Earth and Planetary Science Letters 403 (2014) 225–235

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

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Can a fractionally crystallized magma ocean explain the thermo-chemical evolution of Mars?

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ARTICLE INFO

Article history:
Received 7 May 2013
Received in revised form 22 May 2014
Accepted 24 June 2014
Available online 24 July 2014
Editor: C. Sotin

Keywords:
Mars
magma ocean
mantle reservoirs
mantle overturn
chemical gradient
thermo-chemical convection

ABSTRACT

The impact heat accumulated during the late stage of planetary accretion can melt a significant part or even the entire mantle of a terrestrial body, giving rise to a global magma ocean. The subsequent cooling of the interior causes the magma ocean to freeze from the core–mantle boundary (CMB) to the surface due to the steeper slope of the mantle adiabat compared to the slope of the solidus.

Assuming fractional crystallization of the magma ocean, dense cumulates are produced close to the surface, largely due to iron enrichment in the evolving magma ocean liquid. A gravitationally unstable mantle thus forms, which is prone to overturn. We investigate the cumulate overturn and its influence on the thermal evolution of Mars using mantle convection simulations in 2D cylindrical geometry. We present a suite of simulations using different initial conditions and a strongly temperature-dependent viscosity. We assume that all radiogenic heat sources have been enriched during the freezing-phase of the magma ocean in the uppermost 50 km and that the initial steam-atmosphere created by the degassing of the freezing magma ocean was rapidly lost, implying that the surface temperature is set to present-day values. In this case, a stagnant lid quickly forms on top of the convective interior preventing the uppermost dense cumulates to sink, even when allowing for a plastic yielding mechanism. Below this dense stagnant lid, the mantle chemical gradient settles to a stable configuration. The convection pattern is dominated by small-scale structures, which are difficult to reconcile with the large-scale volcanic features observed over Mars' surface and partial melting ceases in less than 900 Ma. Assuming that the stagnant lid can break because of additional mechanisms and allowing the uppermost dense layer to overturn, a stable density gradient is obtained, with the densest material and the entire amount of heat sources lying above the CMB. This stratification leads to a strong overheating of the lowermost mantle, whose temperature increases to values that exceed the liquidus. The iron-rich melt would most likely remain trapped in the lower part of the mantle. The upper mantle in that scenario cools rapidly and only shows partial melting during the first billion year of evolution. Therefore a fractionated global and deep magma ocean is difficult to reconcile with observations. Different scenarios assuming, for instance, a hemispherical or shallow magma ocean, or a crystallization sequence resulting in a lower density gradient than that implied by pure fractional crystallization will have to be considered.

1. Introduction

Geochemical analyses of the so-called SNC meteorites suggest the existence of three to four separate and isotopically distinct reservoirs in the martian mantle, which have been preserved over the entire planetary evolution (e.g., Jagoutz, 1991; Papike et al., 2009). Two of these reservoirs are depleted in incompatible elements and are most likely situated in the mantle, whereas the third one is enriched and could be located either in the crust or in the mantle (Foley et al., 2005). Until present, most of the dynamical simulations of Mars' mantle convection did not account for the formation or preservation of such reservoirs, and focused mainly on explaining, for instance, the crustal dichotomy or the Tharsis bulge (e.g., Breuer et al., 1998; Harder and Christensen, 1996; Schumacher and Breuer, 2006; Keller and Tackley, 2009; Šramek and Zhong, 2012). Scenarios that consider the early formation of chemical heterogeneities in the Martian mantle assume either a global magma ocean (Elkins-Tanton et al., 2003, 2005a;

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http://dx.doi.org/10.1016/j.epsl.2014.06.034

Debaille et al., 2009), or at least substantial partial melting in the earliest phase of the planet's evolution (Schott et al., 2001; Ogawa and Yanagisawa, 2011; Plesa and Breuer, 2014). It has been suggested that early in the evolution of Mars, the large amount of primordial heat due to accretion, core formation, and possibly short lived radioactive elements can give rise to a magma ocean as a consequence of significant or perhaps even complete melting of the mantle (e.g., Breuer and Moore, 2007). This assumption is confirmed by studies on short-lived radionuclides such as ^{182}Hf suggesting that the separation between silicates and iron occurred in the first million years after accretion (e.g., Kleine et al., 2002). Furthermore, estimates of the timescale of the process of core formation suggest that, to achieve rapid separation, both silicates and metals need to be fluid, at least in the upper part of mantle (e.g., Stevenson, 1990).

