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1 Why might Planets and Moons have Early Dynamos?

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5 **Abstract**

6 Remanent crustal magnetization of martian and lunar crustal rocks plausibly
7 records magnetic fields related to core dynamos active during the first few hun-
8 dred Myr of evolution of these bodies. Such early fields suggest that core dy-
9 namos may occur during or as a result of accretion. We investigate whether
10 the processes governing the segregation and sinking of metallic Fe after a large
11 impact can yield thermal conditions in the core that favour dynamo action on
12 growing planets. Depending on the sizes of the impactor and planet, as well as
13 the temperature-dependence of the viscosity, we identify conditions in which an
14 early transient core dynamo is possible. We also consider the effect of a molten
15 layer surrounding the protocore on the duration of this early dynamo. Our re-
16 sults suggest that dynamos can be initiated in bodies with a radius of 3500 km
17 radius or greater under Earth-like conditions for ohmic dissipation in the core,
18 and in smaller bodies if a less restrictive critical magnetic Reynolds number
19 condition is applied. These dynamos may persist for several kyr to several Myr
20 depending on the heat transfer regime at the protocore-mantle boundary.

21 *Key words:* early dynamos; growing planets; meteoritic impacts; numerical
22 modelling.

23 1. Introduction

24 Among the terrestrial planets and moons of the inner Solar System, Mars
25 and possibly the Moon record evidence for early and short-lived (a few hundred
26 Myr) internally generated magnetic fields. Mars Global Surveyor and Lunar
27 Prospector magnetometer data (*Acuña et al.*, 1999; *Hood et al.*, 2003) confirm
28 the presence of ancient magnetized crust on these bodies. Although later dy-
29 namo fields (beginning around or after 3.9 Ga) have been suggested for both
30 bodies (*Cisowski et al.*, 1983; *Schubert et al.*, 2000), most evidence points to ear-
31 lier dynamos, that ceased by ~ 3.9 Ga (*Johnson and Phillips*, 2005; *Lawrence*
32 *et al.*, 2008; *Lillis et al.*, 2008; *Garrick-Bethell et al.*, 2009). Moreover, recent
33 analyses of paleomagnetism in angrite meteorites suggest that an early and
34 internally generated dynamo occurred within the angrite parent body (*Weiss*
35 *et al.*, 2008). Although there is increasing evidence for the occurrence of these
36 transient dynamos, their origin and timing remain a major enigma. Here we
37 investigate a plausible temporal link between early dynamos and the large im-
38 pacts that characterize the late accretion histories of terrestrial bodies in our
39 solar system. Assuming chondritic bulk compositions, these impacts can lead
40 to extensive mantle melting, iron-silicate separation, iron segregation and ulti-
41 mately core formation (*Tonks and Melosh*, 1992; *Ricard et al.*, 2009; *Monteux*
42 *et al.*, 2009; *Srámek et al.*, 2010). If dynamo action depends primarily on the
43 rate of core heat loss (*Buffett*, 2002), then a crucial issue for explaining early
44 dynamos is to understand the thermal states of both the impacted planet and
45 the nascent core: Under what conditions will the resulting core cooling be suf-
46 ficiently large to favor dynamo action?

47

48 Large meteorite impacts can govern the early stages of planetary formation.
49 Figure 1 shows a schematic temporal evolution of our impact-induced dynamo

50 model from time t_1 (impact time) to time t_5 (dynamo is initiated). A collision
 51 between an impactor and an undifferentiated protoplanet leads to an increase in
 52 mass of the impacted body, the formation of an impact basin, the propagation
 53 of shock waves (Fig. 1, t_1) and a local spherical temperature increase, ΔT_0 ,
 54 below the impact site (Fig. 1, t_2). All these processes are very fast compared
 55 to the following thermo-chemical re-equilibration. The temperature increase is
 56 constant within a spherical volume V_{ic} called isobaric core with radius R_{ic} and
 57 rapidly decreases away from it (*Pierazzo et al.*, 1997; *Senshu et al.*, 2002; *Mon-*
 58 *teux et al.*, 2007). Assuming that meteorites impact at the escape velocity of the
 59 impacted body, this temperature increase scales with the size of the impacted
 60 planet and is superimposed on the initial mean temperature, T_0 , of the planet
 61 before impact. For sufficiently large planets, the temperature increase can over-
 62 come the melting temperature of the metallic phase leading to metal-silicate
 63 separation (*Monteux et al.*, 2009) (Fig. 1, t_3). The dense metallic phase sinks
 64 towards the center of the impacted body. How this occurs is debated. Proposed
 65 mechanisms include percolation through a solid mantle (*Stevenson*, 1990; *Shan-*
 66 *non and Agee*, 1996) and negative diapirism in the form of cm-sized droplets in
 67 a magma ocean [e.g. *Rubie et al.* (2003)] to large mantle diapirs [e.g. *Tonks and*
 68 *Melosh* (1992); *Senshu et al.* (2002); *Ricard et al.* (2009)]. In addition, it has
 69 been proposed that metallic diapirs can sink towards the centre of the planet
 70 via fracturation and diking (*Tonks and Melosh*, 1992; *Solomatov*, 2000; *Steven-*
 71 *son*, 2003b) or viscous deformation (*Honda et al.*, 1993; *Monteux et al.*, 2009)
 72 depending on the thermal state of the planet. Here, we consider a simplified
 73 scenario in which a single large impact into an initially undifferentiated planet
 74 leads to a single Fe-diapir that sinks through an effectively viscous mantle at a
 75 Stokes velocity (Fig. 1, t_4).

76

77 Whether convection and dynamo generation can occur in a liquid iron proto-
78 core depends on the temperature difference between the protocore and the sur-
79 rounding mantle. A protocore sufficiently hotter than the surrounding silicate
80 mantle is one way to explain the presence of an early dynamo on terrestrial plan-
81 ets and moons. Previous studies have examined the following possible causes of
82 an early hot protocore and dynamo: radioactive heating (*Yoshino et al.*, 2003;
83 *Williams and Nimmo*, 2004; *Nimmo*, 2009), strong cooling by mantle convection
84 (*Nimmo and Stevenson*, 2000; *Stegman et al.*, 2003; *Breuer and Spohn*, 2003;
85 *Elkins-Tanton et al.*, 2005) or impact heating (*Monteux et al.*, 2009; *Reese and*
86 *Solomatov*, 2010). Here we focus on whether the thermal conditions resulting
87 from a single large impact and the subsequent differentiation event are also
88 favourable to dynamo initiation. In this case, the temperature difference be-
89 tween the protocore and the overlying mantle depends on the thermal evolution
90 of the metallic diapir as it sinks through a colder mantle towards the centre of
91 the planet (Fig. 1, t_5). Because the heat content and buoyancy of the diapir
92 depend on its volume, one critical parameter will be the radius of the metal-
93 lic diapir R_{Fe} . An additional potentially important contribution to the initial
94 core-mantle temperature difference will be the extent to which the gravitational
95 potential energy released on sinking is converted into heat in the metallic di-
96 apir and in the surrounding mantle via viscous dissipation. The importance
97 of this effect, and where it occurs (i.e., primarily in the diapir or surrounding
98 mantle), will depend critically on the strongly temperature-dependent viscosity
99 $\eta(T)$ of the rocks composing the protoplanetary mantle. For impacted planets
100 of given radii R we use the numerical model of *Monteux et al.* (2009) to identify
101 $R_{Fe} - R - \eta(T)$ conditions favoring dynamo action. In addition, we investigate
102 the dependence of the dynamo duration on the protocore size, and on conditions
103 that lead to melting of the overlying mantle material.

