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# 1 Dynamics of core merging after a martian mega-impact

2 J. Monteux<sup>a,b</sup>, A. M. Jellinek<sup>b</sup>, and C. L. Johnson<sup>b,c</sup>

3 <sup>a</sup>*Laboratoire de Planétologie et de Géodynamique de Nantes, UMR CNRS, Université de*  
4 *Nantes, France.*

5 <sup>b</sup>*Department of Earth and Ocean Sciences, University of British Columbia, Canada.*

6 <sup>c</sup>*Planetary Science Institute, Tucson, AZ*

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## 7 **Abstract**

A number of geophysical arguments point out the possibility that a giant impact occurring within the first 500 Myr of the martian history could be responsible for the dichotomy between the northern lowlands and the southern highlands. This giant impact may also have important consequences on the initiation or cessation of the transient martian dynamo. At the time of the impact both the impactor and Mars were differentiated. Hence, an important volume of metallic material from the impactor probably merged with the martian pre-impact core. In this study, we characterize this core merging dynamics assuming that the impactor's core sank as a single diapir within the martian mantle and we obtain characteristic timescales of the merging time between the two cores as a function of the rheology and the size of the sinking diapir. For large impacts, the coalescence time should take less than 1 Myr even for a very viscous mantle and this time linearly decreases with decreasing mantle viscosity. We also measure the influence of the impact heating and of the viscous heating associated with the core merging on the martian thermal state as a function of the rheology and size of the impactor. Finally, we discuss the influence of such a giant impact on the magnetic history of Mars. We show that, assuming perfect thermal mixing between the hot merging core and the planetary core, dynamo activity might be favoured by large impacts, low mantle viscosity and a thin thermal boundary layer below the CMB. Considering that dynamo longevity depends on

how long it takes to remove the added thermal anomaly, post-impact dynamo can eventually last several hundred Myrs.

- *Key words:* meteoritical impacts; numerical modeling; mega-impacts; Mars;
  - dynamo.
-

## 10 1. Introduction

11 Mega impacts probably played an important role in the late history of ter-  
12 restrial planetary accretion. A Mars-size impact can, for example, explain the  
13 Earth-Moon system (*Hartmann and Davis, 1975*) and a large impact removing  
14 part of the silicate mantle is a hypothesis for the high iron/silicate ratio on Mer-  
15 cury (*Smith, 1979; Benz et al., 1988*). Among the hypotheses for the origin of  
16 the martian dichotomy (*Elkins-Tanton et al., 2003; Roberts and Zhong, 2006*),  
17 an exogenic origin by a mega impact that displaced crustal material from the  
18 northern to the southern hemisphere seems to be the most plausible candidate  
19 (*Wilhelms and Squyres, 1984; Nimmo et al., 2008; Andrews-Hanna et al., 2008;*  
20 *Marinova et al., 2008, 2011*). Models for the martian impact suggest that the  
21 impactor was 200 to 1300 km in radius and hit the planet with a speed com-  
22 parable to or larger than the martian escape velocity (i.e.  $v_{imp} > 5 \text{ km.s}^{-1}$ )  
23 within the first 500 Myr of Martian history (*Frey, 2006*).

24 In addition to a history of large impacts, Earth, Mercury and Mars have, or  
25 have had, an internally generated magnetic field. Mercury and the Earth have  
26 active core dynamos, whereas Mars had only a transient internally generated  
27 magnetic field (*Acuña et al., 1999; Hood et al., 2003; Lillis et al., 2008a*) that  
28 ceased by around 4.0 Ga (*Acuña et al., 1999; Johnson and Phillips, 2005; Lillis*  
29 *et al., 2008b*). The timing of the initiation of the martian dynamo is difficult to  
30 constrain and strongly depends on the differentiation processes that occurred  
31 during the first million years of the martian history (*Monteux et al., 2011*). The  
32 cessation of the martian dynamo is also still currently debated. Recent models  
33 have measured the influence of large impacts on the dynamo generation process  
34 and in particular on the cessation of core convection induced by the reduction  
35 of the core-mantle boundary (CMB) heat-flux (*Roberts et al., 2009; Watters*  
36 *et al., 2009; Arkani-Hamed and Olson, 2010; Roberts and Arkani-Hamed, 2012*).

37 However, recent models show that the thermal anomaly induced by a large  
38 impact impact and the formation of a hot molten iron layer from the impactor's  
39 core cumulating at the CMB can also favor dynamo generation (*Reese et al.*,  
40 2010).

41 At the time of the proposed giant impact, both Mars and the impactor were  
42 probably differentiated (*Yoshino et al.*, 2003). Models suggest that although  
43 some material was ejected far from Mars, the majority of the mass of the im-  
44 pactor's core was retained within Mars and merged with the pre-impact Martian  
45 core (*Canup, 2004*) (*Marinova does not consider the impactor's core*). The aim  
46 of this study is to characterize the dynamics of core merging as a result of a  
47 diapiric descent of molten iron (*Monteux et al.*, 2009) on a Mars-size planet. In  
48 addition, we investigate how the processes of impact and core merging might in-  
49 fluence the thermal regime of Mars' core and, in turn, magnetic field generation  
50 (cf., *Monteux et al.* (2011)).

## 51 **2. Thermo-chemical state before the martian mega-impact**

### 52 *2.1. Pre-impact interior of Mars*

53 The initial structure and thermal state of a growing planet depends particu-  
54 larly on the characteristics of its accretion from chondritic material (*Safronov*,  
55 1978; *Kaula, 1979; Agee, 1997*). During accretion, heating driven by a combina-  
56 tion of the dissipation of impact energy and the decay of short lived radionuclides  
57 such as  $^{26}\text{Al}$  and/or  $^{60}\text{Fe}$  (*Yoshino et al.*, 2003; *Monteux et al.*, 2007) increases  
58 the mean internal temperature and gives rise to a radial temperature gradient  
59 that depends on the accretion rate relative to the rate of radiative cooling to  
60 space (*Kaula, 1979; Senshu et al.*, 2002). If the growth rate is very high in com-  
61 parison to surface cooling, this heating can ultimately cause partial or complete  
62 melting of the chondritic material (*Yoshino et al.*, 2003) and lead to extensive

63 metal/silicate separation (*Tonks and Melosh, 1992; Senshu et al., 2002; Monteux*  
64 *et al., 2009*).

65 Hf/W chronology suggests that core formation happened during the first 10-  
66 30 million years of Mars' history (*Lee and Halliday, 1997; Nimmo and Kleine,*  
67 *2007*). Such a rapid process involves extensive melting potentially enhanced by  
68 radiogenic heating as a result of the decay of short-lived radionuclides (*Yoshino*  
69 *et al., 2003*), impact heating (*Tonks and Melosh, 1992; Senshu et al., 2002;*  
70 *Monteux et al., 2009*) and gravitational energy conversion during metal/silicate  
71 separation (*Stevenson, 1989; Ricard et al., 2009*). Metal/silicate separation can  
72 occur via a wide range of phenomena such as percolation (*Shannon and Agee,*  
73 *1996*), the sedimentation of metallic rain through a magma ocean (*Rubie et al.,*  
74 *2003; Höink et al., 2005*) or a large diapir sinking through a solid mantle after  
75 an impact (*Tonks and Melosh, 1992; Monteux et al., 2009*). Whatever the mech-  
76 anism, Mars' internal structure characterized by a  $\sim 1700$  km diameter Fe-core  
77 was mostly established within  $\sim 10$  Myr of the planet's formation (*Yoder et al.,*  
78 *2003*) (cf Fig.1a).

79 The gravitational heat released during martian core formation was parti-  
80 tioned between the planet's core and mantle. The fraction of gravitational heat  
81 taken up by the metal or the silicate fraction depends strongly on the rheology  
82 of the planet and on the segregation mechanisms (*Samuel and Tackley, 2008;*  
83 *Monteux et al., 2009; Ke and Solomatov, 2009*). The combined processes leading  
84 to core formation yield a wide range of possible early thermal states. In partic-  
85 ular, the core could initially have had a temperature close to the deep mantle  
86 temperature if thermal equilibration was efficient or it could have been hotter  
87 than the mantle if the energy of core formation was mainly retained within the  
88 iron. In the latter case, the transition between the hot core and the relatively  
89 cooler mantle would occur within a thermal boundary layer with a thickness,  $\delta$ ,

90 inversely proportional to the efficiency of core cooling by convection (Fig.1a).

