Possible crippling of the core dynamo of Mars by Borealis impact

Jafar Arkani-Hamed

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1. Introduction

[1] Assuming that the northern lowland of Mars is created by a giant Borealis impact, I investigate its consequences on the thermal state of the mantle and the core dynamo for four different impact models using the scaling laws of crater formation, the shock pressure model of Pierazzo et al. (1997), and the “foundering” shock heating model of Watters et al. (2009). The impact heating enhances the temperature of the mantle by 1000–3000 K down to \( \sim 1000 \) km depth. The superheated upper mantle ascends rapidly as a giant plume and develops a strong convection in the entire mantle of the subimpact hemisphere, while the antipodal hemisphere remains almost undisturbed for the period of 100 Myr considered in this study. The upwelling of the plume rapidly sweeps up the impact-heated base of the mantle and replaces it with the cold surroundings, reducing the effects of the impact-heated mantle on the heat loss of the core. However, direct shock heating stratifies the core and effectively suppresses a preexisting thermal convection in the core. This cripples a preexisting thermally driven core dynamo. It takes 35–85 Myr for the stratified core model to exhaust impact heat and resume global convection and possibly regenerate a strong dynamo. Adding the superheated iron content of an impactor on the core does not create an appreciable dynamo but elongates the heat exhaustion time and delays the regeneration of a strong dynamo by an additional \( \sim 40 \) Myr. It is concluded that Borealis impact, if it occurred, could have crippled the core dynamo of Mars for up to \( \sim 70–120 \) Myr.

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in section 3. The direct shock heating of the core causes thermal stratification and suppresses a thermally driven core dynamo [Arkani-Hamed and Olson, 2010a, 2010b], as demonstrated in section 4. The subsequent cooling of the stratified core is addressed in section 5, and the generation of a core dynamo in the convecting outer core is investigated in section 6. Reese and Solomatov [2010] suggested that the iron content of the impactor may descend and create a superheated layer of iron on the core. Section 7 explores the effect of such an iron layer on subsequent core cooling and regeneration of the core dynamo. The lack of magnetic signature of the lowland as a whole and the presence of some weak anomalies over the lowland are explained in section 8. Concluding remarks are relegated to section 9.

2. Impact Heating of Mars

[4] This section estimates the shock pressure distribution and temperature increase in the interior of Mars induced by a Borreals impact, closely following the procedure used by Arkani-Hamed and Olson [2010a]. An axisymmetric spherical Mars model is adopted and it is assumed that the impact occurred vertically and created a circular basin. The axis of symmetry passes through the impact site and the center of Mars. Although the axes of the ellipse fitted to the lowland are ~8500 and 10,600 km [Andrews-Hanna et al., 2008] I consider a circular basin diameter of either 7000 or 8000 km to be conservative and also to account for possible subsequent modifications of the basin. Two impact velocities of 10 and 6 km/s are used. The former is the average of the impact velocities of Mars proposed by Neukum and Wise [1976] and the upper limit of the impact velocity in the sweet spot of the parameter space investigated by Marinova et al. [2008]. The latter is the lower limit of the velocity in the sweet spot. The 10 km/s impact velocity is probably more realistic (see below).

[5] The impactor size is estimated on the basis of empirical scaling laws of Holsapple [1993] and Melosh [1989]. Briefly, the diameter of the model basin is related to the diameter of its transient cavity, which provides a means to calculate the kinetic energy of the impactor. Using an impact velocity and the impactor density, I then determine the impactor size. Note that Holsapple’s [1993] scaling law probably underestimates the transient diameter [Stewart, 2010], because the formation of a giant impact basin is largely controlled by dynamic weakening along impact-induced faults and the transient diameter is closer to the final diameter of the basin than that derived from the scaling laws [e.g., Senft and Stewart, 2009]. The direct relationship between the transient diameter and the energy of the impactor indicates that the calculated impact effects are also underestimated. The shock pressure distribution in a Mars model is calculated using the average model of Pierazzo et al. [1997] and assuming constant but different acoustic velocities and densities in the core and the mantle. It is worth mentioning that Pierazzo et al. [1997] considered impact velocities of 10 km/s or higher. Their model may be less accurate for the low impact velocity of 6 km/s considered in the present study. As a shock wave passes, the target decompresses adiabatically and a large percentage of the deposited energy is released as kinetic energy (i.e., excavation of the basin). The remaining energy goes into heat [Bjorkman and Holsapple, 1987], increasing the temperature of the impactor and target. The corresponding impact-induced temperature increase is estimated using the “founding” shock heating model of Watters et al. [2009]. Note that the founding model yields the least temperature increase in the lower mantle among the three shock heating models investigated by Watters et al. [2009], emphasizing that the temperature increases presented in this paper are probably further underestimated. Table 1 lists the physical parameters used in this study, and Table 2 gives the results of the scaling laws applied to the four Borreals impact models. At a reasonable impact velocity of 10 km/s, the mass of an impactor capable of creating the giant northern lowland is an order of magnitude smaller than that of the impactors considered during the accretion of Mars [e.g., Ke and Solomatov, 2009].

