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On the cooling of a deep terrestrial magma ocean

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ABSTRACT

Several episodes of complete melting have probably occurred during the first stages of the Earth's evolution. We have developed a numerical model to monitor the thermal and melt fraction evolutions of a cooling and crystallizing magma ocean from an initially fully molten mantle. For this purpose, we numerically solve the heat equation in 1D spherical geometry, accounting for turbulent heat transfer, and integrating recent and strong experimental constraints from mineral physics. We have explored different initial magma ocean viscosities, compositions, thermal boundary layer thicknesses and initial core temperatures.

We show that the cooling of a thick terrestrial magma ocean is a fast process, with the entire mantle becoming significantly more viscous within 20 kyr. Due to the slope difference between the adiabats and the melting curves, the solidification of the molten mantle occurs from the bottom up. In the meantime, a crust forms due to the high surface radiative heat flow, the last drop of fully molten silicate is restricted to the upper mantle. Among the studied parameters, the magma ocean lifetime is primarily governed by its viscosity. Depending on the thermal boundary layer thickness at the core–mantle boundary, the thermal coupling between the core and magma ocean can either insulate the core during the magma ocean solidification and favor a hot core or drain the heat out of the core simultaneously with the cooling of the magma ocean. Reasonable thickness for the thermal boundary layer, however, suggests rapid core cooling until the core–mantle boundary temperature results in a sluggish lowermost mantle. Once the crystallization of the lowermost mantle becomes significant, the efficiency of the core heat loss decreases. Since a hotter liquidus favors crystallization at hotter temperatures, a hotter deep mantle liquidus favors heat retention within the core. In the context of an initially fully molten mantle, it is difficult to envision the formation of a basal magma ocean or to prevent a major heat depletion of the core. As a consequence, an Earth's geodynamo sustained only by core cooling during 4 Gyr seems unlikely and other sources of motion need to be invoked.

1. Introduction

Geochemical evidence (Touboul et al., 2012; Rizo et al., 2013) suggests that the Earth's mantle has experienced several episodes of global melting during its early evolution, leading to the formation of the early continental crust and facilitating the core formation (Kleine et al., 2009). These episodes were probably enhanced by giant impacts occurring during the late stages of planetary formation (Agnor et al., 1999). Although not yet clearly established, it is likely that these giant impacts, such as the one that is thought to have formed the Earth–Moon system, could have melted 30 to 100% of the Earth's mantle depending on the impactor/target mass ratio and on the pre-impact thermal state of the target (Canup, 2012; Ćuk and Stewart, 2012; Nakajima and Stevenson, 2015). During the cooling and the subsequent crystallization of a magma ocean (MO), compatible elements (e.g. Mg, Cr) were preferentially collected in the solid phase while the incompatible elements (e.g. Al, Na, Fe) selectively partitioned into melts. In addition to temperature, the degree of solid–melt fractionation is highly sensitive to a variety of physical parameters, including pressure (Nomura et al., 2011; Andrault et al., 2012). Hence, characterizing the cooling of a deep terrestrial magma ocean and in particular the timescale and depth at which the last drop of melt solidifies are of first importance to understand the current chemical composition of the Earth's mantle and the dating of its major differentiation events (Boyet and Carlson, 2005).

The composition and the rheology of such a magma ocean directly affect its lifetime, but remain poorly constrained (Solomatov, 2007). The magma ocean is composed of low viscosity molten sili-

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cate material but its chemical composition remains uncertain, with a MgO/SiO_2 ratio around those of chondritic or peridotitic compositions (Ringwood, 1966; Allègre et al., 1995; Javoy et al., 2010). Recent high-pressure laboratory measurements report the solidus and liquidus of both a chondritic and peridotitic mantle compositions up to pressures that are compatible with the Earth's lowermost mantle conditions (Fiquet et al., 2010; Andrault et al., 2011). Moreover, recent shock experiments now provide important constraints on the thermodynamic parameters used to determine the adiabatic profiles in the magma ocean up to 140 GPa (Mosenfelder et al., 2009; Thomas et al., 2012; Thomas and Asimow, 2013). Since the difference between their slopes governs the depth at which crystallization is initiated, both the liquidus and the adiabats play a key role in the cooling of the magma ocean. If the adiabat had a steeper slope than the liquidus in the mid-mantle (Mosenfelder et al., 2007; Stixrude et al., 2009), solidification would start at mid-mantle depth. In this case, a lowermost magma ocean would cool and solidify much more slowly because of the thermal blanketing of the overlying solid mantle (Labrosse et al., 2007). However, if the mantle liquidus had a steeper slope than the adiabat through the whole mantle (Thomas et al., 2012), solidification would start from the CMB thus reducing the likeliness of a basal magma ocean, unless invoking an enrichment in dense incompatible elements in the residual liquid. In any case, the important dynamical change does not occur when the adiabat crosses the liquidus, because the mantle keeps its liquid behavior, but rather when the degree of partial melting decreases below a critical value from which the mantle behaves as a solid. Therefore, the recent determination of melting curves and elastic parameters of silicate melts up to core–mantle boundary (CMB) conditions offers a great opportunity to improve our knowledge of the cooling dynamics of a deep terrestrial magma ocean.