## 91 2.2. Impactor interior

92 Assuming that both the impactor and the impacted body had chondritic  
93 compositions, their volumetric metal fractions  $f_0$  should be close (we consider  
94 that the impactor has the same metal content as Mars and we use  $f_0 = 12.5\%$   
95 (*Stevenson, 2001*)). Hence, for 200-1300 km diameter impactors, an additional  
96 volume of core material with a radius between 100 km and 700 km merges  
97 with the preexisting core (Fig.1). The interior temperature of the impactor is  
98 unconstrained. However, as the impactor is smaller in radius than Mars, it was  
99 probably cooler, on average than the martian interior.

## 100 3. Thermo-chemical state after a mega-impact

### 101 3.1. Mantle heating and melting

102 Although part of the kinetic energy of the impactor is dissipated as a result  
103 of the irreversible work done by shock waves to damage crustal rocks, another  
104 part of this energy is dissipated thermally within a spherical region (the isobaric  
105 core) with a volume  $V_{ic}$  that is typically taken to be 3 times larger than the  
106 volume of the impactor (*O'Keefe and Ahrens, 1977; Croft, 1982; Pierazzo et al.,*  
107 *1997*). In this region the temperature increases uniformly from  $T_0$  to  $T_0 + \Delta T_0$ ,  
108 where  $T_0$  is the temperature of the impacted mantle before the impact and  $\Delta T_0$   
109 is the temperature increase due to impact heating and melting. Assuming that  
110 part of this thermal energy is also consumed to melt the isobaric core and the  
111 impactor's iron core with volume  $V_{Fe} = f_0 V_{ic}/3$ , this temperature increase is  
112 (*Monteux et al., 2011*):

$$\Delta T_0 = \frac{4/9\pi\gamma\rho_0^2GR^2 - L_{Si}\rho_{Si} - \frac{f_0}{3}L_{Fe}\rho_{Fe}}{h_m\rho_{Si}C_{p,Si}}, \quad (1)$$

113 Here,  $G$  is the gravitational constant,  $\rho_{Si}$ ,  $L_{Si}$  and  $C_{p,Si}$  are the density, latent  
114 heat and specific heat of the silicate material,  $R$  is the radius of the impacted  
115 planet and  $\gamma$  is the fraction of the kinetic energy of the impactor ultimately  
116 dissipated to heat up the mantle (*O’Keefe and Ahrens, 1977; Monteux et al.,*  
117 *2007*).  $L_{Fe}$  and  $\rho_{Fe}$  are the density and latent heat of the impactor’s core.  
118 As  $f_0$  is  $O(10^{-1})$  then the contribution to  $\Delta T_0$  from the third term in Eq.1 is  
119 negligible and we can simplify Eq.1 to:

$$\Delta T_0 = \frac{4/9\pi\gamma\rho_0^2GR^2 - L_{Si}\rho_{Si}}{h_m\rho_{Si}C_{p,Si}}, \quad (2)$$

120 The effect of the energy lost as a result of the irreversible work done by shock  
121 waves outside the isobaric core is accounted for in the geometric parameter  $h_m$ ,  
122 which is the volume of rock damaged and heated inside and outside the isobaric  
123 core normalized to the volume of the isobaric core itself (see *Monteux et al.*  
124 *(2011)* for details). From empirical fits to the fall off in pressure with distance  
125 from the edge of the isobaric core  $h_m \approx 2 - 3$  (*Senshu et al., 2002*). Using the  
126 set of parameters in Table 1 and making the conservative hypothesis that the  
127 martian mantle is solid before the impact (i.e. its temperature is just below the  
128 melting temperature of the mantle silicates),  $\Delta T_0 \sim 400$  K.

129 Outside the isobaric core, the shock pressure and the excess temperature de-  
130 crease rapidly with distance  $r$  from the isobaric core as approximately  $\Delta T_0(R_{ic}/r)^m$   
131 (*Senshu et al., 2002; Monteux et al., 2009*). Hence, the volume effectively heated  
132 and damaged after the impact is larger than the isobaric core (see Fig.2). Con-  
133 sequently, when  $R_{imp} > 500$  km, impact heating occurs within the planetary  
134 core. Because the shockwave velocity is lower in liquid than in solid material,  
135 the dissipation rate and the impact heating will be weaker in the liquid core  
136 than in the surrounding silicate mantle (*Arkani-Hamed and Olson, 2010*). For  
137 simplicity, we will assume that there is no difference in impact heating between



138 the martian core and mantle. Hence, for large impactors, our calculations over-  
139 estimate the direct impact heating within the martian core.

140

### 141 3.2. *The fate of the impactor's material*

142 During the mega-impact, a significant fraction of the impactor's mantle is  
143 ejected far from the impact site and eventually forms a disk of debris. The rest  
144 of the impactor's mantle can eventually be incorporated into the mantle of the  
145 impacted planet. In contrast, most of the impactor's core is retained within the  
146 impacted planet (*Canup, 2004*) (cf Fig.1b-c), and in our models we assume that  
147 100% of the impactor's core is retained within the martian mantle after the gi-  
148 ant impact. Once incorporated in the impacted planet, the impactor's core can  
149 overcome shearing and thinning processes (especially if the impact is oblique)  
150 as well as turbulent motions in a magma ocean that can break up large volumes  
151 of liquid metal into droplets (*Rubie et al., 2003; Deguen et al., 2011; Samuel,*  
152 *2012*). However, the impactor's core may merge with the core of the impacted  
153 planet, without emulsification on a droplet lengthscale, if the impactor's core  
154 is larger than the thickness of the magma ocean and if the impact is nearly  
155 vertical (*Rubie et al., 2003; Reese et al., 2010; Dahl and Stevenson, 2010*). As  
156 mentioned earlier, we consider the pre-impact mantle to be solid, and we also  
157 take the impact trajectory to be nearly vertical. Thus, we hypothesize that the  
158 impactor's core retains its spherical shape after the impact and that its centre  
159 is buried within the martian mantle.

160

## 161 4. Dynamic models of diapir sinking

162 Once buried below the surface, the dense metallic phase from the impactor's  
163 core sinks towards the centre of the impacted planet (Fig.1). The large devia-

164 toric stress generated by the metallic diapir can lead to a non linear rheology  
 165 (*Samuel and Tackley, 2008*), elasto-plastic deformation (*Gerya and Yuen, 2007*)  
 166 or even to fracturing processes (*Davies, 1982*). These produce dissipation in  
 167 the mantle surrounding the sinking core (*Turcotte and Emerman, 1983; Steven-*  
 168 *son, 2003*). If the iron/silicate viscosity contrast is small, viscous dissipation  
 169 (*i.e.* heating) occurs in both the silicate mantle and the sinking core. Alter-  
 170 natively, if the iron/silicate viscosity contrast is large, the viscous dissipation is  
 171 concentrated in the silicate mantle, but heat can then diffuse into the sinking  
 172 iron core (*Samuel and Tackley, 2008; Monteux et al., 2009*).

#### 173 4.1. Physical model: Descent of a single diapir

174 To investigate the dynamics of the metallic diapir, we adapt the numeri-  
 175 cal finite volume model in spherical axisymmetric geometry of *Monteux et al.*  
 176 (2009). We assume that the mantle is deforming in a diffusion creep limit. The  
 177 conservation of energy applied to a planet of radius  $R$  leads to

$$\frac{DT}{Dt} = \frac{\nabla^2 T}{Ra_\chi} + Di \left( \frac{\eta}{\eta_0} \Gamma \Omega - v_r \left( T + \frac{T_0}{\Delta T_0} \right) r \right), \quad (3)$$

178 where  $T$ ,  $t$  and  $r$  are dimensionless temperature, time and radius.  $v_r$  is the  
 179 dimensionless radial velocity.  $Ra_\chi = \frac{\Delta \rho_0 g_0 R^3}{\kappa \eta_0}$  is the compositional Rayleigh  
 180 number and  $Di = \frac{\alpha \rho_0 g_0 R}{\rho_0 C_p}$  is the dissipation number (with  $\kappa$  the heat diffusivity  
 181 and  $\overline{\rho_0 C_p}$  the average specific heat of the impacted body).  $\Delta \rho_0$  is the density  
 182 difference between metal and silicates. Following *Monteux et al. (2009)*, we in-  
 183 troduce the buoyancy ratio  $\Gamma = \Delta \rho_0 / (\rho_0 \alpha \Delta T_0)$  where  $\rho_0$  and  $\alpha$  are the mean  
 184 density and thermal expansion of the impacted material (Tab.1). Gravity de-  
 185 pends on radius  $r$ . Within the impacted planet mantle, we consider that gravity  
 186 is constant  $g_0$  with  $g_0 = \frac{4}{3} G \pi \rho_0 R$ . Within the core, gravity is a linear function  
 187 of the radius and  $g(r) = g_0 \frac{r}{R_c}$ .  $\Omega$  is the dimensionless dissipation function and