Figure 1 shows the impact-induced temperature increase in Mars for impact model A. The results for the other models have similar characteristics in spatial variations, though different temperature values. The shock wave propagates as a spherical wave centered on the isobaric sphere located directly beneath the impact site. It is reflected at the surface with a 180° phase change and no loss of

<p>| Table 1. Physical Parameters Common Among the Models |</p>
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value (Unit)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mars radius ($R_\text{M}$)</td>
<td>3390 (km)</td>
</tr>
<tr>
<td>Core radius ($R_\text{C}$)</td>
<td>1700 (km)</td>
</tr>
<tr>
<td>Gravity at the surface ($g$)</td>
<td>3.72 (m/s²)</td>
</tr>
<tr>
<td>Gravity at the core-mantle boundary ($g_c$)</td>
<td>3.56 (m/s²)</td>
</tr>
<tr>
<td>Mantle density ($\rho_m$)</td>
<td>3500 (kg/m³)</td>
</tr>
<tr>
<td>Core density ($\rho_c$)</td>
<td>7500 (kg/m³)</td>
</tr>
<tr>
<td>Melt density ($\rho_m$)</td>
<td>2700 (kg/m³)</td>
</tr>
<tr>
<td>Impactor density ($\rho_{\text{imp}}$)</td>
<td>3000 (kg/m³)</td>
</tr>
<tr>
<td>$P'$ wave velocity in the mantle ($V_{p}'_\text{m}$)</td>
<td>7.24 (km/s)</td>
</tr>
<tr>
<td>$P'$ wave velocity in the core ($V_{p}'_\text{c}$)</td>
<td>4 (km/s)</td>
</tr>
<tr>
<td>$S$ wave velocity in the mantle ($V_{s}'_\text{m}$)</td>
<td>4 (km/s)</td>
</tr>
<tr>
<td>Impact velocity ($V_{\text{imp}}$)</td>
<td>10 (km/s)</td>
</tr>
<tr>
<td>The constant in Pierazzo et al.’s [1997] model for the mantle ($S_m$)</td>
<td>1.25</td>
</tr>
<tr>
<td>The constant in Pierazzo et al.’s [1997] model for the core ($S_c$)</td>
<td>1.6</td>
</tr>
<tr>
<td>Specific heat of the mantle ($C_{p_m}$)</td>
<td>1200 (J/kg/K)</td>
</tr>
<tr>
<td>Specific heat of the core ($C_{p_c}$)</td>
<td>600 (J/kg/K)</td>
</tr>
<tr>
<td>Thermal expansion coefficient of the mantle at the surface ($\alpha_{m_s}$)</td>
<td>$3 \times 10^{-5}$ (1/K)</td>
</tr>
<tr>
<td>Thermal expansion coefficient of the core ($\alpha_c$)</td>
<td>$10^{-4}$ (1/K)</td>
</tr>
<tr>
<td>Melting expansion coefficient of the mantle ($\Delta \alpha_m$)</td>
<td>0.024</td>
</tr>
<tr>
<td>Thermal conductivity of the core ($K_c$)</td>
<td>40 (W/m/K)</td>
</tr>
<tr>
<td>Thermal conductivity of the mantle at 273 K ($K_m$)</td>
<td>4 (W/m/K)</td>
</tr>
<tr>
<td>The preimpact mean heat flux at the core-mantle boundary ($J_{\text{pm}}$)</td>
<td>40 (mW/m²)</td>
</tr>
<tr>
<td>Latent heat of melting ($L$)</td>
<td>400 kJ/kg</td>
</tr>
<tr>
<td>Surface temperature ($T_s$)</td>
<td>230 K</td>
</tr>
<tr>
<td>Present average uranium content of the mantle ($\text{U}_\text{avg}$)</td>
<td>16 ppb</td>
</tr>
<tr>
<td>Present average potassium content of the mantle ($\text{K}_\text{avg}$)</td>
<td>305 ppm</td>
</tr>
<tr>
<td>Present average thorium content of the mantle ($\text{Th}_\text{avg}$)</td>
<td>56 ppb</td>
</tr>
<tr>
<td>Reference viscosity at the core-mantle boundary ($\nu_\text{ref}$)</td>
<td>$10^{15}$ Pa s</td>
</tr>
<tr>
<td>Kinetic viscosity of the core ($\nu_c$)</td>
<td>1 m²/s</td>
</tr>
</tbody>
</table>

