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On the cooling of a deep terrestrial magma ocean

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Abstract

We have developed numerical models to monitor the thermo-chemical evolution of a cooling and crystallizing magma ocean from an initially fully molten mantle. For this purpose, we use a 1D spherical approach accounting for turbulent convective heat transfer. Our numerical model benchmarked with analytical solutions solves the heat equation with extremely fine grid spacing (down to the mm), which is a strong requirement for resolving properly thin thermal boundary layers. This model also integrates recent and strong experimental constraints from mineral physics.

1 Introduction

During its early evolution, geochemical evidences suggest that the Earth's mantle has experienced several episodes of global melting leading to the formation of the early continental crust [1] or facilitating the core formation [2]. The late stages of planetary formation probably involved several large to giant impacts [3]. Although not yet clearly established, it is likely that these giant impacts, such as the one that probably 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 [4]. Hence, the likelihood of episodic fully molten mantle is important.

The dynamics of such a thick magma ocean is very turbulent because of the small viscosity of the fully molten mantle material [5]. Studies of the magma ocean cooling are often restricted to the first 1000-2000 km. The aim of our study is to characterize the cooling dynamics within a fully molten magma ocean. We explore different initial magma ocean thermal states as well as different initial core temperatures. Our model integrates recent and strong experimental constraints such as the melting curves (solidus and liquidus) that have been determined up to core-mantle boundary conditions [6]. Our model also benefits from the recent advances in the determination of the equation of state of silicate liquids and of the thermal conductivity of deep mantle material.

2 Numerical models

We model the secular cooling of a fully molten magma ocean by convective transport of heat in a 1-D spherically symmetric geometry. In vigorously convecting systems such as magma oceans, the temperature distribution is nearly adiabatic and isentropic [5]. Hence as the initial condition we assume an adiabatic temperature profile with a surface temperature of 3000 K. The solidus and liquidus may play a major role in the early thermal evolution of the magma ocean. Recent laboratory experiments now constrain the liquidus and solidus of chondritic material up to pressures compatible with the core mantle boundary conditions [6] (Fig.1).

Because of the low viscosity of the molten magma ocean, Rayleigh numbers can reach values ranging from 10^{20} to 10^{30} [5]. We model the thermal evolution of a deep magma ocean using a spherically symmetric one-dimensional single-phase flow model. We solve the following equation of heat transfer:

$$\rho C_p \frac{\partial T}{\partial t} = \nabla.(k\nabla T) + \rho H \quad (1)$$

with ρ is the density, C_p is the mantle heat capacity, T is the temperature, t is the time, k is the thermal conductivity and H is the radiogenic heating.

Thermal energy is efficiently transferred by thermal convection in the region where the temperature gradient is steeper than the adiabatic temperature gradient. In Eq. 1, the thermal conductivity k is the sum of the intrinsic thermal conductivity of the mantle material k_c and the effective thermal conductivity due to thermal convection k_v. We estimate the effective thermal conductivity as follows [8]:

$$k_v = F_{conv} L / \Delta T$$

$$F_{conv} = 0.089 k_c \Delta T Ra^{1/3} / L \quad (2)$$

$$Ra = \frac{\alpha g(r) C_p \rho^2 \Delta T L^3}{k\eta}$$

where L is the thickness of the magma ocean, F_{conv} is the convective heat flux, ΔT is the temperature excess relative to the adiabatic temperature profile, Ra is the Rayleigh number, α is the thermal expansion coefficient of the magma ocean, $g(r)$ is the gravity at radius r and η is the local dynamic viscosity.

3 Results

Our preliminary results (Fig. 1) show that the temperature rapidly decreases down to the liquidus temperature of the mantle and solidification occurs first from the surface where the temperature is imposed to 273 K leading to the formation of a thin crust in the thermal boundary layer [5]. Then, solidification occurs at the bottom as the liquidus is steeper than the adiabatic temperature profile. The solid fraction increases up to a critical value separating the turbulent liquid regime from the solid-state viscous regime [5,7]. As soon as this critical value is reached (within $\sim$ 1000 yr) the efficiency of mantle cooling becomes limited.

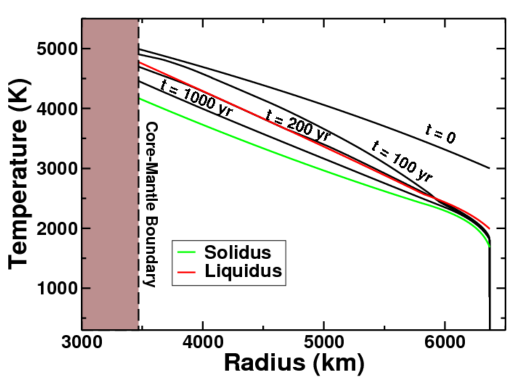

Figure 1: Preliminary calculations of the cooling of a fully molten Earth, using a 1D model to solve the heat equation at each depth.

4 Conclusion

We show that a deep magma ocean starts to cristallise rapidly after its formation. Then, once the melt fraction reaches a critical value, the cooling efficiency becomes limited. This decrease in the cooling rate can have important consequences for the deep thermal state of the mantle, the chemical fractionation of compatible/incompatible elements and, hence for the formation of a dense basal magma ocean [9].

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