188 expresses the conversion of potential energy into heat :

$$\Omega = 2 \underline{\underline{\varepsilon}} : \underline{\underline{\varepsilon}}. \quad (4)$$

189 where  $\underline{\underline{\varepsilon}}$  is the dimensionless shear strain rate tensor. Prior to impact, we  
 190 assume a homogenous temperature  $T_0$  in the martian mantle. The viscosity is  
 191  $\eta = \eta_0 \lambda_0^T$  where  $\lambda_0$  is the viscosity ratio (less than 1) between the hot impacted  
 192 material and cold surrounding material far from the impact site,  $\eta_0$ , at the start  
 193 of the experiment (*Ratcliff et al., 1997; Ziethe and Spohn, 2007*).  
 194 The other dimensionless governing equations are continuity

$$\nabla \cdot \mathbf{v} = 0, \quad (5)$$

195 and momentum conservation, assuming infinite Prandtl number

$$-\nabla P + \nabla \cdot \left( \frac{\eta}{\eta_0} \left[ \nabla \mathbf{v} + [\nabla \mathbf{v}]^T \right] \right) + \left( \frac{T}{\Gamma} - f \right) r \mathbf{e}_r = 0, \quad (6)$$

196 where  $\mathbf{v}$  and  $P$  are the non-dimensional velocity and pressure and  $\mathbf{e}_r$  is the radial  
 197 unit vector. The buoyancy force that drives the flow of the diapir towards the  
 198 centre of the protoplanet increases with the metallic volume fraction  $f$  that  
 199 varies between 1 (pure metal) and 0 (pure silicates). The metal volume fraction  
 200  $f$  is then simply advected by the flow :

$$\frac{Df}{Dt} = 0. \quad (7)$$

#### 201 4.2. Numerical model

202 We implement a finite volume numerical model to solve Eq.3, Eq.5, Eq.6 and  
 203 Eq.7 in axisymmetric spherical geometry. We use a stream function formulation  
 204 for the equations of motion with a direct implicit inversion method (*Schubert*

205 *et al.*, 2001). Eq.3 and Eq.7 are solved by an Alternating Direction Implicit  
 206 (ADI) scheme (*Peaceman and Rachford*, 1955; *Douglas*, 1955). The stream  
 207 function, temperature and compositional fields are described by a second-order  
 208 approximation in space. To limit numerical diffusion when solving the transport  
 209 equations, especially for the compositional field, we use a Total Variation Dimin-  
 210 ishing Superbee scheme (*Roe*, 1986; *Laney*, 1998) implemented in an implicit  
 211 way (*Srámek et al.*, 2010) which enables a high resolution of pure advective  
 212 fields. We use at least  $200 \times 400$  grid points. Velocity boundary conditions are  
 213 free-slip at the surface and along the symmetry axis. Thermal boundary condi-  
 214 tions are isothermal at the surface and insulating along the symmetry axis.

#### 215 4.3. Analytical model of the core merging process

216 Buoyancy-driven interactions between two deformable viscous drops has  
 217 been widely studied both experimentally and numerically (*Davis et al.*, 1989;  
 218 *Yiantsios and Davis*, 1990; *Manga and Stone*, 1993). Building on this body  
 219 of work, we characterize the interaction between the preexisting core and the  
 220 sinking diapir by monitoring the gap distance,  $h$ , between the surfaces of the  
 221 impacting core and the planetary core along the axis of symmetry (Fig.1c-d).  
 222 We compare the results from our numerical models with analytical predictions.  
 223 We consider that the initial time  $t_0$  is the time at which the impact occurs.

224 During sinking, a diapir will deform the surrounding mantle material, driving  
 225 a circulation with a spatial extent that depends on the diameter of the diapir  
 226 and on the rheology of the mantle. When the diapir is small and far from  
 227 the core-mantle boundary, the presence of the preexisting core has no effect on  
 228 the diapir motion and the diapir sinks with a classical Rybczynski-Hadamard  
 229 velocity  $V$  (*Hadamard*, 1911; *Rybczynski*, 1911) (Fig.3, left)

$$V = -\frac{dh}{dt} = \frac{a_0 \Delta \rho_0 g_0 R_{Fe}^2}{\lambda \eta_0}. \quad (8)$$

230  $a_0$  is a geometrical constant and  $\lambda$  is the viscosity ratio between the metallic  
 231 phase and the mantle. Integration of Eq.8 leads to

$$h(t) = h_0 \left( 1 - \frac{t}{t_s} \frac{R_{Fe}}{h_0} \right) \quad (9)$$

232 where  $h_0$  is the initial gap thickness and  $t_s$  the characteristic time given by

$$t_s = \frac{R_{Fe}}{V} = \frac{\lambda \eta_0}{a_0 \Delta \rho_0 g_0 R_{Fe}} \quad (10)$$

233 The Rybczynski-Hadamard velocity gives the terminal velocity of a spherical  
 234 viscous drop sinking through an infinite viscous fluid. Hence it differs from the  
 235 Stokes velocity which is the terminal velocity of a solid sphere sinking through  
 236 an infinite viscous fluid. In our models, the mechanical boundary conditions are  
 237 applied at a finite distance: the planetary surface and the core-mantle bound-  
 238 ary, CMB. Thus, the Rybczynski-Hadamard velocity is only an approximation  
 239 (*Honda et al.*, 1993; *Samuel and Tackley*, 2008) and we consider in our models  
 240 that the time during which this regime occurs is negligible.

241

242 When the viscosity of the preexisting planetary core is comparable to or less  
 243 than that of the mantle, then the planetary core offers little resistance to the  
 244 radial flow generated by the sinking core. Hence, the CMB acts as a mobile  
 245 interface. When  $h(t) \sim R_{Fe}$  and as the diapir approaches the planet's core,  
 246 the merging core slows down and the planetary core starts to deform while the  
 247 sinking diapir is weakly deformed (Fig.3, middle). As the distance between the  
 248 diapir and the CMB decreases, the normal stress difference becomes more local-  
 249 ized (*Lee and Leal*, 1982). When the diapir is at a distance  $h(t) \ll R_{Fe}$  a late  
 250 lubrication regime is established in the thin gap between the two cores (Fig.3,  
 251 right).

252

253 The dynamics of interaction between two drops merging in a fluid with dif-  
254 ferent viscosity is governed by the viscosity ratio between the two drops and the  
255 surrounding material and can be treated in two regimes: a short-time regime  
256 (referred here as "early times") (Fig.3, middle) during which the preexisting  
257 core deforms to accommodate the merging between the two cores and a lubrica-  
258 tion long-time regime (referred here as "late times") (Fig.3, right) during which  
259 the narrowing of the gap between the two cores tends to slow down the rate of  
260 fluid drainage within the gap. These two regimes are separated in time by a  
261 characteristic coalescence time  $t_c$ .