In general, the magma ocean freezes from the bottom (i.e. from the CMB, if it comprises the entire mantle) to the surface because of the steeper slope of the mantle adiabat compared to the slope of the solidus (e.g., Solomatov, 2000). Solomatov (2000) discusses two main freezing mechanisms: equilibrium and fractional crystallization, depending on the size and settling velocity of crystals that form upon solidification. In the case of equilibrium crystallization, crystal size and settling velocities are so small that freezing takes place before crystal–melt separation occurs. The magma ocean solidifies without differentiating and a chemically homogeneous mantle is formed. In the case of fractional crystallization, instead, denser cumulates are formed while the crystallization of the magma ocean proceeds from the bottom, mainly because of iron enrichment in the evolving magma ocean liquid (Elkins-Tanton et al., 2003). The consequence is the formation of an unstably stratified density profile. Whether equilibrium or fractional crystallization in a magma ocean is the dominant process also depends on the time scale of freezing and may change with time. A large degree of melting of the mantle implies a very low viscosity of the magma ocean. The viscosity of a melt/crystal mixture increases abruptly near a critical crystal fraction as suggested by theoretical and experimental studies of concentrated suspensions (e.g., Mooney, 1951; Roscoe, 1952; Campbell and Forgacs, 1990) and by experiments with partial melts (e.g., Arzi, 1978; Lejeune and Richet, 1995). This sudden increase in viscosity is defined by the so-called "rheological transition", which depends on the crystal size distribution, the crystal shape and orientation, and other factors. The critical crystal fraction for this transition varies in the literature. Saar et al. (2001), using numerical models, found that for plagioclase crystals, the critical crystal fraction that allows a crystal network to form lies between 8 and 20%. Other studies place this transition at 30% to 50% depending on the basalt composition (Philpotts et al., 1996). In Solomatov (2000), the rheological transition is assumed to take place at ∼60%. Solomatov (2000) argues for instance that freezing of the Earth's lower mantle lasts only a few hundred years, until the crystal fraction becomes larger than the critical value marking the rheological transition. For larger crystal fractions, the time scale for further crystallization slows down by several orders of magnitudes (∼10 to 100 Ma) as the convection speed and cooling efficiency of the partially molten mantle is significantly reduced. Fractional crystallization may become dominant at this stage. If the crystals do not remain in suspension, a crystal network forms instead and efficiently compacts, ultimately resulting in a fractionation between crystals and melt. Therefore, depending on how the magma ocean freezes, i.e. via fractional or equilibrium crystallization, either a chemically homogeneous or stratified mantle is obtained and different evolutionary paths can be expected. So far, two end-member scenarios have been suggested in the literature for Mars:

(1) In the first scenario, it is assumed that the early martian mantle is homogeneous apart from an upper depleted layer (Schott et al., 2001; Ogawa and Yanagisawa, 2011). Such a structure can be realized if Mars did not experience a deep magma ocean but a large amount of melting causing secondary differentiation early in its evolution. Alternatively, if a deep magma ocean existed, the largest part of the mantle, from the CMB up to a certain depth, freezes rapidly by equilibrium crystallization and only melt from a remaining shallow layer rises via porous flow due to compaction of the silicate matrix or through cracks and channels. This early melt produces a primordial crust and marks the start of the differentiation process. As the mantle residue is in general less dense than the primordial mantle, chemically distinct reservoirs can form depending on the density contrast between depleted and primordial mantle (Plesa and Breuer, 2014). Typically, a depleted and compositionally buoyant upper mantle can develop, which overlies an undepleted and compositionally denser lower mantle (Schott et al., 2001; Ogawa and Yanagisawa, 2011). These two layers convect separately with the lower one that tends to become thinner with time because of erosion, while its composition remains almost primordial. However, whether or not this scenario can actually explain the isotopic characteristics of the martian meteorites is unclear and, to the best of our knowledge, it is a problem that still needs to be investigated.

(2) In the second scenario, the mantle freezes by fractional crystallization resulting in a stratification with Fe-rich cumulates close to the surface and Mg-rich cumulates at the CMB, a configuration that is gravitationally unstable. Being highly incompatible, radioactive elements are enriched in the evolving residual liquid of the magma ocean. Therefore, a substantial amount of radioactive heat-sources is concentrated in the uppermost layer of Mars' mantle (Elkins-Tanton et al., 2005b). The unstable layering results then in an global mantle overturn, which has been estimated to take place within ∼10–100 Ma (Elkins-Tanton et al., 2003, 2005b, 2008; Debaille et al., 2009). As a consequence of the overturn, the dense surface material sinks to the core–mantle boundary, while the light material from the CMB rises to the surface, thus leading to a stable, chemically layered mantle. It is further assumed that, during the overturn, rising upwellings can melt by adiabatic decompression and produce an early crust. The residual mantle of this process has been associated with the source region of the nakhlites (Debaille et al., 2009). The chemical layering following the overturn suppresses subsequent thermally driven convection and early-formed reservoirs are preserved throughout the planet's history. In this way, distinct reservoirs can be maintained stable over the rest of the thermal evolution of Mars. In a recent study, Scheinberg et al. (2014) examined the mantle overturn following magma ocean solidification using convection models in a 3D spherical geometry and a temperature dependent viscosity but neglecting the decay of radioactive elements and the cooling of the core, i.e. two factors that contribute to render convection less and less vigorous as the mantle evolution proceeds. Their simulations indicate that the cumulate overturn tends to be characterized by a short-wavelength convective planform. In addition, the authors argue that, despite the stable density gradient resulting from the initial overturn, convection and mixing may bring near the surface small-scale reservoirs, which, upon melting, could account for the differences in the isotopic signatures found among the SNC meteorites.

An important feature of the mantle overturn is also the subduction of the dense surface material. Most models employed to investigate this process (Elkins-Tanton et al., 2003, 2005a, 2005b; Scheinberg et al., 2014) assumed a small viscosity contrast between the interior and the surface, implicitly allowing for the mobilization of the uppermost layers. A small viscosity contrast can be the consequence of high surface temperatures maintained during the solidification phase of the magma ocean. In fact, a large amount of volatiles is expected to be outgassed and to form a steam atmosphere able to keep the surface temperature even