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Giant impacts, heterogeneous mantle heating and a past hemispheric dynamo on Mars

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Abstract

The martian surface exhibits a strong dichotomy in elevation, crustal thickness and magnetization between the southern and northern hemispheres. A giant impact has been proposed as an explanation for the formation of the Northern Lowlands on Mars. Such an impact probably led to strong and deep mantle heating which may have had implications on the magnetic evolution of the planet. We model the effects of such an impact on the martian magnetic field by imposing an impact induced thermal heterogeneity, and the subsequent heat flux heterogeneity, on the martian core-mantle boundary (CMB). The CMB heat flux lateral variations as well as the reduction in the mean CMB heat flux are determined by the size and geographic location of the impactor. A polar impactor leads to a north-south hemispheric magnetic dichotomy that is stronger than an east-west dichotomy created by an equatorial impactor. The amplitude of the hemispheric magnetic dichotomy is mostly controlled by the horizontal Rayleigh number Ra_h which represents the vigor of the convection driven by the lateral variations of the CMB heat flux. We show that, for a given Ra_h , an impact induced CMB heat flux heterogeneity is more efficient

28 than a synthetic degree-1 CMB heat flux heterogeneity in generating strong hemispheric
29 magnetic dichotomies. Large Ra_h values are needed to get a dichotomy as strong as the
30 observed one, favoring a reversing paleo-dynamo for Mars. Our results imply that an
31 impactor radius of ~ 1000 km could have recorded the magnetic dichotomy observed in
32 the martian crustal field only if very rapid post-impact magma cooling took place.

33 Keywords: Mars, dynamo, magnetic field, core-mantle boundary, impact, heat flux.

34 **1 Introduction**

35 Giant impacts have strongly influenced the internal structure and dynamics of the terrestrial
36 planets during the primordial stages of their evolutions (Hartmann and Davis, 1975; Benz
37 et al., 1988; Asphaug et al., 2006; Andrews-Hanna et al., 2008; Marinova et al., 2008; Nimmo
38 et al., 2008; Jutzi and Asphaug, 2011). These events are plausible explanations for remarkable
39 features of the solar system such as the small volume of Mercury’s mantle relative to its core
40 (Benz et al., 1988; Gladman and Coffey, 2009), the Earth-Moon system (Canup, 2004) and the
41 topographic martian and lunar hemispheric dichotomies (Marinova et al., 2008; Nimmo et al.,
42 2008; Jutzi and Asphaug, 2011). Giant impacts have also been invoked to explain the initiation
43 or cessation of the dynamos of the terrestrial planets and moons (Roberts et al., 2009; Arkani-
44 Hamed and Olson, 2010; Reese and Solomatov, 2010; Monteux et al., 2013; Monteux and
45 Arkani-Hamed, 2014). In these models, the impactors’ radii typically range between 100 and
46 1000 km. These impacts deliver a large amount of heat to the deep mantle, which is likely to
47 strongly affect the efficiency of core cooling and in turn the dynamo activity. Although there
48 is a higher probability that a giant impact will fall on low-latitudes of the planetary surface
49 (Le Feuvre and Wieczorek, 2011), true polar wander events can ultimately place the resulting
50 thermal anomaly at high-latitudes of the CMB. Moreover, large impacts could be responsible
51 for significant resurfacing and reset the magnetization of the pre-impact material (Langlais and
52 Thébault, 2011; Lillis et al., 2013).

53 On Earth, the influence of lower mantle thermal heterogeneity on core magnetohydro-
54 dynamics has been extensively studied using numerical dynamos with imposed non-uniform
55 outer boundary conditions. It has been shown that heterogeneous core-mantle boundary (CMB)
56 heat flux causes a deviation from axisymmetry in the core flow (Aubert et al., 2007), in the

57 time-average paleomagnetic field (Olson and Christensen, 2002) and in locations of intense
58 magnetic flux patches on millennial time-scales (Bloxham, 2002; Amit et al., 2010). It may
59 also explain the emergence of intense magnetic flux patches in the equatorial region (Amit and
60 Choblet, 2012) and may even yield field locking (Gubbins et al., 2007; Willis et al., 2007). Het-
61 erogeneous CMB heat flux may also recover the lateral variations in the inner-core boundary
62 seismic properties (Aubert et al., 2008; Amit and Choblet, 2009). Finally, reversal frequency
63 and the trajectory of the paleomagnetic dipole axis during reversals may also be governed
64 by the heterogeneous lower mantle (Glatzmaier et al., 1999; Kutzner and Christensen, 2004;
65 Olson et al., 2010, 2013; Olson and Amit, 2014).

66 Heterogeneous mantle control has also been proposed to explain some features of planetary
67 magnetic fields. Cao et al. (2014) found that high equatorial CMB heat flux breaks the core
68 flow symmetry and produces north-south asymmetric magnetic fields which may explain the
69 observed field of Mercury (Anderson et al., 2012). Stanley (2010) argued that temperature
70 differences in the surrounding envelope of the convective zone of Saturn axisymmetrize its
71 magnetic field. It has also been proposed that CMB heterogeneity may have controlled the
72 shape of the current Martian magnetic field (Stanley et al., 2008). Mars is characterized by a
73 striking magnetic field dichotomy, which is correlated with the topographic dichotomy. The
74 northern lowlands are mostly devoid of significant magnetic fields. In contrast the southern
75 highlands exhibit large and in some places intense magnetic field anomalies, up to 1500 nT at
76 90 km altitude as measured by Mars Global Surveyor (Acuña et al., 1998). This is two orders
77 of magnitude larger than the crustal magnetic field on Earth. In terms of magnetized material,
78 this suggests a thick (40 km) and intensely magnetized (up to 12 A/m) lithosphere to produce
79 the observed magnetic field (Langlais et al., 2004), or any combination of a thinner lithosphere
80 and a more intense magnetization (e.g. Parker, 2003).

81 The martian magnetic dichotomy can be explained using two end-members scenarios. In
82 the external scenario, the dynamo was equally strong in both hemispheres, and the resulting
83 magnetization was equally strong in both hemispheres. Then the magnetization of the northern
84 hemisphere was removed or erased after the dynamo cessation, e.g., by a giant impact (Nimmo
85 et al., 2008) or volcanic activity (Lillis et al., 2008). Alternatively a significant magnetization
86 was never recorded in the northern hemisphere because surface conditions, lithological or

87 alteration processes were different from those in the southern hemisphere (Rochette, 2006;
88 Quesnel et al., 2009; Chassefière et al., 2013). In the internal scenario, the magnetization is
89 strong only in the southern hemisphere because the dynamo was hemispheric to begin with
90 (Langlais and Amit, 2008; Stanley et al., 2008; Amit et al., 2011).

91 Such an hemispheric dynamo could have been driven by CMB heat flux heterogeneity
92 possibly caused by a very large-scale mantle convection pattern (Harder and Christensen, 1996;
93 Zhong and Zuber, 2001; Elkins-Tanton et al., 2003, 2005; Ke and Solomatov, 2006; Roberts
94 and Zhong, 2006) or by a giant impact (Roberts et al., 2009). In this study we propose a
95 model for the magnetic field dichotomy in which the dynamo hemisphericity (internal origin)
96 is related to a large impact (external origin) (Stanley et al., 2008; Amit et al., 2011; Dietrich and
97 Wicht, 2013). For that purpose, we model heterogeneous CMB heat flux resulting from giant
98 impact heating and investigate its influence on the core dynamo by imposing it as a static,
99 laterally-varying outer boundary condition on numerical dynamo models. In this approach,
100 the CMB heat flux pattern and amplitude, as well as the reduction in the mean heat flux with
101 respect to a reference pre-impact value, are determined by the impactor size, using a synthetic
102 description of the impact heating zone. In section 2 we describe our method. The results are
103 presented in section 3. Discussion, post-impact time evolution and applicability of our results
104 to Mars are given in section 4. Conclusions and possible planetary applications are highlighted
105 in section 5.