262

263 At "early times" (where  $h \geq R_{Fe}$  and  $t < t_c$ ) (Fig.3, middle), the flow  
264 circulation induced by the sinking diapir and the deformation of the preexisting  
265 core tends to facilitate the drainage of silicate fluid from the region between the  
266 merging cores. According to *Yiantsios and Davis* (1990) the flow speed is given  
267 by

$$V = -\frac{dh}{dt} = \frac{a_1 \Delta \rho_0 g_0 R_{Fe}^{3/2} h^{1/2}}{\lambda \eta_0} \quad (11)$$

268 where  $a_1$  is a geometrical constant that we determine from our numerical models.  
269 Integration of Eq.11 leads to

$$h(t) = h_0 \left(1 - \frac{t}{t_c}\right)^2 \quad (12)$$

270 and

$$t_c = \frac{2h_0^{1/2} \lambda \eta_0}{a_1 \Delta \rho_0 g_0 R_{Fe}^{3/2}} \quad (13)$$

271

272 At "late times" (where  $h < R_{Fe}$  and  $t > t_c$ ) (see Fig.3, right), the rate at which

273 the gap thins out as a function of time depends whether the CMB act as a rigid  
 274 boundary or a mobile interface. According to *Jones and Wilson* (1978) and  
 275 *Yiantsios and Davis* (1990), for two free boundaries the gap thickness should  
 276 scale with:

$$h(t) = a_2 t^{-1/3} \quad (14)$$

277 where  $a_2$  is a relative deformation scale ( $a_2 \propto BoR_{Fe}/h_0$ ) that can be deter-  
 278 mined from our numerical models and  $Bo = \Delta\rho_0 g_0 R_{Fe}^2 / \sigma$  is the Bond number  
 279 which is the dimensionless ratio of the buoyancy forces to the surface tension  
 280  $\sigma$  (*Yiantsios and Davis*, 1990). In our models,  $Bo$  is theoretically infinite since  
 281 surface tension forces are neglected but has a numerical finite value that can  
 282 be determined fitting our numerical results with theoretical predictions from  
 283 Eq.14.

#### 284 4.4. Numerical results

##### 285 4.4.1. Initial setup

286 We use the numerical model described earlier to characterize the dynamics  
 287 of core merging after a giant impact on a Mars size body. In this section, the  
 288 martian core is homogeneously hot with  $T = T_{core} = 2000$  K (*Roberts and*  
 289 *Arkani-Hamed*, 2012) (i.e.  $\delta = 0$ ). The initial thermo-chemical conditions  
 290 are shown schematically in Fig.2 in which the hatched material represent the  
 291 metallic phase from the impactor. The initial temperature is represented with  
 292 a grey scale.

##### 293 4.4.2. Evolution of the gap thickness

294 Figure 4 represents the thermal and chemical evolution after a 600 km im-  
 295 pact (i.e.  $R_{Fe} = 300$  km) on a differentiated Mars with homogeneous viscosity  
 296 ( $\eta_0 = 10^{22}$  Pa.s). To simplify our study, we fixed the initial gap thickness  $h_0$

297 (i.e. the depth at which the impactor’s core is buried after impact) to a value  
 298 that does not vary with  $R_{imp}$ . We choose  $h_0 = 580$  km so that the impactor’s  
 299 core is initially always within the isobaric core for the whole range of impactor  
 300 radii studied here. The four rows correspond to real time snapshots at 0, 0.88,  
 301 2.2, 82.4 Myr respectively. As illustrated in Figure 4, the core from the impactor  
 302 rapidly sinks toward the pre-existing core. During the sinking, heating occurs at  
 303 the interface between the diapir and the surrounding mantle (second line, left).  
 304 Simultaneously, deformation of the pre-existing core accommodates the merging  
 305 (second line, right). At the end of the chemical merging, the impactor’s core is  
 306 significantly hotter than it was initially (third line). If the end of the merging  
 307 occurs rapidly (within 6 Myrs), the thermal readjustment needs a longer time  
 308 and the extra-heat from impact and viscous dissipation is then removed by a  
 309 large thermal plume that starts later from the CMB (fourth line).

310

311 We monitored the gap thickness  $h$  between the front of the merging core  
 312 and the CMB as a function of time for different sizes of impact  $R_{imp}$  (equiv-  
 313 alent to different sizes of merging core,  $R_{Fe}$ , with  $R_{Fe} = R_{imp}/2$ ). We used  
 314  $R_{imp}$  ranging from 200 to 800 km. Fig.5 illustrates the temporal evolution of  
 315 the gap thickness (black solid line) for the case illustrated in Fig.4 ( $R_{imp} = 600$   
 316 km,  $R_{Fe} = 300$  km,  $\eta_0 = 10^{22}$  Pa.s and  $h_0 = 580$  km). We compare the ana-  
 317 lytical models and the numerical models for both the short early time regime  
 318 (red dashed line, Eq.12) and the long late time regime (green dashed line Eq.14).

319

320 From our numerical results, we obtain the characteristic coalescence times  
 321  $t_c$  and the numerical value for  $a_2$  for the range of  $R_{imp}$  used in this study.  
 322 As predicted from Eq.13, our numerical values for  $t_c$  decrease in proportion to  
 323  $R_{Fe}^{-3/2}$  (red circles, Fig. 6). We fit the numerical  $t_c$  values with the analytical



324 prediction (black dashed line, Fig. 6) from Eq.13 and obtained  $a_1 = 1/9$ .

325 *4.4.3. Influence of the temperature-dependance of the mantle viscosity on the*  
326 *core merging timescale*

327 The viscosity contrast between the sinking diapir and its environment is a  
328 key parameter in understanding the core merging dynamics. As the impactor's  
329 core is sinking, shear heating occurs at the interface of the diapir (*Samuel et al.*,  
330 2010) and both the temperature of the diapir and the surrounding mantle in-  
331 crease. Depending on the rheological contrast between the metallic diapir and  
332 the mantle, the mean temperature increase within the sinking diapir can reach  
333  $\approx 2\Delta T_0$  (*Monteux et al.*, 2009). This thermal modification can influence the  
334 sinking dynamics especially if the rheology is temperature dependant. More-  
335 over, as illustrated in Fig.3, viscous dissipation may also occur in the planetary  
336 core as a result of the shear flow produced at the CMB during the late stages  
337 of core merging especially in the non realistic case where the planetary core  
338 and its mantle have the same viscosity (i.e., the core viscosity is way too large  
339 compared to the real problem). However, this heating process is negligible com-  
340 pared to the temperature increase induced by the merging of a hot diapir in the  
341 preexisting core.

342  
343 We introduced a temperature-dependent viscosity in our core-merging mod-  
344 els. We chose the same initial thermal state as previously used with a core hotter  
345 than the surrounding mantle ( $T_{core} > T_0$ ). As the temperature of the impacted  
346 mantle and of the impactor's core are initially equal, the impactor's core initially  
347 starts to sink in a medium with an homogenous temperature. Hence, our results  
348 illustrate the effect of viscous heating of the surrounding material on the diapir  
349 dynamics. In our models, the viscosity varies with  $\eta = \eta_0 \lambda_0^T$  (with  $T$  the nor-  
350 malized temperature), hence the viscosity within the impacted mantle will be

351  $\approx \eta_0 \lambda_0$  while the mean viscosity of the metallic diapir at the end of the sinking  
 352 will be  $\approx \eta_0 \lambda_0^2$  (since its temperature increases by  $\sim 2\Delta T_0$ ). Consequently, a  
 353 decrease of  $\lambda_0$  by one order in magnitude in our models should decrease the vis-  
 354 cosity contrast between the diapir and the surrounding material and the sinking  
 355 time by one order of magnitude. However, the viscosity contrast between the  
 356 diapir and the cold mantle far from the impact site will reach approximately  
 357 two orders of magnitude.

358

359 To measure the effect of the rheology on the coalescence time, we plot  $h(t)$   
 360 in Figure 7 for four different viscosity parameters: ( $\eta_0 = 10^{22}$  Pa.s,  $\lambda_0 = 1$ ),  
 361 ( $\eta_0 = 10^{20}$  Pa.s,  $\lambda_0 = 1$ ), ( $\eta_0 = 10^{22}$  Pa.s,  $\lambda_0 = 0.1$ ) and ( $\eta_0 = 10^{22}$  Pa.s,  
 362  $\lambda_0 = 0.01$ ) with  $R_{Fe} = 300$  km ( $R_{imp} = 600$  km) and  $h_0 = 580$  km (see Figure  
 363 7). When the viscosity is homogenous, the preexisting core has an influence on  
 364 the core merging dynamics and Eq.12 is valid. However, where viscous dissipa-  
 365 tion heats the sinking diapir and softens the surrounding mantle, core merging  
 366 is faster and the sinking dynamics is close to a classic Stokes regime with a ter-  
 367 minal Rybczynski-Hadamard velocity in agreement with Eq.9. We fit uniform  
 368 viscosity evolutions for  $h(t)$  with Eq.12 and temperature dependent evolutions  
 369 for  $h(t)$  with Eq.9 and obtain the characteristic sinking times listed in Tab.2.