104 **2. Thermo-chemical model**

105 *2.1. Initial state*

106 The initial structure and thermal state of a growing planet depends par-
107 ticularly on the characteristics of its accretion (*Safronov, 1978; Kaula, 1979*).
108 Indeed, differentiated terrestrial planets probably formed rapidly from accumu-
109 lation of chondritic material (*Agee, 1997*). In more detail, during accretion,
110 heating driven by a combination of the dissipation of impact energy and the
111 decay of short lived radionuclides such as ^{26}Al and/or ^{60}Fe (*Senshu et al., 2002;*
112 *Yoshino et al., 2003; Monteux et al., 2007*) increases the mean internal tem-
113 perature and gives rise to a radial temperature gradient that depends on the
114 accretion rate relative to the rate of radiative cooling to space. Where this
115 growth rate is very high in comparison to surface cooling, this heating can ul-
116 timately cause partial or complete melting (*Yoshino et al., 2003*) and extensive
117 internal differentiation. In contrast, where the growth rate is very low in com-
118 parison to surface cooling, little additional heating occurs (*Wood et al., 2006*)
119 and the growing planet remains an undifferentiated mechanical mixture of the
120 accreted impactors. Hence, depending on its growth rate, the frequency of
121 impacts and the size of the impactors, a protoplanet can potentially grow to
122 a Mars size radius in ≈ 1 Myr without undergoing any significant episodes
123 of melting and any subsequent global differentiation processes (*Senshu et al.,*
124 *2002*). Short-lived isotopic constraints restrict the timing of core formation on
125 terrestrial planets to be within the first tens of Myr following accretion (*Kleine*
126 *et al., 2002; Yin et al., 2002; Touboul et al., 2007*). From these models, the
127 martian core formation occurred faster (within 30 Myr) than the Earth's core
128 formation (within 50 Myr). Consequently, between 1Myr (the growing planet
129 reaches a Mars size radius with no or little differentiation) and 50 Myr (the core
130 is fully formed), large impacts can lead to episodic large differentiation events.

131 Depending on the thermal states of the resulting protocore and surrounding
132 mantle, subsequent cooling may drive a core dynamo that generates a magnetic
133 field observable at the surface.

134

Whether an impact onto an undifferentiated planet leads to melting and core formation depends critically on the fraction γ of the initial kinetic energy E_{kin} of the impactor that is dissipated within the planet itself as a result of work done by shock waves (*Tonks and Melosh, 1992; Monteux et al., 2009*). E_{kin} is proportional to the impact velocity v_{imp}^2 , which can be decomposed into two contributions:

$$v_{imp}^2 = v_{esc}^2 + v_{\infty}^2. \quad (1)$$

Here, v_{esc} is the escape velocity of the impacted planet and v_{∞} is the velocity of the impactor at a distance much greater than that over which the gravitational attraction of the impacted planet is important. A conservative and well-constrained estimate of E_{kin} is possible if we take $v_{\infty} = 0$ and assume $v_{imp} = v_{esc}$. We note that in the case of an impact on a moon orbiting around a larger central body, the gravitational attraction of the central body can increase this impact velocity and we address this issue in section 5. Laboratory experiments and modelling studies suggest that the energy released within the planet is mainly dissipated by frictional heating and melting within a spherical region (an “isobaric core”) with a volume V_{ic} that is at most 3 times larger than the volume of the impactor V_{imp} (*O’Keefe and Ahrens, 1977; Croft, 1982; Pierazzo et al., 1997; Senshu et al., 2002*). Within this region the temperature increases uniformly from T_0 to $T_0 + \Delta T_0$ where T_0 is the temperature of the impacted body before the impact and ΔT_0 is the temperature increase due to impact heating. Experiments suggest also that there is an additional thin spherical shell of damaged and heated material outside the isobaric core. Assuming

that the work done by shock waves in this shell causes no melting, the excess temperature ΔT_0 decreases rapidly and smoothly with distance r across the shell as approximately $\Delta T_0(R_{ic}/r)^m$. Following *Senshu et al.* (2002), and fitting the decay of peak pressure with distance away from the edge of the isobaric core $m \approx 4.4$ (*Monteux et al.*, 2007). Hence, the amount of energy E_{ext} dissipated as heat outside the isobaric core is:

$$E_{ext} = 2\pi \overline{\rho_0 C_p} \int_0^\pi \int_{R_{ic}}^{+\infty} \Delta T_0 \left(\frac{R_{ic}}{r}\right)^m r^2 \sin \theta d\theta dr \quad (2)$$

with θ the zenith angle from the impact site and $\overline{\rho_0 C_p}$ is the average specific heat of the impacted body. For the axisymmetric geometry of the isobaric core, the previous equation can be reduced to:

$$E_{ext} = \overline{\rho_0 C_p} V_{ic} \Delta T_0 \left(\frac{3(2m-5)}{2(m-3)(m-2)} \right) \quad (3)$$

Assuming that $v_{imp} = v_{esc}$ (*Kokubo and Ida*, 1996) and that $V_{ic} = 3V_{imp}$ (*Senshu et al.*, 2002), a balance between the kinetic energy delivered to heat the growing planet (γE_{kin}) and the energy used to heat up and melt the isobaric core and to heat up the surrounding material without melting is

$$\frac{4\pi}{9} \gamma G \rho_0^2 V_{ic} R^2 = \overline{\rho_0 C_p} \Delta T_0 V_{ic} + E_{ext} + V_{ic} (\rho_{Fe} f_0 L_{Fe} + \rho_{Si} (1 - f_0) L_{Si}). \quad (4)$$

After some algebra we obtain the temperature increase at the impact site as a function of the impacted planet radius (*Monteux et al.*, 2007)

$$\Delta T_0 = \frac{1}{h_m \overline{\rho_0 C_p}} \left[\frac{4\pi}{9} \gamma \rho_0^2 G R^2 - (\rho_{Fe} f_0 L_{Fe} + \rho_{Si} (1 - f_0) L_{Si}) \right]. \quad (5)$$

Here, G is the gravitational constant, ρ_0 is the density of the undifferentiated

planetary material, R is the radius of the impacted planet. L_{Fe} and L_{Si} are the latent heat required for melting respectively the metallic and the silicate phases. f_0 is the volume fraction of metal in the impacted planet. From equations (3), (4), and (5), h_m represents the amount of heat that is used to increase the temperature inside and outside the isobaric core relative to the amount of heat used to increase the temperature by ΔT_0 within the core and (*Monteux et al.*, 2007)

$$h_m = 1 + \frac{3(2m - 5)}{2(m - 3)(m - 2)} \approx 2.7 \quad (6)$$

135

136

137 Because ΔT_0 depends on γ , the value for γ requires some discussion. *O’Keefe*
 138 *and Ahrens* (1977) show that γ depends on the impact velocity and that the
 139 fraction of energy converted into heat increases with impact velocity. Our as-
 140 sumption that $v_{imp} = v_{esc}$ implies that our modelled impactors are in a “slow
 141 impactor regime”. From the measurements of *O’Keefe and Ahrens* (1977) a rea-
 142 sonable value for γ is 0.3. We apply this value in our calculations and discuss
 143 the influence of the uncertainty of this choice in section 5. An additional issue
 144 that requires mention is the assumption that there is no melting in the spherical
 145 shell around the isobaric core. For $\gamma = 0.3$, calculation of ΔT_0 as a function
 146 of planet radius shows that this approximation is strictly appropriate for plan-
 147 etary radii less than about 4000 – 4500 km. For larger planets, latent heat may
 148 be consumed outside the isobaric core and so ΔT_0 calculated with equation 5
 149 may be overestimated. The magnitude of the overestimate is, however, unclear
 150 because the effect of melting on the presence and structure of the spherical shell
 151 region bounding the isobaric core is unknown. Nevertheless, assuming that the
 152 geometry of the region is similar to the case without melting, calculations show
 153 that the uncertainty in the value for γ will have a greater influence on our

154 estimates for ΔT_0 .

The magnitude of the temperature increase in Eq.5 depends on the two terms in square brackets: the left-hand term describes the kinetic energy released to heat the mantle on impact and the right-hand term is the latent heat consumed to melt the silicate and iron components of the mantle. For iron-silicate segregation to occur in the heated volume, ΔT_0 must be positive and a sketch of the thermal state following impact is shown in Fig. 1 (t_2). Thus, equating the two terms in parentheses in Eq.5 gives a critical impacted planet radius above which segregation is possible:

$$R_{crit} > \left[\frac{\rho_{Fe} f_0 L_{Fe} + \rho_{Si} (1 - f_0) L_{Si}}{4\pi/9\gamma\rho_0^2 G} \right]^{1/2} \quad (7)$$

155 For parameter values listed in Tab.1, $R_{crit} = 1620$ km. Hence, impact heating
156 on a Moon to a Mars-sized body will result in the separation of the dense iron
157 phase from the silicate phase as a metallic diapir (Fig. 1). We note that whereas
158 our assumption that $v_{imp} = v_{esc}$ is appropriate for most planets, for moons this
159 simplification underestimates the impact velocity, which will be governed by the
160 gravitational potential of the central body and leads to a lower bound on ΔT_0 .

