

# 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. Because of the slope difference between the adiabat 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. We show that 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

28 core cooling until the core-mantle boundary temperature results in a sluggish  
29 lowermost mantle. Once the crystallization of the lowermost mantle becomes  
30 significant, the efficiency of the core heat loss decreases. Since a hotter liquidus  
31 favors crystallization at hotter temperatures, a hotter deep mantle liquidus fa-  
32vors heat retention within the core. In the context of an initially fully molten  
33 mantle, it is difficult to envision the formation of a basal magma ocean or to  
34 prevent a major heat depletion of the core. As a consequence, an Earth's geo-  
35dynamo sustained only by core cooling during 4 Gyr seems unlikely and other  
36 sources of motion need to be invoked.

37

38 *Keywords:* Early Earth, thermal evolution, magma ocean, numerical modeling

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## 39 **1. Introduction**

40 Geochemical evidence (*Touboul et al.*, 2012; *Rizo et al.*, 2013) suggest that  
41 the Earth's mantle has experienced several episodes of global melting during its  
42 early evolution, leading to the formation of the early continental crust and facil-  
43 itating the core formation (*Kleine et al.*, 2009). These episodes were probably  
44 enhanced by giant impacts occurring during the late stages of planetary forma-  
45 tion (*Agnor et al.*, 1999). Although not yet clearly established, it is likely that  
46 these giant impacts, such as the one that is thought to have formed the Earth-  
47 Moon system, could have melted 30 to 100% of the Earth's mantle depending  
48 on the impactor/target mass ratio and on the pre-impact thermal state of the  
49 target (*Canup*, 2012; *Ćuk and Stewart*, 2012; *Nakajima and Stevenson*, 2015).  
50 During the cooling and the subsequent crystallization of a magma ocean (MO),  
51 compatible elements (e.g. Mg, Cr) were preferentially collected in the solid  
52 phase while the incompatible elements (e.g. Al, Na, Fe) selectively partitioned  
53 into melts. In addition to temperature, the degree of solid-melt fractionation  
54 is highly sensitive to a variety of physical parameters, including pressure (*No-*

55 *mura et al.*, 2011; *Andrault et al.*, 2012). Hence, characterizing the cooling of  
56 a deep terrestrial magma ocean and in particular the timescale and depth at  
57 which the last drop of melt solidifies are of first importance to understand the  
58 current chemical composition of the Earth's mantle and the dating of its major  
59 differentiation events (*Boyet and Carlson*, 2005).

60

61 The composition and the rheology of such a magma ocean directly affect its  
62 lifetime, but remain poorly constrained (*Solomatov*, 2007). The magma ocean  
63 is composed of low viscosity molten silicate material but its chemical compo-  
64 sition remains uncertain, with a MgO/SiO<sub>2</sub> ratio around those of chondritic  
65 or peridotitic compositions (*Ringwood*, 1966; *Allègre et al.*, 1995; *Javoy et al.*,  
66 2010). Recent high-pressure laboratory measurements report the solidus and  
67 liquidus of both a chondritic and peridotitic mantle compositions up to pres-  
68 sures that are compatible with the Earth's lowermost mantle conditions (*Fiquet*  
69 *et al.*, 2010; *Andrault et al.*, 2011). Moreover, recent shock experiments now  
70 provide important constraints on the thermodynamic parameters used to deter-  
71 mine the adiabatic profiles in the magma ocean up to 140 GPa (*Mosenfelder*  
72 *et al.*, 2009; *Thomas et al.*, 2012; *Thomas and Asimow*, 2013). Since the differ-  
73 ence between their slopes governs the depth at which crystallization is initiated,  
74 both the liquidus and the adiabat play a key role in the cooling of the magma  
75 ocean. If the adiabat had a steeper slope than the liquidus in the mid-mantle  
76 (*Mosenfelder et al.*, 2007; *Stixrude et al.*, 2009), solidification would start at  
77 mid-mantle depth. In this case, a lowermost magma ocean would cool and so-  
78 lidify much more slowly because of the insulating effect of the overlying solid  
79 mantle (*Labrosse et al.*, 2007). However, if the mantle liquidus had a steeper  
80 slope than the adiabat through the whole mantle (*Thomas et al.*, 2012), so-  
81 lidification would start from the CMB thus reducing the likeliness of a basal

82 magma ocean, unless invoking an enrichment in dense incompatible elements  
83 in the residual liquid. In any case, the important dynamical change does not  
84 occur when the adiabat crosses the liquidus, because the mantle keeps its liq-  
85 uid behavior, but rather when the degree of partial melting decreases below  
86 a critical value from which the mantle behaves as a solid. Therefore, the re-  
87 cent determination of melting curves and elastic parameters of silicate melts  
88 up to core-mantle boundary (CMB) conditions offers a great opportunity to im-  
89 prove our knowledge of the cooling dynamics of a deep terrestrial magma ocean.

90  
91 The magma oceans such as the one generated by the Moon-forming im-  
92 pact participated to the core-formation process. The early thermal state of  
93 the core remains poorly constrained. It results from the contribution of the  
94 accretionary processes (*Safronov*, 1978; *Kaula*, 1979), including giant impact  
95 (*Tonks and Melosh*, 1992) and radiogenic heating (*Yoshino et al.*, 2003) as well  
96 as the conversion of potential energy into heat via viscous dissipation during  
97 the metal/silicate separation (*Ke and Solomatov*, 2009; *Monteux et al.*, 2009;  
98 *Ricard et al.*, 2009; *Samuel et al.*, 2010). The combined processes leading to  
99 core formation can yield a wide range of possible early thermal states, depend-  
100 ing on the nature and timescale of core formation processes. The core could  
101 initially have had a temperature close to the deep mantle temperature if ther-  
102 mal equilibration was efficient. Alternatively, it could have been hotter than the  
103 mantle if the gravitational potential energy released during core formation was  
104 largely retained within the core itself, a situation which would be followed by  
105 a strong heating of the lowermost mantle from this superheated core (*Samuel*  
106 *et al.*, 2010). In turn, the thermo-mechanical properties of the magma ocean  
107 can have a strong influence on the early evolution of the heat repartition be-  
108 tween the core and the mantle. A key question is to determine how much a deep

109 magma ocean can enhance core cooling. This can have severe consequences on  
110 the duration and the generation of the Earth’s dynamo (*Monteux et al.*, 2011).

111

112 The low magma ocean viscosities resulting from the high early temperatures  
113 imply that the cooling of such a deep molten mantle was highly turbulent (*Solo-*  
114 *matov*, 2007). Studies of the early mantle have either characterized the cooling  
115 of a magma ocean restricted to the first 1000 km (*Abe*, 1997) or did not con-  
116 sider the presence of a molten layer just above the core-mantle boundary, and  
117 its effect of the CMB heat flow (*Nakagawa and Tackley*, 2014). However, the  
118 hypothesis of an early largely molten mantle combined with the determination  
119 of solidus/liquidus and thermodynamical properties of silicate melts up to 140  
120 GPa now allow a more accurate characterization of the cooling of a deep terres-  
121 trial magma ocean and the thermal coupling with its underlying core. The aim  
122 of this work is to constrain the lifetime of a deep magma ocean and to determine  
123 the pressure at which the magma ocean crystallization finished. To achieve these  
124 goals, we have developed a numerical model to characterize the early evolution  
125 of (i) the temperature and melt fraction of an initially fully molten isochemical  
126 mantle and (ii) the temperature of the core. We incorporate in our models the  
127 recent and strong experimental constraints on the solidus and liquidus profiles  
128 and on the thermodynamical properties of silicate melts up to  $\sim 140$  GPa. We  
129 explore different core temperatures, magma ocean compositions and viscosities.