370

371 For a uniform viscosity, the decrease of  $\eta_0$  from  $10^{22}$  to  $10^{20}$  Pa.s reduces  
 372 the characteristic sinking time from 1.56 Myr to 15.6 kyr. For a temperature  
 373 dependant viscosity, the characteristic sinking time decreases as  $\lambda$  decreases. As  
 374 predicted from Eq.10 and Eq.13, the relation between the characteristic sinking  
 375 time and the rheology contrast is almost linear (Table 1).

## 376 5. Effect on the preexisting dynamo

377 We now discuss the consequences of a giant-impact and the subsequent core  
378 merging process on the martian dynamo generation. We first list the conditions  
379 required for dynamo generation and then measure the influence of the core  
380 merging on the efficiency of core cooling for different structures of the thermal  
381 boundary layer within the planetary core before the impact.

### 382 5.1. Conditions required for dynamo generation

383 As discussed in *Monteux et al.* (2011), three conditions are required to gen-  
384 erate a dynamo on a growing planet:

- 385
- 386 1. The mean heat flux  $q_{CMB}$  out of the core must exceed the adiabatic value  
387  $q_A$  such that convection can occur (*Stevenson et al.*, 1983). This condition  
388 is

$$q_{CMB} > q_A = \frac{k_{Fe} \alpha_{Fe} g_c T_c}{C_{p,Fe}} \quad (15)$$

389 where  $k_{Fe}$ ,  $\alpha_{Fe}$ ,  $T_c$  and  $C_{p,Fe}$  are respectively the thermal conductiv-  
390 ity, the thermal expansion, the temperature and the heat capacity of the  
391 metallic phase and  $g_c$  is the gravity at the surface of the protocore. The  
392 largest uncertainties in the calculation of  $q_A$  concern  $T_c$  and  $k_{Fe}$ . Assum-  
393 ing reasonable values, convection in the martian core should start once  
394  $q_{CMB} > q_A = 5 - 19 \text{ mW.m}^{-2}$  (*Nimmo and Stevenson*, 2000) (see Tab.1  
395 for values).

- 396
- 397 2. The ratio of the rate at which gravitational potential energy is released by  
398 convection to the rate of ohmic dissipation,  $\Phi$ , must exceed a critical value  
399 (*Buffett*, 2002):

$$\frac{4\pi R_c^2 q_{CMB}}{\Phi} > \frac{1}{\epsilon_T}. \quad (16)$$

400 Here,  $\epsilon_T$  is the Carnot-style efficiency for thermal convection. Here, we do  
 401 not consider the effect of chemical convection or the presence of an inner  
 402 core. Assuming that the characteristic length scale of the flow leading to  
 403 magnetic field generation is the radius of the protocore,  $\Phi$  can be approxi-  
 404 mated as (*Buffett, 2002*):

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

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

$$\epsilon_T = \frac{0.8\pi}{3} \frac{\alpha_{Fe} G \rho_{Fe} R_c^2}{C_{p,Fe}} \left( 1 - \frac{q_A}{q_{CMB}} \right), \quad (18)$$

411 with  $\rho_{Fe}$  the density of the metallic phase and  $G$  the gravitational constant.

412

413 Hence Eq.16, Eq.17 and Eq.18 lead to:

$$q_{CMB} - q_A > \frac{\nu \bar{B}^2 C_{p,Fe}}{0.8\pi G \mu \alpha_{Fe} \rho_{Fe} R_c^3} \approx 8. \times 10^{-2} \text{ mW.m}^{-2} \quad (19)$$

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

417 and Aubert, 2006)

$$Re_m = \frac{UL}{\nu} > Re_m^{crit} = O(10 - 10^2) \quad (20)$$

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

$$U \sim \left( \frac{(q_{CMB} - q_A) \alpha_{Fe} g_c R_{Fe}}{\rho_{Fe} C_{p,Fe}} \right)^{1/3}, \quad (21)$$

426 Taking  $L = R_c$  and  $Re_m^{crit} = 10^2$ , the combination of Eq.20 and Eq.21  
 427 leads to the condition

$$q_{CMB} - q_A > \frac{\rho_{Fe} C_{p,Fe}}{\alpha_{Fe} g_c R_c} \left( Re_m^{crit} \frac{\nu}{R_c} \right)^3 \approx 10^{-4} \text{ mW.m}^{-2} \quad (22)$$

428 Among the three criteria above, the first is typically considered a necessary  
 429 condition for a thermally-driven dynamo (Subadiabatic dynamos are possible  
 430 if *e.g.*, compositional gradients also drive convection.). However it is not a  
 431 sufficient condition for dynamo action, as indicated by the other two conditions.  
 432 To get the two last conditions, the heat flux  $q_{CMB}$  at the CMB has to be larger  
 433 than the adiabatic heat flux  $q_A$  but the difference required is small (between  
 434  $8. \times 10^{-2}$  and  $10^{-4} \text{ mW.m}^{-2}$ ). Hence we will consider from now that any heat  
 435 flux larger than  $q_A$  can potentially lead to a dynamo generation (*Nimmo and*

436 *Stevenson, 2000; Williams and Nimmo, 2004).*

437 *5.2. Structure of the thermal boundary layer within the core before the impact*

438 To understand the thermal influence of the merging process on the thermally-  
439 driven dynamo action, we need to have an idea of the structure of the thermal  
440 boundary layer on the core side of the core-mantle boundary prior to impact,  
441 which depends on the previous accretion and core formation histories (*Monteux*  
442 *et al.*, 2009; *Samuel et al.*, 2010). Two end-member models for the initial inte-  
443 rior thermal state of Mars are the isothermal case with the core and the mantle  
444 having the same temperature, and a situation where the core is much hotter  
445 than the mantle.

446

447 In the case of a pre-impact core hotter than the mantle, the thickness  $\delta$  of  
448 the TBL is governed by the mode of heat loss to the mantle. Whereas strong  
449 cooling to the mantle leads to thin thermal boundary layers and vigorous con-  
450 vection, weak cooling leads to relatively more sluggish flow or no convection at  
451 all. To simplify this problem we consider three pre-impact core thermal states:  
452 an isothermal core and mantle, and two models where the core is hotter than  
453 the mantle with  $\delta = 0.1R_{Fe}$  (strong core cooling) and  $\delta = 0.4R_{Fe}$  (weak core  
454 cooling). In the last two models the temperature decreases linearly in the ther-  
455 mal boundary layer from the planetary core where we impose  $T = T_{core}$  to the  
456 CMB where  $T = T_0$ .

457

458 *5.3. Evolution of the CMB heat flow after impact: Numerical results.*

459 Following a mega-impact extensive mantle heating causes the average core  
460 heat flow to decline or reverse direction. This phenomenon has been inferred  
461 to explain the cessation of the martian dynamo (*Roberts et al.*, 2009; *Watters*

462 *et al.*, 2009). For large impactors, the resulting core heating can lead to strati-  
463 fication within the core and subsequent dynamo cessation (*Arkani-Hamed and*  
464 *Olson*, 2010). However lower mantle heating from the impact can drive and  
465 focus the formation of mantle plumes, which in turn increases core heat loss  
466 and can enhance the strength and longevity of a dynamo (*Reese et al.*, 2010).

467

468 Depending on the pre-impact core formation and accretionary heating his-  
469 tories, as well as the energetics of post-impact sinking discussed in section 3,  
470 a wide range of early martian thermal states are plausible. We monitor  $q_{CMB}$   
471 for different size of impactors (i.e. merging core radii), viscosity contrasts and  
472 efficiencies of core cooling (i.e. thermal boundary layer thickness). Assuming  
473 perfect thermal mixing of the pre- and post-impact cores, Fig.8 shows that core  
474 merging leads to a significant increase in the mean heat flux  $q_{CMB}$ . Our results  
475 show that a heat flux sufficiently large to drive a dynamo is favored for large  
476 impacts, low viscosities and for strongly-cooled pre-impact cores. In addition,  
477 dynamo longevity is enhanced for low core viscosities.

