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## The deep Earth may not be cooling down

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#### 9 Abstract

10 The Earth is a thermal engine generating the fundamental processes of geomagnetic field, plate 11 tectonics and volcanism. Large amounts of heat are permanently lost at the surface yielding the classic 12 view of the deep Earth continuously cooling down. Contrary to this conventional depiction, we 13 propose that the temperature profile in the deep Earth has remained almost constant for the last ~4.3 14 billion years. The core-mantle boundary (CMB) has reached a temperature of ~4400 K in probably 15 less than 1 million year after the Moon-forming impact, regardless the initial core temperature. This 16 temperature corresponds to an abrupt increase in mantle viscosity atop the CMB, when ~60% of 17 partial crystallization is achieved, accompanied with a major decrease in heat flow at the CMB. Then, 18 the deep Earth underwent a very slow cooling until it reached ~4100 K today. This temperature at, or 19 just below, the mantle solidus is suggested by seismological evidence of ultra-low velocity zones in 20 the D"-layer. Such a ~steady thermal state of the CMB temperature excludes thermal buoyancy from 21 being the predominant mechanism to power the geodynamo over geological time.

22 An alternative mechanism to sustain the geodynamo is mechanical forcing by tidal distortion and 23 planetary precession. Motions in the outer core are generated by the conversion of gravitational and 24 rotational energies of the Earth-Moon-Sun system. Mechanical forcing remains efficient to drive the 25 geodynamo even for a sub-adiabatic temperature gradient in the outer core. Our thermal model of the 26 deep Earth is compatible with an average CMB heat flow of 3.0 to 4.7 TW. Furthermore, the regime 27 of core instabilities and/or secular changes in the astronomical forces could have supplied the 28 lowermost mantle with a heat source of variable intensity through geological time. Episodic release 29 of large amounts of heat could have remelted the lowermost mantle, thereby inducing the dramatic 30 volcanic events that occurred during the Earth's history. In this scenario, because the Moon is a 31 necessary ingredient to sustain the magnetic field, the habitability on Earth appears to require the 32 existence of a large-satellite.

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35	
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37	geomagnetic field.
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39	
40	<u>Highlights</u> :
41	- We review the early and the present thermal states of the deep Earth
42	- We highlight the paradoxes implied by the classical scenario based on secular cooling
43	- We propose a 4.3 Gy steady thermal state scenario for the lowermost mantle
44	- We suggest the Moon as a necessary ingredient to sustain the Earth's magnetic field
45	
46	
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### 52 **1. Introduction**

53 Our knowledge of the present-day thermal state of the deep Earth has largely improved based on comparisons between seismic observations and experimental and theoretical characterisations of the 54 55 Earth's materials, including studies of phase transitions and melting curves. In addition, we now have 56 a more precise idea of the early evolution of the deep Earth based on paleomagnetic and geochemistry 57 data as well as numerical modelling. However, the path the deep Earth has followed from its early 58 formation to its current state remains puzzling. In particular, the Earth's heat budget over the  $\sim 4.5$ 59 billion years (Gy) of its existence remains difficult to balance satisfactorily. In this article, we 60 challenge the classical view that the Earth is continuously cooling. We propose an alternative model 61 for the thermal and magnetic evolution of the deep Earth. First, we review the early and the present 62 thermal states of the deep Earth (sections 2 and 3), then we highlight the major paradoxes implied by 63 the classical scenario based on secular cooling (section 4). Finally, we propose a steady thermal state 64 scenario (section 5) involving an alternative source of energy to power the geodynamo: precession 65 and tides (sections 6 and 7). The implications of this scenario are presented in the final section.

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#### 67 **2.** The early temperature profile in the deep Earth

### 68 2.1 The primordial core

69 The initial core temperature is related to a variety of processes but primarily results from the 70 mechanism of core-mantle segregation (Stevenson, 1990). Metal/silicate separation is a rapid event (<60 My) contemporaneous with the Earth's accretion from many planetary embryos (Kleine et al., 71 72 2002; Rudge et al., 2010). After a meteoritic impact, the Fe-droplets produced by fragmentation of the impactor's core descended and equilibrated thermally in a highly turbulent magma ocean (Deguen 73 74 et al., 2011; Samuel, 2012; Wacheul et al., 2014). The resulting iron layer/pond that formed at the 75 bottom of the magma ocean then descended through the underlying, more viscous mantle in the form 76 of diapirs or, alternatively, through channels (Stevenson, 1990). In both cases, the corresponding 77 gravitational potential energy released was dissipated into heat, but the heat partitioning between the 78 iron and the silicate depends on the segregation mechanism (Rubie et al., 2015). The diapir 79 mechanism tends to favor heat transfer to the viscous mantle (Monteux et al., 2009; Samuel et al., 80 2010), while channels favor a hotter core (Ke and Solomatov, 2009), yielding a wide range of 81 plausible thermal states at the core-mantle boundary. If metal-silicate thermal equilibration had been 82 efficient, the initial core temperature would be similar to that of the lowermost mantle. Taking into 83 consideration the contributions of heating from large impacts and the decay of short-lived 84 radionuclides, the early core was likely to be hotter than the mantle liquidus soon after its formation 85 (Rubie et al., 2015).