161

162 This temperature increase is superimposed on the pre-impact thermal regime
163 of the protoplanet. The background thermal state depends on the accretionary
164 process, which can lead to a temperature profile that increases with radius to-
165 wards the surface (*Senshu et al.*, 2002), and on the strong early rate of radiogenic
166 heating, which augments the mean internal temperature of the planet (*Yoshino*
167 *et al.*, 2003). For simplicity we assume that T_0 is uniform and close to the eu-
168 tectic temperature of the Fe-FeS system at 1 bar (*Fei et al.*, 1997) and that the
169 impacted body before the impact is a homogenous mixture of silicate and metal.

170

171 For a given initial bulk composition, the radius of the diapir separated after
 172 a large impact R_{Fe} is a function of the impactor size R_{imp} and f_0 in the im-
 173 pacted planet and scales as $R_{Fe} = (3f_0)^{1/3}R_{imp}$ (Monteux *et al.*, 2009). The
 174 metal diapir sinks towards the centre of the planet over time-scales from several
 175 kyr to several Myr depending on the rheology of the protoplanetary mantle.
 176 Once at the center of the undifferentiated planet, this "metallic diapir" will be
 177 referred to as a "protocore".

178

179 2.2. Heat partitioning during the sinking of the diapir

180 The initial thermal state of the protocore is governed by its volume and the
 181 rate and location of viscous dissipation during its sinking, which depends on the
 182 viscosity ratio λ (lower than 1) between the hot protocore and the relatively cold
 183 undifferentiated surrounding mantle (Ke and Solomatov, 2009; Samuel *et al.*,
 184 2010). For a given λ the rate of viscous dissipation depends on the shear strain
 185 rate $\dot{\epsilon}^2$, where $\dot{\epsilon} \sim V_S/R_{Fe}$ and V_S is the sinking velocity. Monteux *et al.* (2009)
 186 showed that for small diapirs (i.e., $R_{Fe} < 70$ km), the rate of viscous dissipation
 187 is negligible and the metallic phase only cools during its sinking. In contrast,
 188 for $R_{Fe} > 70$ km and for mantle and diapir viscosities that are equal (uniform
 189 viscosity), viscous deformation within the diapir and the surrounding mantle
 190 leads to dissipation that can increase the temperature of the metal diapir over
 191 its initial value from T_0 to $T_{core} > T_0 + \Delta T_0$.

192 Fig. 1 (top left panel) shows where the viscous heating occurs when a spheri-
 193 cal metallic diapir sinks within a Rybczynski-Hadamard analytical velocity field
 194 (Rybczynski, 1911; Hadamard, 1911) for 2 different viscosity ratios λ . For a
 195 uniform viscosity ($\lambda = 1$), viscous coupling between the metallic diapir and the
 196 surrounding material occurs and a fraction of gravitational energy is converted
 197 to heat up the diapir. However, for small viscosity ratios ($\lambda < \mathcal{O}(10^{-2})$), vis-

198 cous heating is restricted to the surrounding mantle and is concentrated at the
 199 diapir’s poles where strain rates are greatest (Fig. 1, top left panel and (*Samuel*
 200 *et al.*, 2010)). In that case, the initial temperature of the protocore is lower
 201 than in the isoviscous case and remains close to the initial temperature of the
 202 diapir (i.e. $T_{core} = T_0 + \Delta T_0$).

203 2.3. Dynamical model for the sinking of an iron diapir

204 2.3.1. Physical model

To investigate the dynamics of the metallic diapir separated after a large impact, we adapt the numerical finite volume model in spherical axisymmetric geometry of *Monteux et al.* (2009). In particular, neglecting the effects of adiabatic compression, conservation of energy applied to an undifferentiated planet of radius R leads to

$$\frac{DT}{Dt} = \frac{\nabla^2 T}{Ra_\chi} + D_\chi \frac{\eta}{\eta_0} \phi. \quad (8)$$

205 where T and t are dimensionless temperature and time (T and t are respectively
 206 normalized by ΔT_0 and by a Stokes time $t_{Stokes} = \eta_0 / \Delta \rho g_0 R$). $Ra_\chi = \frac{\Delta \rho_0 g_0 R^3}{\kappa \eta_0}$
 207 is the compositional Rayleigh number and $D_\chi = \frac{\Delta \rho_0 g_0 R}{\rho_0 \overline{C_p} \Delta T_0}$ is the dissipation
 208 number (with κ the heat diffusivity and $\overline{C_p}$ the average specific heat of the
 209 impacted body). $\Delta \rho_0$ is the density difference between metal and silicates. g_0 is
 210 the gravity at the surface of the impacted protoplanet. ΔT_0 is the initial impact
 211 induced temperature increase below the impact site. ϕ is the dimensionless
 212 dissipation function and expresses the conversion of potential energy into heat:

$$\phi = \tau : \nabla \mathbf{v} \quad (9)$$

213 where τ is the dimensionless deviatoric stress tensor and \mathbf{v} is the non-dimensional
 214 velocity (normalized with $v_{Stokes} = R / t_{Stokes}$). Prior to impact, we assume a
 215 homogenous temperature T_0 of the growing planet. The viscosity is $\eta = \eta_0 \lambda^T$

216 where $\eta_0 = \eta(T_0)$ is the viscosity of the cold undifferentiated material at the
 217 start of the experiment (*Ratcliff et al., 1997; Ziethe and Spohn, 2007*).

218

Assuming the protoplanet is incompressible, the other dimensionless governing equations are continuity

$$\nabla \cdot \mathbf{v} = 0, \quad (10)$$

and momentum conservation assuming infinite Prandtl number

$$-\nabla P + \nabla \cdot \left(\frac{\eta}{\eta_0} \left[\nabla \mathbf{v} + [\nabla \mathbf{v}]^T \right] \right) + \left(\frac{T}{B} - f \right) r \mathbf{e}_r = 0, \quad (11)$$

219 where P and r are the non-dimensional pressure and radius (P and r are re-
 220 spectively normalized by η_0/t_{Stokes} and R). \mathbf{e}_r is the radial unit vector and
 221 f represents the volume fraction of metal. Following *Monteux et al. (2009)*,
 222 we introduce the buoyancy ratio $B = \Delta\rho_0/(\rho_0\alpha\Delta T_0)$ where ρ_0 and α are the
 223 density and thermal expansion of the undifferentiated material (see Tab.1 for
 224 values). Gravity depends on the radial position r and $g(r) = \frac{4}{3}G\pi\rho_0r = g_0\frac{r}{R}$.

225

After an impact and the subsequent differentiation, the metallic phase and the purely silicate phase are separated within the thermal anomaly (Fig.1, t_3). Hence a chemical readjustment occurs leading to the sinking of the dense iron though the lighter undifferentiated material. The buoyancy force that drives the flow of the diapir towards the centre of the protoplanet increases with the volume fraction of metal f . In our models, we consider only two chemical phases meaning that f varies between 1 (pure metal) and 0.17 (undifferentiated material with chondritic composition). During sinking, chemical diffusion is

negligible. The metal volume fraction f is then simply advected by the flow :

$$\frac{\partial f}{\partial t} + v \cdot \nabla f = 0. \quad (12)$$

2.3.2. Simplifications of the governing equations

We make a number of approximations to simplify the analysis of the initial thermal state of the protocore. We neglect the thermal buoyancy T/B in the momentum equation as the buoyancy number B is very large (the density difference between metal and silicates is 140 to 560 times larger than the thermal density variations). In addition, in treating the metal-silicate separation, our model neglects multiphase dynamics associated with core-mantle segregation (Golabek *et al.*, 2008; Srámek *et al.*, 2010).

234

Prior to the impact, the planet is undifferentiated meaning that the impact and the subsequent local differentiation leads to the first episode of core formation. Moreover, as this study focuses on the thermal state of the metallic phase once at the centre of the planet, we do not consider the dynamical effects of the purely silicate phase that spreads beneath the impacted planet surface while the diapir sinks to the centre of the planet (Reese and Solomatov, 2006; Monteux *et al.*, 2009).