#### 478 5.4. Consequences for the dynamo generation

479 Among the recent models that correlate giant-impact and martian dynamo  
480 activity (*Roberts et al.*, 2009; *Watters et al.*, 2009; *Arkani-Hamed and Olson*,  
481 2010; *Reese et al.*, 2010), only the last one considers the thermal consequences  
482 of the core merging and assumes the formation of a hot layer at the CMB.  
483 However, none of these models consider the viscous dissipation during the core  
484 merging and the subsequent thermal perturbation at the CMB. Even if our  
485 model does consider this effect, the results from the previous section have to be  
486 considered carefully. The dynamo generation process is not a straightforward  
487 consequence of the core merging process and some assumptions on the thermal  
488 mixing within the martian core are needed to correlate core merging and dy-

489 namo histories. Indeed, if thermal mixing is not perfect and the merging core  
490 forms a hot layer at the CMB this thermal stratification may kill a thermally  
491 driven dynamo (*Arkani-Hamed and Olson, 2010*). The heat flux at the CMB  
492 might be greater than the critical value for a dynamo but the heat flux into  
493 the core might be large enough to make it go subadiabatic. However, if the hot  
494 layer mixes efficiently (Fig.1f), our results may give a picture of the dynamo  
495 generation after a giant impact.

496

497       Assuming perfect thermal mixing, our results show that dynamo initiation  
498 might be favoured by large impacts, low viscosity of the preexisting core and  
499 strong core cooling (i.e. thin thermal boundary layer). Assuming that dynamo  
500 longevity depends on how long it takes to remove the added thermal anomaly,  
501 post-impact dynamo can last several hundred Myrs. Whether or not the new  
502 core material will mix as it spreads is not certain. The answer depends on the  
503 density contrast between the merging core and the preexisting core and in par-  
504 ticular how hot is the merging core. The mixing also depends on the strength of  
505 the preexisting core convection. Whether ambient convective motions plus the  
506 shear related to the spread of the hot current cause mixing has to be calculated  
507 and depends on the balance between driving inertia and stabilizing buoyancy  
508 forces.

509

510       In our models, the mean heat flux  $q_{CMB}$  is overestimated because all the  
511 hot merging material stays at the CMB. However, we treat the core and mantle  
512 as having the same viscosity so the spreading of the new core material is in  
513 the wrong dynamical limit. Considering the effect of core merging on the core  
514 dynamics in a more realistic way (i.e. with finite Reynolds number) is currently  
515 beyond the scope of this paper and will be developed in a separated study in



516 which our results will constrain the initial thermal conditions within the core.  
517 Mixing will be enhanced at very high Reynolds number, depending on the den-  
518 sity difference between the intruded and pre-existing cores (see e.g. *Linden*  
519 (1977, 1979)).

520

521 On the other hand, our results are also quite conservative since we assume  
522 that the giant impact occurs with the martian escape velocity. When the two  
523 planetary embryos have collided near the end of accretion, the impact velocity  
524 was probably higher than that of the planetesimals during the accretion period.  
525 Considering larger impact velocities can severely increase the post-impact tem-  
526 perature in the mantle and in the merging core before the sinking from Eq.1  
527 and Eq.2. Hence, we give only a lower bound of the temperature increase in the  
528 core after the diapir merging.

## 529 **6. Conclusions**

530 Giant impacts may have played an important role in the thermal and mag-  
531 netic evolutions of terrestrial planets. The giant impact forming the martian  
532 dichotomy could have supplied a significant volume of iron that has merged  
533 with the preexisting core. The size of the impact and the rheology strongly  
534 influenced the dynamics of core merging which probably occurred in less than a  
535 million years. During the sinking, the mantle and the merging core have over-  
536 come viscous heating which have modified the thermal state at the core mantle  
537 boundary. Hence, depending on the mixing efficiency in the core, the heat sup-  
538 ply might have enhanced a dynamo activity.

539

540 Our models better constrain the dynamics of the sinking of a large metallic  
541 diapir within a silicate mantle and especially the influence of the deformation of

542 the preexisting core on the merging process. We obtain characteristic timescales  
543 of the coalescence time between the two cores as a function of the viscosity of  
544 the mantle and the size of the sinking diapir. Our results are in agreement with  
545 theoretical and experimental models of buoyancy-driven motions sinking toward  
546 a deformable surface. We also show that the thermal consequences of the merg-  
547 ing process can influence the generation of a martian dynamo. Assuming perfect  
548 thermal mixing, the dynamo generation might be favoured by large impacts, low  
549 viscosity of the preexisting core and strong preexisting core cooling. This dy-  
550 namo can eventually last several hundred million years depending on the mixing  
551 process within the core. Our results also illustrate the importance of better con-  
552 straining the thermal states of the cores, the size of the biggest diapir that can  
553 remain consistent after a giant impact and the depth at which the impactor's  
554 core is buried before experiencing viscous deformation and core merging process.

555

556 The heating induced by the impact within the mantle and the viscous dis-  
557 sipation during diapir sinking have enhanced a thermal anomaly within the  
558 martian mantle and, hence, a dichotomy of the core cooling. Indeed, the core  
559 heat flux at the CMB below the impact site was probably different than the  
560 heat flux at the opposite side of the core. This mantle heterogeneity might  
561 affect the dynamo and the paleomagnetic field recorded at the surface (*Stan-*  
562 *ley et al., 2008*). Moreover, recent results show that the efficiency of a mantle  
563 heterogeneity centred at the geographical pole in producing a south-north di-  
564 chotomy is much higher than that of an heterogeneity centred at the equator in  
565 producing an east-west dichotomy (*Amit et al., 2011*). Hence, our models may  
566 have interesting implications in the understanding of the structure of the past  
567 martian dynamo.

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Table 1: Typical parameter values for numerical models

Mars radius	$R$	3400 km
Mars core radius	$R_c$	1700 km
Impactor radius	$R_{imp}$	200 - 800 km
Impactor core radius	$R_{Fe}$	100 - 400 km
Density difference	$\Delta\rho_0$	4500 kg m <sup>-3</sup>
Average density	$\rho_0$	4060 kg m <sup>-3</sup>
Iron density	$\rho_{Fe}$	8000 kg m <sup>-3</sup>
Silicate density	$\rho_{Si}$	3500 kg m <sup>-3</sup>
Mean thermal expansion	$\alpha$	$4.5 \times 10^{-5}$ K <sup>-1</sup>
Iron thermal expansion	$\alpha_{Fe}$	$1.5 \times 10^{-5}$ K <sup>-1</sup>
Silicate thermal expansion	$\alpha_{Si}$	$5 \times 10^{-5}$ K <sup>-1</sup>
Iron heat capacity	$C_{p,Fe}$	800 J K <sup>-1</sup> kg <sup>-1</sup>
Silicate heat capacity	$C_{p,Si}$	1000 J K <sup>-1</sup> kg <sup>-1</sup>
Iron latent heat	$L_{Fe}$	$2.7 \times 10^5$ J kg <sup>-1</sup>
Silicates latent heat	$L_{Si}$	$4 \times 10^5$ J kg <sup>-1</sup>
Pre-impact mantle temperature	$T_0$	1600 K
Heat diffusivity	$\chi$	$10^{-6}$ m <sup>2</sup> s <sup>-1</sup>
Thermal conductivity	$k$	4 W m <sup>-1</sup> K <sup>-1</sup>
Metal content	$f_0$	12.5%
Reference viscosity	$\eta_0$	$10^{20}$ - $10^{22}$ Pa s
Impact energy conversion coefficient	$\gamma$	0.3
Volume effectively heated by impact	$h(m)$	2.7
Gravitational constant	$G$	$6.67 \times 10^{-11}$ m <sup>3</sup> kg <sup>-1</sup> s <sup>-2</sup>
Average magnetic field strength	$\overline{B}$	2.5 mT
Magnetic diffusivity	$\nu$	2 m <sup>2</sup> s <sup>-1</sup>
Magnetic permeability	$\mu$	$4\pi \times 10^{-7}$ H m <sup>-1</sup>

Table 2: Sinking time values obtained fitting numerical experiments with theoretical predictions ( Eq.9 and Eq.12) for different values of  $\eta_0$  and  $\lambda$  (with  $R_{imp} = 600$  km,  $h_0 = 580$  km)

	$\eta_0 = 10^{22}$ Pa.s $\lambda = 1$	$\eta_0 = 10^{20}$ Pa.s $\lambda = 1$	$\eta_0 = 10^{22}$ Pa.s $\lambda = 0.1$	$\eta_0 = 10^{22}$ Pa.s $\lambda = 0.01$
$t$	1.56 Myr	$1.56 \times 10^{-2}$ Myr	$9.5 \times 10^{-2}$ Myr	$9.5 \times 10^{-3}$ Myr