The magma oceans such as the one generated by the Moon-forming impact participated to the core-formation process. The early thermal state of the core remains poorly constrained. It results from the contribution of the accretionary processes (Safronov, 1978; Kaula, 1979), including giant impact (Tonks and Melosh, 1992) and radiogenic heating (Yoshino et al., 2003) as well as the conversion of potential energy into heat via viscous dissipation during the metal/silicate separation (Ke and Solomatov, 2009; Monteux et al., 2009; Ricard et al., 2009; Samuel et al., 2010). The combined processes leading to core formation can yield a wide range of possible early thermal states, depending on the nature and timescale of core formation processes. The core could initially have had a temperature close to the deep mantle temperature if thermal equilibration was efficient. Alternatively, it could have been hotter than the mantle if the gravitational potential energy released during core formation was largely retained within the core itself, a situation which would be followed by a strong heating of the lowermost mantle from this superheated core (Samuel et al., 2010). In turn, the thermo-mechanical properties of the magma ocean can have a strong influence on the early evolution of the heat repartition between the core and the mantle. A key question is to determine how much a deep magma ocean can enhance core cooling. This can have important consequences on the duration and the generation of the Earth's dynamo (Monteux et al., 2011).

The low magma ocean viscosities resulting from the hot early temperatures imply that the cooling of such a deep molten mantle was highly turbulent (Solomatov, 2007). Studies of the early mantle have either characterized the cooling of a magma ocean restricted to the first 1000 km (Abe, 1997) or did not consider the presence of a molten layer just above the core–mantle boundary, and its effect of the CMB heat flow (Nakagawa and Tackley, 2014). However, the hypothesis of an early largely molten mantle combined with the determination of solidus/liquidus and thermodynamical properties of silicate melts up to 140 GPa now allow a more accurate characterization of the cooling of a deep terrestrial magma ocean and the thermal coupling with its underlying core. The aim of this work is to constrain the lifetime of a deep magma ocean and to determine the pressure at which the magma ocean crystallization finished. To achieve these goals, we have developed a numerical model to characterize the early evolution of (i) the temperature and melt fraction of an initially fully molten isochemical mantle and (ii) the temperature of the core. We incorporate in our models recent and strong experimental constraints on the solidus and liquidus profiles and on the thermodynamical properties of silicate melts up to ∼140 GPa. We explore different core temperatures, magma ocean compositions and viscosities.

2. Convective cooling of the magma ocean

Miller et al. (1991) characterized the cooling and the subsequent crystallization of a magma ocean with a chondritic composition as a sequence of isentropes with decreasing potential temperature. Later on, Abe (1997) investigated the thermal evolution of magma ocean using a one-dimensional heat transfer model. However, these studies were restricted to the first 1000 km and did not integrate the mutual influence of the magma ocean and its underlying material on the cooling. Labrosse et al. (2007) studied the cooling of a stable dense molten layer above the CMB overlaid by a solid mantle. In their model they consider the crystallization of a single-component (forsterite) magma ocean assuming a solidification proceeding from the top to the bottom according to Mosenfelder et al. (2007). More recently, Nakagawa and Tackley (2014) characterized the coupled thermal evolution of Earth's early mantle and core considering a 2900 km thick viscous mantle but ignoring the potential presence of a molten layer just above the core–mantle boundary, and its effect of the CMB heat flow. Here, we model the secular cooling of an initially fully molten magma ocean by convective transport of heat in a 1-D spherically symmetric geometry. We assume a multicomponent chemically homogeneous magma ocean made of a combination of forsterite, enstatite, fayalite, anorthite and diopside. In the following sections, we describe the model setup and equations.

2.1. Physical model for planetary thermal evolution

We model the thermal evolution of a 2900 km thick isochemical silicate mantle overlying an iron core by solving the conservation of energy in a one-dimensional, spherically symmetric domain (with a radius ranging from 3500 to 6400 km):

$$\rho C_p \frac{\partial T}{\partial t} = \nabla.(k\nabla T), \tag{1}$$

with ρ the density, C_p the mantle heat capacity, T the temperature, t the time and k the thermal conductivity. Among the heat sources that have potentially delivered the energy required for significant melting in the early Earth, the decay of short-lived radioactive isotopes such as ^{26}Al and ^{60}Fe have probably played a major role especially for 10 to 100 km size objects (Yoshino et al., 2003). However, their half-life times (0.73 My and 1.5 My respectively) (Carlson and Lugmair, 2000) are much shorter than the time at which the Moon forming impact is supposed to have occurred (between 30 and 100 Myrs after the formation of the first solids of the Solar System) (Kleine and Rudge, 2011). Concerning the long-lived radioactive elements such as ^{40}K, Th or U, their concentrations were certainly significant at the time of the Moon-forming impact, but their heat production rates are much smaller. Hence the contribution form the long-lived radio-active elements during the magma ocean lifetime is negligible. Thus, we can reasonably neglect radiogenic heating in our models.