m^{2}sec, but is likely between 0.05 and 0.02 J/m^{2}sec, using equation 3 from [14].

Results for core heat flow following cumulate overturn: Assuming that the Martian silicate mantle begins with a chondritic trace element composition, the final one percent of liquid contains a trace element concentration of about 65 times chondritic, similar to bulk Earth crust. An evolved liquid fraction is spread over a radius of about 250 km by a critical crystal network and possible stagnant crust at the surface of the planet. This ~700°C radiogenically-enriched material falls during overturn to the core-mantle boundary, which remains at the silicate solidus temperature of ~2,100°C. The radiogenic layer is sufficiently thick that even a stagnant lid on the planet will not prevent some quantity of the material from foundering. Core heat flux is therefore driven by the initially steep temperature gradient between the core and the cold fallen cumulates, but increasingly limited both by core cooling and by radiogenic heating in the lowermost cumulates. The preferred initial temperature difference across the core-mantle boundary is therefore 2,100-700°C = 1,400°C.

Models vary in the distribution of radiogenic elements: some are distributed exponentially with depth and others have homogeneous layers at the core-mantle boundary (fig. 1). Fig. 2 shows the mantle temperature evolution for each model. Heat flux from the core begins for almost all models at ~0.6 J/m^{2}sec, dropping off exponentially and falling below 0.05 J/m^{2}sec between 40 and 100 million years later. The Martian core therefore has sufficient superadiabatic heat flux in all of these models to initiate a core dynamo for 40 to 100 million years after magma ocean crystallization.

If the temperature difference across the core-mantle boundary due to cumulate overturn is only 1,000°C, heat flux remains above the necessary 0.05 J/m^{2}sec for ~30 Ma. To prevent heat flux from ever exceeding the 0.05 J/m^{2}sec value the temperature difference across the core-mantle boundary must be 150°C or less.

Discussion and Conclusions: Convective instability of the lower mantle could result in a second episode of magnetic field generation, though this has not been observed. Those who favor an origin of long-lived Tharsis volcanism by decompression melting in a deep mantle plume likewise must question whether this solid-state convection could result in sufficiently rapid heat transfer to generate a magnetic field.

All the models run in this suite produced superadiabatic core heat flux thought to be sufficient to produce a core dynamo magnetic field for 40 to 100 million years after magma ocean cumulate overturn. The deepest cumulate are heated radiogenically to the point that they may melt.

Overturning moves cold cumulates to the core-mantle boundary, producing core heat flux, and also moves hot cumulates nearer the surface where they can melt adiabatically and produce an early crust. These models predict the formation of tens of kilometers of new crust during the time the magnetic field is active. This new crust is cooling to and then below the Curie temperature of approximately 600°C, available to record the magnetic field.

References: [1] Safronov (1978) *Icarus 33*, 3. [2] Kaula (1979) *JGR 84*, 999. [3] Stevenson (1987) *Ann. Rev. EPS 15*, 271. [4] Halliday (2001) *Space Sci. Rev 96*, 197. [5] Hess (2001) *32nd LPSC*, 1319. [6] Zaranek (2004) *35th LPSC*. [7] Elkins-Tanton (2003) *MAPS 38*, 1753. [8] Solomatov (2000) in *Origin of the Earth and Moon*. [9] Bertka (1997) *JGR 102*, 5251. [10] Longhi (1992) In *Mars*. [11] Draper (2003) *PEPI 139*, 149. [12] Green (2000) *Lithos 53*, 165. [13] Skulski (1994) *Chem. Geo. 117*, 127. [14] Stevenson (2003) *EPSL 208*, 1.

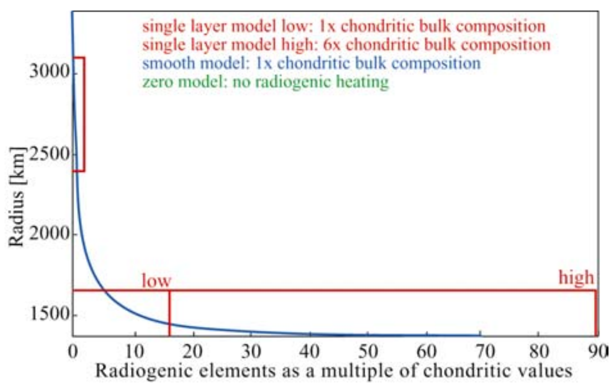

Figure 1. Mantle profile of radiogenic elements immediately after cumulate overturn.

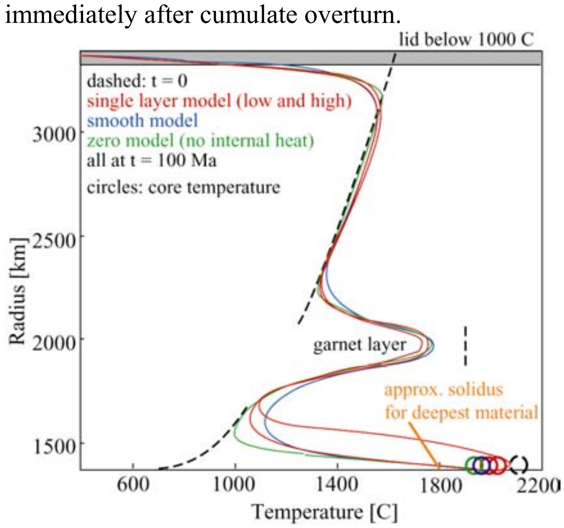

Figure 2. Temperature profiles in the Martian mantle immediately following overturn, and at 100 Ma.