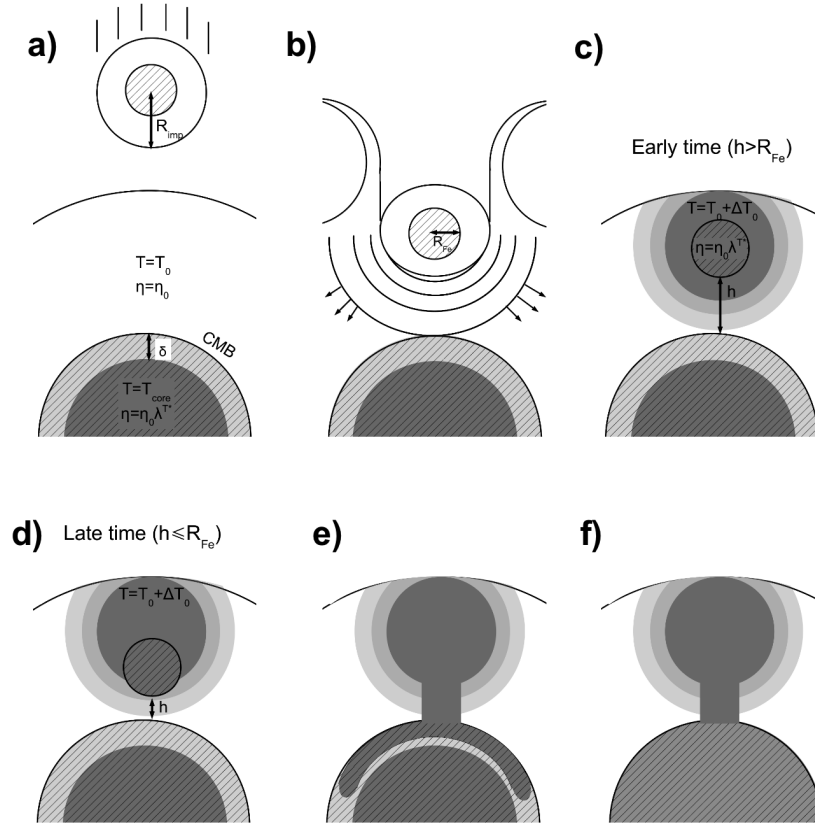


Figure 1: Schematic representation of the thermo-chemical evolution following a mega-impact on a Mars-size body. After the impact (a) and the subsequent shockwave propagation (b), a large thermal anomaly remains and the impactor's core is buried within the martian mantle (c). Then, this dense metallic material rapidly sinks towards the pre-impact martian core. During the sinking, potential energy is converted as heat in the mantle via viscous dissipation. Two sinking regimes occur (c and e) depending on the distance  $h$  between the impactor and the martian core (see section 4). Finally the intruded metallic material merges with the preexisting core, spreads at its top (e) and eventually mixes with it later (f) (which we don't model here). The hatched volumes represent the metallic material from the core and the impactor.

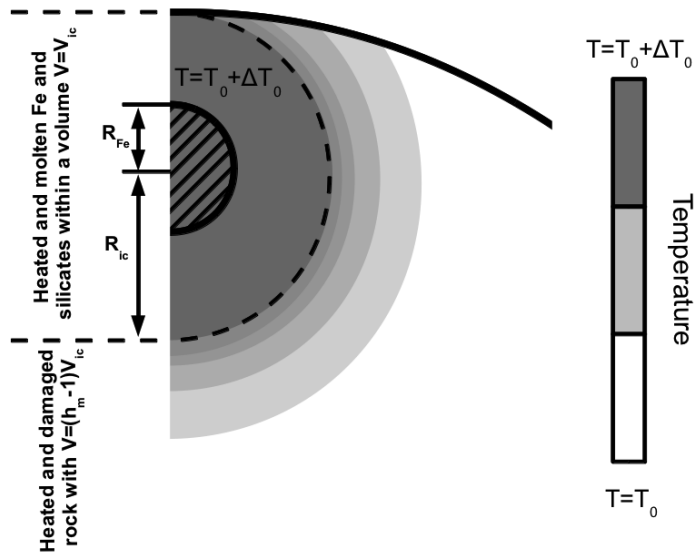


Figure 2: Sketch to show repartitioning of the post impact heating in the martian mantle. The solid black line represents the planetary surface. Melting is restricted to the material limited by the dashed line. The impactor's core is represented hatched lines.

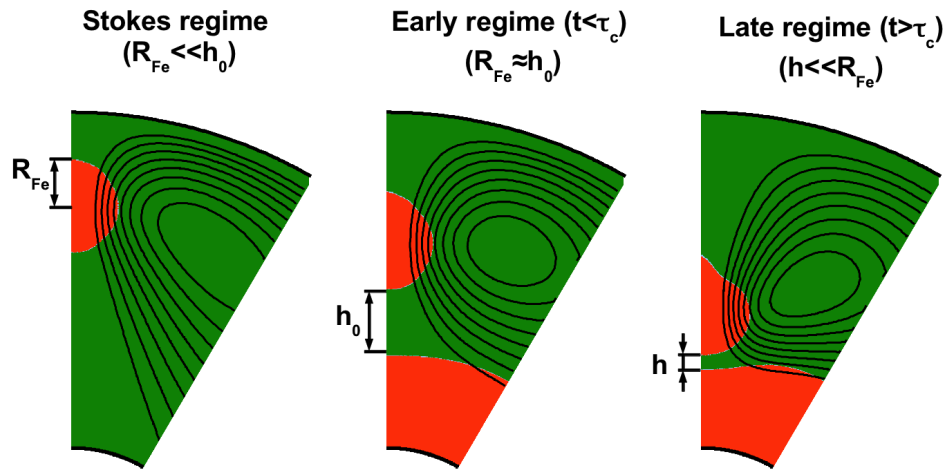


Figure 3: Dynamics of core merging for the three regimes described in section 4.3 and computed for  $R_{Fe} = 300$  km sinking in an isoviscous differentiated Mars size planet (silicate material is represent in green and metallic material is represented in red). Solid black lines illustrate the streamlines resulting from the diapir sinking. In the left figure, where we do not consider the presence of a planetary core, the viscous stress scales with  $1/R_{Fe}$ . In the middle figure, interaction with the core start to matter and the viscous stress scales with  $1/h_0$ . In the right figure where  $h < R_{Fe}$ , interactions between the diapir and the merging core increase and the lubrication force scales with  $1/h$ .



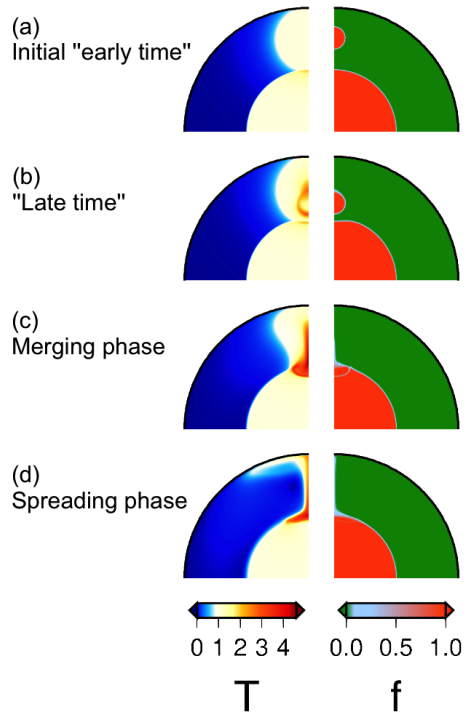


Figure 4: Non-dimensional temperature (left) and composition (right) at times  $t = 0$  (a),  $t = 0.6$  Myr (b),  $t = 6$  Myr (c) and  $t = 38.8$  Myr (d) (computed for a uniform viscosity with  $R = 3400$  km,  $R_{imp} = 600$  km and  $300 \times 600$  grid points). As the diapir sinks, its velocity decreases because of the increasing influence of the pre-existing core on the dynamics (see section 4.3 for details). After the merging (c), the hot metallic fraction from the impactor spreads at the top of the martian core (d).