130

## 131 **2. Convective cooling of the magma ocean**

132 *Miller et al.* (1991) characterized the cooling and the subsequent cristallisa-  
133 tion of a magma ocean with a chondritic composition as a sequence of isentropes  
134 with decreasing potential temperature. Later on, *Abe* (1997) investigated the

135 thermal evolution of magma ocean using a one-dimensional heat transfer model.  
 136 However, these studies were restricted to the first 1000 km and did not inte-  
 137 grate the mutual influence of the magma ocean and its underlying material on  
 138 the cooling. *Labrosse et al. (2007)* studied the cooling of a stable dense molten  
 139 layer above the CMB overlaid by a solid mantle. In their model they consider  
 140 the crystallization of a single-component (forsterite) magma ocean assuming a  
 141 solidification proceeding from the top to the bottom according to *Mosenfelder*  
 142 *et al. (2007)*. More recently, *Nakagawa and Tackley (2014)* characterized the  
 143 coupled thermal evolution of Earth’s early mantle and core considering a 2900  
 144 km thick viscous mantle but ignoring the potential presence of a molten layer  
 145 just above the core-mantle boundary, and its effect of the CMB heat flow. Here,  
 146 we model the secular cooling of an initially fully molten magma ocean by con-  
 147 vective transport of heat in a 1-D spherically symmetric geometry. We assume  
 148 a multicomponent chemically homogeneous magma ocean made of a combina-  
 149 tion of forsterite, enstatite, fayalite, anorthite and diopside. In the following  
 150 sections, we describe the model setup and equations.

### 151 2.1. Physical model for planetary thermal evolution

152 We model the thermal evolution of a deep iso-chemical silicate mantle over-  
 153 lying an iron core by solving the conservation of energy in a one-dimensional,  
 154 spherically symmetric domain (with a radius ranging from 3500 to 6400 km):

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

155 with  $\rho$  the density,  $C_p$  the mantle heat capacity,  $T$  the temperature,  $t$  the time  
 156 and  $k$  the thermal conductivity. Among the heat sources that have potentially  
 157 delivered the energy required for significant melting in the early Earth, the decay  
 158 of short-lived radioactive isotopes such as  $^{26}\text{Al}$  and  $^{60}\text{Fe}$  have probably played  
 159 a major role especially for 10 to 100 km size objects (*Yoshino et al., 2003*).

160 However, their half-life times (0.73 My and 1.5 My respectively) (*Carlson and*  
161 *Lugmair*, 2000) are much shorter than the time at which the Moon forming im-  
162 pact is supposed to have occurred (between 30 and 100 Myrs after the formation  
163 of the first solids of the Solar System) (*Kleine and Rudge*, 2011). Concerning  
164 the long-lived radioactive elements such as  $^{40}\text{K}$ , Th or U, their concentrations  
165 were certainly significant at the time of the Moon-forming impact, but their heat  
166 production rates are much smaller. Hence the contribution from the long-lived  
167 radio-active elements during the magma ocean lifetime is negligible. Thus, we  
168 can reasonably neglect radiogenic heating in our models.

169

170 Thermal energy is transferred by convection in the region where the temper-  
171 ature gradient is steeper than the adiabatic temperature gradient, or by conduc-  
172 tion elsewhere. To account for the heat transfer within a vigorously convecting  
173 magma ocean, in Eq. 1, the thermal conductivity  $k$  is the sum of the intrinsic  
174 thermal conductivity  $k_c$  and an effective conductivity due to thermal convection  
175  $k_v$ . Following *Neumann et al.* (2014), the latter is estimated as follows:

$$k_v = F_{conv}L/\Delta T, \quad (2)$$

176 where  $L$  is the thickness of the magma ocean at time  $t$ ,  $F_{conv}$  is the convec-  
177 tive heat flux at radius  $r$  and time  $t$  and  $\Delta T$  is the difference between the  
178 temperature profile and the adiabatic profile  $T_{ad}$  with a potential temperature  
179 corresponding to the temperature of the surface of the magma ocean.

180

181 At radius  $r$ , the convective heat flux  $F_{conv}$  depends on the local Rayleigh  
182 number  $Ra$ :

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

183 where  $\alpha$  is the thermal expansion coefficient of the magma ocean,  $g$  is the grav-  
 184 itational acceleration assumed to be constant through the whole mantle and  $\eta$   
 185 is the local dynamic viscosity.

186

187 In the context of an initially fully molten mantle, the convective velocities  
 188 are estimated to  $u_0 \sim 10$  m/s (*Solomatov, 2007*) leading to Reynolds number  
 189 values ( $Re = \rho u_0 L / \eta$ ) in the order of  $10^9$  (with  $\eta = 100$  Pa.s and  $\rho = 4000$   
 190  $\text{kg.m}^{-3}$ ) and, hence, to highly turbulent convective cooling. The low magma  
 191 ocean viscosities induce Rayleigh numbers as large as  $10^{30}$  (*Solomatov, 2007*). In  
 192 such a context, two regimes arise depending on  $Ra$  (*Solomatov, 2007; Neumann*  
 193 *et al., 2014*): the "soft-turbulent" regime (if  $Ra < 10^{19}$ ) where

$$F_{conv} = 0.089 \frac{k_c \Delta T}{L} Ra^{1/3}, \quad (4)$$

194 and the "hard-turbulent" regime (if  $Ra \geq 10^{19}$ ) where

$$F_{conv} = 0.22 \frac{k_c \Delta T}{L} Ra^{2/7} Pr^{-1/7} \lambda^{-3/7}, \quad (5)$$

195 where  $Pr$  is the local Prandtl number ( $= C_p \eta / k_c$ ) and  $\lambda$  is the aspect ratio for  
 196 the mean flow. We assume that  $\lambda = 1$  and that rotation does not have any  
 197 significant effect on the heat flow (*Solomatov, 2007*).

198

199 Depending on their ability to migrate either towards the solid phase (com-  
 200 patible) or towards the liquid phase (incompatible), the relative abundance of  
 201 some chemical elements can potentially modify the buoyancy of both the liquid  
 202 and solid phases during the crystallization of a magma ocean. This chemical  
 203 fractionation process could be enhanced by the fractional crystallization of the  
 204 magma ocean at least in the upper mantle (*Solomatov, 2007*) and would lead



205 to a liquid residual phase that is more or less buoyant than the solid phase. A  
 206 dense liquid phase could favor a basal magma ocean (*Labrosse et al.*, 2007) or  
 207 large scale cumulate overturns (*Elkins-Tanton et al.*, 2003, 2005) while a dense  
 208 solid phase would enhance volcanic activity (*Moyen and Martin*, 2012; *Martin*  
 209 *et al.*, 2014). However, the values of the partition coefficients at high pressures  
 210 of the elements that have a key contribution in the density of each phase (e.g.  
 211 Fe) are still debated (*Andrault et al.*, 2012). In addition, crystals may also  
 212 gravitationally separate with the dense cumulates sinking and the lighter ones  
 213 floating toward the surface leading to a supplementary segregation process. In  
 214 a highly turbulent environment such as a thick magma ocean, the vigor of the  
 215 convection probably prevents from any chemical segregation especially during  
 216 the early stages of the magma ocean solidification. Moreover, *Tonks and Melosh*  
 217 (1990) have shown that the Rayleigh number for a planetary scale magma ocean  
 218 was so high that crystals would remain entrained in the magma which would  
 219 effectively preclude crystal separation. Therefore, we do not consider any spa-  
 220 tial variation in the chemical composition and we currently consider neither the  
 221 mass flux of melt owing to gravitational separation nor the mass flux of melt  
 222 owing to convective mixing.