2.3.3. Numerical model

We implement a finite volume numerical model to solve Eq.8, Eq.10, Eq.11 and Eq.12 in axisymmetric spherical geometry in a 200×400 grid. We use a stream function formulation for the equations of motion with a direct implicit inversion method (Schubert *et al.*, 2001). Eq.8 and Eq.12 are solved by an Alternating Direction Implicit (ADI) scheme (Peaceman and Rachford, 1955; Douglas, 1955). The stream function, temperature and compositional fields

249 are described by a second-order approximation in space. To limit numerical
 250 diffusion when solving the transport equations, especially for the compositional
 251 field, we use a Total Variation Diminishing Superbee scheme (Roe, 1986; Laney,
 252 1998) implemented in an implicit way (Srámek *et al.*, 2010) which enables high
 253 resolution of purely advective fields. Mechanical boundary conditions are free-
 254 slip at the surface and along the symmetry axis. Thermal boundary conditions
 255 are isothermal at the surface and insulating along the symmetry axis.

256 3. Magnetic field considerations

257 3.1. Temperature of the protocore

Once at the centre of the planet, core cooling depends on the temperature of the protocore and the thermal state and melting temperature of the silicate at the protocore-mantle boundary. As discussed above, the temperature of the protocore depends on the post-impact temperature difference ΔT_0 (cf. Eq.5). The maximal temperature increase after sinking is (Monteux *et al.*, 2009) is :

$$\Delta T \approx \Delta T_0 \left(1 + a(\lambda) \left(\frac{r_0}{R} \right)^2 \frac{\Delta \Theta}{\Delta T_0} \right) \quad (13)$$

where $a(\lambda)$ is the fractional amount of viscous dissipation in the metal during the sinking of the iron towards the centre of the planet. r_0 is the initial distance between the inertial centre of the metallic diapir (noted i.c. in (Fig. 1, \mathbf{t}_3)) and the centre of the planet before the diapir starts to sink. $\Delta \Theta$ is the maximum temperature increase in the unrealistic case in which all the potential energy released by the diapir is used to heat up the metal phase. Assuming the impact velocity is the escape velocity of the planet (cf., section 2) then $\Delta \Theta / \Delta T_0$ is constant, independent of the planet size and equal to 11.8 (Monteux *et al.*,

2009) and the temperature of the metallic phase at the end of the sinking is

$$T_{core} \approx T_0 + \Delta T = T_0 + \Delta T_0 \left(1 + a(\lambda) \left(\frac{r_0}{R} \right)^2 \frac{\Delta \Theta}{\Delta T_0} \right). \quad (14)$$

258 During the sinking of the metallic diapir, its pressure increases and leads to
 259 additional heating by compressional effects. This heating process is inversely
 260 proportional to the buoyancy ratio B (Monteux *et al.*, 2009). In our models,
 261 B is large enough to consider that this thermal effect is negligible compared to
 262 viscous heating (Samuel and Tackley, 2008).

263 3.2. Conditions for an early magnetic field.

264 Three conditions are required to drive a dynamo on a growing planet:

265

- 266 1. The heat flow Q out of the core must exceed the adiabatic value Q_A such
 267 that convection can occur (Stevenson *et al.*, 1983). This condition is

$$Q > Q_A = \frac{k_{Fe} \alpha_{Fe} g_c T_{Fe}}{C_{p,Fe}} 4\pi R_{Fe}^2 \quad (15)$$

268 where k_{Fe} , α_{Fe} , T_{Fe} and $C_{p,Fe}$ are respectively the thermal conductivity,
 269 the thermal expansion, the temperature and the heat capacity of the metal-
 270 lic phase and g_c is the gravity at the surface of the protocore (see Tab.1
 271 for values). In our model, once the metallic phase has reached the center
 272 of the impacted planet, T_{Fe} decreases monotonically with time due to heat
 273 conduction to the surrounding mantle.

274

- 275 2. The ratio of the rate at which gravitational potential energy is released by
 276 convection to the rate of ohmic dissipation, Φ , must exceed a critical value
 277 (Buffett, 2002):

$$\frac{Q}{\Phi} > \frac{1}{\epsilon_T}. \quad (16)$$

Here, ϵ_T is the Carnot-style efficiency for thermal convection. We note that a dynamo can be generated even with a subadiabatic heat flow if the compositional convection is important (*Buffett, 2002; Christensen and Wicht, 2008*). Here, we do not consider the effect of chemical convection or the presence of an inner core. Assuming that the characteristic length scale of the flow leading to magnetic field generation is the radius of the protocore, Φ can be approximated as (*Buffett, 2002*):

$$\Phi = \left(\frac{\nu \overline{B}^2}{\mu} \right) \frac{4}{3} \pi R_{Fe}, \quad (17)$$

where \overline{B} is the average strength of the magnetic field within the core, ν is the magnetic diffusivity and μ is the magnetic permeability (see Tab.1 for values). In the absence of constraints on \overline{B} for early planets, we assume a current Earth-like value of 2.5 mT (*Kuang and Bloxham, 1997*) that is independent of the protocore size or the planetary radius. The efficiency of thermal convection is given by (*Buffett, 2002*)

$$\epsilon_T = \frac{0.8\pi}{3} \frac{\alpha_{Fe} G \rho_{Fe} R_{Fe}^2}{C_{p,Fe}} \left(1 - \frac{Q_A}{Q}\right), \quad (18)$$

278 with ρ_{Fe} the density of the metallic phase and G the gravitational constant.

279

3. The structure of the convective motions carrying magnetic field lines must be sufficiently complicated to favor self-sustaining dynamo action. A measure of this complexity is that the magnetic Reynolds number (*Christensen*

and Aubert, 2006)

$$Re_m = \frac{UL}{\nu} > Re_m^{crit} = O(10 - 10^2) \quad (19)$$

Here, L and U are the characteristic length and velocity scales for the flow within the protocore and ν is the magnetic diffusivity of the metal phase. Whereas the natural length scale in the problem is the depth of the convecting iron layer, the choice of an appropriate velocity scale depends on the leading order force balance (Christensen, 2010). As the rotation rate of growing planets is potentially time-dependent and poorly constrained a convenient and reasonable choice is based on a balance between inertial and buoyancy forces and is (Stevenson, 2003a):

$$U \sim \left(\frac{q\alpha_{Fe}g_c R_{Fe}}{\rho_{Fe}C_{p,Fe}} \right)^{1/3}, \quad (20)$$

where q is the heat flux out of the core ($q = Q/(4\pi R_{Fe}^2)$). Taking $L = R_{Fe}$, $g_c = 4/3\pi G\rho_{Fe}R_{Fe}$, the combination of Eq.19 and Eq.20 leads to the condition

$$Re_m = \left(\frac{4\pi Gq\alpha_{Fe}R_{Fe}^2}{3C_{p,Fe}} \right)^{1/3} \frac{R_{Fe}}{\nu} > O(10 - 10^2) \quad (21)$$

280 Among the three criteria above, the first is typically considered a necessary
 281 condition for a thermally-driven dynamo. (Subadiabatic dynamos are possi-
 282 ble if *e.g.*, compositional gradients also drive convection.). As we shall see in
 283 Section 4, this condition ($Q > Q_A$) is easily met in our models. However it
 284 is not a sufficient condition for dynamo action, as indicated by the other two
 285 conditions. The second condition assumes an Earth-like dynamo: in particular
 286 it assumes an Earth-like value for the root mean square magnetic field in the

287 core, a particular efficiency for conversion of gravitational potential energy into
 288 magnetic energy and a particular toroidal-to-poloidal field conversion efficiency.
 289 The third condition is a more general criterion for dynamo action, and whereas
 290 scalings other than Eq.20 for the velocity field are possible, we have chosen one
 291 that does not depend upon knowing the early rotation rate of the planet or
 292 moon.

