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1 Dynamics of core merging after a martian mega-impact

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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}Al and/or ^{60}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 ~ 1700 km diameter Fe-core
77 was mostly established within ~ 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, δ ,

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 ΔT_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*):

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

Here, G is the gravitational constant, ρ_{Si} , L_{Si} and $C_{p,Si}$ are the density, latent heat and specific heat of the silicate material, R is the radius of the impacted planet and γ is the fraction of the kinetic energy of the impactor ultimately dissipated to heat up the mantle (*O’Keefe and Ahrens, 1977; Monteux et al., 2007*). L_{Fe} and ρ_{Fe} are the density and latent heat of the impactor’s core. As f_0 is $O(10^{-1})$ then the contribution to ΔT_0 from the third term in Eq.1 is negligible and we can simplify Eq.1 to:

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

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

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

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

3.2. The fate of the impactor's material

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

4. Dynamic models of diapir sinking

Once buried below the surface, the dense metallic phase from the impactor's 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 Emmerman, 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 κ 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 ρ_0 and α 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 $g(r) = g_0 \frac{r}{R_c}$. Ω 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 λ_0 is the viscosity ratio (less than 1) between the hot impacted
 192 material and cold surrounding material far from the impact site, η_0 , at the start
 193 of the experiment (*Ratcliff et al.*, 1997; *Ziethé 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

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

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 200×400 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 λ 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

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

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

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

4.4. Numerical results

4.4.1. Initial setup

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

4.4.2. Evolution of the gap thickness

Figure 4 represents the thermal and chemical evolution after a 600 km impact (i.e. $R_{Fe} = 300$ km) on a differentiated Mars with homogeneous viscosity ($\eta_0 = 10^{22}$ 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 λ_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 η_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 λ 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:

- 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} , α_{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).

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

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

Here, ϵ_T is the Carnot-style efficiency for thermal convection. 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*):

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

where \bar{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 \bar{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*)

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

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

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

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

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*

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 ν 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 δ 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 ⁻³
Average density	ρ_0	4060 kg m ⁻³
Iron density	ρ_{Fe}	8000 kg m ⁻³
Silicate density	ρ_{Si}	3500 kg m ⁻³
Mean thermal expansion	α	4.5×10^{-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
Iron latent heat	L_{Fe}	2.7×10^5 J kg ⁻¹
Silicates latent heat	L_{Si}	4×10^5 J kg ⁻¹
Pre-impact mantle temperature	T_0	1600 K
Heat diffusivity	χ	10 ⁻⁶ m ² s ⁻¹
Thermal conductivity	k	4 W m ⁻¹ K ⁻¹
Metal content	f_0	12.5%
Reference viscosity	η_0	10 ²⁰ -10 ²² Pa s
Impact energy conversion coefficient	γ	0.3
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

Table 2: Sinking time values obtained fitting numerical experiments with theoretical predictions (Eq.9 and Eq.12) for different values of η_0 and λ (with $R_{imp} = 600$ km, $h_0 = 580$ km)