106 **2 Method**

107 **2.1 Impact heating at the CMB**

108 Large impacts brought to Mars a formidable amount of energy that is a function of the impactor
109 mass and velocity, the latter strongly depending on the impacted planet radius R . After a large
110 collision on a Mars-size body, a significant fraction of this energy is deeply buried as heat
111 within the mantle and leads to a local temperature increase ΔT_0 below the impact site. The
112 size and the shape of the post-impact thermal anomaly depend on several parameters such
113 as the size of the impactor, the impact velocity and angle, and the structure of the martian

114 mantle. Increasing the size of the impactor leads to an increase of the heated volume while
115 increasing the impact angle from 0 (head-on impact) to larger values (oblique impacts) reduces
116 the maximal depth reached by the post-impact thermal anomaly (Pierazzo et al., 1997; Pierazzo
117 and Melosh, 2000). Here for simplicity, we consider that the volume of the thermal anomaly
118 only scales with the size of the impactor and we consider the case of a head-on impact. Hence,
119 the post-impact thermal anomaly in our models is approximately uniform within a spherical
120 volume (termed isobaric core) with radius R_{ic} that is 1 to 1.44 times larger than the radius of
121 the impactor R_{imp} (Pierazzo et al., 1997; Senshu et al., 2002; Monteux et al., 2013).

122 On Mars, the impactor size invoked to explain the topographic dichotomy ranges between
123 320 and 1350 km (Marinova et al., 2008; Nimmo et al., 2008). This has to be compared to the
124 size of the martian mantle. Based on solar tidal deformations, the martian core radius has been
125 estimated between 1520 and 1840 km (Yoder et al., 2003). For simplicity, we assume a core
126 radius of 1700 km, which implies a mantle thickness of about 1700 km. Hence, considering
127 that $R_{ic} = 1.44R_{imp}$, the post-impact spherical thermal anomaly is likely to overlap the CMB
128 for $R_{imp} > 500$ km. For an impactor radius of $R_{imp} = 1200$ km, the disruption of the impacted
129 planet will only occur when the impact velocity reaches values of ~ 100 km/s which is much
130 larger than the impact velocity v_{imp} considered here ($v_{imp} = 5$ km/s) (Tonks and Melosh,
131 1992; Reese et al., 2010). In our models, we consider that the impactor radius ranges between
132 600 and 1000 km bearing in mind that larger impactors with larger impact angles could have
133 similar thermal consequences at the CMB (Pierazzo et al., 1997; Pierazzo and Melosh, 2000).

134 As the volume of the isobaric core is governed by the size of the impactor, the magnitude
135 of the temperature increase can be directly related to the impact velocity. Making the conser-
136 vative hypothesis that the impact velocity is close to the martian escape velocity and that the
137 volume of the isothermal sphere is 3 times larger than the impactor (Senshu et al., 2002; Mon-
138 teux et al., 2013), the energy balance accounting for heating and melting of both the impactor
139 and impacted material may lead to a uniform spherical temperature increase of ~ 400 K in the
140 martian mantle (Monteux et al., 2013). Away from the isothermal sphere, the temperature de-
141 creases rapidly with distance r as $(R_{ic}/r)^m$ with m typically ranging between 4 and 5 (Senshu
142 et al., 2002; Monteux et al., 2007).

143

144 Geochemical evidence and crater densities indicate that the martian topographic dichotomy
145 could have formed within the first 50 Myr of Solar System formation and that the martian
146 northern hemisphere has been low and stable for nearly all of Mars' history (Zuber, 2001;
147 Frey et al., 2002; Solomon et al., 2005; Marinova et al., 2008). Hence, the impact-driven tem-
148 perature increase is superimposed to the pre-impact thermal state of the martian mantle that
149 strongly depends on the short-lived radiogenic heating, the accretion processes and the dissi-
150 pation of gravitational energy during the core formation (Senshu et al., 2002; Golabek et al.,
151 2009; Šrámek et al., 2010). The uncertainties on the relative importance of these processes as
152 well as the diversity of the processes involved in the core formation lead to a wide range of
153 plausible early thermal states after the full differentiation of Mars. For simplicity, we assume
154 here a 1D radially dependent pre-impact mantle temperature field. The choice of this specific
155 temperature field is not crucial as long as the impact heating of the mantle is predominant. As
156 shown later, the main parameter affecting the dynamo is the amplitude of the heat-flux hetero-
157 geneity at the CMB. Before the impact heating, we consider a simplified temperature profile
158 as in Monteux et al. (2013) with a CMB temperature of $T = 2000$ K and a convective mantle
159 temperature of ~ 1600 K (Roberts and Arkani-Hamed, 2012). It should be noted that in reality,
160 the pre-impact thermal state of the martian mantle was probably much more complicated than
161 the one used in our models, with lateral heterogeneities as well as radial variations (including
162 thermal boundary layers and pressure dependence). For simplicity, we do not consider the
163 changes of mantle properties with depth such as the pressure increase and the corresponding
164 adiabatic heating. Because we consider here that the mantle temperature above the CMB is
165 uniform and equal to 1600 K while the core temperature is 2000 K, adding more complex-
166 ity should slightly decrease the amplitude of the heat flux heterogeneity. However, since the
167 early martian thermal state is poorly constrained, our simple model may be considered as a
168 reasonable first step to understand the influence of giant impacts on planetary dynamos.

169 The superposition of the large impact-driven temperature increase leads to a significant
170 perturbation of the pre-impact homogeneous CMB mean heat flux q_0^h . To obtain the post-
171 impact heat flux at the martian CMB for a given impactor radius, we used an impact heating
172 model similar to the one described in Monteux et al. (2013). A uniform spherical temperature
173 anomaly rapidly decreasing with distance is superimposed on the martian pre-impact temper-

174 ature field. The impact heating is followed by a thermal re-adjustment within a characteristic
 175 time that is governed by the rheology of the mantle surrounding the impact-induced thermal
 176 anomaly. This characteristic time is $\tau_{imp} \sim 10$ Myr (Monteux et al., 2007). The characteristic
 177 magnetic diffusion time is $\tau_\lambda = r_o^2/\lambda$ where r_o is the core radius and λ is the magnetic diffu-
 178 sivity (e.g. Bloxham and Jackson, 1991). For Mars $r_o = 1700$ km and combined with updated
 179 estimates of the electrical conductivity of molten iron in Earth’s core conditions of $\lambda \sim 0.5$
 180 m^2/s (Pozzo et al., 2012), these estimates give $\tau_\lambda \sim 180$ kyr. Since $\tau_{imp} \gg \tau_\lambda$, the post
 181 impact mantle temperature field may be considered constant in our numerical dynamo models.

182 The shock wave also leads to a temperature increase within the core of the impacted planet,
 183 much stronger in the region directly beneath the impact site. The low-viscosity rotating liquid
 184 core cannot sustain lateral variations of temperature and the core overturns, resulting in a stably
 185 stratified temperature which increases with radius. In the case of a homogeneous CMB heat
 186 flux, the thermal stratification occurs within a few kyr (Arkani-Hamed and Olson, 2010). This
 187 stratification kills the possible pre-existing core convection, and hence the core dynamo. Then
 188 it can take up to several tens of Myr to remove this stratification by conductive core heat loss
 189 (Arkani-Hamed and Olson, 2010). However, a significant fraction of the mantle right above
 190 the CMB may have experienced melting which facilitated the core cooling. The time needed
 191 to remove the core excess heat by convection is $10^3 - 10^4$ yr when considering a molten layer
 192 above the CMB (Monteux et al., 2011). It is therefore likely that after the impact the martian
 193 dynamo died during the time needed to generate and remove the post-impact core stratification
 194 (i.e. $\sim 10^3 - 10^4$ yr). Then, the dynamo probably re-started, while the impact-driven lower
 195 mantle anomaly was still in place (during $\tau_{imp} \sim 10$ Myr). For simplicity, we do not consider
 196 here the core impact heating and the subsequent rapid thermal readjustment. Our dynamo
 197 models are convectively unstable throughout the shell, corresponding to the state of the system
 198 after core stratification has been removed and while the CMB heterogeneity was still in place.
 199 We discuss this aspect in the conclusion section.