293 3.3. Heat flow across the protocore: Analytic considerations

The three conditions required for an early dynamo (Eq.15, Eq.16 and Eq.21)
 all impose conditions on the heat flow out of the protocore. At the protocore-
 mantle boundary there are two asymptotic heat transfer regimes. The first is a
 conductive transfer across a thermal boundary layer growing on the mantle side
 where the corresponding heat flow Q_{cond} is :

$$Q_{cond} \sim \frac{4\pi k \Delta T R_{Fe}^2}{\delta} \quad (22)$$

294 where $\delta \approx 2\sqrt{\kappa t}$ is the thickness of the conductive boundary layer at any time
 295 t . As t becomes large $\delta = c_2 R_{Fe}$ (where c_2 is an order 1 constant (*Monteux*
 296 *et al.*, 2009)). k is the thermal conductivity of the surrounding material. For
 297 computational reasons, in our numerical models, we make the approximation
 298 that thermal conductivities for metallic and undifferentiated material are the
 299 same (i.e., $k = k_{Fe} = k_{Si}$, Table 1). In reality, the thermal conductivity of the
 300 core material is ≈ 10 times bigger than the conductivity for the silicates. Hence,
 301 this simplification underestimates the protocore heat flow and our criteria for
 302 dynamo action is more restrictive than in the more realistic case of unequal
 303 silicate and metal thermal conductivities.

304

As illustrated in Fig.2, the temperature of the protocore might overcome

the solidus temperature of the surrounding material depending on its chemical composition and on the impacted planetary radius. Hence, a melt layer may form above the protocore (see also section 5.2). As this melt layer grows in thickness (and assuming it is greater than a few tens of meters thick), vigorous convection will enhance core cooling over the conductive case and lead to a heat flow that is independent of the thickness of the melt layer and of the form (*Turner, 1973*):

$$Q_{conv} \sim 0.8\pi R_{Fe}^2 \left(\frac{\rho_{Si}^2 g_c \alpha_{Si} k^2 C_{p,Si}}{\mu_{melt}} \right)^{1/3} (T_{core} - T_{melt,Si})^{4/3} \quad (23)$$

where ρ_{Si} , α_{Si} , $C_{p,Si}$ and μ_{melt} are respectively the density, the thermal expansion, the heat capacity and the viscosity of the molten silicate layer (see Tab.1 for typical values). Eventually, the heat flux out of the core will, of course, become balanced by the heat flux across the overlying solid mantle and the melt layer will have a quasi-steady thickness.

4. Numerical results

4.1. Diapir descent: The initial protocore temperature

In our models, where we specify the viscosity of the impacted planet as $\eta_0 = 10^{22}$ Pas, metallic diapirs descend to the centre of the planet relatively quickly (e.g. in less than 10 Myr for $R_{Fe} = 400$ km) (*Monteux et al., 2009*). During the sinking of the diapir the temperature difference between the protocore and the mantle increases rapidly in response to viscous heating depending on the fraction of gravitational energy converted into heat ($a(\lambda)$). Then this temperature difference decreases as the diapir cools once it reaches the centre of the planet.

320

321 We monitor T_{core} for diapirs in the range of 150 – 500 km and planet radii
 322 in the range of 2000 – 4000 km. For the diapir radii range studied here, the
 323 protocore temperature at the end of sinking is given by Eq.13 and Eq.14. The
 324 results from our numerical models for uniform viscosity contrasts are shown in
 325 Fig. 2 and are in agreement with Eq.14. Theoretical predictions of T_{core} are
 326 represented in Fig. 2 with and without taking into account the effect of melting
 327 in the isobaric core after impact (grey and black solid lines respectively). The
 328 region in between defines the range of initial temperature and the difference
 329 between these two theoretical models is within the range of uncertainties of γ .
 330 For a uniform viscosity, viscous coupling between metal and undifferentiated
 331 material is important and $a(\lambda) \approx 15\%$ (*Monteux et al.*, 2009). As the viscosity
 332 contrast between the hot metallic phase and the surrounding mantle increases,
 333 the fraction of gravitational potential energy converted into heat in the metallic
 334 phase decreases (see section 2.2). That is, as the mantle/core viscosity ratio
 335 becomes very large $a(\lambda) \rightarrow 0$ and viscous dissipation is concentrated in the
 336 mantle (*Monteux et al.*, 2009). In this limit, the initial temperature of the pro-
 337 tocore after the sinking is $T_{core} = T_0 + \Delta T_0$. The predictions of Eq.14 for this
 338 asymptotic case are shown in Fig. 2 (dashed line). As expected, the reduced
 339 dissipation in the metallic diapir leads to a lower predicted T_{core} .

340

341 A high temperature contrast between the protocore and the surrounding
 342 undifferentiated mantle can eventually lead to the formation of a molten layer
 343 above the protocore. In Fig. 2, we include the chondrite solidus temperature
 344 from *Agee* (1997) and the dry peridotite solidus from *Zerr and Boehler* (1994)
 345 at the centre of the impacted planet as a function of the planetary radius. This
 346 effect will be more precisely discussed in section 5.2.

347

348 Because of the practical difficulty in obtaining accurate solutions for very
 349 large viscosity variations, we restrict our numerical studies to the isoviscous case,
 350 which is well-explained by our theory. The agreement between our isoviscous
 351 theory and the numerical solutions give us confidence that application of the
 352 theory for the variable viscosity case is justified. From Figure 2 these isoviscous
 353 solutions will overestimate the core temperature by $\approx 10 - 40\%$, depending
 354 on the protoplanetary radius. Thus, quantitatively, the uncertainties of this
 355 enhanced heat flow for predicting dynamo action are not so large. From Eq.13,
 356 assuming $a(\lambda) = 15\%$, the temperature difference between the protocore and the
 357 cold surrounding material in the isoviscous model is overestimated by a factor
 358 ≈ 2.8 . From Eq.21 $Re_m \propto Q^{1/3} \propto \Delta T^{1/3}$ and, thus, the magnetic Reynolds
 359 number will be a factor of about 1.4 too large (i.e., it is still accurate to $O(1)$).

360 4.2. *Dynamo initiation in a conductive heat transfer case*

361 To determine whether or not a dynamo is initiated in the protocore, we mon-
 362 itor the average heat flow Q across the protocore-mantle boundary as a function
 363 of time for a wide range of $R - R_{Fe}$ conditions. We stop calculations 50 Myr
 364 after the impact because at this time the core is probably fully formed and be-
 365 cause of the potential onset of mantle convection, the thermal consequences of
 366 which are beyond the scope of this paper (*Behoukova and Choblet, 2009*).

367
 368 $R - R_{Fe}$ conditions studied, we find that $Q \gg Q_A$ after impact, which
 369 means that core convection can occur (see Eq.15). For example, once a metal-
 370 lic diapir with $R_{Fe} = 300$ km has reached the centre of a planet with radius
 371 $R = 3500$ km, it starts to cool and over the 50 Myr period studied, Q/Q_A de-
 372 creases from 32.5 to 20. For $R = 2000$ km, Q/Q_A decreases from 15 to 10 over
 373 the same time interval (in the calculation of Q_A we consider that $k_{Fe} = 40$ W
 374 $m^{-1} K^{-1}$). Assuming that the thickness of the thermal boundary layer around

375 the protocore scales with the size of the protocore, this corresponds to a temper-
 376 ature difference of ≈ 300 to 2700 K between the metallic phase and the overlying
 377 mantle material (see Fig. 2). However, as discussed in Section 3, this condition
 378 alone does not guarantee dynamo action, and so we next examine the additional
 379 requirements for a dynamo.