## 87 *2.2 From the magma ocean to the mantle*

88 The giant Moon forming impact (MFI) that occurred ~60 million years (My) after the Earth's 89 formation (Touboul et al., 2007), likely re-melted the entire mantle (Nakajima and Stevenson, 2015) 90 and significantly heated up the Earth's core (Herzberg et al., 2010) (Fig. 1a). The MFI, through the 91 release of energy induced by gravitational segregation of the impactor's core, could have potentially 92 increased the core temperature further by 3500-4000 K (Rubie et al., 2015). This could have resulted 93 in an initial CMB temperature on the order of 6000 K (Nakagawa and Tackley, 2010), which is well 94 above the mantle solidus of ~4150 K (Andrault et al., 2011; Figuet et al., 2010). Therefore, the MFI 95 could have caused intensive melting in the lowermost mantle.

96 Upon cooling of this giant-impact induced magma ocean, a thin crust rapidly formed at the 97 surface, within the cold upper thermal boundary layer (Solomatov, 2015) (Fig. 1b). Just below, the 98 magma ocean is expected to have convected vigorously. Hence, its internal temperature would follow 99 an adiabatic profile undergoing a progressive decrease in potential surface temperature with time (Abe, 1997; Solomatov, 2000). Due to the fact that the liquidus and solidus curves present P-T slopes 100 101 steeper than the magma ocean adiabats for the chondritic-type composition (Andrault et al., 2011; 102 Thomas and Asimow, 2013), the magma ocean should solidify from the bottom up (Fig. 1b). The 103 heat flux at the surface could have been as high as  $\sim 10^6$  W/m<sup>2</sup>, which suggests crystallization of most of the magma ocean within  $\sim 10^3$  years (Solomatov, 2015). However, the situation may have been 104 105 complicated by physical and chemical processes such as suspension, turbulence, nucleation, and 106 percolation or by the formation of an opaque atmosphere at the Earth's surface. These effects could 107 delay the complete crystallization of the upper mantle up to  $10^8$  years after the magma ocean began cooling (e.g. after the MFI) (Fig. 1c and 1d) (Lebrun et al., 2013; Sleep et al., 2014). 108

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## 110 **3.** The present-day temperature profile in the deep Earth

## 111 *3.1 Upper mantle, transition zone and the lower mantle*

112 The temperature profile from the shallow mantle to a few hundred kilometers above the CMB is 113 relatively well documented. The most robust constraints originate from the phase transformations at 114 the 410, 520 and 660 km discontinuities; the depth of the phase transformations must be compatible with the phase diagram of the major upper mantle minerals, mainly olivine, which is well constrained 115 116 experimentally. When including the effect of entropy variations between the different polymorphs, a 117 temperature discontinuity of a few hundred degrees is induced at the seismic discontinuities (Katsura 118 et al., 2010; Stacey and Davis, 2008). Complications in the temperature determination using the 119 olivine phase diagram may arise from the uncertainties in mantle concentrations of FeO, water and 120 ferric iron (e.g. (Frost and Dolejs, 2007)), because they modify the pressure of the phase transitions slightly. Lateral temperature variations are also expected from colder temperatures in subduction 121

122 zones to hotter temperatures in regions of upwelling mantle (ocean ridges, for example). Altogether, 123 the uncertainty is less than a couple of hundred degrees. Then, the temperature profile in the lower 124 mantle is classically extrapolated from anchor points in the transition zone using an adiabatic gradient, 125 which yields additional uncertainties. Slightly different temperature profiles can be obtained, 126 depending on the equations of states used for the mantle (Brown and Shankland, 1981; Stacey and 127 Davis, 2008). Other predictions give a significantly higher temperature profile when refining the seismic profiles (Vp, Vs,  $\rho$ ) from the mineral equations of states ( $\rho$ , K, G) (Matas et al., 2007) (Fig. 128 129 2a).

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#### 131 *3.2 The lowermost mantle*

132 The thermal state of the lowermost mantle is not directly correlated to the surface potential 133 temperature, but is rather tied to the temperature of the core and the heat flux at the CMB. The seismic observations of thermochemical heterogeneities and partial melting in the D"-region provide 134 135 additional information to anchor the CMB temperature (Herzberg et al., 2013; Rost et al., 2005; Wen 136 and Helmberger, 1998). It has been argued that the thermochemical piles present in this region could 137 be interpreted as patches of post-bridgmanite (PBg) embedded in bridgmanite (Bg), which would require a double crossing of the Bg-PBg phase transition (Hernlund et al., 2005). Using the P-T 138 139 Clapeyron slope of the polymorphic transition, this situation is possible if a sharp temperature change occurs when approaching the CMB, for a CMB temperature higher than 4000 K. This method based 140 141 on the phase diagram of an *unrealistically* pure MgSiO<sub>3</sub> end-member has been subsequently 142 challenged by experiments performed on the geophysically relevant Al-bearing (Mg,Fe)SiO<sub>3</sub> Bg 143 (Andrault et al., 2010; Catalli et al., 2009). Still, the argument for a double crossing may stand for a 144 partial and progressive transition in Bg (Hernlund, 2010).