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

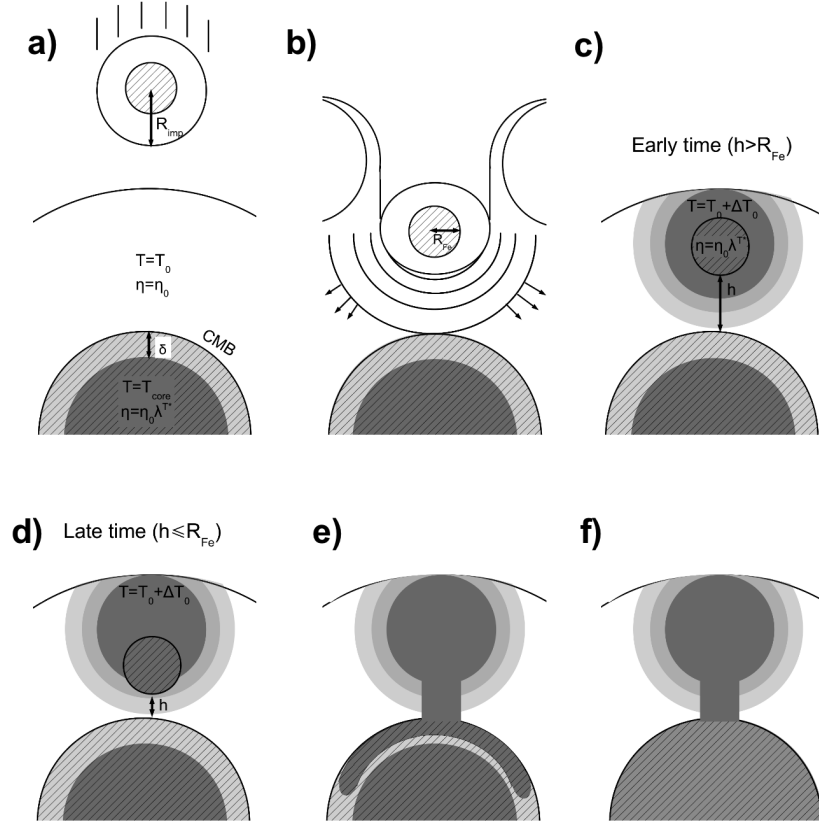


Figure 1: Schematic representation of the thermo-chemical evolution following a mega-impact on a Mars-size body. After the impact (a) and the subsequent shockwave propagation (b), a large thermal anomaly remains and the impactor's core is buried within the martian mantle (c). Then, this dense metallic 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.

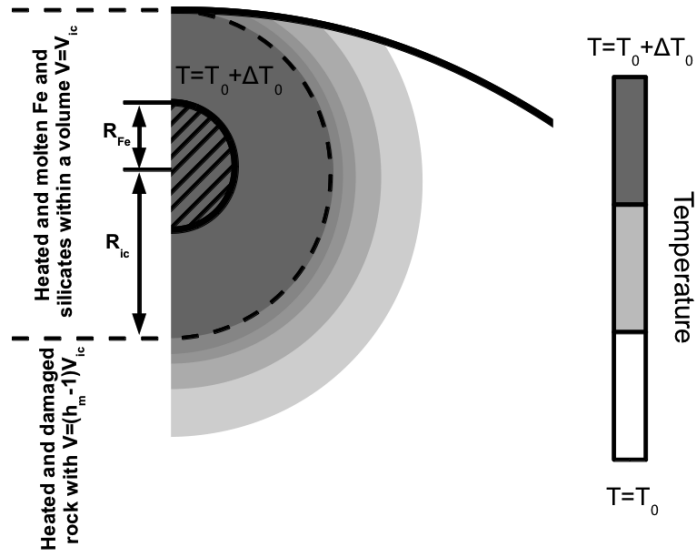


Figure 2: Sketch to show repartitioning of the post impact heating in the martian mantle. The solid black line represents the planetary surface. Melting is restricted to the material limited by the dashed line. The impactor's core is represented hatched lines.

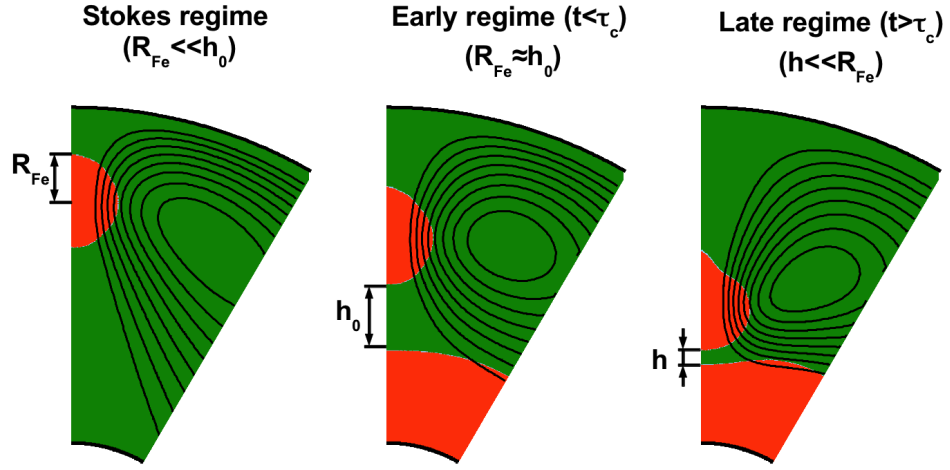


Figure 3: Dynamics of core merging for the three regimes described in section 4.3 and computed for $R_{Fe} = 300$ km sinking in an isoviscous differentiated Mars size planet (silicate material is represent in green and metallic material is represented in red). Solid black lines illustrate the streamlines 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 $1/h$.