200 As the post-impact mantle temperature reaches the temperature of the core, the heat flux
 201 q is nearly zero where the isobaric core overlaps with the CMB. Away from the isobaric core
 202 and along the CMB, the heat flux increases rapidly to its pre-impact mean value q_0^h (Fig. 1).
 203 The reduction in the mean CMB heat flux due to the impact corresponds to the relative CMB

204 surface that is heated by the impactor and is defined by q_0^r as

$$q_0^r = \frac{q_0^h - q_0}{q_0^h} \quad (1)$$

205 where q_0 is the post-impact mean heat flux. The amplitude of the heat flux heterogeneity is
 206 commonly given by q^* (Olson and Christensen, 2002) with

$$q^* = \frac{q_{max} - q_{min}}{2q_0} \quad (2)$$

207 where q_{max} and q_{min} are the maximal and minimal values of the post-impact heat flux respec-
 208 tively. In Fig. 1, we show q_0^r and q^* obtained from our post-impact heating model as a function
 209 of the impactor size. As the impactor size increases, the extension of the isobaric core on the
 210 CMB increases, i.e. q_0 decreases and therefore q_0^r and q^* increase. An impactor with a radius
 211 smaller than $R_{imp} \sim 500$ km has a negligible effect on the heated surface of the CMB. For
 212 small impactors a small portion of the CMB is heated so $q_0 \simeq q_0^h$, and since $q_{max} = q_0^h$ and
 213 $q_{min} = 0$ always hold, for small impactors $q^* \simeq 0.5$. An impactor radius of $R_{imp} \sim 1000$ km
 214 decreases the mean CMB heat flux by 26.5% and produces an heterogeneity with amplitude
 215 $q^* \sim 0.67$. In our models, we limit the impactor radius to 1000 km and consider a head-on
 216 impact which may represent cases with larger impactors and smaller impact angles (Pierazzo
 217 et al., 1997; Pierazzo and Melosh, 2000).

218

219 Next we expand the impact-driven CMB heat flux pattern in terms of spherical harmonic
 220 coefficients (Fig. 2). In order to use it as an outer boundary condition for heat flux in our
 221 dynamo models, we performed a spherical harmonic expansion truncated at $\ell_{max} = 20$ (see
 222 Fig. 2 for $R_{imp} = 800$ km). First we fit the CMB heat flux with the following analytical
 223 expression in terms of the angular distance ω from the impactor's center:

$$q_\omega = \exp \left[-\frac{1}{n} \left(\frac{\omega}{\omega_0} \right)^n \right] \quad (3)$$

224 The Gauss-like function (3) avoids undesirable Gibbs effects associated with discontinuous
 225 gradients on the edge of the imprint of the isobaric core on the CMB. The best fit parameters
 226 found for the three impactor radii are $\omega_0 = 15^\circ$ and $n = 3$ for $R_{imp} = 600$ km, $\omega_0 = 30^\circ$ and
 227 $n = 7$ for $R_{imp} = 800$ km and $\omega_0 = 48^\circ$ and $n = 7$ for $R_{imp} = 1000$ km.

2.2 Numerical dynamo models

We solve the set of self-consistent non-dimensional Boussinesq magnetohydrodynamics equations for dynamo action due to thermal convection of an electrically conducting fluid in a rotating spherical shell (for governing equations and more details see Amit et al., 2011). We use the code MAGIC by Johannes Wicht (Wicht, 2002). We analyze numerical dynamos with rigid insulating boundary conditions. The models differ in the imposed outer boundary heat flux pattern and the amplitude of its variation. A summary of model parameters, outer boundary heat flux patterns, geometries and main results is given in Table 1.

It is likely that Mars has no solid inner core up to the present day (e.g. Schubert and Spohn, 1990; Breuer et al., 2010) and, as a consequence, convection in the early martian dynamo was purely thermal (Amit et al., 2011; Dietrich and Wicht, 2013), driven by secular cooling and perhaps by radioactive heating. This convection mode is highly sensitive to CMB heat flux heterogeneity and may thus break internal dynamo symmetries with relatively moderate heterogeneity amplitudes (Hori et al., 2014). Due to numerical singularity at the center of the planet, we retain in our dynamo models a small inner core with a radius $r_i/r_o = 0.2$ of the outer core radius. We impose zero heat flux on the inner boundary so that the inner core is convectively passive. Overall, results by Aubert et al. (2009) and Hori et al. (2010) suggest that such a relatively small and passive inner core has little effect on the dynamo models.

Four internal non-dimensional parameters control the dynamo action. The heat flux Rayleigh number (Olson and Christensen, 2002) represents the strength of buoyancy force driving the convection relative to retarding forces

$$Ra = \frac{\alpha g_0 q_0 D^4}{k \kappa \nu} \quad (4)$$

where α is thermal expansivity, g_0 is gravitational acceleration on the outer boundary at radius r_o , q_0 is the mean heat flux across the outer boundary, D is shell thickness, k is thermal conductivity, κ is thermal diffusivity and ν is kinematic viscosity. The Ekman number represents the ratio of viscous and Coriolis forces

$$E = \frac{\nu}{\Omega D^2} \quad (5)$$

253 The Prandtl number is the ratio of kinematic viscosity to thermal diffusivity

$$Pr = \frac{\nu}{\kappa} \quad (6)$$

254 and the magnetic Prandtl number is the ratio of kinematic viscosity to magnetic diffusivity λ

$$Pm = \frac{\nu}{\lambda} \quad (7)$$

255 A fifth non-dimensional number is the amplitude of the outer boundary heat flux heterogeneity
256 which is expressed by q^* (see Eq. 2).

257 In all cases a volumetric homogeneous heat source ϵ compensates for the loss of heat
258 through the outer boundary according to

$$-4\pi r_o^2 Pr \left[\frac{\partial T}{\partial r}(r_o) \right] = \frac{4}{3} \pi (r_o^3 - r_i^3) \epsilon \quad (8)$$

259 where [...] denotes averaging over the outer boundary surface S . In terms of the non-dimensional
260 variables $\left[\frac{\partial T}{\partial r}(r_o) \right] = 1$, so for $r_i/r_o = 0.2$ (the geometry used in the study) the source term is
261 $\epsilon \simeq 2.42$. We use moderate amplitudes of CMB heat flux heterogeneity to avoid violation of
262 the Boussinesq approximation on which the dynamo code relies.

263 Most dynamo models fall into two categories. In the first, the radial field on the CMB is
264 dominated by an axial dipole component, but the field does not reverse. In the second, the field
265 is multipolar and dipole reversals occur (Kutzner and Christensen, 2002). Earth-like models
266 that are both dipole-dominated and reversing are only found in a narrow transitional regime of
267 parameters space (Olson, 2007; Wicht et al., 2009). Following Amit et al. (2011), we consider
268 cases from both regimes.

269 **2.3 Hemispheric magnetic dichotomy monitoring**

270 The kinetic energy of the impactor is dissipated as a result of the irreversible work done by
271 shock waves in damaging crustal rocks as well as heating and melting the target material (Pier-
272 azzo et al., 1997; Senshu et al., 2002; Reese and Solomatov, 2006; Monteux et al., 2011). After
273 the excavation process, a significant fraction of the material molten by impact is redistributed
274 heterogeneously at the surface of the impacted planet (Marinova et al., 2008, 2011). The distri-
275 bution of the molten material is governed by the impact parameters such as the impact velocity

276 and angle. At low impact velocities (6 – 10 km/s) and oblique impact angles (30 – 60°),
277 50 – 70% of the impact-induced melt distribution might be contained within the area of impact
278 and 25 – 30% might be deposited at the antipode of the impact site (Marinova et al., 2008).
279 The impact-induced molten material might be redistributed over a thickness that ranges be-
280 tween 30 and 50 km (Marinova et al., 2008, 2011). The cooling and the crystallization of the
281 molten material leads to the formation of the impact induced crust potentially recording the
282 anomalous dynamo.

283 We consider two end member crust formation scenarios (Amit et al., 2011; Langlais and
284 Thébault, 2011) as illustrated in Fig. 3. In the case of continuous and homogeneous crust
285 formation, each part of the martian crust is formed by a large number of incremental and su-
286 perposed additions (e.g. lava flows, sills, dykes) over an extended period of time. Each new
287 layer records the magnetic field at its time of cooling below the Curie Temperature, and the
288 present crustal field at a specific location results from the vertical superposition of the mag-
289 netization vectors of the various layers (Fig. 3, bottom line). If this crust formation scenario
290 occurs while the dynamo reverses (Fig. 3, bottom right), the present local crustal field would
291 thus represent the intensity of a long-term time-average martian paleomagnetic field. In the
292 extreme case of periodic inversions and layers with equal thicknesses, the superposed opposite
293 sign magnetization vectors could eventually cancel each other. The other end-member model
294 assumes a random crust formation where crustal units are formed in relatively rapid events
295 (Fig. 3, top line). Individual crustal blocks are created randomly both in space and time and
296 each block acquires a magnetization which only depends on the dynamo field at the time of
297 cooling. The present local field in this scenario thus results from adjacent magnetization vec-
298 tors, or in a probabilistic way, from the time-average of the paleofield intensity. It is likely
299 that neither one of these end-member scenarios represent what actually occurred on Mars, and
300 rather that the actual way in which the magnetized part of the martian crust formed may be
301 intermediate between the two scenarios.