*Sparks and Parmentier [1994]
energy. The interference of the direct and reflected waves reduces the resulting shock pressure near the surface. Upon impinging the core-mantle boundary, the direct shock wave is partly reflected back to the mantle and partly transmitted to the core. The reduction of temperature in the mantle near the core-mantle boundary due to the interference of the direct and reflected waves is small and ignored. Because the shock wave velocity in the liquid iron core is smaller than that in the overlying solid silicate mantle, the refraction angle, calculated on the basis of Snell’s law, is always smaller than the incident angle at the core-mantle boundary. Consequently, parts of the mantle and core do not receive shock waves. Table 2 includes the colatitude relative to the subimpact point on the core where the shock ray is tangent to the core, in degrees.

Table 2. Characteristics of the Scaling Laws

<table>
<thead>
<tr>
<th>Model</th>
<th>D_b</th>
<th>D_tr</th>
<th>D_iso</th>
<th>D_imp</th>
<th>V_imp</th>
<th>M</th>
<th>P_s</th>
<th>E_imp</th>
<th>E_m</th>
<th>E_c</th>
<th>E_heat/E_imp</th>
<th>(\theta_c)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>8000</td>
<td>4498</td>
<td>1184</td>
<td>1614</td>
<td>10</td>
<td>7.7</td>
<td>2.4</td>
<td>3.8</td>
<td>1.0</td>
<td>0.03</td>
<td>0.27</td>
<td>52.6</td>
</tr>
<tr>
<td>B</td>
<td>7000</td>
<td>3977</td>
<td>1012</td>
<td>1381</td>
<td>10</td>
<td>4.8</td>
<td>2.4</td>
<td>2.4</td>
<td>0.78</td>
<td>0.02</td>
<td>0.33</td>
<td>53.9</td>
</tr>
<tr>
<td>C</td>
<td>8000</td>
<td>4498</td>
<td>1414</td>
<td>2148</td>
<td>6</td>
<td>18.2</td>
<td>1.2</td>
<td>3.3</td>
<td>0.4</td>
<td>0.04</td>
<td>0.13</td>
<td>50.7</td>
</tr>
<tr>
<td>D</td>
<td>7000</td>
<td>3977</td>
<td>1210</td>
<td>1838</td>
<td>6</td>
<td>13.3</td>
<td>1.2</td>
<td>2.0</td>
<td>0.4</td>
<td>0.04</td>
<td>0.22</td>
<td>52.4</td>
</tr>
</tbody>
</table>

* \(D_b\) is basin diameter, \(D_{tr}\) is the transient cavity diameter, \(D_{iso}\) is the diameter of the isobaric sphere, and \(D_{imp}\) is the impactor diameter, all in km. \(V_{imp}\) is the impact velocity in km/s and \(M\) is the mass of the impactor in \(10^{11}\) kg. \(P_s\) is the shock pressure in the isobaric sphere in \(10^{11}\) Pa s. \(E_{imp}\) is the kinetic energy of the impactor, \(E_m\) is the heat energy partitioned to the mantle, \(E_c\) is the heat energy partitioned to the core, \(E_{heat}\) is the total heat energy input to Mars, and \(\theta_c\) is the colatitude relative to the subimpact point on the core where the shock ray is tangent to the core, in degrees.