380

In the special case of the impacted planet being Earth-like from a magnetic
 perspective, convection within the protocore has to be sufficiently vigorous to
 supply energy at a rate that compensates for the loss due to ohmic decay. We
 monitor the ratio of the heat flow to the ohmic dissipation as a function of time.
 Assuming a conductive heat transfer across the protocore mantle boundary,
 $k = k_{Si}$ and $\delta \sim R_{Fe}$, from Eq.13, Eq.17 and Eq.22 we obtain

$$\left(\frac{Q_{cond}}{\Phi} \right) \sim \frac{3\mu k \Delta T_0}{\nu \bar{B}^2} \left(1 + a(\lambda) \frac{\Delta \Theta}{\Delta T_0} \right). \quad (24)$$

381 which is a conservative bound on whether a dynamo will occur (we consider
 382 here that $r_0/R \approx 1$). $a(\lambda)$ and $\Delta \Theta / \Delta T_0$ are independent of R . However, from
 383 Eq.5, $\Delta T_0 \propto R^2$. Hence, (Q_{cond}/Φ) varies with R^2 . Figure 3 shows Q/Φ as a
 384 function of the planet radius for different iron diapir sizes and different times
 385 (1 Myr and 50 Myr), after the diapir reaches the centre of the impacted planet.
 386 The dashed line is the theoretical prediction from Eq.24 for a uniform viscosity
 387 case (i.e., $a(\lambda) = 15\%$). Numerical results are consistent with Eq.24 at very
 388 early time (just after the metallic diapir has reached the centre of the planet).
 389 At larger time, the protocore cools down and Q/Φ decreases. The temperature
 390 of the metallic phase after sinking and hence the initial heat flow across the
 391 protocore is a sensitive function of the viscosity contrast between the metallic
 392 phase and the surrounding mantle (see section 3.2). The theoretical value of
 393 the initial conductive heat flow when the viscosity of the hot metallic fraction is

394 much smaller than the viscosity of the relatively cold undifferentiated material
 395 is given by Eq.13 and Eq.22 with $a(\lambda) = 0\%$. The black dotted line in Fig.3
 396 shows the corresponding theoretical prediction for (Q_{cond}/Φ) immediately after
 397 sinking as a function of R for $a(\lambda) = 0\%$. In this case, the theoretical value for
 398 (Q_{cond}/Φ) is different by a factor of 2.8 compared with the uniform viscosity
 399 case, consistent with expectations from Rybczynski-Hadamard effects.

400

401 In all sets of planetary and metallic diapir radii studied here $Q \gg Q_A$.
 402 Consequently, the condition for the presence of a dynamo from the combination
 403 of Eq.16 and Eq.18 simplifies to:

$$\frac{Q}{\Phi} > \frac{3}{0.8\pi} \frac{C_{p,Fe}}{G\alpha_{Fe}\rho_{Fe}R_{Fe}^2} \quad (25)$$

404 Figure 4 shows Q/Φ as a function of the protocore radius resulting from a single
 405 impact for different planet sizes in uniform viscosity cases. For $t = 1$ Myr after
 406 the end of the sinking (black filled symbols), and for a given planetary radius,
 407 different trends appear as a function of R_{Fe} . The ratio Q/Φ is initially larger
 408 for larger bodies because $Q \propto \Delta T_o \propto R^2$. However, as mentioned in section
 409 2.2 when $R_{Fe} < 70$ km, no viscous heating occurs during the sinking (*Mon-*
 410 *teux et al., 2009*) as cooling by diffusion is faster than viscous heating. As R_{Fe}
 411 increases from 70 km radius, viscous heating becomes more and more efficient
 412 and leads to an increase of the initial heat flux and as a consequence of Q/Φ .
 413 When R_{Fe} reaches a critical value (≈ 300 km), the simplification $r_0/R = 1$ is
 414 less and less valid since the diapir is initially buried deeper. Hence, for a given
 415 R , as R_{Fe} increases r_0/R decreases so the initial temperature of the protocore,
 416 the heat flux and Q/Φ also decrease. With subsequent cooling, the reduction
 417 in core temperature over a given time period Δt , say, scales as $Q\Delta t/(\overline{\rho C_p}R_{Fe}^3)$.
 418 Because heat flow will decline with the core-mantle temperature difference, and

419 the incremental change in core temperature $\propto (1/R_{Fe}^3)$, small protocores cool
 420 faster than big ones. Hence, for a given R , and for $t = 50$ Myr after the end of
 421 the sinking (white filled symbols), Q/Φ increases as a function of R_{Fe} .

422

423 In Fig.4, the black dashed line represents the condition required to get a
 424 dynamo from Eq.25. Results show that a dynamo is difficult to initiate with a
 425 small volume of metallic phase (i.e. small impactors). Once the protocore ra-
 426 dius is larger than 400 km (i.e. $R_{imp} > 500$ km), however, an early dynamo can
 427 be initiated in the first 50 Myr in a planet with a radius greater than 2500 km
 428 and in the absence of a molten layer surrounding the protocore. As the planet
 429 size increases, the initial protocore temperature increases and the duration of
 430 the early dynamo is longer (over 50 Myr). As the protocore size increases, its
 431 characteristic cooling time increases and the duration of the early dynamo is
 432 also longer. However, a large viscosity contrast between hot iron and cold un-
 433 differentiated material will decrease the initial heat flow by a factor of 2.8 (see
 434 Fig. 3), and increase the dynamo threshold from Eq.25. The dotted line in Fig.4
 435 illustrates this theoretical effect. Fig.4 shows that a dynamo can still be initi-
 436 ated after an impact, but an increase of the impactor and impacted planet radii
 437 are needed relative to the isoviscous case. Hence an impact with $R_{imp} > 625$
 438 km (i.e. $R_{Fe} > 500$ km) on a Mars size body (i.e. $R \sim 3500$ km) can initiate an
 439 early magnetic field. The duration of this early dynamo increases with the size
 440 of the impactor.

441

442 Dynamo generation requires a magnetic Reynolds number larger than a crit-
 443 ical value (see Eq.21). We monitored the mean heat flux q as a function of time
 444 and calculated the corresponding magnetic Reynolds numbers. Fig.5 shows the
 445 results for the two extreme planetary radii (i.e. $R = 2000$ km and $R = 4000$ km)

446 and the range of protocore sizes studied here. The two lines (solid and dashed)
 447 correspond to $q = q_A = Q_A/(4\pi R^2)$ with $R = 2000$ km and $R = 4000$ km
 448 respectively. Three regimes emerge. Below, the solid line $q < q_A$ and no con-
 449 vection occurs. Above the solid line and below Re_m^{crit} , convection occurs but no
 450 dynamo is generated. Above these criteria, the thermally-driven convective flow
 451 within the protocore can generate a dynamo. As mentioned above, the mag-
 452 netic Reynolds number will be a factor of about 1.4 too large in the isoviscous
 453 compared with the temperature-dependent viscosity case. The dotted line rep-
 454 represents the values for Re_m expected from the temperature-dependent viscosity
 455 model (i.e. Re_m from isoviscous models divided by a factor 1.4). This figure
 456 illustrates three results. First, temperature-dependent viscosity has a weak in-
 457 fluence on the Re_m values obtained in this study. Second, as the heat flux out of
 458 the protocore obtained from our numerical models is ~ 20 larger than the adia-
 459 batic heat flux q_A (see Eq.15 and discussion above), then $Re_m(q) \gg Re_m(q_A)$.
 460 Third, Re_m easily exceeds the critical value needed to generate a dynamo in the
 461 whole range of planet and protocore radii studied here.

462

463 5. Discussion

464 These results provide an important link between the accretion and differen-
 465 tiation histories of the terrestrial planets and their potential for early magnetic
 466 fields. They suggest that the presence or absence of an early magnetic field on
 467 a planet could be related to its accretionary history.

468 5.1. Application to Solar System objects

469 Our results indicate that for bodies in the range $R = 2000$ – 4000 km and pro-
 470 tocore radii in the range 300 – 650 km, the core heat flow exceeds the adiabatic
 471 heat flow, and Re_m easily exceeds Re_m^{crit} for 50 Myr after the iron diapir reaches

472 the center of the planet and in the absence of additional mantle melting at the
473 protocore-mantle boundary. These conditions suggest that for the range of R -
474 R_{Fe} conditions studied, dynamo action is possible. A more restrictive criterion
475 is that the heat flow must also exceed the ohmic dissipation under Earth-like
476 conditions. In this case, and assuming no mantle melting, we find that a large
477 impact (a 625 km radius impactor, or $R_{Fe} = 500$ km) onto a Mars-size (*i.e.*, R
478 = 3500 km or greater) undifferentiated planet is needed to initiate a dynamo
479 that would persist at least for ~ 50 Myr (Fig. 4).

480

481 As noted earlier the ohmic dissipation criterion has several underlying as-
482 sumptions about the dynamo action. In particular, we note that the Earth-like
483 value of the magnetic field in the core \bar{B} may not be applicable to other bodies.
484 For example, Mercury, the only other inner solar system body to have a present-
485 day dynamo, has a magnetic field at its surface that is 100 times weaker than
486 Earth's surface field. While the origin of this weak field is not well understood,
487 it is at least reasonable to propose that Mercury's core field could be weaker
488 than Earth's. A factor of two decrease in the core field results in a factor of four
489 increase in Q/Φ , allowing dynamo action for smaller impactors and impacted
490 bodies.

