THE EFFECTS OF MAGMA OCEAN DEPTH AND INITIAL COMPOSITION ON PLANETARY DIFFERENTIATION. L. T. Elkins-Tanton, E. M. Parmentier, and P.C. Hess, Brown University, Department of Geological Sciences, 324 Brook St., Providence RI 02912, Lindy@alum.mit.edu, EM_Parmentier@brown.edu.

Introduction: One or more magma oceans are hypothesized to have formed on the terrestrial planets during their early evolution [e.g. 1-4]. The Moon is conspicuously marked by compositional differentiation in the form of its anorthosite highlands, while the other terrestrial planets appear to lack early plagioclase flotation. One of the most conspicuous features on Mars, in contrast, is its crustal dichotomy. These ancient large-scale features may be linked to the dynamic processes of initial planetary differentiation. First-order planetary characteristics such as radius and volatile content can influence differentiation sufficiently to produce bodies with surface features as dissimilar as those of Mars and the Moon.

Models: Magma ocean (MO) fractional solidification is calculated assuming convection in a low- viscosity liquid magma ocean maintains a homogeneous composition [5, 6] and solidification proceeds from the bottom upward. Incompatible elements are enriched in the evolving liquids as solidification proceeds. Subsequent cumulate overturn to gravitational stability is calculated using a thermochemical convection code.

Planetary Radius and a Heterogeneous Mantle: The magma ocean generally produces cumulates that increase in density with radius as it solidifies from the bottom upward. Models indicate that cumulates do not form a monotonic density gradient: Cumulates in portions of the lower and upper mantle have identical densities but different bulk compositions.

Numerical experiments and theory indicate that non-monotonic density profiles readily overturn to form mantles that are laterally heterogeneous compositionally [7, 8]. After overturn density decreases monotonically with radius. The densest materials from near the surface sink to the core-mantle boundary and buoyant mid-mantle material rises to the surface. Upgoing and downgoing plumes with identical densities, however, stall at mutual neutral buoyancy, creating a radius interval with large-scale lateral heterogeneity in temperature and major and trace elements. This lateral heterogeneity provides a range of magmatic source regions at a given depth, and will influence the initiation and pattern of thermal convection. Post-overturn temperature profiles are shown in Fig. 1.

In a fractional crystallization model, following overturn to stability, the Moon has a laterally-heterogeneous mantle over ~40% of its 1,000 km initial magma ocean depth. Mars by comparison is laterally heterogeneous over only ~30% of its 2,000-km cumulate mantle (Fig. 1) (this example assumes a whole-mantle magma ocean; see below for partial mantle treatments). On the Moon this result is consistent with the many compositions of picritic basalts that appear to have originated at similar depths.

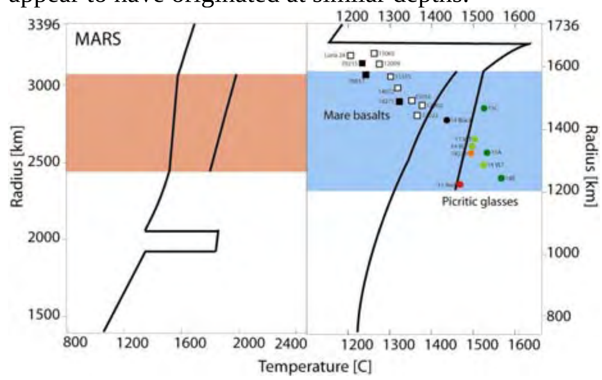

Figure 1: Mantle temperatures of Mars (left) and the Moon (right) following overturn of the cumulate mantle to gravitational stability. The shaded regions indicate the depth ranges over which the mantle is laterally heterogeneous in composition and temperature. Lunar multiple saturation points from references listed in [9].

Magma Ocean Depth and Crustal Compositions: The temporal distribution of heat sources in the early solar system likely produced partial-mantle magma oceans, either instead of or in addition to one or more whole-mantle magma oceans. Righter *et al.* [10] found that Martian meteorites' trace element abundances are consistent with a 700 to 800 km magma ocean. Rubie *et al.* [11] estimated the depth of a terrestrial magma ocean to be 550 km based on Ni partitioning.

Fractional crystallization of a magma ocean necessarily creates early cumulates with lower densities (higher MgO content) than the undifferentiated progenitor, and late cumulates from evolved liquids that are denser (higher FeO content) than any undifferentiated material. If viscosity is low enough to allow flow during solid-state cumulate overturn, cumulates and/or the primitive material will rise to a level of neutral buoyancy in the mantle, perhaps sufficiently shallow to allow adiabatic melting and form an earliest crust.

On Mars, if the partial magma ocean is as deep as ~1,550 km (~ 19 GPa) then the layer of differentiated majorite + ringwoodite that crystallizes between 19 GPa and 14 GPa is thick enough to rise during gravitational overturn above its solidus and melt to produce an early basaltic crust. In a partial magma ocean between ~1,550 and ~1,250 km (~15 GPa) deep neither the primitive material nor the majorite + ringwoodite cumulates rise sufficiently to melt. The only magma produced for an early crust is from the shallowest pyroxene + olivine composition. If the magma ocean is less than ~1,150 km (~14 GPa) deep, the primitive material rises sufficiently during overturn to melt adiabatically. These experiments indicate that different initial magma ocean depths will produce distinct compositions of earliest planetary crust, and will in some cases create two distinct compositions (Fig. 2).