223

224 Viscosity is strongly dependent on the local melt fraction  $\phi$ , which is calcu-  
 225 lated as follows:

$$\phi = \frac{T - T_{sol}}{T_{liq} - T_{sol}}, \quad (6)$$

226 where  $T_{liq}$  and  $T_{sol}$  are the liquidus and solidus temperatures, respectively.  
 227 Following *Abe* (1997) the viscosity of partially molten silicates  $\eta$  varies between  
 228 a fully molten end-member  $\eta_m$  (assumed to be constant) and a solid end-member

229  $\eta_s$  that is temperature dependent:

$$\eta = \text{MIN} \left[ \frac{(1-\phi)\rho_m\eta_s + \phi\rho_s\eta_m}{(1-\phi)\rho_m + \phi\rho_s}, \eta_m \left\{ \frac{(1-\phi)\rho_m + \phi\rho_s}{(1-A)(1-\phi)\rho_m + \phi\rho_s} \right\}^{2.5}, 10^{21} \right], \quad (7)$$

230 where  $\rho_m$  is the density of the molten material,  $\rho_s$  is the density of the solid  
 231 material and  $A = 1.67$  (*Abe, 1997*). When the temperature,  $T$ , is lower than the  
 232 solidus temperature of mantle, the viscosity of the solid mantle  $\eta_s$  is estimated  
 233 as follows (*Abe, 1997*):

$$\eta_s = \eta_{s,0} \exp \left( B \frac{T_{liq}}{T} \right). \quad (8)$$

234 We used  $\eta_{s,0} = 256$  Pa s, and  $B = 25.17$  based on the olivine rheology (*Karato*  
 235 *and Wu, 1993; Abe, 1997*). In the above equation, as for an Arrhenius relation,  
 236 the viscosity of the solid mantle increases with the pressure. Assuming an adia-  
 237 batic temperature profile with a potential temperature of 1600K (*Tackley, 2012*)  
 238 leads to a viscosity value of  $\sim 10^{23}$  Pa.s in the lowermost mantle compatible  
 239 with estimates of the present-day mantle viscosity profiles (*Čížková et al., 2012*).  
 240 Considering a relationship that involves the solidus rather than the liquidus in  
 241 Eq. 8 would not affect our results since both the liquidus and the solidus used in  
 242 our models have a similar trend. Since we consider here the evolution of a fully  
 243 molten to partially molten magma ocean, the cooling timescale of the magma  
 244 ocean is mainly governed by  $\eta_m$  and is weakly dependent on  $\eta_s$  as we will show  
 245 later.

246 A strong increase in viscosity occurs when the melt fraction equals 40% (*Abe,*  
 247 *1997*). Hence, when the melt fraction approaches this critical value, the pres-  
 248 ence of crystals drastically reduces the efficiency of the magma ocean cooling.  
 249 In our models, the mantle is considered as part of the magma ocean as long  
 250 as its melt fraction is larger than 50% (*Neumann et al., 2014*) and we stop our  
 251 simulations when the thickness of the magma ocean drop below 100 km.

252

253 The pressure profile  $P$  is obtained fitting the PREM model (*Dziewonski and*  
254 *Anderson, 1981*) with a quadratic function of the radius  $r$  and is assumed to  
255 remain constant with time:

$$P = 4.0074 \times 10^{11} - 91862r + 0.0045483r^2. \quad (9)$$

## 256 *2.2. Melting curves and adiabat*

257 Due to the uncertainties related to the chemical composition of the magma  
258 ocean, we consider in our study two models for melting curves and elastic pa-  
259 rameters: the F-peridotitic model and the A-chondritic model. We describe these  
260 two models in the following sections.

### 261 *2.2.1. Melting curves*

262 The solidus and liquidus play a major role in the early thermal evolution of  
263 the magma ocean. Recent laboratory experiments now constrain the liquidus  
264 and solidus of mantle-like material up to pressures compatible with the CMB  
265 conditions (*Fiquet et al., 2010; Andraut et al., 2011*). However, the differences  
266 between these two studies, in particular the liquidus temperatures, appear too  
267 large to be solely due to the difference in chemical compositions between the  
268 two types of mantle materials. Regardless of the controversy, we performed cal-  
269 culations using the melting curves from both reports, leading to a F-peridotitic  
270 model (*Fiquet et al., 2010*) and a A-chondritic model (*Andraut et al., 2011*).  
271 The solidus and liquidus profiles are obtained fitting experimental results with  
272 a modified Simon and Glatzel equation (*Simon and Glatzel, 1929*). For pres-  
273 sures below  $P = 20$  GPa, we use experimentally determined solidus and liquidus  
274 temperatures of chondritic mantle from *Herzberg and Zhang (1996)*:

$$T_{sol} = 1661.2 \left( \frac{P}{1.336 \times 10^9} + 1 \right)^{(1/7.437)}, \quad (10)$$

$$T_{liq} = 1982.1 \left( \frac{P}{6.594 \times 10^9} + 1 \right)^{(1/5.374)}, \quad (11)$$

275 with  $T_{liq}$  is the liquidus temperature and  $T_{sol}$  the solidus temperature.

276

277 Since the difference between the F-peridotitic and A-chondritic solidus for  
 278 pressures larger than  $P = 20$  GPa is not large, we use the experimentally  
 279 determined A-chondritic solidus from *Andrault et al.* (2011):

$$T_{sol} = 2081.8 \left( \frac{P}{101.69 \times 10^9} + 1 \right)^{(1/1.226)}. \quad (12)$$

280 For pressures above 20 GPa, the difference between the F-peridotitic and the  
 281 A-chondritic liquidus is more important. We use the following expression

$$T_{liq} = c_1 \left( \frac{P}{c_2} + 1 \right)^{(1/c_3)}, \quad (13)$$

282 with ( $c_1 = 78.74$ ,  $c_2 = 4.054 \times 10^6$ ,  $c_3 = 2.44$ ) for F-peridotitic liquidus (*Fiquet*  
 283 *et al.*, 2010) and ( $c_1 = 2006.8$ ,  $c_2 = 34.65 \times 10^9$ ,  $c_3 = 1.844$ ) for A-chondritic  
 284 liquidus (*Andrault et al.*, 2011).

### 285 2.2.2. Thermodynamical parameters

286 The thermodynamical parameters for the molten magma ocean are closely  
 287 related to its chemical composition. Volume and elastic parameters of silicate  
 288 liquids has been recently characterized up to 140 GPa using shock compression  
 289 experiments (*Mosenfelder et al.*, 2007, 2009; *Thomas et al.*, 2012; *Thomas and*  
 290 *Asimow*, 2013). We assume here two multicomponent systems for (i) a A-  
 291 chondritic composition (62% enstatite + 24% forsterite + 8% fayalite + 4%

292 anorthite + 2% diopside) and (ii) a F-peridotitic composition (33% enstatite +  
 293 56% forsterite + 7% fayalite + 3% anorthite + 0.7% diopside). Using fourth-  
 294 order Birch-Murnaghan/Mie-Grüneisen equation of state fits for molten silicate  
 295 liquids from *Thomas and Asimow (2013)*, we obtain the melt density  $\rho_m$ , the  
 296 volumetric thermal expansion  $\alpha$  as a function of pressure as well as the specific  
 297 heat  $C_p$  of the molten material for these two multicomponent assemblages. The  
 298 density of the solid phase is then calculated as:

$$\rho_s = \rho_m + \Delta\rho, \quad (14)$$

299 with  $\Delta\rho$  the density difference between solid and liquid (see Tab. 2 for values).

300

301 For a zone of partial melting, the density  $\rho'$ , the coefficient of volumetric  
 302 thermal expansion  $\alpha'$  and the specific heat  $C'_p$  are given as follows (*Solomatov,*  
 303 2007) :

$$\frac{1}{\rho'} = \frac{1 - \phi}{\rho_s} + \frac{\phi}{\rho_m}, \quad (15)$$

$$\alpha' = \alpha + \frac{\Delta\rho}{\rho(T_{liq} - T_{sol})}, \quad (16)$$

$$C'_p = C_p + \frac{\Delta H}{T_{liq} - T_{sol}}, \quad (17)$$

304 where  $\Delta H$  is the latent heat released during solidification.