145 The CMB temperature can also be constrained using the melting curve of the silicate mantle. The non-ubiquitous character of the seismic features in the D"-layer forces the CMB temperature to be 146 147 lower than the mantle solidus. Otherwise, there would be a continuous melting line below which the 148 mantle would be partially molten, due to higher temperatures in the thermal boundary layer when 149 approaching the CMB. The solidus of chondritic-type material plots at 4150 (+/-150) K at a CMB 150 pressure of 135 GPa (Andrault et al., 2011), very close to that of the peridotitic-type mantle (Figuet 151 et al., 2010). This solidus temperature should actually remain valid for any reasonable mineralogical system composed of Bg, CaSiO<sub>3</sub>-perovskite and (Mg,Fe)O-ferropericlase, because of the pseudo-152 153 eutectic behavior. In contrast, the solidus temperature could be lowered in the presence of a high FeO-154 content (Mao et al., 2005), high volatile contents (Nomura et al., 2014) or when the excess mantle 155 ferropericlase is replaced by an excess SiO<sub>2</sub>, *e.g.*, for a basaltic composition (Andrault et al., 2014). 156 Water can have a dramatic effect, lowering the solidus temperature to ~3570 K, but it is unlikely that

the lower mantle contains a very high water content (Bolfan-Casanova et al., 2003). In contrast, the descent of slabs toward the CMB is clearly imaged by seismic tomography (Grand et al., 1997), and slabs may very well reach the CMB. The solidus temperature of a mid-ocean ridge basalt at the CMB was reported to be 3800 (+/-150) K (Andrault et al., 2014), which suggests that a CMB temperature of 4000 (+/-200) K would produce non-ubiquitous regions of partial melt, in agreement with seismic observations (Fig. 2a).

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#### *3.3 The core*

165 The melting curve of pure iron was a long-running source of controversy until recent 166 experimental measurements using laser-heated diamond anvil cells fell in perfect agreement with 167 shock-wave data and ab-initio calculations (Anzellini et al., 2013) (Fig. 2b). The originality of this 168 experiment is fast heating, to prevent the sample pollution from C diffusing out of the diamond anvils, 169 together with the *in situ* detection of sample melting using X-ray diffraction. It suggests a melting 170 temperature of pure Fe at the inner core boundary (ICB) of 6230 (+/- 500) K. The light elements 171 present in the outer core at a level of 10 wt% should lower this melting point. The melting-temperature 172 depletion induced by the presence of S, O and Si are 100, 50 or 30 K/wt%, respectively. Unfortunately, the nature and combination of light elements in the outer core remain subject to debate. 173 174 A reasonable composition could be 2.5, 5.0 and 5.0 wt% of S, O and Si, respectively, in agreement with the geochemical constraints (Dreibus and Palme, 1996), the density jump at the ICB (Alfè et al., 175 176 2002) and the seismic profiles (Vp,  $\rho$ ) in the outer core (Morard et al., 2013). For this Fe-allov 177 composition, the melting-temperature depletion can be estimated to 650 (+/- 100) K. This yields an 178 ICB temperature of 5580 (+/-600) K.

179 When this anchor point is extrapolated to the CMB using the equation of state of Fe, it yields a 180 CMB temperature of 4100 K, if we assume a constant Grüneisen parameter of 1.51 (Vocadlo et al., 181 2003). The relative changes in the cocktail of light elements would not drastically change this 182 extrapolated CMB temperature. We note that regardless of the thermal model considered for the core, 183 the heat flux at the CMB remains moderate. Thermal boundary layer(s) with a large temperature jump 184 are unlikely to develop inside the liquid outer core. Therefore, the CMB temperature is a good proxy 185 to discuss the core temperatures, using a relevant adiabatic profile. Based on the melting diagram of the Fe-alloy, the inner core should disappear for CMB temperatures above ~4250 K. This temperature 186 is only a couple hundred degrees above the current CMB temperature and is also just above the mantle 187 188 solidus of ~4150 K. This indicates that the onset of the inner core crystallization is expected to happen 189 before the lowermost mantle completely solidified.

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#### 191 4. Secular cooling of the deep Earth? Major unresolved paradoxes

These different lines of reasoning converge to the remarkable conclusion that today the CMB temperature is precisely at, or just below, the solidus of the silicate mantle, at 4100 K (+/- 200) K. If the Earth has been cooling for the last ~4.5 Gy, the early core would need to be significantly hotter in the past and, hence, overlaid by molten mantle, a primordial basal magma ocean (BMO) giving birth later to the D"-layer (Labrosse et al., 2007). In this article, we challenge this classic view of Earth's secular cooling based on three major paradoxes:

(i) Since when (and for how much additional time) the CMB temperature has (and will) remain
 *precisely* just below the solidus of the silicate mantle? It would be very unusual if this peculiar
 situation were a pure coincidence, since this temperature corresponds to a major change in the mantle
 state through a first-order phase transformation (the onset of melting at the solidus).

202 (ii) Producing the geodynamo by a combination of thermal buoyancy and compositional 203 convection (Buffett, 2000) requires a heat flux through the CMB of possibly up to ~10-15 terawatts 204 (TW), depending on the controversial values of electrical and thermal conductivities of the outer core 205 (Pozzo et al., 2012; Zhang et al., 2015). Three ingredients have been advanced to explain the 206 persistence of a significant CMB heat flux from the oldest (4.2 Gy ago) evidences of paleo-magnetic 207 field (Tarduno et al., 2015) to the present day: (a) The first is a tremendous initial core temperature 208 of 5000-7000 K (Davies et al., 2015; Labrosse, 2015; Nakagawa and Tackley, 2010), well above the 209 mantle solidus and even above the liquidus. Such high temperatures would delay the crystallization 210 of the inner core, possibly to a period as late as 0.7 Gy ago (Labrosse, 2015), even if such a late start 211 is still controversial (Biggin et al., 2015). (b) The second ingredient is a high concentration of 212 radiogenic K in the core. Experimental determination of the K partitioning between metal and silicate 213 suggests a maximum of 250 ppm K in the core (Bouhifd et al., 2007), while values below 50 ppm K 214 appear more likely (Corgne et al., 2007; Watanabe et al., 2014). However, models of core cooling 215 with values as high as 400-800 ppm K are not capable of sustaining the geodynamo until the present 216 (Nakagawa and Tackley, 2010). (c) It was suggested that layered structures in the lowermost mantle 217 (Nakagawa and Tackley, 2014) or at the top of the core (Buffett, 2014) may help to retain heat within 218 the core, but the density contrast needs to be large to maintain a gravitationally stable liquid layer in 219 the highly turbulent flows expected in a very hot Earth. Such a global density stratification in the 220 mantle or in the core has not been observed seismically.