Figure 1. The temperature increase in the mantle and the core of model A (Table 2) by shock heating following the Borealis impact. No attempt is made to incorporate the latent heat of melting and evaporation near the impact site. The black areas do not receive any shock wave. The heating of the mantle by the shock wave emerging out of the core in the antipodal region is insignificant and ignored. Note that the maximum temperature increase in the antipodal part of the core is about 50 K. Upon emerging out of the core the shock wave suffers energy loss due to partial reflection into the core. Also, because of the higher specific heat of the silicate mantle compared to that of the iron core, the maximum impact temperature increase directly above the core is in the order of 25 K.
The jump increases the outward heat flux of the core and does not allow the impact-heated lowermost mantle to reduce the heat flux.

Mantle heating by the shock wave emerging out of the core in the antipodal region is small and ignored in Figure 1, partly because the shock wave traveling the entire core loses a major amount of its energy and partly because a fraction of the shock wave impinging the antipodal region of the core-mantle boundary is reflected back into the core. Figure 2b shows that the impact temperature increase at the antipo-

**Figure 2.** (a) Shock pressure and (b) impact-induced temperature increase along the axis of symmetry for models A–D (see Table 2). A negative radius means a radius in the antipode direction. The mantle in this antipodal region is not shown in the horizontal axis. Note that shock pressure and temperature increases in the antipodal region of the mantle are not calculated, as explained in the Figure 1 caption, which explains the Figure 2 cut off at the −1700 km radius. The ray theory [Aki and Richards, 2002] is adopted for calculation of shock wave transmission across the core-mantle boundary.
3. Impact-Induced Mantle Dynamics

Borealis impact drastically changes the mantle dynamics, by introducing a huge amount of impact heat in the upper mantle as seen in Figure 1. This section is concerned with the mantle dynamics caused by the impact heating. The effects of a large impact on the mantle dynamics of terrestrial planets have been investigated on the basis of convection in a rectangular coordinate [Watters et al., 2009], in an axisymmetric cylindrical coordinate [Monteux et al., 2007], in an axisymmetric spherical coordinate [Ghods and Arkani-Hamed, 2007] and (Ghods, A., and J. Arkani-Hamed, Effects of the Borealis impact on the thermal evolution of Mars, submitted to Physics of the Earth and Planetary Interiors, 2010), and in a three-dimensional spherical coordinate [Roberts et al., 2009; Ke and Solomatov, 2009]. I follow the technique used by Ghods and Arkani-Hamed (2007) in their studies of the effects of Aiken impact on the thermal evolution of the Moon, partly because the heating of the mantle by a vertical impact is axisymmetric and the geometry of the axisymmetric spherical coordinate is more relevant. Briefly, I adopt a self-gravitating spherical mantle, with temperature-dependent and melt depletion-dependent viscosity, temperature-dependent thermal conductivity, and a pressure-dependent thermal expansion coefficient. The mantle dynamic equations are solved in a spherical shell with an inner radius $R_e = 1700$ km and an outer radius $R_o = 3390$ km. The temperature at the surface is kept unchanged and a cooling core-mantle boundary is adopted.