491

492 Independently of its mode of formation, such a scenario is more difficult to
493 envision for the Moon for two reasons. First, the Moon is smaller ($R = 1700$
494 km), and in our models, the heat flow across the core mantle boundary does
495 not reach the critical value required to overcome ohmic dissipation. Second,
496 its volumetric metallic fraction is only 1 – 2% meaning that the impactor size
497 needed to differentiate a large volume of metal is unrealistically large. However,
498 our models do not consider any contribution of the central body to the impact

499 velocity. v_{esc} for the Earth is 4-5 times bigger than v_{esc} for the Moon. Consid-
 500 ering that the impact velocity on the Moon is equal to the escape velocity on
 501 the Earth can increase the initial temperature of the Moon's protocore and the
 502 protocore-mantle boundary heat flow by a factor ≈ 20 . Moreover, our choice of
 503 γ provides a lower bound on ΔT_0 . If V_∞ is not zero, heating within and outside
 504 the core will increase. Finally, radiogenic heating (*Williams and Nimmo, 2004;*
 505 *Nimmo, 2009*), latent heat release and compositional convection within the core
 506 (*Labrosse et al., 1997*) can also lead to a significant increase of the heat flow
 507 across the core-mantle boundary. The solidification of a 100 km radius inner
 508 core can release $\sim 10^{10}$ W which is 1000 times larger than Φ for a 200 km core
 509 radius planet. Under the assumption that a heat flow high enough to overcome
 510 Earth-like ohmic dissipation is required, these sources of heat and/or composi-
 511 tional buoyancy could explain an early dynamo on the Moon or on small radii
 512 objects.

513 *5.2. The effect of a molten lower mantle layer on dynamo duration*

If the rocks at the protocore-mantle boundary are at sufficiently high tem-
 perature and pressure to be close to their melting temperature, heat transfer
 from the protocore can cause the progressive growth of a mantle melt layer (see
 Eq.5 and Fig.2). Assuming that the undifferentiated mantle surrounding the
 protocore is homogenous, that gravity decreases linearly with depth and that
 $(R_{Fe}/R)^2 \ll 1$, the pressure at the core mantle boundary is:

$$P_{CMB} \approx \frac{2}{3}\pi G\rho_0^2 R^2 \quad (26)$$

514 For the impacted planet radii range in our models, P_{CMB} ranges from 10 to 50
 515 GPa. The formation of a molten layer at the top of the core depends on the
 516 initial temperature of the protocore which strongly depends on the impacted

517 planet radius R (see Eq.5). Melting occurs when the initial protocore temper-
518 ature is larger than the solidus of the surrounding material. In the isoviscous
519 case, Fig.2 shows that the protocore temperature overcome the solidus temper-
520 ature of the surrounding mantle only when $R > 3500$ km with a chondritic
521 composition (*Agee, 1997*) or when $R > 4500$ km with a dry peridotite compo-
522 sition (*Zerr and Boehler, 1994*) . When the viscosity of the diapir is smaller
523 than the surrounding undifferentiated solid mantle, the initial temperature of
524 the protocore is smaller and it is more difficult to form a molten layer unless
525 the pre-impact temperature T_0 at the centre is increased by ≈ 500 K. As men-
526 tioned in section 2.1, the pre-impact temperature is a function of the accretion
527 rate and of the early radiogenic heating. In our models the pre-impact thermal
528 state is homogenous and $T_0 = 1250$ K. Hence, in our single impact scenario,
529 the formation of a molten layer at the top of the protocore without additional
530 silicate differentiation would require a pre-impact temperature that increases
531 with depth.

532

The conductive model and condition given by Eq.25 is not appropriate if
heat transfer from the core melts the lower mantle. Vigorous convection within
a growing melt layer will enhance the heat transfer from the protocore. Thus,
mantle melting can increase the likelihood of dynamo action over the conditions
identified in Fig.4 but, in return, can decrease the duration of an early dynamo
as heat will be removed more quickly. Hence, a key issue is to know whether
convection in the melt layer will cause the core to freeze. In the case of a molten
layer, the duration of a dynamo is limited by the time needed to remove the
excess heat (the specific heat plus the latent heat of crystallization) from the

protocore with a heat flow Q . In that case, the duration of the dynamo is :

$$t_{dynamo} \approx \frac{4\pi R_{Fe}^3 \rho_{Fe} [C_{p,Fe}(T_{core} - T_{melt,Fe}) + L_{Fe}]}{3Q} \quad (27)$$

533 where $T_{melt,Fe}$ and L_{Fe} are the melting temperature and the latent heat of the
 534 metallic phase (see Tab.1 for typical values). Figure 6 shows the time needed
 535 to remove the heat excess as a function of the planetary radius for 3 different
 536 heat transfer types: solid state conduction, time-dependent conduction and
 537 convection within a molten layer. Our results show that this time can vary with
 538 several orders of magnitude. In the case of a conductive layer, the time to remove
 539 the heat excess is very long ($\approx 10^9$ yrs) and the duration of an early dynamo is
 540 essentially governed by the time during which vigorous convection occurs within
 541 the protocore ($\approx 10^6 - 10^7$ yrs). However, in the case where a molten convecting
 542 layer forms above the protocore, core freezing occurs in $10^3 - 10^4$ yrs, potentially
 543 limiting the duration of the dynamo to < 10 kyr. As underlined in Fig..2, the
 544 extent of melting depends on the composition of the surrounding rocks and on
 545 the easiness to cross the solidus. Hence, the composition of the early mantle is
 546 also a key parameter in the duration of the early dynamo.

547 5.3. Future investigations

548 This dynamo initiation model is a first step toward a more general model
 549 of core formation. From the results of our models, several questions need to be
 550 addressed:

- 551 1. How do hyper-velocity impacts affect the heat repartition after an impact
 552 and how does this in turn enhance an impact-induced early dynamo? The
 553 fraction of kinetic energy (γ) that is retained as heat below the impact site
 554 and the spatial heat repartition after an impact (h_m) are key parameters
 555 to characterize the early inner thermal state of growing planets. Better

556 constraints on these parameters would greatly assist any models that in-
557 vestigate the thermal consequences of impacts early in a planet’s history.

558

559 2. How multiple diapirs from multiple impacts will affect the early protocore
560 thermal state? The sinking of multiple metallic diapirs will strongly af-
561 fect the thermal state of both the core and the mantle and may play a
562 key role on the temperature contrast between these two reservoirs (*Golabek*
563 *et al.*, 2009; *Samuel et al.*, 2010). This prospect is currently beyond the
564 scope of this study because of the geometry used in our numerical models
565 (axisymmetric spherical). However, our study gives a parametrization of
566 the inner thermal consequences after a single large metallic diapir sinking.
567 The scalings proposed here could be used to compute more accurate one
568 dimensional thermal evolution models of growing planets.

569

570 3. What is the influence of the timescale between two successive impacts? If
571 a second impact occurs shortly after the first (i.e., within one thermal dif-
572 fusion timescale for first metallic diapir diameter), the thermal effect of the
573 first impact will still be present within the mantle and the second diapir will
574 sink into a warmer and less viscous mantle. The second diapir will descend
575 more quickly. Moreover, the viscosity contrast between the second diapir
576 and the hotter mantle will decrease and supplementary viscous heating will
577 occur within the second metallic diapir (see section 2.2).

578

579 4. In the framework of a late giant impact between two differentiated bodies,
580 what is the effect of the core merging on the dynamo activity? The thermal
581 effect of giant impact on dynamos has been recently studied (*Roberts et al.*,
582 2009; *Reese and Solomatov*, 2010) but the dynamics of merging between
583 the preexisting core and the impactor’s core and the thermal consequences

584 have also to be considered in future models. Within the framework of our
585 model, a giant impact might kill a preexisting dynamo depending on size
586 of the impactor as the thermal effect of both the impact and the viscous
587 heating during the sinking may reduce the total heat flow across the CMB.