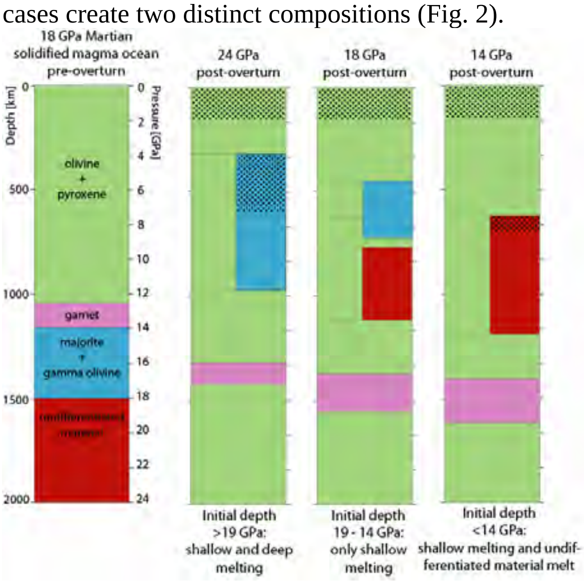

Figure 2: Solidified pre-overturn Martian 18 GPa MO (left). *Right-hand three columns*: Post-overturn profiles for MOs of three depths. Where the column is divided vertically the resulting mantle is heterogeneous laterally. Approximate regions of adiabatic melting are stippled.

Volatiles and Plagioclase Flotation: The Moon is likely a dry body formed from heated material following the giant impact with the Earth [12]. Its low mantle pressure range and dry magma ocean composition both encourage plagioclase formation. Mars was a wetter planet in its past, and its high mantle iron content indicates more oxidizing conditions of formation than were present on the Moon or Earth. Water in silicate melt suppresses plagioclase formation, and the higher pressure range of the Martian interior further limits its stability field. Water, then, is one of the dominant variables in the outcome of magma ocean crystallization.

On the Earth plagioclase may become stable at pressures from 1.5 to 0.9 GPa [*e.g.*, 13]. The depth of Mars at 1.5 GPa is 125 km, and corresponds to 90% solidification of a 2,000-km deep magma ocean. On the Moon, 1.5 GPa is reached at 360 km and represents 56% solidification of a 1,000-km deep lunar magma ocean. Experimental results show the plagioclase stability field is lowered by 70 to 100 degrees with the addition of 2 to 3 wt% of water [7, 14]. Plagioclase crystallization on Mars is therefore suppressed until late in solidification by both the planet's pressure gradient and by any water in the evolving liquids.

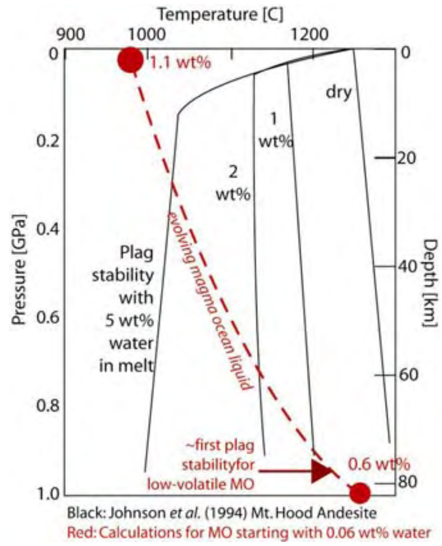

Figure 3: Possible Martian MO liquid evolution. *Solid lines*: Changes in plag stability in a Mt. Hood Andesite with addition of water [14]. *Dashed line*: MO liquid evolution from 0.06 wt% H_2O initial bulk composition.

Plagioclase crust formation is controlled by the magma's volatile content and the size of the body. Plagioclase flotation may be one of the few ways to create a lasting crust on a liquid magma ocean (quench crusts will founder [7]). Such a crust, in the absence of an insulating atmosphere, will slow the cooling of the planet significantly. The Moon may therefore have remained partially molten far longer than Mars, which likely cooled through convection with little or no conductive crust.

Conclusions: Following magma ocean solidification and cumulate overturn, no body will have a simple depth-composition relationship, but all will be stably stratified with respect to density, inhibiting the onset of thermal convection. The initial depth of the magma ocean predicts the compositions of the earliest crusts resulting from adiabatic melting during overturn. Formation of a plagioclase crust may be the only way to significantly slow magma ocean crystallization, and its presence is more likely for smaller and drier bodies. Thus smaller bodies with magma oceans may solidify more slowly than larger bodies.

References: [1] Wood *et al.* (1970) Proc. Apollo 11 LPSC, 965-988. [2] Smith *et al.* (1970) Proc. Apollo 11 LPSC, 897-925. [3] Warren (1985) *Ann. Rev. Earth Planet. Sci. 13*, 201-240. [4] Tonks and Melosh (1993) *JGR 98*, 5319-5333. [5] Solomatov (2000) *Origin Earth and Moon*, U. Ariz.. [6] Elkins-Tanton *et al.* (2003) *MAPS*, *38*, 1753-1771. [7] Elkins-Tanton et al. (2005) *JGR* doi:10.1029/2005JE002480, 2005. [8] Zaranek and Parmentier (2004) *JGR* doi: 10.1029/2003JB002462. [9] Elkins-Tanton *et al.* (2004) *EPSL*, *222*, 17-27. [10] Righter *et al.* (1998) *GCA 62*, 2167-2177. [11] Rubie *et al*. (2003) *EPSL*, *205*, 239-255. [12] Hartman and Davis (1975) *Icarus 24*, 504-515. [13] Presnall *et al.* (2002) GCA 66, 2073-2090. [14] Johnson *et al*. (1994) Rev. in Mineralogy 30, 281-330.