302 We follow the statistical measures proposed by Amit et al. (2011) corresponding to these
303 two end-member crust formation scenarios. In the context of a continuous homogeneous crust
304 formation scenario, we calculate the ratio of intensities of the time-average field at the planet
305 surface

$$SN_{cnt} = [|\langle \vec{B} \rangle|]_{sh} / [|\langle \vec{B} \rangle|]_{nh} \quad (9)$$

306 where SN denotes the ratio between the rms surface average in the southern and northern
 307 hemispheres and the subscript 'cnt' denotes *continuous* crust formation. In the context of a
 308 random crust formation scenario, we calculate the magnetic dichotomies based on the time-
 309 average of the magnetic field intensity

$$SN_{rnd} = [\langle |\vec{B}| \rangle]_{sh} / [\langle |\vec{B}| \rangle]_{nh} \quad (10)$$

310 where the subscript 'rnd' denotes *random*. Eqs. (9)-(10) are applied for the east-west di-
 311 chotomies EW_{cnt} and EW_{rnd} by replacing the summations with the appropriate hemispheres.

312 The relevance of the continuous or random crust formation scenario depends on the rela-
 313 tion between the cooling time of the Impact Induced Molten Material (IIMM) and the typical
 314 magnetic timescales. The cooling time of the IIMM strongly depends on the presence of an
 315 atmosphere that may prevent the IIMM from rapid cooling (e.g. Lebrun et al., 2013). In the
 316 case of molten material induced by a giant impact, a significant fraction of the atmosphere
 317 (if any) can be eroded from the impacted body (Shuvalov, 2009) which strongly enhances the
 318 excess heat removal and decreases the cooling and solidification times. Hence, the duration
 319 of the partially molten stage decreases from more than 1 million years with an atmosphere
 320 to ~ 1000 years when no atmosphere is present (Lebrun et al., 2013). The two crust forma-
 321 tion scenarios give identical dichotomies for non-reversing dynamos (Amit et al., 2011). In
 322 the reversing case, if the crust was formed over a period much longer than a typical magnetic
 323 timescale, the continuous scenario is relevant (Dietrich and Wicht, 2013). Conversely, if the
 324 crust was formed faster than a typical magnetic timescale (i.e. without any atmosphere), the
 325 random scenario is relevant (see Fig. 3).

326 **3 Influence of impact heating on core dynamo**

327 Various CMB heat flux models corresponding to varying impactor sizes were examined. We
 328 considered two extreme geographic locations of the impactor, with a center falling on the geo-
 329 graphical pole or on the equator. The impactor introduces warm material to the lower mantle,

330 thus locally decreases the CMB heat flux. Apart from pattern dependence, the reduction in
331 mean CMB heat flux (expressed by the Ra number compared to its reference homogeneous
332 case) as well as the amplitude of the heat flux heterogeneity q^* are all dependent in a self-
333 consistent manner on the radius of the impactor. A larger impactor warms a larger part of the
334 CMB, thus reduces the mean CMB heat flux q_0 more and produces larger q^* (see Table 1).

335 Fig. 4 shows the time-average rms radial field at the CMB and the time-average field
336 intensity at the surface of Mars for case I2 with an impactor of radius 800 km falling on
337 the north geographic pole. Here and elsewhere all magnetic field values are given in units
338 of $\sqrt{\rho\mu_0\lambda\Omega}$ where ρ is the fluid density and μ_0 is permeability of free space. The impactor
339 yields an hemispheric field similar to those obtained with synthetic Y_1^0 CMB heat flux patterns
340 (Stanley et al., 2008; Amit et al., 2011; Dietrich and Wicht, 2013).

341 Fig. 5 shows the time-average zonal temperature and flow in case I2. The reduced CMB
342 heat flux in the north pole region caused by the impactor results in a relatively warmer fluid
343 there. The colder fluid in the more vigorously convecting southern hemisphere is associated
344 with fluid downwelling at high-latitudes that concentrates the magnetic field. This produces a
345 south-north magnetic hemispheric dichotomy. In addition, the boundary driven thermal wind
346 flow exhibits a large one cell meridional circulation with surface flow going southward, carry-
347 ing weak magnetic flux from north to south. These dynamical features are also in agreement
348 with that obtained with a synthetic Y_1^0 CMB heat flux pattern (Stanley et al., 2008; Amit et al.,
349 2011).

350 When the impactor is falling on the equatorial plane, some magnetic dichotomy may be
351 expected between eastern and western hemispheres. Indeed Fig. 6 shows that in case I5
352 such a dichotomy is obtained. Note that the east-west dichotomy in this case is significantly
353 weaker than the south-north dichotomy in the corresponding polar impactor case I2 with the
354 same impactor radius of 800 km. However, compared to the east-west dichotomy obtained
355 with a synthetic Y_1^1 CMB heat flux pattern (Amit et al., 2011), the impactor driven east-west
356 dichotomy is significantly stronger (compare cases I5 and Y3 in Table 1).

357 Fig. 7 shows two snapshots of the radial field at the CMB and the intensity at the surface
358 of Mars for case I3 with the largest impactor studied here (radius of 1000 km) falling on the

359 north geographic pole as well as the corresponding long term time-average surface intensity.
360 The differences between the two snapshots attest to the chaotic time dependence of this model.
361 Nevertheless, in both snapshots the relatively small-scale radial field at the CMB is strongest at
362 high latitudes of the southern hemisphere. The surface intensity is large scale, and also peaks
363 at the polar region of the southern hemisphere. In this case, the time-average south-north
364 dichotomy is 2.25 (see Table 1), very close to the lower bound estimation of 2.4 obtained by
365 Amit et al. (2011) based on observations of the martian crustal magnetic field (Langlais et al.,
366 2004).

367 Finally we examine the influence of the strength of the internal core convection. Case
368 I4 is identical to case I3 except for its Ra value which is larger. The south-north magnetic
369 dichotomy in this case is 2.58 (Table 1), demonstrating that more vigorous core convection
370 produces more hemispherical fields.

371 Our models show that a polar impactor leads to a stronger north-south hemispheric mag-
372 netic dichotomy than an equatorial impactor to an east-west dichotomy. We also find that in
373 the non-reversing regime of parameters, the magnetic field dichotomy that can be recorded
374 in the cooling IIMM is independent of the crust formation scenario (random or continuous).
375 Reversing dynamos that convect stronger produce stronger magnetic dichotomies than non-
376 reversing dynamos. These results are in agreement with those obtained by Amit et al. (2011)
377 for degree-1 heat flux patterns.

378 The amplitude of the CMB heat flux heterogeneity (Fig. 2) is commonly measured by the
379 peak to peak lateral variation (Olson and Christensen, 2002). However, when the pattern is not
380 smooth and very localized, with imbalance between areas of positive and negative anomalies,
381 q^* is inadequate. This is exactly the case in the impact driven heat flux patterns considered here.
382 For example, for very small impacts the amplitude is negligible but q^* approaches 0.5 (Fig.
383 1b). We quantify the impact driven heterogeneity amplitude by q_0^r (Eq. 1) which measures
384 the extent of the CMB surface affected by the impact heating. Therefore, in order to compare
385 the efficiency of magnetic hemispheric dichotomy generation by impact driven patterns with
386 the efficiency by synthetic degree-1 patterns, q_0^r in the first must be compared with q^* in the
387 latter. In addition we propose that the governing parameter controlling the amplitude of the

388 hemispheric magnetic dichotomy is the horizontal Rayleigh number Ra_h (Willis et al., 2007).
 389 We define Ra_h as:

$$Ra_h = \begin{cases} q^* Ra & \text{in Y cases} \\ q_0^r Ra & \text{in I cases} \end{cases} \quad (11)$$

390 To adequately compare the synthetic Y_1^0 CMB heat flux cases from Amit et al. (2011) and
 391 the polar impact driven CMB heat flux cases from this study, we plot in Fig. 8 the increase
 392 of the south-north dichotomy $SN - 1$ as a function of Ra_h (11). This figure shows that
 393 for a given lateral forcing, an impact induced CMB heat flux heterogeneity is significantly
 394 more efficient than a synthetic Y_1^0 CMB heat flux heterogeneity in generating hemispheric
 395 magnetic dichotomies. It also shows that the impactor size needed to generate the observed
 396 martian dichotomy (Amit et al., 2011) is slightly larger than 1000 km, which is within the
 397 estimated range of impactor sizes (Marinova et al., 2008). The better efficiency of impact
 398 driven hemispheric dynamos over the synthetic cases is even more pronounced in the equatorial
 399 cases (see Table 1).