A preimpact model is calculated to establish a reasonable temperature distribution prior to the impact. The model starts at 4.6 Ga, assumed to be after the accretion of the planet and the core formation and magma ocean solidification, with an adiabatic initial temperature distribution in the mantle and core that reaches 2000 K at the core-mantle boundary [Hauck and Phillips, 2002]. The temperature decreases linearly in the upper 100 km from ~1600 K to the surface temperature of 230 K. The core is assumed to be liquid with an adiabatic temperature distribution. Radiative heat generation is taken into account in the mantle and the crust, while no internal heating is considered for the core. The radioactive elements concentration at present follows that of Wanke et al. [1994] with K, Th, and U values of 305 ppm, 56 ppb, and 16 ppb, respectively. Thirty percent of the radioactive elements are allocated to the crust and the rest are uniformly distributed in the mantle. There is good evidence that a stagnant lid has existed on Mars almost from day one [e.g., Breuer and Spohn, 2003; Williams and Nimmo, 2004]. Note that a stagnant lid is produced due to both the low temperature near the surface which increases the strength of the material, and the low density of the crust which increases the buoyancy of the lid. The estimates of the crustal thickness range in a very wide spectrum from 50 to 250 km [Breuer and Spohn, 2003]. The interpretation of the gravity and topography suggests crustal thicknesses of 40–80 km [Zuber et al., 2000]; the thickest crust underlies the oldest regions. Because of very early chemical differentiation [e.g., Halliday et al., 2001], it is plausible that a major part of the crust was formed while the magma ocean was solidifying. The initial thickness of the crust is taken to be 50 km in the models. A stagnant lid convection regime is adopted in the calculations of mantle dynamics. The initial thickness of the stagnant lid is taken to be 60 km, assuming that the base of the stagnant lid is determined by an elastic-to-ductile transition temperature of ~1070 K [e.g., Stevenson et al., 1983; Schubert and Spohn, 1990]. Because of inefficient heat transfer by thermal conduction through the stagnant lid, the radioactive heating gradually increases the mantle temperature in the early history of the planet [e.g., Breuer and Spohn, 2003]. Detailed study of the radioactive heating of Martian mantle, considering different amounts of partitioning of the radioactive elements to the crust, suggests that the temperature in the mantle may increase by up to 250 K (e.g., see the nominal model of Arkani-Hamed [2005a]). Adopting a higher initial temperature in the mantle, such as solidus [Ke and Solomatov, 2009] will guarantee that the mantle remains partially molten for a long time or even up to the present. Note that Ke and Solomatov [2009] did not include radioactive heating in their models.

The mantle is allowed to melt as its temperature surpasses the solidsus. I use McKenzie and Bickle’s [1988] batch melting model for fertile dry peridotite to calculate the fractional melting at the end of each time step. Because of its lower density compared to that of solid residue, melt separates from the solid residue and rapidly moves to the surface. Only a small few percent of melt that is trapped in the disconnected pores remains with the solid residue. In this study I assume that melt remains in the solid matrix and moves with the matrix if it is less than or equal to 3% of a volume element. The excess melt is extracted and placed on the surface instantaneously as a newly forming part of the crust. I also take into account the enthalpy of melting in the energy equation and calculate the temperature and melt depletion fields accordingly. A melt depletion field specifies the solid residue after a certain amount of melting. For example, a melt depletion of 0.20 indicates a solid residue after 20% melting. Melting is a chemical reaction which depletes the iron content of the solid residue, reducing its density and increasing its viscosity [e.g., Hirth and Kohlstedt, 1996]. Therefore the density of the solid residue is a function of temperature and the fractional melting. I consider both melt depletion buoyancy, due to the low-density solid residue, and melt retention buoyancy, due to the presence of low-density melt in the pores, in the momentum equations. The density of the solid residue $\rho_s$ is calculated by

$$\rho_s = \rho_m[1 - \alpha_m(T - T_o) - \beta_mM],$$ (1)
where \( \rho_m \) is the density of the pristine mantle at surface temperature \( T_o (= 230 \text{ K}) \), \( T \) is the temperature in kelvin, \( \alpha_m \) and \( \beta_m \) are coefficients of thermal expansion and melt expansion of the mantle, respectively, and \( M \) denotes the melt depletion. The effective density of the matrix (i.e., a volume element) \( \rho \) is determined by

\[
\rho = \rho_o + \varphi (\rho_l - \rho_o),
\]

where \( \varphi \) is the volumetric content of melt and \( \rho_l (= 2700 \text{ kg/m}^3) \) is the density of the melt.