(iii) The large initial temperature required for the primordial core implies large fractions of melt in the lowermost mantle. However, it seems very difficult to maintain a basal magma ocean (BMO) at a temperature significantly above the mantle solidus (or, more precisely, above the viscous transition in the mantle, which correspond to ~40% of partial melting (Abe, 1997). This issue is detailed below) for a long period of time (see (Monteux et al., 2016)). (a) The first reason for this is that the silicate melt viscosity is orders of magnitude lower than that of the solid mantle. To show its 227 effect, let's assume a temperature jump of 1000 K in a super-adiabatic thermal boundary layer lying 228 above the CMB. For a silicate melt, the boundary layer would be thinner than one meter (Solomatov, 2015), which would drive the heat flux to more than  $10^6$  TW. This implies a negligible temperature 229 230 jump at the CMB when the silicate is liquid above the CMB. A detailed calculation of the cooling of a hot primitive core with or without a molten layer above the CMB suggests core cooling 10<sup>6</sup> times 231 232 faster in the presence of a BMO (Monteux et al., 2011). (b) Another argument is based on the fact 233 that the adiabatic profiles of solid, liquid and partially molten chondritic-type mantle present P-T 234 slopes less steep than the mantle solidus (Andrault et al., 2011; Thomas and Asimow, 2013). This 235 implies that the degree of mantle partial melting should increase with elevation from the CMB, not 236 the contrary. For this reason, having a CMB temperature significantly above the mantle solidus would 237 yield major mantle instabilities. If vertical chemical segregation would eventually produce a BMO 238 with a composition different than the average mantle, the situation would not be drastically different. 239 The temperature profile in the BMO would follow a guasi-adiabatic profile from the hot CMB to the 240 interface between the BMO and the overlaying solid mantle. Then, all arguments raised above remain 241 valid, but with a dominant interface for heat exchange being located between the BMO and the solid 242 mantle, instead of exclusively at the CMB. (c) A last argument is that the hot melt could be unstable 243 in the BMO because it is buoyant. It would travel through the mantle towards the Earth's surface. 244 Unfortunately, this issue remains controversial (Andrault et al., 2012; Nomura et al., 2011).

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## 246 **5.** A quasi-constant CMB temperature over geological time

247 The most likely thermal state after the mantle has achieved a ~stable character is that the BMO 248 became significantly viscous, thus below the typical temperature threshold corresponding to 60% of 249 crystallization. This happens at a CMB temperature of ~4400 K for a chondritic-type mantle 250 (Andrault et al., 2011). We note that a peridotitic mantle would become viscous at a slightly higher 251 temperature (Figuet et al., 2010), however, this composition is much less relevant to the primitive 252 mantle at the CMB. As discussed above, this final temperature could be achieved in less than  $10^8$ 253 years and it is weakly dependent on the initial CMB temperature (Lebrun et al., 2013; Sleep et al., 254 2014; Solomatov, 2015). This gives rise to a new paradox: while the CMB temperature reached ~4400 255 K early in the Earth's history, it is only a few hundred degrees below today, at a temperature of ~4100 256 K. We note that the uncertainty on the temperature difference of ~300 K is independent of the 257 uncertainty on the experimental determination of the solidus temperature. Indeed, the early and 258 present-day CMB temperatures (slightly above, and just below, respectively) are determined relative 259 to a same reference that is the solidus of the average mantle at the CMB. The uncertainty on the  $\sim 300$ 260 K secular cooling of the CMB is estimated to  $\pm 100$  K.

261 Such a stable CMB temperature is compatible with geological constraints on the time evolution 262 of the mantle potential temperature (MPT, i.e. the extrapolation to the planetary surface of the 263 mantle's adiabatic temperature profile). For example, petrological analyses of Archean and 264 Proterozoic basalts (between 1.5 and 3.5 Gy old) preserved at the Earth's surface show primary 265 magma compositions compatible with an MPT only ~200 K greater than today (Herzberg et al., 2010). 266 A similar temperature change is reported between Archean tonalite-trondhjemite-granodiorite 267 associations of 4.0 to 2.5 Gy old (Martin and Moyen, 2002). We note that the CMB temperature and 268 the MPT are not formally linked to each other, due to an adjustable temperature jump in the thermal 269 boundary layer above the CMB. Still, they both refer to the thermal state of the deep Earth.