### 305 2.2.3. *Adiabats*

306 In vigorously convecting systems such as magma oceans, the temperature  
 307 distribution is nearly adiabatic and isentropic (*Solomatov, 2007*). In one-phase  
 308 systems, such as a completely molten or a completely solid layer, the equation

309 for an adiabat is

$$\left(\frac{\partial T}{\partial r}\right)_S = -\frac{\alpha g}{C_p} T. \quad (18)$$

310 In two-phase systems, the effects of phase changes need to be considered (*Solomatov*, 2007). The equation for such adiabat is given by:

$$\left(\frac{\partial T}{\partial r}\right)_S = -\frac{\alpha' g}{C_p'} T. \quad (19)$$

312 This leads to a two-phase adiabat that is steeper than the purely liquid or  
313 solid one-phase adiabats (*Solomatov*, 2007). The adiabatic temperature profiles  
314 are calculated by numerical integration of Eq. 18 and Eq. 19 using a fourth-  
315 order Runge-Kutta method (*Press et al.*, 1993). These adiabatic temperature  
316 profiles are used to calculate at each depth and when superadiabatic the temper-  
317 ature difference  $\Delta T$  from Eq. 3. The liquidus and solidus profiles as well as the  
318 adiabatic profiles obtained from Eq. 18 for temperatures ranging between 1400  
319 and 4000 K are shown in Fig. 1. We start our models assuming an adiabatic  
320 temperature profile with a potential surface temperature  $T_p = 3200$  K.

### 321 2.3. Boundary conditions

322 Large impacts can generate a rock vapor atmosphere that can last for some  
323 years until its energy is radiated to space (*Svetsov*, 2005). The presence of an  
324 atmosphere is expected to slow down the radiation of heat to space (*Hamano*  
325 *et al.*, 2013; *Lebrun et al.*, 2013). However, most of the pre-impact atmosphere is  
326 likely to be eroded after a giant impact (*Shuvalov*, 2009). Thus, in our models,  
327 we neglect the effect of thermal blanketing, and impose a radiative heat flux  
328 boundary condition at the surface:

$$F_{surf} = \sigma (T_{surf}^4 - T_{eq}^4), \quad (20)$$

329 with  $T_{surf}$  the temperature at the surface of the MO,  $\sigma$  the Stefan-Boltzman  
 330 constant and  $T_{eq} = 273$  K the expected equilibrium surface temperature.

331

332 At the base of the silicate mantle, core-mantle thermal coupling is accounted  
 333 for via a conductive heat flux imposed at the core-mantle boundary (CMB):

$$F_{core} = \frac{k_c(T^{core} - T_{CMB}^{mantle})}{e_{TBL}}, \quad (21)$$

334 where  $T^{core}$  is the average core temperature at the CMB (i.e. we neglect the  
 335 thermal boundary layer within the core) and  $T_{CMB}^{mantle}$  is the mantle temperature  
 336 right above the CMB.  $e_{TBL}$  is the thickness of the thermal boundary layer at the  
 337 bottom of the mantle where the heat is extracted from the core by conduction.  
 338  $T_{CMB}^{mantle}$  is obtained solving Eq. 1 while  $T^{core}$  is obtained from:

$$V_{core}\rho_{Fe}C_{p,Fe}\frac{dT^{core}}{dt} = S_{core}F_{core}, \quad (22)$$

339 where  $V_{core}$  is the core volume,  $S_{core}$  is the core surface,  $\rho_{Fe}$  is the core density,  
 340  $C_{p,Fe}$  is the core heat capacity and  $F_{core}$  is the heat flux through the CMB. This  
 341 formulation allows to follow the evolution of core temperature as a function of  
 342 time, based on the CMB heat flux.

343

#### 344 2.4. Numerical model

345 Eq. 1 is discretized using a semi-implicit predictor-corrector Finite Difference  
 346 scheme, of second-order in both space and time (*Press et al.*, 1993). Our scheme  
 347 was successfully benchmarked against steady and unsteady analytical solutions  
 348 for diffusion problems (*Crank*, 1975). The mantle is discretized using  $n = 2900$   
 349 equally spaced grid points resulting in a constant spatial resolution  $\Delta r = 1km$ .  
 350 Non-linear effects are handled via a fixed-point/picard iteration procedure. The

351 variable time step is determined as  $\Delta t = \min(\Delta r^2/\kappa)$ , where  $\kappa(r) = k/(\rho C_p)$  is  
352 the effective diffusivity.

### 353 **3. Results**

#### 354 *3.1. Thermal evolution of a deep magma ocean*

355 We follow the thermal evolution of deep magma ocean with an initially adi-  
356 abatic temperature profile with  $T_p = 3200$  K. Before we study the effect of each  
357 magma ocean parameter, we consider the following model as a reference case: a  
358 A-chondritic model, a 1 m thick bottom thermal boundary layer ( $e_{TBL}$ ), a melt  
359 viscosity value of  $\eta_m = 100$  Pa.s and an initial core temperature  $T_0^{core} = 5000$   
360 K. Fig. 2 shows that the temperature rapidly decreases from the surface where  
361 heat is efficiently removed by radiative cooling even if a thin solid crust is formed  
362 within this upper thermal boundary layer. In the deepest part of the mantle,  
363 the temperature profile remains adiabatic but the cooling is slower. After 5000  
364 yr, solidification occurs from the CMB where the liquidus is steeper than the  
365 adiabatic profile. As cooling proceeds, the melt fraction decreases and the last  
366 parcel with 100% melt starts to solidify in the upper mantle (see Fig. 3, left).  
367 Finally, the whole magma ocean drops below a 50 % melt value in  $t_{MO} \sim 150$   
368 kyr with  $t_{MO}$  being the magma ocean lifetime. This time ranges between the  
369 two characteristic timescales mentioned in *Solomatov (2000)*:  $10^3$  yr when crys-  
370 tallization starts from the bottom and  $10^8$  yr when crystallization of the last  
371 drop of melt occurs in the shallow magma ocean. Clearly, our magma ocean  
372 lifetime is much shorter than the cooling timescale of  $\sim 4$  Gyr proposed by  
373 *Labrosse et al. (2007)*. This is due to the fact that, in our model, the solidifi-  
374 cation occurs from the bottom-up which prevents our magma ocean from being  
375 overlaid by a thick insulating solid mantle. In *Abe (1997)*, the magma ocean  
376 was restricted to a 1000 km-deep domain and the melting curves were steeper



377 than the one used in our model. Hence, within 150 kyr, most of the mantle  
378 temperature profile computed from *Abe* (1997) is well below the solidus.

379

380 We monitored the thickness of the magma ocean (i.e. the thickness of the  
381 material having a melt fraction larger than 50 %) as a function of time. Fig.  
382 4 (left, black line) shows that after a short period ( $\sim 20$  kyr) where the man-  
383 tle remains mostly molten, the thickness of the magma ocean rapidly decreases  
384 from 2900 km to 200 km, with a change of slope for a thickness of  $\sim 2000$  km  
385 for the A-chondritic model. At the change of slope, the melt fraction reaches  
386 the critical value of 40% in the lowermost mantle, which induces an abrupt  
387 increase of its viscosity. Then the melt fraction progressively decreases at all  
388 mantle depths until we stop our simulations when the magma ocean thickness  
389 drops below 100 km. Fig. 4 (right, black line) shows the core temperature as  
390 a function of time for our reference case. In this model, the core temperature  
391 decreases monotonically from its initial value of 5000 K to a value of 4430 K,  
392 which corresponds to a melt fraction  $\phi \sim 40\%$  at  $P=140$  GPa (i.e. when the  
393 abrupt change in viscosity occurs).