400 **4 Implication for crustal magnetic dichotomy**

401 Our results show that large Ra_h (and therefore large Ra) are needed in order to get a dichotomy
 402 as strong as the observed. As large Ra are likely to be associated with reversing dynamos (e.g.
 403 Kutzner and Christensen, 2002), the answer depends on the duration of a chron relative to the
 404 duration of the crystallization of the IIMM induced by the impact. If a slow crystallization rate
 405 was coupled with short magnetic chrons (Dietrich and Wicht, 2013), then our models cannot
 406 explain the observed martian dichotomy. However, if a relatively rapid magma cooling was
 407 coupled with a relatively low reversal frequency then the impact induced magnetic dichotomy
 408 may have been recorded within the martian crust and explain the observed signal. Hence, the
 409 relevant magnetic timescale is a duration of a typical chron.

410 In our non-reversing dynamo models the chron duration is effectively infinite and the hemi-
 411 spheric magnetic dichotomies may be recorded by the crustal magnetization (Fig. 3, left).
 412 However, the amplitudes of the dichotomies in these cases are too low to explain the observed

413 hemispheric crustal magnetic dichotomy on Mars (Table 1). Dynamo models with large am-
414 plitude hemispheric dichotomies tend to reverse frequently, in agreement with the findings of
415 Dietrich and Wicht (2013). For example, in case I3 a typical chron persists for ~ 18 kyr. In
416 order for such a model to record an hemispheric magnetic dichotomy at the crust, very fast
417 crust formation is required.

418 Dietrich et al. (2013) studied hemispherical dynamos in the framework of classical mean
419 field theory. For $q^* > 0.6$ Dietrich and Wicht (2013) found hemispheric $\alpha\Omega$ dynamos with
420 fast oscillations over periods of ~ 10 kyrs. In our models the amplitude of CMB heat flux
421 heterogeneity is moderate with $q^* \sim 0.6$, and more importantly $q_0^r \ll 0.6$. Our dynamos are
422 therefore of the α^2 -type with chaotic (reversing or non-reversing) behaviour. In addition, the
423 moderate q_0^r does not change the dynamo regime from stable to reversing. In these dynamos
424 the duration of a chron (or the reversal frequency) depends on the level of turbulence in the core
425 (Olson and Amit, 2014), which is in general unknown (even for the Earth). Chron duration
426 varied immensely over Earth's history between 40 kyrs - 40 Myrs (Merrill et al., 1998).

427 Fig. 9 illustrates the interplay between the relevant time-scales: the crystallization time τ_c ,
428 the spreading time τ_{imp} and the duration of a chron. In the context of a martian giant impact,
429 the associated molten thickness δ ranges between 30 and 50 km and the melt fraction is \sim
430 20 % (Marinova et al., 2008). Hence, the crystallization and cooling times of the shallow
431 IIMM induced by one impact should be rapid: even if the complete solidification timescale
432 ultimately depends on the poorly constrained evolution of the post-impact transient atmosphere
433 (Abe, 1997), crystallization of this local molten material should occur in less than 1000 yr
434 (Reese and Solomatov, 2006). The subsequent evolution of the partially molten material could
435 involve isostatic readjustment of a deep, initially hemispheric, retained melt region and lateral
436 spreading as a gravity current (Reese et al., 2011). Overall, the surface temperature of the
437 IIMM falls below the Curie temperature in a timescale that strongly depends on factors such
438 as the characteristics of the above atmosphere or the viscosity of the solid mantle if isostatic
439 rebound is involved. This instant characterizes the beginning of the time interval when the
440 solidified crust starts to record the magnetic field. The cold front then propagates downwards
441 from the surface and a secondary front most probably develops at the base of the IIMM whose
442 nature depends on the thermal state of the pre-impact crust on top of which the IIMM is

443 superimposed (Fig. 9). The crustal material originating in the impact event stops recording the
444 magnetic field precisely when the innermost region is cooler than the Curie temperature. In
445 summary, given these somewhat overlapping ranges of timescales for the crust formation and
446 magnetic chrons, we argue that both end-member crust formation scenarios are in principle
447 possible, bearing in mind that reality may be somewhere in between.

448 Our study also emphasizes the importance of the duration of the post-impact thermal
449 anomaly at the CMB (τ_{imp}). If this timescale is shorter than the time needed by the IIMM
450 to cool down to temperatures smaller than the Curie temperature τ_c , the hemispheric dynamo
451 has ended before the crust was able to record it. However, if τ_{imp} is larger than τ_c , the post-
452 impact magnetic field can be recorded within the crust. The beginning of this record starts at
453 $\tau_{c,top} \sim 1$ kyr and ends at $\tau_{c,int} \sim 10 - 100$ Myr, this time interval being mostly controlled by
454 heat diffusion. The deep post-impact thermal re-adjustment occurs within a characteristic time
455 that is governed by the rheology of the mantle surrounding the post-impact thermal anomaly.
456 For a mantle viscosity comparable to that of the present day Earth, say around 10^{21} Pa.s, this
457 characteristic spreading time is $\tau_{imp} \sim 10 - 100$ Myr (Monteux et al., 2007; Watters et al.,
458 2009) which is much larger than the characteristic magnetic diffusion time $\tau_\lambda \sim 30$ kyrs (e.g.
459 Bloxham and Jackson, 1991) or the time needed by the top of the IIMM to reach the Curie
460 temperature $\tau_{c,top}$ (Fig. 9).

461 5 Conclusion

462 Can the observed martian hemispheric magnetic dichotomy be the consequence of a giant
463 impact that has led to an internal hemispheric magnetic field? Our results show that a ~ 1000
464 km radius impactor can generate a sufficiently large scale thermal anomaly at the CMB so
465 that a hemispheric dynamo is generated. An impact induced CMB heat flux heterogeneity
466 is more efficient than a synthetic degree-1 CMB heat flux heterogeneity in generating strong
467 hemispheric magnetic dichotomies. This magnetic dichotomy is stronger for a polar impact
468 than for an equatorial impact. This result reconciles the giant impact induced scenario evoked
469 to explain the martian topographic dichotomy (Andrews-Hanna et al., 2008; Marinova et al.,
470 2008; Nimmo et al., 2008) and the heterogeneous CMB heat flux scenario proposed to explain

471 why the magnetization is strong only in the southern hemisphere (Stanley et al., 2008) via
472 a more realistic CMB heat flux pattern than the previously used synthetic degree-1. Hence,
473 we propose here that an external event at the martian surface may have produced an internal
474 hemispheric dynamo.

475 From our results, three relevant timescales arise: the cooling time of the impact induced
476 molten material, the duration of a magnetic chron and the duration of the post-impact thermal
477 anomaly at the CMB. The first is mainly governed by the characteristics of the martian atmo-
478 sphere, the second is governed by the core dynamics while the third is governed by the mantle
479 dynamics. If a relatively rapid magma cooling was coupled with a relatively low reversal
480 frequency and with a stable CMB heat flux heterogeneity, then the impact induced magnetic
481 dichotomy may have been recorded within the martian crust and explain the observed signal.

482 According to some studies (e.g. Arkani-Hamed and Olson, 2010), a giant impact might
483 have led to a thermal stratification at the top of the core that terminated the dynamo. How-
484 ever, core stratification can be removed much faster by convection of molten material, over a
485 timescale orders of magnitude shorter than the longevity of mantle thermal anomalies. Hence
486 our scenario of giant impact leading to heterogenous mantle heating and to a hemispheric
487 dynamo on Mars may follow an episode of dynamo shutdown.

488 Our impact driven CMB heat flux heterogeneity model may also be applied to model dy-
489 namos of other planets. Indeed, giant impacts were common in the later stage of accretion of
490 terrestrial planets. The Earth is likely formed by accretion of a few dozen moon to Mars-size
491 planetary embryos (see review by Chambers, 2004). A Mars-size impact on Earth may have
492 resulted in the formation of the Moon (Hartmann and Davis, 1975; Cameron and Ward, 1976;
493 Canup, 2004). The Moon was also probably hit by a large planetesimal at the end of its forma-
494 tion (Jutzi and Asphaug, 2011). Finally, an oblique collision of a large body with a mass about
495 one sixth of Mercury's has likely stripped away a significant part of its mantle (Smith, 1979;
496 Benz et al., 1988). Furthermore, these three terrestrial objects have or have had an internally
497 generated magnetic field (Stevenson, 2003). Hence, giant impacts have potentially strongly
498 influenced their internal dynamics and dynamo activities.