270 Because the CMB temperature is intimately linked to the core thermal state, a steady CMB 271 temperature over billions of years excludes core cooling as a major ingredient for driving the 272 geodynamo during this period. There are two alternative sources that can induce the turbulent fluid 273 motion in the outer core needed to produce the geomagnetic field: (i) Chemical buoyancy occurs 274 when light elements (mainly O (Alfè et al., 2002)) are released at the ICB due to inner core growth. 275 This effect becomes significant when the temperature drops below  $\sim$ 4250 K, thus at only  $\sim$ 250 K 276 above the present-day CMB temperature (Fig. 2b). Previous work dedicated to the analysis of the 277 relative effects of compositional and thermal convection suggests that the same magnetic field can be 278 generated with approximately half the heat throughput needed if the geodynamo was purely thermally 279 driven (Gubbins et al., 2004). Still, none of the recent studies suggest that chemical buoyancy could 280 drive alone the geomagnetic field for billions of years (Davies et al., 2015; Labrosse, 2015). 281 Alternatively, (ii) mechanical forcing induced by precession and tidal distortions of the CMB (Dwyer 282 et al., 2011; Le Bars et al., 2015; Tilgner, 2005) could have been a major ingredient to maintain the 283 geomagnetic field, since the formation of the Moon. It could still operate today.

284

## 285 **6.** Precession and tides, an alternative mechanism to drive the geodynamo

286 Precession and tidal distortions of a planet's CMB induced by gravitational interactions with a 287 companion (e.g. Earth and Moon) are both capable of generating core turbulence and of sustaining a 288 dynamo with critical magnetic Reynolds numbers comparable to thermal and compositional 289 dynamos (Cebron and Hollerbach, 2014; Tilgner, 2005). Indeed, planetary cores, as any rotating 290 fluid, permit eigenmodes of oscillation called "inertial modes", whose restoring force is the Coriolis 291 force. Precession and tides, seen from the mantle frame of reference as small periodic perturbations 292 of the rotating fluid core, are capable of resonantly exciting those inertial modes, leading to fluid 293 instabilities, turbulence and dynamo action. More specifically, two types of instabilities have been 294 described in the literature, the same generic mechanisms working both for precession and tidal 295 excitations (see details in (Le Bars et al., 2015) and references therein): (1) the direct resonance of 296 one given inertial mode, whose non-linear interactions produce a localized geostrophic shear layer, 297 which can then destabilize and lead to turbulence (Malkus, 1968; Sauret et al., 2014); (2) the triadic 298 resonance of two inertial modes with the harmonic forcing (Kerswell, 1993, 2002), which can either 299 lead to sustained turbulence or to cycles of growth, saturation and collapse (Le Bars et al., 2010). In 300 either case, it is important to recognize that a resonance is involved: even if the excitation amplitude 301 is small, the resulting flows may be intense, draining their energy from the mechanism sustaining 302 the excited waves, i.e. from the spin-orbit rotational energy of the considered system. Such 303 mechanisms provide an appealing alternative explanation for planetary dynamos when the classical 304 convective model does not apply. For instance, they have been proposed to explain the brief dynamos 305 of the Moon and Mars (Stevenson, 2003), the size of these planets being insufficient to sustain long-306 lived thermally-driven dynamos. The disappearance of Mars' orbiting companion after it collided 307 into the young planet (Arkani-Hamed, 2009) and the recession of the Moon (Dwyer et al., 2011) 308 accompanied by a decrease in the precession intensity, could explain the end of their magnetic 309 histories.

310 The question then remains to determine (1) whether or not mechanically forced instabilities are 311 present in the Earth's core, and (2) whether or not the generated flows are sufficiently powerful to 312 explain the geomagnetic field. Regarding the first point, the recent literature indeed shows that the 313 present Earth, in the absence of a convectively imposed magnetic field (as would be the case in our 314 non-conventional model), is subject to both tidal and precession instabilities generating turbulence 315 (see e.g. (Cebron et al., 2012)). The second point has been the subject of a long debate between the 316 seminal work of (Malkus, 1968; Malkus, 1963) and the subsequent studies of Rochester et al. (1975) 317 and Loper (1975). This debate was resolved by (Kerswell, 1996): even if the laminar flows considered 318 by (Rochester et al., 1975) and (Loper, 1975) are insufficient to sustain a dynamo, the expected 319 turbulent states in the Earth's core are largely sufficient, and the huge amount of energy stored in the 320 Earth-Moon-Sun system (spin and orbit) provides very large source of energy to sustain the magnetic 321 field over geological time (see e.g. (Le Bars et al., 2015)).

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## 323 7. The energy budget to sustain the thermal steady state of the deep Earth

#### 324 *7.1 The core budget*

Additional arguments to support this proposal are provided by considering the energy budget of the Earth's rotational dynamics as a whole. Models supported by precise measurements coming from lunar laser ranging indicate that 3.7 TW is continuously injected from the Earth-Moon-Sun orbital system into the Earth system (Munk and Wunsch, 1998; Wunsch and Ferrari, 2004). Models also indicate that 0.2 TW is dissipated into the Earth's atmosphere and its mantle; direct satellite estimates show that 1 TW is lost to the deep ocean; and the most accurate models indicate additional tidal 331 dissipation in shallow seas of up to 2 TW (Ferrari, 2015). Hence, 0.5 to 1 TW of the dissipated 332 rotational power is still missing in the current energy budget: it may very well be continuously 333 injected into the outer core, where it can fulfill the energy thirst of the geodynamo, estimated to range 334 between 0.1 to 2 TW (Buffett, 2002; Christensen and Tilgner, 2004). The situation was probably even 335 more favorable in the past, when the Moon was closer to the Earth and when the Earth was rotating 336 faster. Indeed, tidal distortion was previously larger and dissipation measured by the Ekman number 337 was smaller, both ingredients being favorable to instability and turbulence (e.g. (Cebron et al., 2012)). 338 One can thus imagine that throughout the history of the Earth-Moon system, turbulent flows and 339 dynamos have been excited by mechanical forcing, the energy dissipated by both ohmic and viscous 340 dissipations participating into the Moon's recession and deceleration (e.g. (Le Bars et al., 2010)).