394

395 We monitored the time evolution of the heat flow coming at the CMB from  
396 the core  $F_{core}$  and the heat loss at the surface  $F_{surf}$  (Fig. 5, a). During the  
397 first 20 kyr, the heat flow at the CMB rapidly increases from  $\sim 10^{16}$  to a nearly  
398 constant value of  $\sim 10^{17}$  W. In the meantime the surface heat flow decreases  
399 from  $\sim 10^{19}$  (in agreement with the value proposed by *Solomatov* (2000)) to  
400  $\sim 10^{17}$  W. Then, both the surface and the core heat flows decrease down to a  
401 value of  $\sim 10^{16}$  W until the end of the magma ocean stage (within  $t_{MO} = 150$   
402 kyr).

403

404 *3.2. Influence initial of core temperature*

405 We monitored the influence of the initial core temperature on the cooling  
406 of the deep magma ocean considering two values: (1)  $T_0^{core} = 5000$  K, which  
407 is equal to the temperature at the bottom of the magma ocean for  $T_p = 3200$   
408 K in the A-chondritic model and (2)  $T_0^{core} = 7000$  K, which corresponds to  
409 a core significantly hotter than the mantle. The comparison of the black and  
410 green curves in Fig. 4 (left) shows that the initial temperature has a negligible  
411 influence on the evolution of the magma ocean thickness as well as on its cooling  
412 timescale. When  $T_0^{core} = 7000$  K, the initial core heat flow is large ( $\sim 10^{18}$  W)  
413 (Fig. 5, a) because of the initial temperature contrast (= 2000 K) with the  
414 lowermost mantle. Then the heat flow decreases as a consequence of both the  
415 progressive solidification of the overlying mantle and the core heat depletion.  
416 When  $T_0^{core} = 5000$  K, the initial CMB heat flow is much smaller than when  
417  $T_0^{core} = 7000$  K, but it increases rapidly due to rapid cooling of the lowermost  
418 mantle. In both cases, the heat flow decreases when the lowermost mantle has  
419 cooled sufficiently to reach the critical melt fraction of 40%. A significantly  
420 hotter initial core leads to an increase of the core heat flow, by a factor of 3  
421 between 5000 K and 7000 K (Fig. 5, a). However, this increase of the initial  
422 core temperature, which should lead to an increase of the surface heat flow, is  
423 not visible on the surface heat flow evolution because of the stronger efficiency  
424 of the surface cooling.

425

426 Also, a core initially 2000 K hotter than the lowermost mantle ends up 170  
427 K hotter at the end of the MO stage (Fig. 4, right). When  $T_0^{core} = 7000$  K  
428 the final core temperature is  $T^{core} = 4600$  K which is slightly larger than the  
429 temperature at which the melt fraction of the lowermost mantle reaches the  
430 40% critical value in the A-chondritic model.

432 *3.3. A-chondritic vs. F-peridotitic model*

433 Here we compare the evolutions of temperature and melt fraction between  
 434 the A-chondritic and F-peridotitic models (using the corresponding liquidus and  
 435 thermodynamical parameters). We focus on cases where  $T_0^{core} = 7000$  K (with  
 436  $T_0^{core} = 5000$  K, the core temperature would be lower than the liquidus for the  
 437 F-peridotitic model at the CMB pressure). Fig. 4 (left) shows that for an initial  
 438 core temperature  $T_0^{core} = 7000$  K, the magma ocean thickness decreases more  
 439 rapidly in the F-peridotitic model (red curve) than in the A-chondritic model  
 440 (green curve). This is the direct consequence of a liquidus being significantly  
 441 higher for a F-peridotitic model than for a A-chondritic model (Fig. 1). Hence,  
 442 during the cooling of an initial fully molten magma ocean, the onset of mantle  
 443 crystallization occurs earlier and the melt fraction decreases more rapidly in the  
 444 F-peridotitic case (Fig. 3, right). The comparison of the A-chondritic and the  
 445 F-peridotitic models shows a peak of the melt fraction in the latter case occur-  
 446 ring at a depth of  $\sim 600$  km. This corresponds to the important slope change  
 447 in the F-peridotitic liquidus that occurs at 20 GPa (see Fig. 1, right). Since the  
 448 melt fraction is a strong function of the liquidus, this discontinuity happens to  
 449 affect the precise depth at which the last drop of melt should solidify. However,  
 450 the last part to solidify should still be located in the shallow mantle, regardless  
 451 the shape of the liquidus in this region. In addition, the magma ocean life-  
 452 time  $t_{MO}$  appears to be weakly affected by the choice of the model and ranges  
 453 between 147 kyr (in the A-chondritic case) to 171 kyr (in the F-peridotitic case).

454

455 Still, the F-peridotitic and A-chondritic models show a significant difference  
 456 on the evolution of the core temperature (Fig. 4, right). For both cases, the  
 457 core temperature decreases asymptotically from its initial value to a value that

458 ranges between 4600 K (A-chondritic case) and 4860 K (F-peridotitic) in about  
459 0.15 Myr. Since the lowermost mantle solidifies more rapidly when considering  
460 a hotter liquidus, a F-peridotitic model for the magma ocean helps to retain  
461 some heat in the core. For both cases, the final core temperature is 100 to  
462 150 K larger than the temperature at which the melt fraction of the lowermost  
463 mantle reaches the 40% critical value.

464

#### 465 3.4. Influence of the bottom thermal boundary layer

466 The thickness of the bottom thermal boundary layer  $e_{TBL}$  governs both the  
467 cooling rate of the core and the energy supplied to the magma ocean (Eq. 21).  
468 In a hard-turbulent context, this thickness is difficult to determine by analogical  
469 models and only theoretical models can constrain this parameter (*Spiegel, 1971*).  
470 Therefore, a rough estimation of  $e_{TBL}$  can be obtained for  $Ra \sim 10^{20}$  based on  
471 the Nusselt number calculation:  $Nu \sim (RaPr)^{1/2}$  (*Spiegel, 1971*). Considering  
472 that  $Nu \sim L/e_{TBL}$ , we obtain  $e_{TBL} \sim L(RaPr)^{-1/2}$ . In our magma ocean  
473 context and assuming that the scaling law is still valid at Rayleigh numbers  
474 up to  $10^{30}$ , this corresponds to a value  $e_{TBL} \sim 10^{-6}$  m. This thickness is ex-  
475 tremely thin, however, it is compatible with previous reports (*Solomatov, 2007*;  
476 *Lebrun et al., 2013*). As a first attempt to understand the influence of bottom  
477 thermal boundary layer in the magma ocean cooling dynamics, we performed  
478 various calculations with  $e_{TBL}$  fixed and ranging from 1 mm to 1 km. The  
479 lower range values can be seen as unrealistic in comparison with, for example,  
480 the core topography. However, the wide range of values considered for  $e_{TBL}$  al-  
481 lows a better illustration of its influence on the magma ocean cooling timescales.