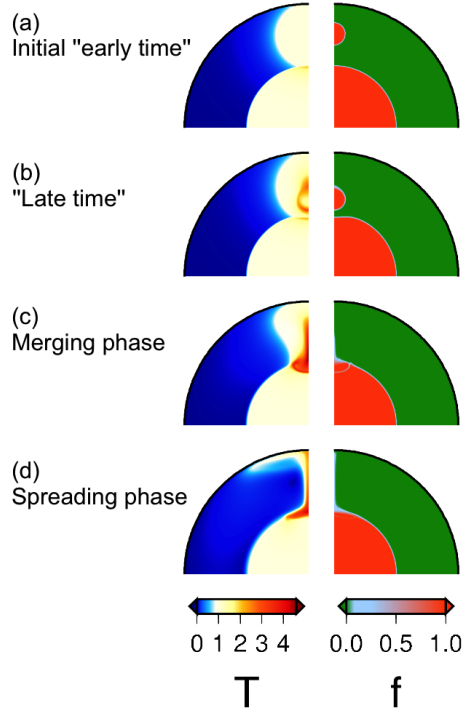


Figure 4: Non-dimensional temperature (left) and composition (right) at times $t = 0$ (a), $t = 0.6$ Myr (b), $t = 6$ Myr (c) and $t = 38.8$ Myr (d) (computed for a uniform viscosity with $R = 3400$ km, $R_{imp} = 600$ km and 300×600 grid points). As the diapir sinks, its velocity decreases because of 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).

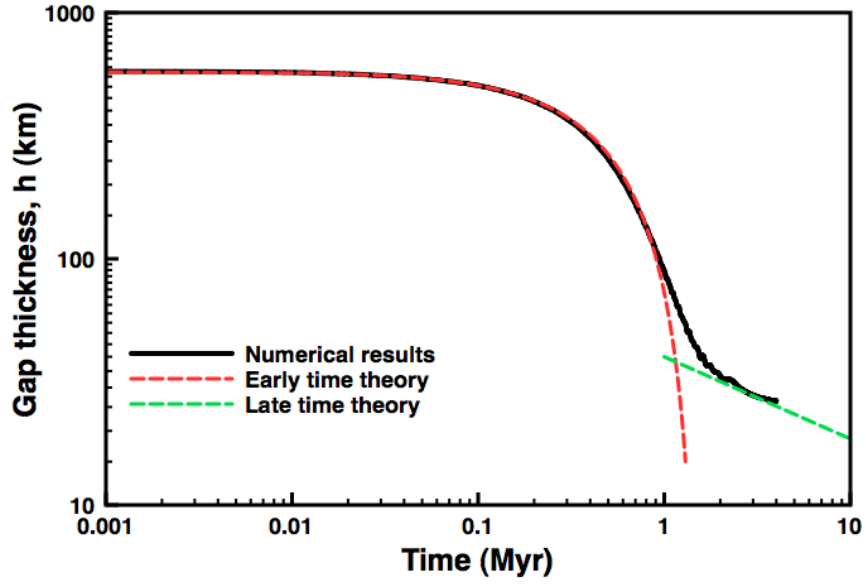


Figure 5: Evolution of the gap thickness between the merging core and the CMB with $R_{imp} = 600$ km (i.e. $R_{Fe} = 300$ km). Theoretical evolutions for early times from Eq.12 and late times from Eq.14 are shown with red and green dashed lines respectively. From this model, we obtain a coalescence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