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Case	Pattern	R_{imp} [km]	Ra	q^*	q_0^r	Rm	SN_{cnt}	SN_{rnd}	EW_{cnt}	EW_{rnd}	τ_{ch}
I1	Polar	600	$2.90 \cdot 10^7$	0.51	0.04	386	-	1.18	-	-	40.9
I2	Polar	800	$2.69 \cdot 10^7$	0.56	0.13	423	-	1.54	-	-	22.6
I3	Polar	1000	$2.21 \cdot 10^7$	0.66	0.26	439	-	2.25	-	-	18.3
Y1	Y_1^0	-	$3 \cdot 10^7$	0.8	-	488	-	2.54	-	-	12.2
I4	Polar	1000	$4 \cdot 10^7$	0.66	0.26	572	-	2.58	-	-	15.4
Y2	Y_1^1	-	$2 \cdot 10^7$	0.8	-	412	-	2.35	-	-	14.9
I5	Equatorial	800	$2.69 \cdot 10^7$	0.56	0.13	350	-	-	-	1.22	28.2
Y3	Y_1^1	-	$3 \cdot 10^7$	0.8	-	405	-	-	-	1.14	17.5
I6	Polar	800	$8.97 \cdot 10^6$	0.56	0.13	99	1.12	1.13	-	-	∞
I7	Polar	1000	$7.35 \cdot 10^6$	0.66	0.26	182	1.59	1.57	-	-	∞
Y4	Y_1^0	-	$1 \cdot 10^7$	0.3	-	114	1.16	1.16	-	-	∞
Y5	Y_1^0	-	$1 \cdot 10^7$	0.5	-	157	1.49	1.50	-	-	∞
I8	Equatorial	800	$8.97 \cdot 10^6$	0.56	0.13	104	-	-	1.01	1.01	∞
Y6	Y_1^1	-	$1 \cdot 10^7$	0.5	-	133	-	-	1.01	1.01	∞

Table 1: Summary of numerical dynamo models parameters and resulted magnetic hemispheric dichotomies. In all cases $r_i/r_o = 0.2$, $Pr = 1$ and $Pm = 3$. The upper part includes reversing dynamos with $E = 3 \cdot 10^{-4}$; The lower part includes non-reversing dynamos with $E = 1 \cdot 10^{-4}$. The imposed CMB heat flux patterns are either degree-1 single harmonics (Y_1^0 or Y_1^1) denoted as cases Y, or impact driven centered at the pole or at the equator denoted as cases I. All Y cases are from Amit et al. (2011) except case Y2 which is from this study. The radius of the impactor is R_{imp} . The conventional definition of the amplitude of the heat flux anomaly q^* is defined as the ratio of the peak-to-peak difference to twice the mean (Olson and Christensen, 2002). The reduction in the mean heat flux due to the impact is q_0^r . Rm is the magnetic Reynolds number based on the rms velocity in the volume of the shell. SN and EW denote *south-north* and *east-west*, 'cnt' and 'rnd' subscripts denote *continuous* and *random* crust formation. East-west ratios are calculated with respect to the hemisphere centered at longitude 0° (center of large heat flux anomaly for Y_1^1 and equatorial I cases) to the hemisphere centered at longitude 180° . For reversing cases only ratios based on random crust formation are given, for Y_1^0 or polar I cases only south-north ratios are given, for Y_1^1 or equatorial I cases only east-west ratios are given. As reference cases, the homogeneous CMB heat flux models from Amit et al. (2011) give SN and EW values of unity. The average chron duration τ_{ch} is given in kyrs for the reversing cases.

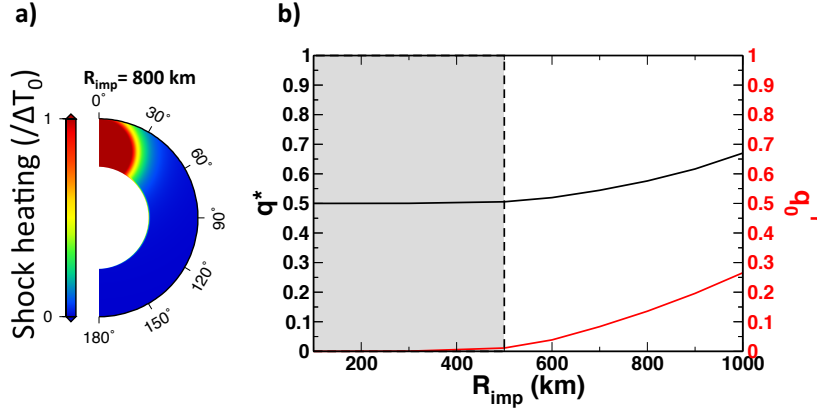


Figure 1: a) martian mantle post-impact temperature increase for $R_{imp} = 800$ km. b) Corresponding conventional amplitude of heat flux heterogeneity on the CMB q^* (black line, Eq. 2) and the mean heat flux reduction q_0^r (red line, Eq. 1) vs. impactor size R_{imp} . In the grey domain ($R_{imp} < 500$ km), the impactor is too small for isobaric core to reach the CMB.

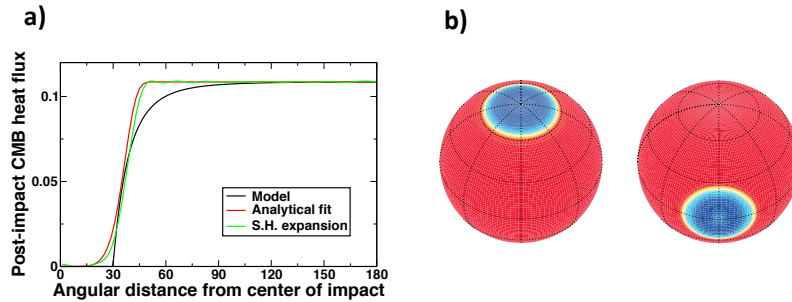


Figure 2: a) Post-impact CMB heat flux (black line), analytical fit (red line) and spherical harmonic expansion (green line) vs. angular distance from the center of the impact driven heated area. b) Imposed CMB heat flux anomalies resulting from an impactor of radius 800 km falling on the north pole (left) or the equator (right).