482

483 We find that the magma ocean lifetime does not depend largely on  $e_{TBL}$  and  
484 its value remains close to 150 kyr for the whole range of  $e_{TBL}$  values assumed

485 here. Fig. 5 shows the CMB and the surface heat flows for four different values  
 486 of  $e_{TBL}$ . For  $e_{TBL} = 1000$  m (Fig. 5, b), the surface heat flow is several orders  
 487 of magnitude larger than the core heat flow and the thermal coupling between  
 488 these two reservoirs is inefficient. For values of thermal boundary layer thick-  
 489 nesses below 1 m (Fig. 5, c and d), the heat flows become comparable and the  
 490 thermal coupling between the core and the molten overlying mantle becomes  
 491 efficient. Ideally,  $e_{TBL}$  should be an adjustable parameter in our calculation,  
 492 related to the effective value of the Rayleigh number. However, this would result  
 493 in extremely small  $e_{TBL}$  values of the order of  $10^{-6}$  m. We show in Fig. 5 that  
 494 decreasing the value of  $e_{TBL}$  from 100 mm to a value of 1 mm does influence  
 495 neither the shape of the surface heat flow nor the magma ocean lifetime (Fig.  
 496 5, c and d). In that cases, the core heat flow rapidly reaches a maximal value  
 497 and decreases within the first 20 kyr, which corresponds to the time needed by  
 498 the melt fraction of the bottom of the magma ocean to reach the critical value  
 499 of 40% (Fig. 6). During this brief period of time, although heat is efficiently re-  
 500 moved from the core (Fig. 6), the CMB heat flow remains considerably smaller  
 501 than the surface heat loss. This is the reason why  $e_{TBL}$  does not significantly  
 502 influence the magma ocean lifetime. Hence, decreasing  $e_{TBL}$  to a smaller value  
 503 (i.e.  $e_{TBL} \ll 10^{-6}$  m) as suggested by our theoretical estimation of the TBL  
 504 thickness or to relate  $e_{TBL}$  to the Rayleigh number that is initially extremely  
 505 high should not influence the results of our study.

506

507 Such a behavior is confirmed by the time evolution of  $T^{core}$ , which is a strong  
 508 function of  $e_{TBL}$  (Fig. 6). As long as  $e_{TBL}$  is larger than 100 m, the initial core  
 509 heat is efficiently retained and the core cooling is not influenced by the cooling of  
 510 the overlying magma ocean. However, for  $e_{TBL} \leq 100$  m the thermal coupling  
 511 between the core and the MO becomes important. For  $e_{TBL} < 1$  m, the core

512 rapidly cools down to  $\sim 4400$  K which corresponds to the temperature where  
 513 the lowermost mantle reaches the critical melt fraction value of 40%. Then the  
 514 core cooling efficiency strongly decreases as the lowermost mantle is becoming  
 515 much more viscous. Finally, the core temperature ends up at a temperature of  
 516  $\sim 4370$  K for  $e_{TBL}$  values ranging between 1 mm to 10 cm.

517

### 518 *3.5. Influence of the magma ocean viscosity*

519 Measurements (*Liebske et al.*, 2005) and ab initio calculations (*Karki and*  
 520 *Stixrude*, 2010) estimate that the dynamic viscosity  $\eta_m$  of peridotitic melt is in  
 521 the range  $10^{-2} - 10^{-1}$  Pa.s. At low degrees of partial melting of a peridotite, the  
 522 viscosity of the generated liquid can eventually increase up to 100 Pa.s (*Kushiro*,  
 523 1986). The viscosity of molten mafic silicate should range between  $10^{-2} - 10^2$   
 524 Pa.s (*Rubie et al.*, 2003). To take into account the effect of this uncertainty on  
 525 the magma ocean lifetime, we perform numerical simulations considering that  
 526 the fully molten magma ocean viscosity  $\eta_m$  ranges between  $10^{-2} - 10^2$  Pa.s.

527

528 Dimensional analysis of Eq. 1 indicates that  $t_{MO}$  is inversely proportional  
 529 to  $F_{conv}$ . In the hard turbulent regime relevant to a thick MO context this term  
 530 scales as  $\eta_m^{-3/7}$ . Consequently, the lifetime of a magma ocean should scale as  
 531  $\eta_m^{3/7}$ . This is confirmed by our numerical results (Fig. 7), and consistent with  
 532 previous work (*Solomatov*, 2007):

$$t_{MO}(\text{Myr}) = 0.018 \eta_m^{3/7}. \quad (23)$$

533 Most importantly, for realistic viscosities of the fully molten early mantle, the  
 534 melt fraction drops below 50% at all mantle depths in less than 1 Myr. For the  
 535 lower range of  $\eta_m$ , this characteristic time scale can decrease down to several

536 kyrs, rather than 1 Myr.

#### 537 **4. Conclusion**

538 The cooling of a thick terrestrial magma ocean is a fast process. The magma  
539 ocean lifetime is principally governed by its viscosity and, for the mid range of  
540 realistic viscosities, the magma ocean reaches a 50% melt fraction at all mantle  
541 depths within 20 kyr. Depending on the thermal boundary layer at the CMB,  
542 the thermal coupling between the core and MO can either insulate the core  
543 during the MO solidification and favor a hot core (for thick TBL), or drain the  
544 heat out of the core. However, as suggested by theoretical calculations, an ex-  
545 tremely thin thermal boundary layer in an ultra-turbulent environment implies  
546 that much of the heat is removed from the core during the cooling of the over-  
547 lying mantle. In this context, our F-peridotitic model (with a higher liquidus)  
548 is more willing to retain the core heat than our A-chondritic model, yielding a  
549 difference of temperature of  $\sim 170$  K after the magma ocean has crystallized.  
550 In addition, the final core temperature increases by a couple hundred degrees as  
551 its early temperature increases. However, for all cases, the average core temper-  
552 ature at the CMB ( $T^{core}$ ) ends up close to the 40% melt fraction temperature  
553 of the silicate magma ocean:  $\pm 100$  K depending on the initial core temperature  
554 and on the thickness of the bottom thermal boundary layer.

555  
556 In contrast with previous reports (*Labrosse et al., 2007*), our model shows  
557 that the crystallization occurs relatively rapidly at the CMB and, after some  
558 crystallization has proceeded, the highest amount of partial melting is found at  
559 intermediate depth between the surface and the CMB (see Fig. 3). One could  
560 argue that this result is dictated by the fact that our calculation neglects the  
561 possible effects of chemical segregation during mantle cooling. On the contrary,

562 we believe that the segregation of a melt above the core mantle boundary would  
563 not help to retain heat in the core (*Labrosse, 2015; Davies et al., 2015*). A melt  
564 that would accumulate just above the CMB by gravitational segregation would  
565 be depleted in refractory elements, thus with a liquidus lower than the rest of  
566 the mantle. The temperature at which this part of the mantle becomes viscous  
567 (at 40% of partial melting) would be lowered and thus the core heat would es-  
568 cape more easily in the presence of a basal magma ocean (*Ulvrová et al., 2012;*  
569 *Nakagawa and Tackley, 2014*), in agreement with our results displayed in Figure  
570 4.

571

572 These results have important consequences for the magnetic history of the  
573 Earth. Indeed, if at some point, a full magma ocean has existed on the Earth,  
574 it is likely that most of the core heat has been removed rapidly. In less than  
575 1Myr, a tremendous heat flow may have lead to a significant decrease of the core  
576 temperature until it reached a value that is close or slightly above the tempera-  
577 ture at which the melt fraction of the lowermost mantle reaches the 40% critical  
578 value (i.e.  $\sim 4400$  K). Assuming that the Earth-Moon system was formed by  
579 a giant impact 100 Myrs after the first solids of the Solar System (*Kleine and*  
580 *Rudge, 2011*), and that this giant impact has completely molten the Earth's  
581 mantle (*Nakajima and Stevenson, 2015*), it is difficult to envision that a large  
582 amount of heat could be retained in the core to sustain the geomagnetic field  
583 by thermal convection for several Gyr (*Andrault et al., 2016*), in contrast with  
584 the proposition of the most recent reports (*Labrosse, 2015; Davies et al., 2015*).