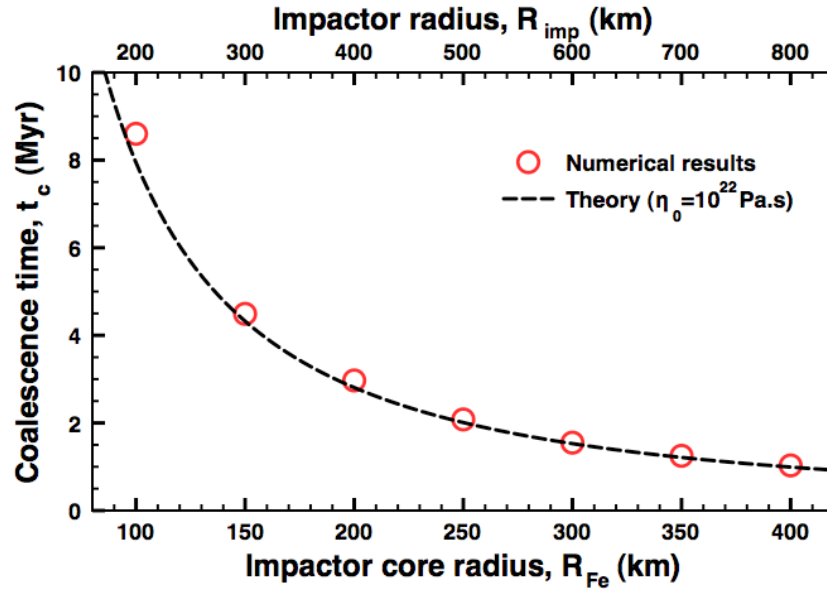


Figure 6: Characteristic coalescence time, t_c , as a function of the impactor core radius R_{Fe} . Results from numerical experiments (with uniform viscosity $\eta_0 = 10^{22}$ Pa.s and $h_0 = 580$ km) are represented with red circles. Theoretical fit from Eq.13 is shown with the black dashed line ($a_1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

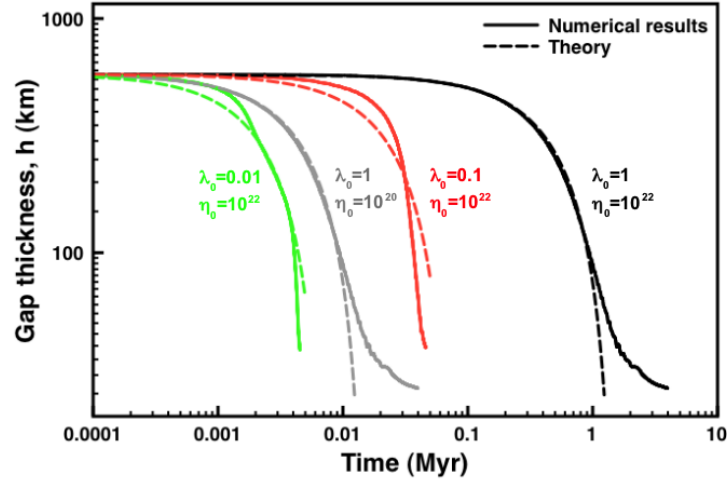


Figure 7: Time evolution of the gap thickness between the sinking core and the pre-existing core for four different rheologies. Solid black line and solid grey line represent results from uniform viscosity models with $\eta_0 = 10^{22}$ and $\eta_0 = 10^{20}$ respectively. Solid red line and solid green 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).

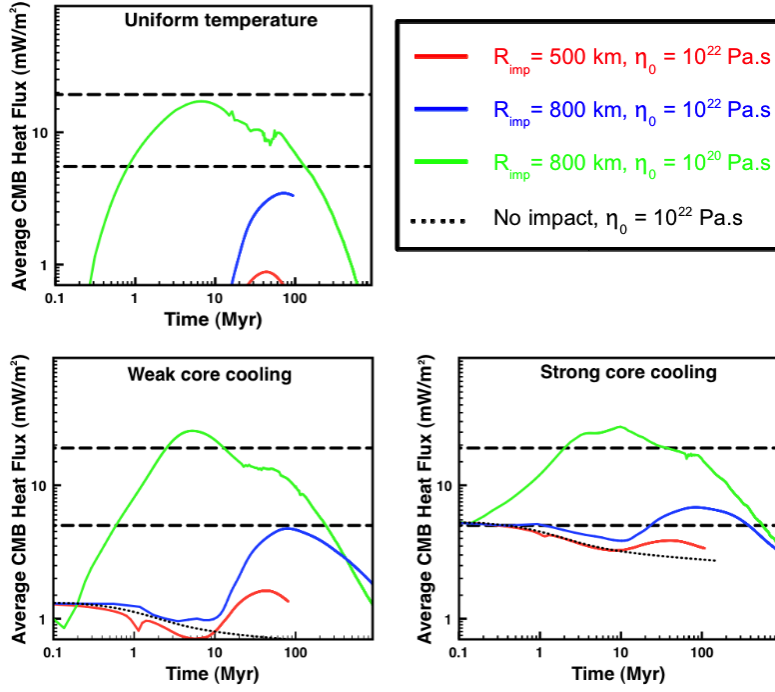


Figure 8: Time evolution of the mean CMB heat flux as a function of the rheology and impactor size. In the three panels different initial thermal states before impact are considered. The two horizontal black dashed lines represent the theoretical dynamo criterion for q_{CMB} ($q_{CMB} > q_A = 5 - 19 \text{ mW.m}^{-2}$). 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.