341 In addition to the 0.5 to 1 TW rotational power injected into the core for dynamo action, part of 342 which is ultimately transformed into heat by viscous and Joule dissipation, three sources could 343 significantly contribute to the heat budget: (i) Radioactive disintegration of potassium (<sup>40</sup>K) could provide between 0.2 to 1.4 TW today (Bouhifd et al., 2007; Buffett, 2002; Corgne et al., 2007; 344 345 Watanabe et al., 2014), (ii) the latent heat of inner core crystallization contributes 0.3 TW, assuming 346 that its crystallization occurs over a period of ~4.3 Gy. (iii) Finally, we note that despite a global 347 steady state, the core temperature may have decreased from ~4400 K originally to ~4100 K today 348 (Fig. 3). Core cooling by ~300 K over 4.3 Gy would provide an average CMB heat flux of ~2 TW. 349 Adding all contributions, the average heat flux coming out of the core could range between 3.0 and 4.7 TW. 350

351 On the other hand, the low range of values for the thermal conductivity of the outer core yields 352 a heat flow along the outer-core adiabat between 1.7 to 3.6 TW (Buffett, 2002). For higher values of 353 the conductivity, as suggested recently (Pozzo et al., 2012; Zhang et al., 2015), the adiabatic heat flow 354 would be more than 10 TW (e.g. (Labrosse, 2015)). This range of values appears significantly higher 355 than the 3.0 to 4.7 TW estimated using our model. We note, however, that the temperature profile in 356 the outer core could very well be slightly sub-adiabatic. It would actually facilitate the vertical 357 thermochemical stratification of the outer-core (Helffrich and Kaneshima, 2013). Dynamos excited 358 by mechanical forcing do not require a super adiabatic temperature in the Earth's outer core: it has 359 already been demonstrated that tidal and precession instabilities exist in a stratified environment, 360 theoretically (Cebron et al., 2012), numerically (Cebron et al., 2010) and experimentally in a cylindrical geometry (Guimbard et al., 2010). Instability involves resonances of gravito-inertial 361 362 waves rather than inertial waves, the main effect being to decrease the excited vertical wavelengths, 363 with negligible or even positive effects on the instability threshold and growth. The same conclusion 364 has been reached concerning other types of unstable flow, for instance Taylor-Couette flows (Le Bars and Le Gal, 2007): contrary to intuition, stratification is capable of increasing flow instability, and 365

turbulence may develop while maintaining an overall global stratification. In addition to tides and precession, a dynamo driven by the solidification of the inner core could also lead to an overall subadiabiatic core, as studied for the case of Mercury (Manglik et al., 2010).

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- 370

## 7.2 The thermal boundary layer in the lowermost mantle

371 The way the CMB heat flux, estimated above in the range of 3.0 to 4.7 TW from the core energy 372 budget, is accommodated in the lowermost mantle depends on 3 major parameters: (i) the thickness 373 of the thermal boundary layer (e<sub>TBL</sub>) where conduction is the dominant mechanism of heat transfer. 374 e<sub>TBL</sub> is generally assumed to be the thickness of the D"-layer, thus 100 to 300 km, as reported by 375 seismological studies (e.g. (Lay et al., 2004)). However, we note that the seismic anomalies used to 376 define e<sub>TBL</sub> are likely to be preferentially concentrated in the hottest (thus deepest) part of the TBL 377 where the temperature approaches the mantle solidus. At larger distances from the CMB, the 378 temperature profile could still be steeper than in the adiabatic profile (as expected within a TBL), 379 which would however not produce detectable seismic anomalies because of the relatively lower 380 temperatures. (ii) The thermal conductivity k of the lowermost mantle, which remains subject to 381 controversial reports. A recent study based on *ab initio* calculations proposed k=3.5 W/m/K for 382 bridgmanite-MgSiO<sub>3</sub> (Tang et al., 2014), a value that could be even lowered, by up to 50%, if 383 accounting for the presence of Fe and Al in the Bg-lattice (Manthilake et al., 2011). On the other 384 hand, values up to 16 W/m/K have been proposed for MgSiO<sub>3</sub> Bg and post-Bg, with a relatively 385 higher (but anisotropic) k value for post-Bg (Ohta et al., 2012). Also, intermediate values of 7-8 386 W/m/K were recently proposed, for the MgSiO<sub>3</sub> end-member again (Stackhouse et al., 2015). (iii) 387 The difference ( $\Delta T_{TBL}$ ) between the CMB temperature (4100 +/- 200 K) and the mantle temperature 388 a few hundred kilometers above the CMB as extrapolated from the adiabatic profile (2600 +/- 200 389 K). The temperature jump in the TBL could be  $\sim$ 1500 K (Fig. 2).

390 Using reasonable values of  $e_{TBL}=200$  km, k=5 W/m/K, and  $\Delta T_{TBL}=1500$  K, we calculate a heat 391 flux at the CMB of ~3.6 TW. This value falls well within the validity limit of our model that is 3 to 392 4.7 TW. We acknowledge that lower k values, or a thicker TLB, would induce a lower CMB heat 393 flux. Considering all possible values of e<sub>TBL</sub> from 100 to 300 km and k from 2 to 8 W/m/K results in 394 plausible CMB heat fluxes ranging from 1.0 to 11.5 TW. Unfortunately, this broad range of 395 uncertainties does not provide additional constraints to our model. We note that it has been suggested 396 that the possible presence of a dense viscous layer above the CMB could help reducing the CMB heat 397 flux (Nakagawa and Tackley, 2014).