585

586 Finally, we acknowledge that our current model neglects the effect of vertical  
587 chemical segregation. While this effect is unlikely to dominate the dynamics in  
588 a highly turbulent magma ocean (*Tonks and Melosh, 1990*), it could become



589 more important when the degree of partial melting becomes close or lower than  
590  $\sim 40\%$  (i.e. when viscosity increases). At this point, the knowledge of the den-  
591 sity contrast between the solid at the liquidus (the first crystal to form) and  
592 the ambient liquid becomes of major importance. Whether the melt sinks, or  
593 floats has important ramifications for understanding the first steps in the dy-  
594 namic modeling of the Earth's differentiation. In the near future, a modeling  
595 effort to integrate the compositional contribution in the buoyancy calculation  
596 between liquid and solid will constitute an important step forward towards the  
597 understanding of the earliest stages of Earth's evolution.

598

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Table 1: Constant and fixed parameter values for numerical models

Earth radius	$R$	6370 km
Core radius	$R_{core}$	3470 km
Mantle thickness	$R - R_{core}$	2900 km
<i>Mantle properties</i>		
Solid density	$\rho_s$	$= \rho_m + \Delta\rho$
Density contrast	$\Delta\rho/\rho$	1.5% ( <i>Tosi et al.</i> , 2013)
Specific enthalpy change	$\Delta H$	$4 \times 10^5$ J/kg ( <i>Ghosh and McSween</i> , 1998)
Viscosity of melt phase	$\eta_m$	$1 - 10^4$ Pa.s
Bottom TBL thickness	$e_{TBL}$	$10^{-3} - 10^3$ m
<i>Core properties</i>		
Density	$\rho_{Fe}$	$10\,000$ kg.m <sup>-3</sup>
Heat capacity	$C_{p,Fe}$	$800$ J.kg <sup>-1</sup> .K <sup>-1</sup>

Table 2: Variable and non-dimensional parameter values for numerical models

Melt density	$\rho_m$	A-model: 2684 – 5274 kg.m <sup>-3</sup> F-model: 2679 – 5378 kg.m <sup>-3</sup>	computed from <i>Thomas and Asimow (2013)</i>
Heat capacity	$C_p$	A-model: 1742 F-model: 1800	computed from <i>Thomas and Asimow (2013)</i>
Thermal expansion coefficient	$\alpha$	A-model: $1.3 \times 10^{-5}$ - $7.9 \times 10^{-5}$ F-model: $2 \times 10^{-5}$ - $9.6 \times 10^{-5}$	computed from <i>Thomas and Asimow (2013)</i>
Viscosity of solid phase	$\eta_s$		from Eq. 8 with $\eta_{s,0} = 256$ Pa.s and B=25.17
Viscosity of the magma ocean	$\eta$	$1 - 10^{21}$	from Eq. 7
Total conductivity	$k$	$5 - 10^7$	$= k_c + k_v$
Rayleigh number	$Ra$	at $t = 0$ : $1 \times 10^{27}$ – $3 \times 10^{27}$	computed from Eq. 3
Prandtl number	$Pr$	$350 - 3.6 \times 10^{24}$	$= C_p \eta / k_c$
Reynolds number	$Re$	at $t = 0$ : $Re \sim 10^9$	from <i>Solomatov (2007)</i>

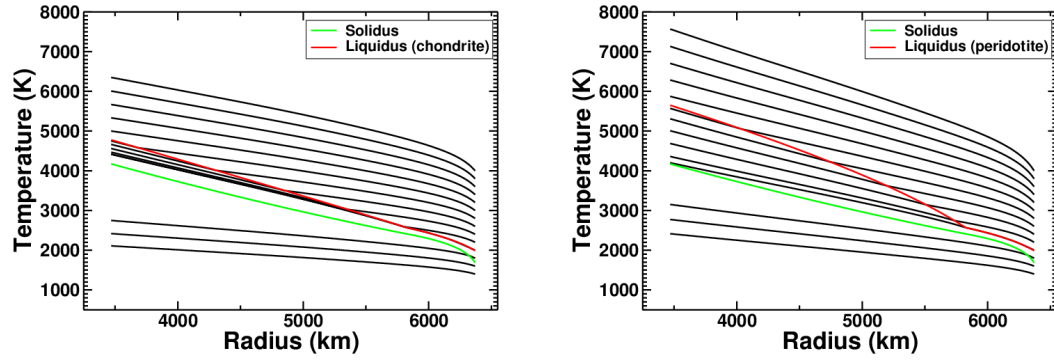


Figure 1: Adiabats (with  $T_p$  ranging between 1400 and 4000 K) computed for the A-chondritic model (left) and the F-peridotitic model (right). The corresponding solidus and liquidus are represented in green and red respectively.

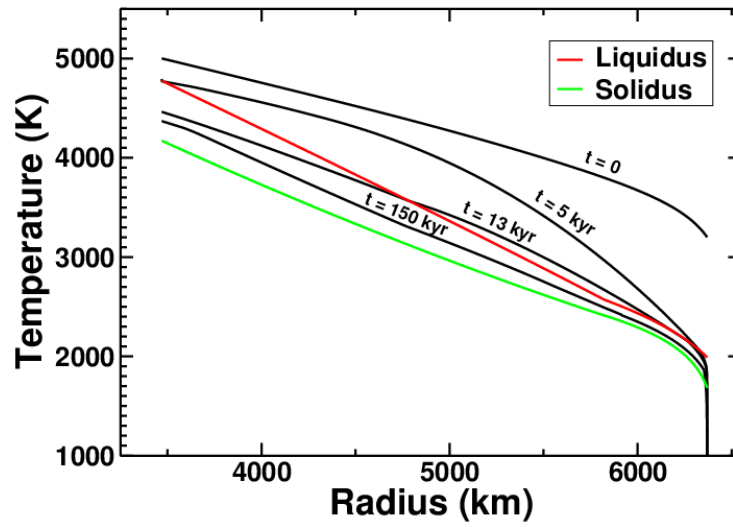


Figure 2: Temperature evolution from an initially adiabatic temperature profile with  $T_p = 3200$  K and  $T_0^{core} = 5000$  K. The liquidus and solidus used in our models are those obtained for the A-chondritic model and are represented respectively with red and green curves. In this model  $e_{TBL} = 1$  m and  $\eta_m = 100$  Pa.s.

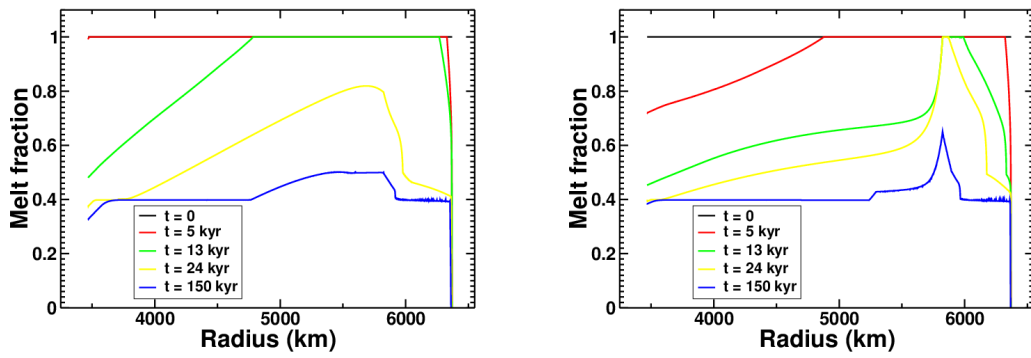


Figure 3: **Left panel:** Melt fraction evolution from an initially completely molten magma ocean and corresponding to the case illustrated in Fig. 2. A melt fraction of 0.4 is a major discontinuity for the magma ocean viscosity (see text). **Right panel:** same with a F-peridotitic model and  $T_0^{core} = 7000$  K