588

589 5. How would a second large impact or merging phenomenon change the Hf-
590 W systematics? Depending on the efficiency of equilibration between the
591 impactor's core and the impacted body's mantle, this merging phenomenon
592 may be a major feature of the current W isotopes composition of planets
593 and influence the Hf-W chronology of accretion (*Jacobsen, 2005*).

594 6. Conclusions

595 The dynamics of terrestrial planetary accretion governs their differentiation
596 histories and their potential for early magnetic fields. We have shown that
597 under Earth-like conditions for ohmic dissipation in the core impactors (with
598 $R_{imp} > 625$ km) leading to major differentiation events during early stages of
599 planetary formations can initiate early dynamos. If the ohmic dissipation crite-
600 rion is relaxed, allowing dynamo action simply if the critical magnetic Reynolds
601 number in the core is exceeded, we find that dynamo action is possible for a
602 larger range of impactors and planets (notably smaller impactors and smaller
603 impacted bodies). Such early dynamos can persist for several kyr to several
604 Myr depending on the size of both the impactor and the impacted body and
605 depending on the heat transfer regime across the protocore. The duration of
606 the early dynamo is also a strong function of the rheology of the early mantle of
607 the impacted body. This rheology determine the sinking velocity of the metallic
608 diapir, the amount of viscous heating during the sinking and the ease to remove
609 the heat from the protocore in the case of a molten layer.

610

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Figure 1: Schematic view of the chemical equilibration following a large impact on an undifferentiated protoplanet. Within the isobaric region that results from the dissipation of the shock wave (time step t_1), the temperature increase (orange circle, t_2) melts the metal (red) that segregates rapidly from silicates (green) (t_3). The initial position of the inertial centre of the iron diapir is indicated with a black dot (i.c.). Then the metallic phase sinks by a diapiric instability (t_4). Subsequent convection within the protocore can initiate an early magnetic field (t_5). The top left panel shows where viscous heating (H_v) occurs for a spherical density anomaly (i.e. a metallic protocore) outlined by the white dashed line within a Rybczynski-Hadamard analytical velocity field (see text for details). $H_v \sim \underline{\underline{\epsilon}}^2$ where $\underline{\underline{\epsilon}}$ is the dimensionless strain rate tensor. The viscous heating is shown for 2 viscosity ratios, λ , between the hot protocore and the cold undifferentiated material. For $\lambda = 1$ (uniform viscosity), viscous heating can occur within the sphere, while for $\lambda \ll 1$, viscous heating only occurs outside the sphere. Here, H_v is normalized by the maximum value of H_v when $\lambda = 1$.

Figure 2: Initial temperature of the protocore once at the centre of the planet as a function of the planetary radius. Each symbol represents a numerical result for a given R_{Fe} (R_{Fe} in the range 150 – 500 km) for a uniform viscosity. Theoretical predictions for T_{core} from Eq.14 are given for uniform viscosity (black solid line, $a(\lambda) = 15\%$) and temperature dependent viscosity (black dashed line, $a(\lambda) = 0\%$). The black solid line is without melting and the grey one is with melting. The region in between defines the range of initial protocore temperature (see Eq.5). The dotted line shows $T = T_0$. The chondritic solidus (Agee, 1997) and the dry peridotite solidus (Zerr and Boehler, 1994) at $P = P_{CMB}$ are represented with dashed-double-dotted lines (see section 5.2 for calculation details).

Figure 3: Temporal evolution (symbols from black to white) of the ratio Q/Φ as a function of the planetary radius. For each planetary radius, we investigated several diapir sizes at times $t = 1$ and 50 Myr for a uniform viscosity. The length of the rectangles represents the variation of Q/Φ for the range of diapir sizes. The width of each rectangle is arbitrary. Theoretical predictions for (Q_{cond}/Φ) at $t = 0$ Myr from Eq.24 for uniform viscosity (black dashed line, $a(\lambda) = 15\%$) and temperature dependent viscosity (black dotted line, $a(\lambda) = 0\%$) are shown. Planetary and moon radii of small solar system objects are indicated on the top x-axis.

Table 1: Typical parameter values for numerical models

Planet radius	R	2000 - 4000 km
Diapir (protocore) radius	R_{Fe}	150 - 500 km
Density difference	$\Delta\rho_0$	4500 kg m ⁻³
Average density	ρ_0	4270 kg m ⁻³
Iron density	ρ_{Fe}	8000 kg m ⁻³
Silicate density	ρ_{Si}	3500 kg m ⁻³
Mean thermal expansion	α	4.4×10^{-5} K ⁻¹
Iron thermal expansion	α_{Fe}	1.5×10^{-5} K ⁻¹
Silicate thermal expansion	α_{Si}	5×10^{-5} K ⁻¹
Iron heat capacity	$C_{p,Fe}$	800 J K ⁻¹ kg ⁻¹
Silicate heat capacity	$C_{p,Si}$	1000 J K ⁻¹ kg ⁻¹
Average specific heat of the impacted body	$\overline{\rho_0 C_p}$	4×10^6 J K ⁻¹ m ⁻³
Initial temperature	T_0	1250 K
Iron melting temperature	$T_{melt,Fe}$	1250 K
Silicate melting temperature	$T_{melt,Si}$	1650 K
Iron latent heat	L_{Fe}	4×10^5 J kg ⁻¹
Silicate latent heat	L_{Si}	2.7×10^5 J kg ⁻¹
Thermal diffusivity	κ	10^{-6} m ² s ⁻¹
Mean thermal conductivity	$k = k_{Si}$	4 W m ⁻¹ K ⁻¹
Iron thermal conductivity	k_{Fe}	40 W m ⁻¹ K ⁻¹
Metal content	f_0	0.17
Reference viscosity	η_0	10^{22} Pa s
Impact energy conversion coefficient	γ	0.3-0.5
Volume effectively heated by impact	h_m	2.7
Gravitational constant	G	6.67×10^{-11} m ³ kg ⁻¹ s ⁻²
Average magnetic field strength	\overline{B}	2.5 mT
Magnetic diffusivity	ν	2 m ² s ⁻¹
Magnetic permeability	μ	$4\pi \times 10^{-7}$ H m ⁻¹

Figure 4: Q/Φ 1 Myr after the diapir reaches the center of the impacted planet as a function of the metallic diapir radius (black symbols). The protocore radius differentiated after a single impact corresponds to an impactor radius indicated on the top x-axis. The symbol shape is related to the radius of the impacted planet. The black dashed line represents the threshold expression from Eq.25 in the isoviscous case ($a(\lambda) = 15\%$). Above this dashed line (white domain) a dynamo is plausible, while below it no dynamo is generated. The dotted line corresponds to the same threshold but assuming a protocore less viscous than the surrounding mantle ($a(\lambda) = 0\%$). If Q/Φ after 50 Myr is within the dynamo domain, a second white filled symbol represents Q/Φ at this time.

Figure 5: Magnetic Reynolds number of the protocore as a function of its size from our numerical models. The protocore radius differentiated after a single impact corresponds to an impactor radius indicated on the top x-axis. The symbol shape is related to the radius of the impacted planet and its color (black or white) to the time after the diapir reaches the center of the impacted planet. The dotted line represents the numerical results expected from a temperature-dependent viscosity model (*i.e.*, Re_m from isoviscous models divided by a factor 1.4). The theoretical Reynolds number values obtained from Eq.21 are represented with a solid line (with ΔT_0 corresponding to $R=4000$ km) and a dashed line (with ΔT_0 corresponding to $R=2000$ km). The grey domain illustrates Re_m^{crit} above which a dynamo is plausible.

Figure 6: Time to remove the excess heat from the protocore for 3 different protocore-mantle boundary heat transfers as a function of the planetary radius. The dotted line represents the case of a conductive boundary layer (Eq.22, $\delta = R_{Fe}$ and Eq.27) and the dashed line represents the case of a time dependent conductive boundary layer (Eq.22, $\delta = 2\sqrt{\kappa t}$ and Eq.27). The black line shows the case where a molten layer surrounds the protocore (Eq.23, $\eta_{melt} = 100$ Pa.s and Eq.27). This time also depends on the size of the protocore. For each case, 2 lines delimit a domain from $R_{Fe} = 150$ km (lower line) to 500 km (upper line).