398 On the other hand, based on the heat carried by plumes ascending from the base of the mantle to 399 the Earth's surface, the core heat loss has been estimated to be  $\sim 2.3$  TW, or perhaps up to 3.5 TW if 400 plumes loose significant amounts of heat during their ascent through the mantle (Davies, 2007). In 401 such thermal budget, it is difficult to take into account the possible deep mantle complexities such as 402 the cooling effect of plate tectonics, the insulating effect of a dense basal layer or also the importance 403 of heat sources available in an enriched deep mantle, because the amplitude of these effects remain 404 highly uncertain. One could argue that our estimated value of the CMB heat flux is much lower than 405 that calculated in recent geodynamic models (e.g. (Nakagawa and Tackley, 2014)). However, such 406 models generally assume a fully viscous mantle, even for an initial CMB temperature of 6000 K that 407 is well above the mantle solidus and the viscous limit of 40% of partial melt. As acknowledged in 408 (Nakagawa and Tackley, 2014), this artificially maintains a hot core for a long period of time, 409 associated with a substantial CMB flux. In a recent study, it was instead shown that an extremely 410 large CMB heat flux prevailed early in the Earth's history, until the viscous transition is reached in 411 the lowermost mantle (Monteux et al., 2016). It yields a CMB temperature of ~4400 K in less than 412 ~1 My after the MFI.

413

#### 414 8. Implications for geodynamics and major geological events

415 In addition to generating the Earth's dynamo, turbulent motions excited by astronomical forcing 416 can induce cycles of growth, saturation and abrupt relaxation of the hydrodynamic instabilities 417 (Kerswell, 1993; Sauret et al., 2014). The collapses could induce an abrupt release of energy, potentially up to 10<sup>9</sup> TW (Kerswell, 1996) over short periods of time, in addition to the resonances 418 419 in the Earth-Moon-Sun spin-orbit system (Greff-Lefftz and Legros, 1999). The pulse duration could 420 vary from a couple periods of rotation (a couple of days) to several hundred years. This corresponds 421 to a broad range of thermal energy release, which could induce core heating by a few to a few hundred 422 of degrees, depending on the integrated pulse amplitude. We note that the 0.5 to 1 TW currently 423 dissipated into the Earth's outer core from the Earth-Moon-Sun orbital system cannot heat the core 424 more than a couple hundred degrees. However, much larger heat pulses could have happened in the 425 past when the Moon was closer to the Earth and when the Earth was rotating faster (Fig. 3).

426 Such fluctuations of the CMB temperature could have two major consequences. (a) Following 427 the adiabatic temperature profile of the Fe-alloy from the CMB to the ICB, they should induce 428 fluctuations in the size of the inner core (Fig. 2b): The abrupt release of hydrodynamic instabilities 429 could reduce the size of the inner core and restore its capability to produce the geodynamo by 430 chemical buoyancy, when the CMB temperature would eventually decrease again by cooling due to 431 weaker dissipation by mechanical forcing. The possibility that an old inner core has undergone several 432 changes in its size, with a rapid decrease and slow increase of its radius could be important for 433 building the inner core anisotropy (Poupinet et al., 1983) as well as a mushy layer at the top of the 434 inner core (Loper and Fearn, 1983). Indeed, both geophysical interpretation are closely related to the mechanism of inner core crystallization (b) On the other hand, partial melting in the lowermost mantle 435

436 could act as an efficient agent for transferring the excess heat of the core to the overlying mantle: 437 Increasing the temperature above the mantle solidus at the CMB would result in an increase in the degree of partial melting in the lowermost mantle, which in turn would induce a larger CMB heat 438 flux. This mechanism could damp the fluctuations in heat production in the turbulent outer core 439 440 yielding a ~fixed CMB temperature, precisely at, or just below, the mantle solidus. This thermal state corresponds well to the present-day view of the D"-layer, where piles of partially molten silicate 441 442 material (the ultra-low velocity zones) interact with mantle convection. Adding heat to the current lowermost mantle would certainly enhance partial melting and the thermal instabilities (inset in Fig. 443 444 3). As a result, one should expect an increase of the volcanic activity at the Earth's surface shortly after the influx of heat at the CMB (Greff-Lefftz and Legros, 1999). If the brutal energy influx is 445 446 important, this could explain dramatic eruptions such as the Deccan Trapps (Courtillot and Fluteau, 447 2010), as well as the periodic growth of continents at the Earth's surface (Arndt and Davaille, 2013; 448 Martin et al., 2014).

Finally, because the Moon appears to be a necessary ingredient to sustain the magnetic field, and because a magnetic field is needed to shield the Earth's atmosphere from erosion by solar wind (*e.g.* (Dehant et al., 2007)), the habitability of Earth-like planet may be subordinated to the existence of a large-satellite. While more than 1000 exoplanets have already been observed, the detection of an accompanying exo-moon is rare (Bennett et al., 2014). Hence, our model could have major implications in future planetary missions as exoplanets with orbiting moons would more likely host extraterrestrial life.