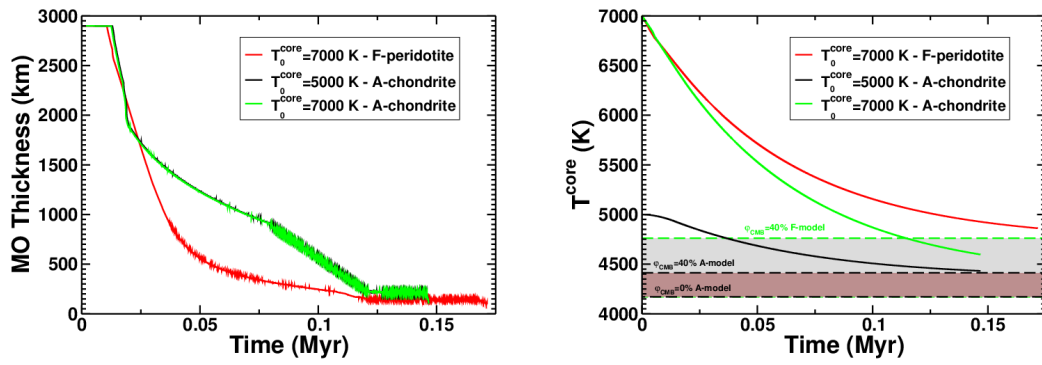


Figure 4: **Left panel** Time evolution of the magma ocean thickness (where the melt fraction is larger than 50%) for different initial core temperatures  $T_0^{core}$  and different initial compositions (with  $e_{TBL} = 1$  m and  $\eta_m = 100$  Pa.s). **Right panel**: Time evolution of the core temperature for different initial core temperatures and different initial compositions (with  $e_{TBL} = 1$  m,  $\eta_m = 100$  Pa.s).

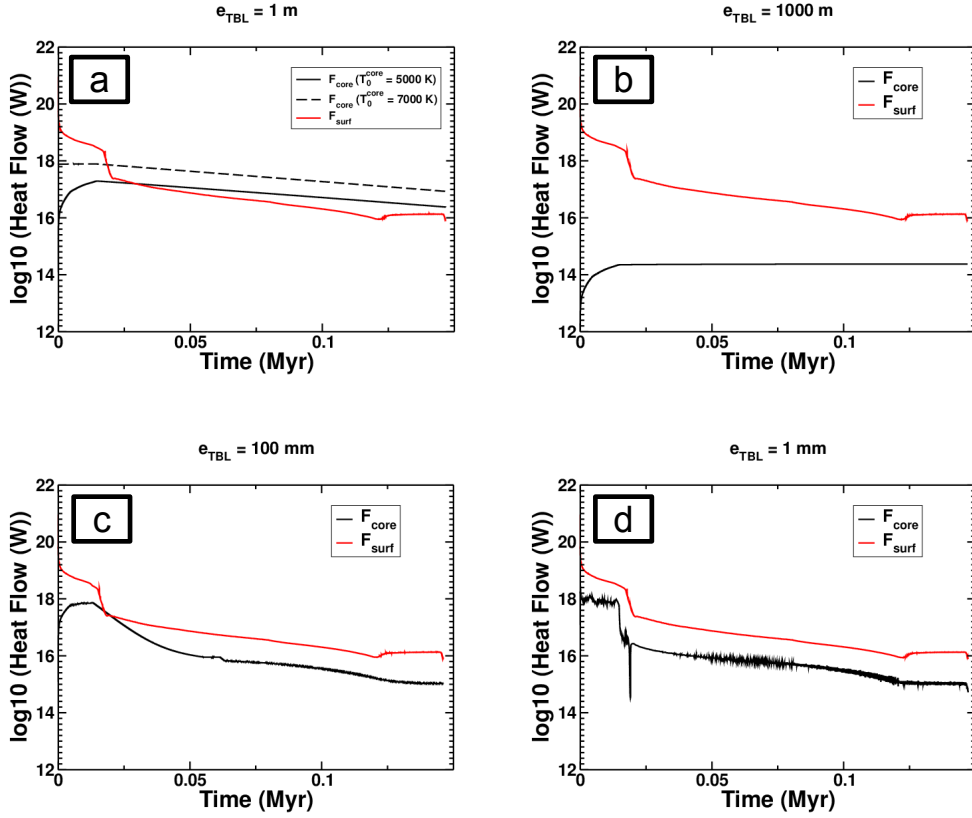


Figure 5: (a): Time evolution of the CMB (black) and surface (red) heat flows for a A-chondritic model (assuming  $e_{TBL} = 1 \text{ m}$ ,  $\eta_m = 100 \text{ Pa.s}$  and  $T_0^{\text{core}} = 5000 \text{ K}$  except for the black dashed line where  $T_0^{\text{core}} = 7000 \text{ K}$ ). (b): Same with  $e_{TBL} = 1000 \text{ m}$ . (c): Same with  $e_{TBL} = 100 \text{ mm}$ . (d): Same with  $e_{TBL} = 1 \text{ mm}$ . In all these models, the magma ocean lifetimes (time at which the plots ends up) are very close



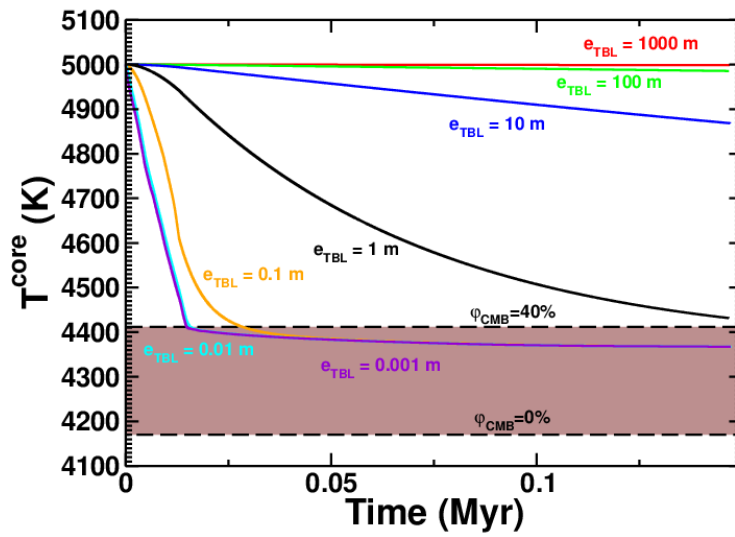


Figure 6: Time evolution of the core temperature for different thicknesses of the bottom thermal boundary layer (assuming a A-chondritic model,  $\eta_m = 100$  Pa.s and  $T_0^{core} = 5000$  K).

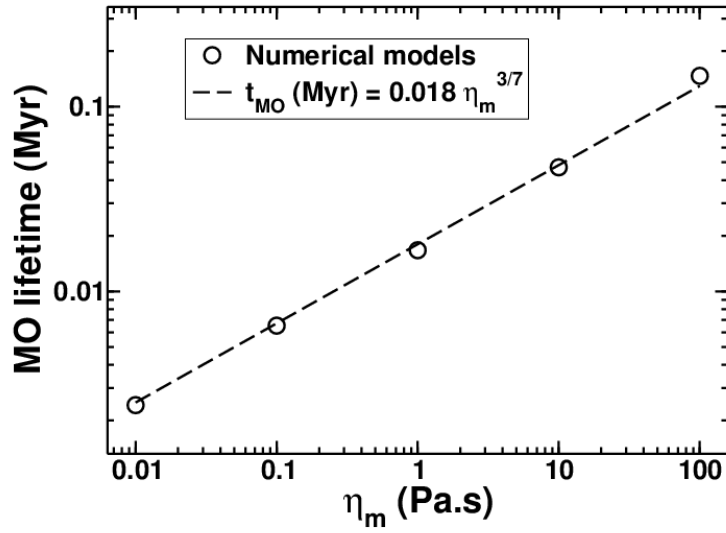


Figure 7: Magma ocean lifetime as a function of the magma ocean viscosity  $\eta_m$  (assuming a A-chondritic model,  $e_{TBL} = 1$  m and  $T_0^{core} = 5000$  K).