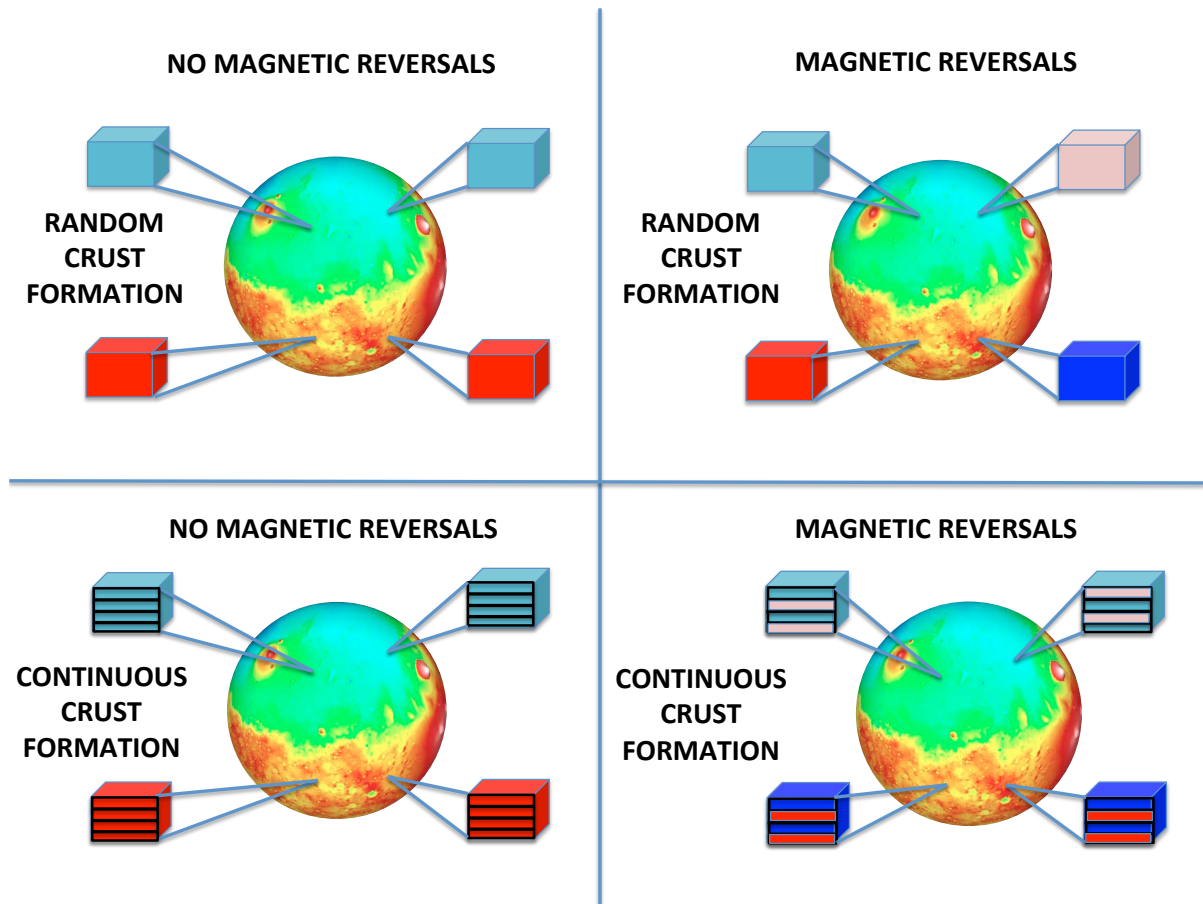


Figure 3: Schematic illustration of the random (top) and continuous (bottom) crust formation scenarios and the corresponding recorded magnetization without (left) or with (right) magnetic reversals. Blue/red denotes negative/positive radial magnetic paleofields respectively, so alternating colors correspond to paleomagnetic reversals. Cubes represent a vertical cut through the martian crust that has recorded strong (deep blue or deep red) or weak (light red or light blue) magnetic paleofields. The weak magnetic paleofield is restricted to the impacted pole while the strong magnetic paleofield is restricted to the opposite one. The current magnetic field observed at the surface results from the vertically integrated magnetization over the cube.

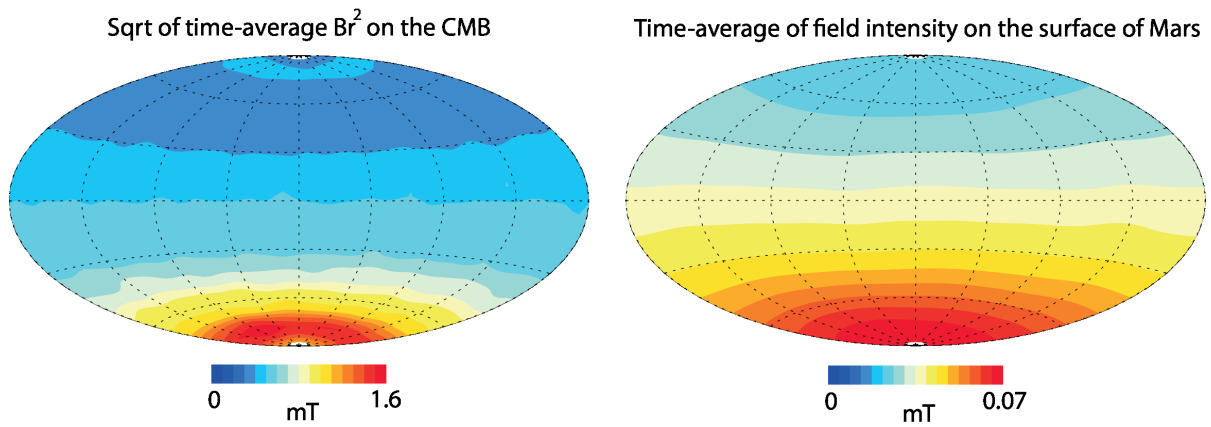


Figure 4: Time-average magnetic field properties on the CMB (left) and on the surface of Mars (right) in case I2. The CMB field is upward continued to the surface of Mars as a potential field.

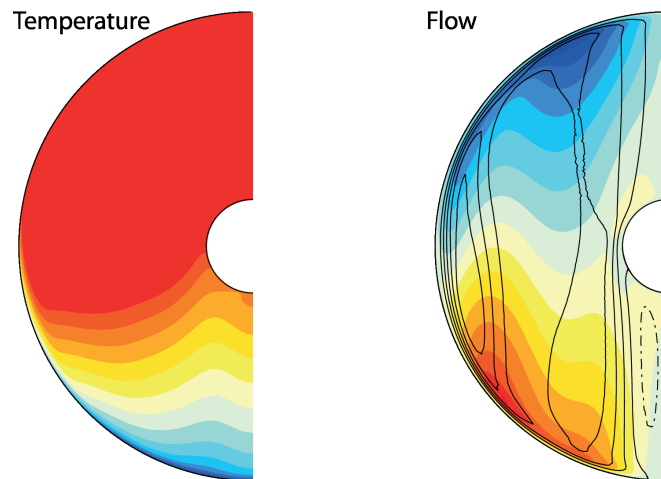


Figure 5: Time-average zonal temperature (left) and flow (right) in case I2. In the right subplot colors denote azimuthal flow and streamlines denote meridional circulation (solid/dashed are anti-clockwise/clockwise respectively).

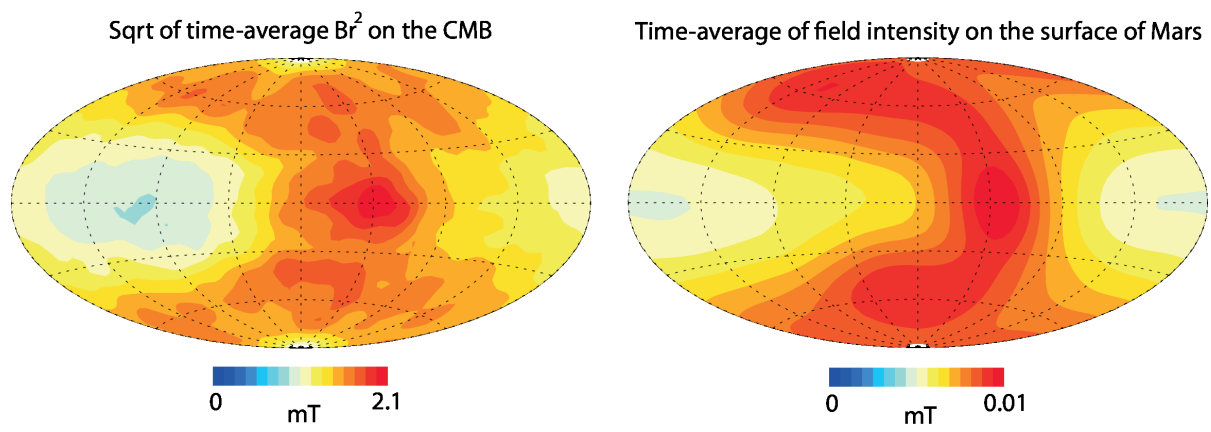


Figure 6: As in Fig. 4 for case I5 (note different color scales).