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## 692 Figure Captions

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694 Figure 1: Schematic representation of the Earth's interior evolution from the Moon forming impact 695 (MFI) to the present. Full crystallization of the molten Earth was probably very complex due to 696 suspension, turbulence, nucleation, and percolation processes (Solomatov, 2015). (a) The MFI 697 occurred 50-100 Myr after the formation of the Ca-Al rich inclusions (CAI, the oldest objects in the 698 solar system). It left the Earth mostly molten with a core temperature potentially above 6000 K (Rubie 699 et al., 2015). (b) In the magma ocean, high temperatures and turbulent state should efficiently cool 700 down the deep mantle and the liquid core. This could favor a thermally driven geodynamo. 701 Progressive decrease of the potential temperature below the viscous threshold of 60 % crystallization took within  $10^3$  to  $\sim 10^6$  years, essentially depending on the magma ocean viscosity (Monteux et al., 702 703 2016). During this time, a thin crust rapidly formed at the Earth's surface. Also, a basal magma ocean 704 could have existed, however, its life time remains controversial (Labrosse et al., 2007; Monteux et 705 al., 2016). Large plumes of hot and/or partially-molten material could have migrated toward the 706 Earth's surface. (c) The final step of mantle solidification could have taken longer, depending on the 707 cooling efficiency at the Earth's surface (Lebrun et al., 2013; Sleep et al., 2014). This period would 708 correspond to progressive decrease of the CMB temperature to the mantle solidus at ~4150 K (Fig. 709 2a), associated with the crystallization and the growth of the inner core (Fig. 2b). This would favor a 710 compositionally driven geodynamo, an ingredient that could still today contribute to sustaining the 711 geodynamo. (d) Later on, the long-term solid-state convection in the mantle, as we currently know it, 712 started. At this period, the CMB temperature could have remained nearly constant for geological 713 times. Because of moderate core cooling and growth of the inner core, a major ingredient to sustain 714 the geodynamo could be mechanical forcing by astronomical forces. In fact, mechanical forcing could 715 have started to induce core motions as soon as the moon was formed. Colors orange to grey 716 (intermediate = yellow) correspond to the mantle encountering a degree of partial melting from 100% 717 to 0% (intermediate=40%), while black and white correspond to liquid (outer) and solid (inner) core, 718 respectively.

Figure 2: Present-day temperature profile in the Earth's (a) mantle and (b) core inferred from

- several experimental arguments. Green circles correspond to the most likely temperature at the
- core-mantle boundary. (a) At the CMB pressure of 135 GPa, melting temperatures of 4180, 4150,
- 723 3800 and 2570 K were reported for peridotite (F-10 (Fiquet et al., 2010)), chondritic-type mantle
- 724 (A-14 (Andrault et al., 2014)), mid-ocean ridge basalt (A-11 (Andrault et al., 2011)) and wet-
- pyrolite (N-14 (Nomura et al., 2014)), respectively. Dashed curves stand for adiabatic profiles (M-

- 726 07 (Matas et al., 2007), H-05 (Hernlund et al., 2005), K-10 (Katsura et al., 2010), BS-81 (Brown
- and Shankland, 1981), B-01 (Bunge et al., 2001)). (b) Melting of pure Fe was reported at 4175 and
- 6230 K for pressure conditions of CMB (135 GPa) and ICB (330 GPa), respectively (A-13
- 729 (Anzellini et al., 2013)). A melting temperature depletion of ~650 K can account for the presence of
- radius from 130 light elements in the core (e.g. (Morard et al., 2013)). The CMB temperature is extrapolated from
- the ICB based on the adiabatic profile in the outer core.
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733 Figure 3: Schematic evolution of the CMB temperature since the Earth's accretion. In the 734 hypothesis of an initial CMB temperature above 6000 K (Rubie et al., 2015), rapid cooling is 735 expected until the drastic increase in mantle viscosity at the CMB (see text and (Monteux et al., 736 2016)). It corresponds to a degree of partial melting of ~40% (Abe, 1997), thus a temperature of ~4400 K for a primordial chondritic mantle (Andrault et al., 2011). Then, the complete mantle 737 738 crystallization could have taken up to more than 1 Gy, as suggested by geodynamic modeling (e.g. 739 (Nakagawa and Tackley, 2010)). After the mantle became significantly viscous, a purely thermally-740 driven dynamo becomes unlikely due to major slow-down of the CMB heat flux. As a result, the 741 CMB temperature remained close to the mantle solidus, at ~4100 K, until today (see Fig. 2). At a 742 period difficult to define precisely based on our model, the appearance of the inner core (indicating 743 a CMB temperature below ~4250 K) provided buoyancy sources from the release of latent heat and 744 light elements. Still, the sources of energy are insufficient to maintain the geodynamo from the first 745 evidence of geomagnetic field, ~4.2 Gy ago (Tarduno et al., 2015) to present day. This strongly 746 suggests that mechanical forcing induced by a combination of astronomical forces (see (Le Bars et al., 2015)) has been a major ingredient to maintain the geodynamo. Due to the intrinsically time-747 748 dependent character of the mechanical forcing, periods of growing instabilities and intense turbulent 749 motions would alternate with cycles of relaxation associated with abrupt releases of large amounts 750 of energy (e.g. (Kerswell, 1993). Abrupt increases of the core temperature, triggering increases in 751 partial mantle melting at the CMB (see inset), could be related to the geological evidence of periods 752 of hot and intense volcanic eruptions (Arndt and Davaille, 2013; Martin et al., 2014). 753



# b) Up to 1Myr after MFI