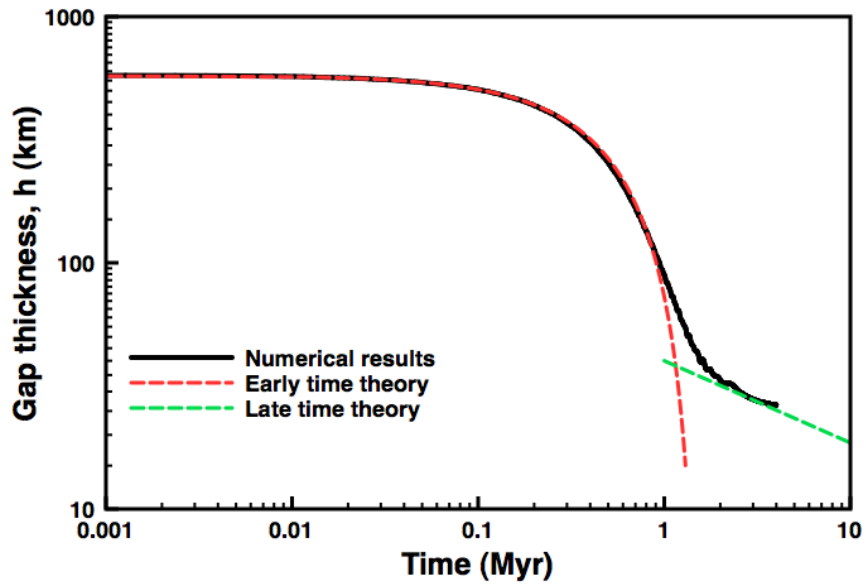


Figure 5: Evolution of the gap thickness between the merging core and the CMB with  $R_{imp} = 600$  km (i.e.  $R_{Fe} = 300$  km). Theoretical evolutions for early times from Eq.12 and late times from Eq.14 are shown with red and green dashed lines respectively. From this model, we obtain a coalescence time,  $t_c = 1.55$ , Myr and  $a_2 \sim Bo = 1.3 \times 10^9$ .

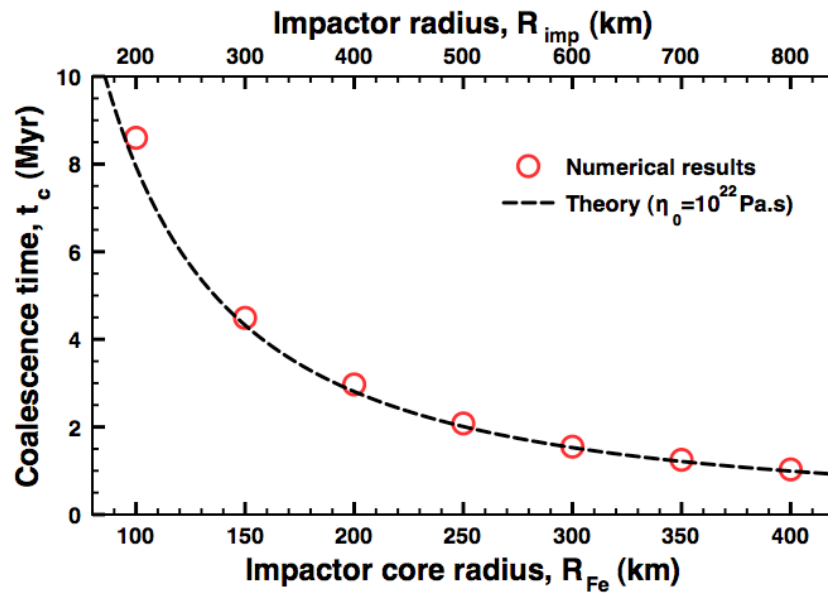


Figure 6: Characteristic coalescence time,  $t_c$ , as a function of the impactor core radius  $R_{Fe}$ . Results from numerical experiments (with uniform viscosity  $\eta_0 = 10^{22}$  Pa.s and  $h_0 = 580$  km) are represented with red circles. Theoretical fit from Eq.13 is shown with the black dashed line ( $a_1 = 1/9$ )

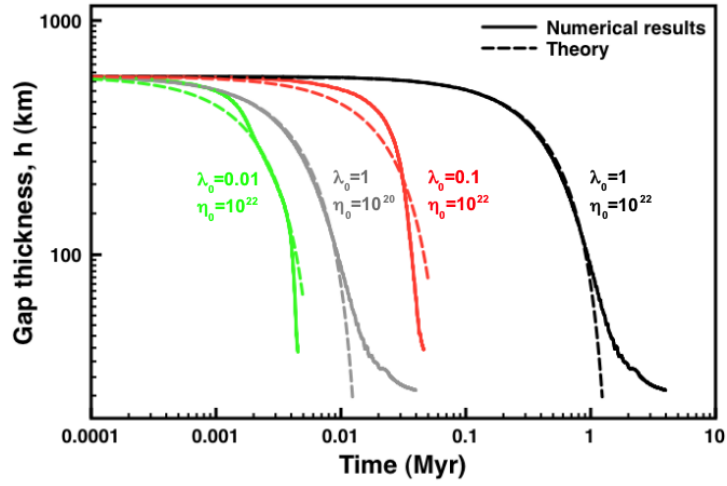


Figure 7: Time evolution of the gap thickness between the sinking core and the pre-existing core for four different rheologies. Solid black line and solid grey line represent results from uniform viscosity models with  $\eta_0 = 10^{22}$  and  $\eta_0 = 10^{20}$  respectively. Solid red line and solid green line represent results from temperature dependent viscosities with  $\lambda = 0.1$  and  $\lambda = 0.01$  respectively (with  $\eta = \eta_0 \lambda^T$  and  $\eta_0 = 10^{22}$  Pa.s). Theoretical predictions are shown in the corresponding color with dashed lines for the coalescence theory (Eq.12) or with dotted lines for the Rybczynski-Hadamard theory (Eq.9).

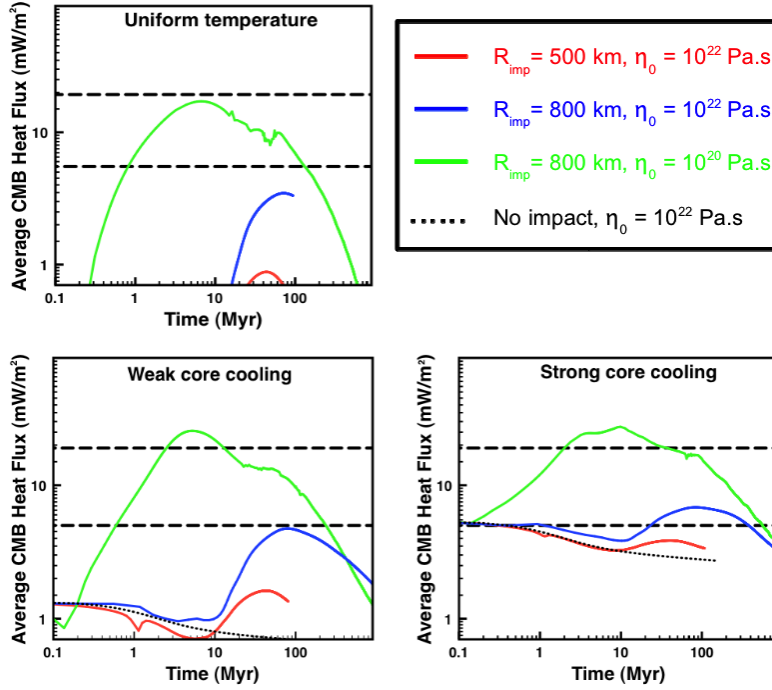


Figure 8: Time evolution of the mean CMB heat flux as a function of the rheology and impactor size. In the three panels different initial thermal states before impact are considered. The two horizontal black dashed lines represent the theoretical dynamo criterion for  $q_{CMB}$  ( $q_{CMB} > q_A = 5 - 19 \text{ mW.m}^{-2}$ ). In the top-left panel, pre-impact temperature is assumed to be constant ( $T = T_0$ ). In the bottom panels, the core is initially hotter than the mantle  $T = T_{core} > T_0$ . In the bottom left panel the CMB thermal boundary layer thickness  $\delta = 0.4R_{Fe}$  and in the bottom right panel  $\delta = 0.1R_{Fe}$ . Red, blue and green solid lines represent the evolution with ( $R_{imp} = 500 \text{ km}$ ,  $\eta_0 = 10^{22} \text{ Pa.s}$ ), ( $R_{imp} = 800 \text{ km}$ ,  $\eta_0 = 10^{22} \text{ Pa.s}$ ), and ( $R_{imp} = 800 \text{ km}$ ,  $\eta_0 = 10^{20} \text{ Pa.s}$ ) respectively. The black dotted line represent  $q_{CMB}$  without any impact.