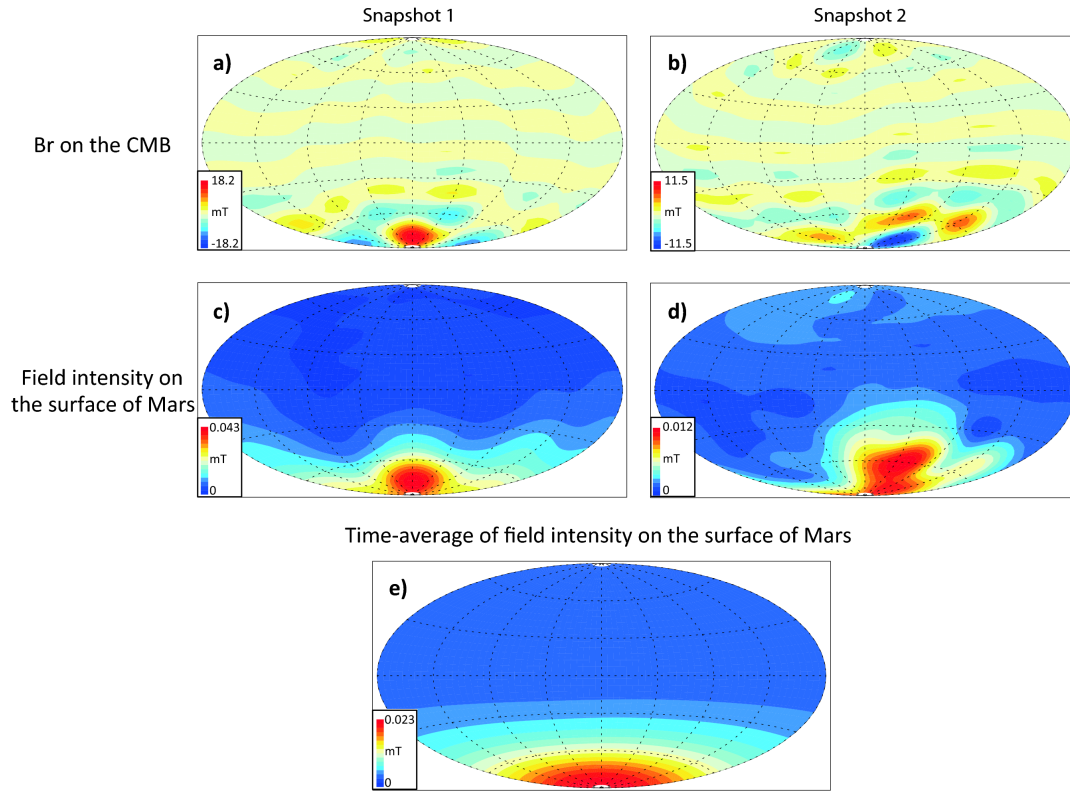


Figure 7: Two arbitrary snapshots (two first lines) of the radial magnetic field B_r on the core-mantle boundary CMB (a and b) and surface intensity of the magnetic field (c and d) in case I3. The time-average surface intensity is shown on the third line (e). Here the spherical harmonic expansion is truncated at $\ell_{max} = 10$. Note differences in color scales among the maps. The CMB field is upward continued to the surface of Mars as a potential field.

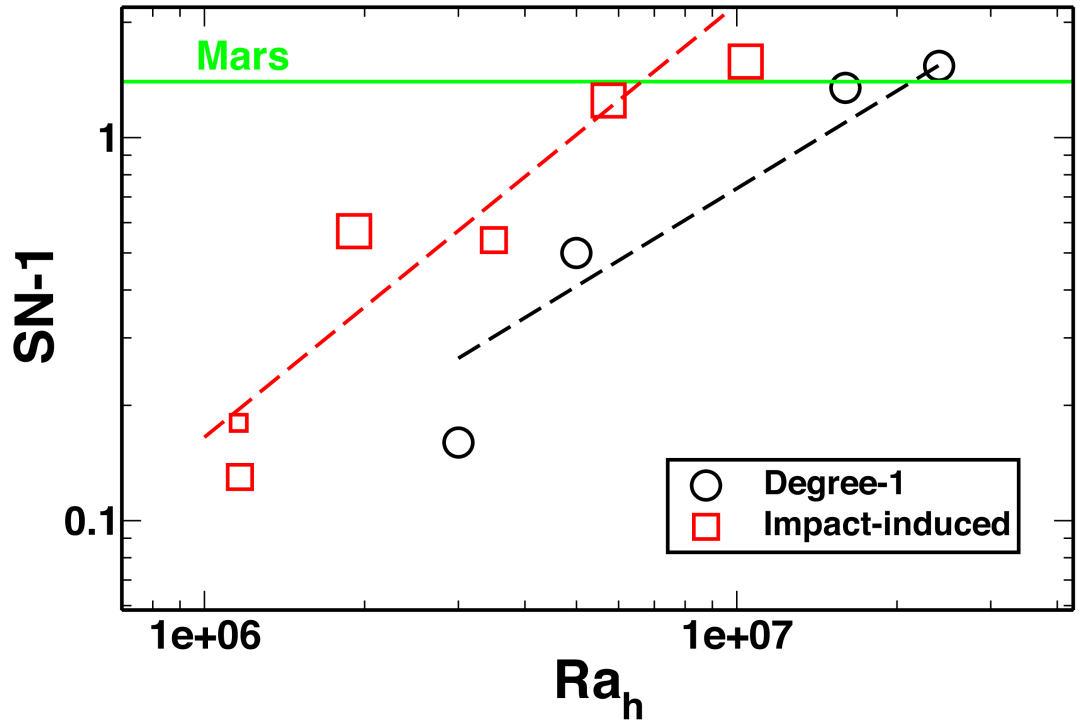


Figure 8: South-north dichotomy as a function of the horizontal Rayleigh number Ra_h (Eq. 11). The black symbols are the values obtained (with $r_i/r_o = 0.2$) for the synthetic Y_1^0 CMB heat flux patterns (with $Ra_h = q^* Ra$). The red symbols are the values obtained here in cases of polar impact driven CMB heat flux patterns (with $Ra_h = q_0^r Ra$). The size of the red symbols increases with the size of the impactor. The corresponding power law fits are plotted with dashed lines. The green horizontal line represents the martian value of $SN - 1$ from Amit et al. (2011) based on observations of the martian crustal magnetic field (Langlais et al., 2004). All degree-1 cases are from Amit et al. (2011) except case Y2 from this study (see also Tab. 1).

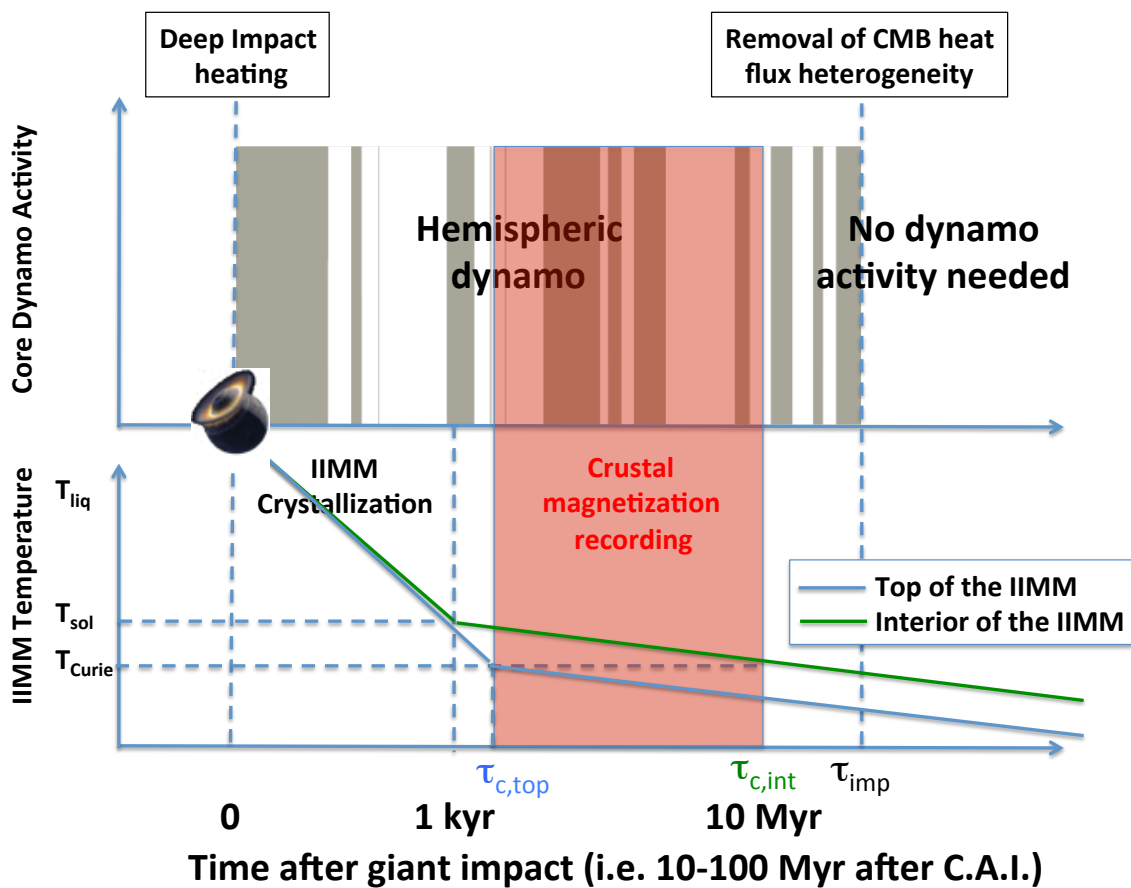


Figure 9: Schematic illustration of temporal evolution of core dynamo activity (top panel) and of the IIMM temperature (bottom panel) after a giant impact. The C.A.I. (Calcium-Aluminium rich Inclusions) are the oldest objects in the solar system. In the bottom panel, the blue line represents the temperature evolution at the interior of the IIMM while the green line represents the top of the IIMM (i.e. the post impact martian surface).