[13] The mantle viscosity is assumed to be temperature and melt depletion dependent. The pristine mantle viscosity is calculated by

\[
\eta = \eta_o \Delta \eta \exp \left( -C (T - T_o) \right)
\]

where the reference viscosity \( \eta_o \) is the viscosity at the core-mantle boundary, \( \Delta \eta \) is the viscosity contrast between the surface and the core-mantle boundary, and \( C = -\ln [\Delta \eta / (T_{\text{CMB}} - T_o)] \) denotes the temperature dependence of the viscosity where \( T_{\text{CMB}} \) is the temperature at the core-mantle boundary. Ghods and Arkani-Hamed (submitted manuscript, 2010) examined reference viscosities of \( 10^3, 10^5 \), and \( 10^7 \text{ Pa s} \), and viscosity contrasts of 1000, 2000, 5000, and 10000. It was found that for a given reference viscosity the main characteristics of the mantle dynamics remain almost unchanged as long as a stagnant lid exists on top, at least during the 100 Myr investigated. A viscosity contrast of 1000 is found sufficient to create a stagnant lithosphere, which is also adopted in this study along with the reference viscosity of \( 10^5 \text{ Pa s} \). Although the vigor of the mantle convection decreases at high reference viscosities, the mantle dynamics is still vigorous especially in the impact-heated region. For the melt depletion dependency, the viscosity of the solid residue is increased linearly, by up to a factor of 10 for a maximum of 25% melting, which is the maximum melting that can be handled by the computer code used in this study. Note the distinction between the melt created by partial melting, a maximum of 25%, and the melt retained in the pores of the matrix, a maximum of 3%. The effect of the pore melt on the viscosity of the matrix is usually small and is ignored in the models, assuming that the evolution of the mantle dynamics is dominated by solid rheology. The viscosity field is time varying and changes both radially and laterally during the evolution because of the time and spatial variations of temperature and melt depletion fields. Figure 3 shows the spatial variations of viscosity immediately after the impact, which is assumed to occur at 4.5 Ga [e.g., Frey et al., 2002]. The viscosity changes significantly in the stagnant lithosphere and moderately in the convecting part of the mantle, as suggested for the stagnant lid convection regime [e.g., Grasset and Parmentier, 1998]. The region directly beneath the impact site which has experienced appreciable melting (see Figure 4b) ends up having a relatively higher viscosity because the melt has already moved up but left behind the high-viscosity solid residue except for a small amount of melt in the pores. The decrease of the solid residue viscosity due to temperature increase is not sufficient to reduce the viscosity of the region appreciably. In the pristine mantle immediately outside this region, the viscosity is decreased below the viscosity of the preimpact model, because the impact heating has elevated the temperature but has not caused melting.

[14] The temperature-dependent thermal conductivity \( K_m \) of the mantle follows Schatz and Simmons [1972] experimental results:

\[
K_m = 414.8/(30.6 + 0.21T); T < 500K
\]

\[
K_m = 414.8/(30.6 + 0.21T) + 0.0023(T - 500); T > 500K.
\]

The thermal expansion coefficient of the mantle \( \alpha_m \) is \( 3 \times 10^{-5} \text{ at the surface and decreases linearly with depth by 50% at the core-mantle boundary.} \)

[15] A postimpact model starts after the very early stages of impact-induced melting [Nimmo et al., 2008; Marinova et al., 2008], when the melt has already moved to the surface and the mantle can be treated as a viscous solid with no
Figure 4
more than 3% melt. However, to account for the thermal effects of these early processes, the temperature is reset to correspond to the temperature of partial melt wherever melting occurred, assuming that temperature in the partially molten region increases linearly from solidus to liquidus. The maximum temperature corresponds to 25% partial melting. This approximation may underestimate the effects of the impact heating on the mantle dynamics.

[16] To better illustrate impact effects on the mantle dynamics, I consider two models, one with impact and the other without impact, and compare their results. For a model with impact it is required to add the preimpact temperature distribution to a preimpact temperature increase and a preimpact temperature increase to a preimpact temperature distribution. The preimpact temperature is required to add the preimpact temperature distribution to the temperature distribution of the preimpact temperature. The resulting temperature is then modified to account for the enthalpy of melting where it surpasses the solidus. The 2-D axisymmetric temperature distribution thus obtained is regarded as the initial temperature of the postimpact model. For the model with no impact the spherically symmetric part of the preimpact temperature is regarded as the initial temperature. Also the preimpact convection velocity is set to zero immediately before the impact in both models. This isolates the impact-induced convection of the postimpact models from that of the preimpact convection. The postimpact thermal evolution models start at the impact time. All models are calculated for a period of 100 Myr, but the results of the first 11 Myr are presented here which is sufficient to demonstrate the development of the impact-induced mantle dynamics and the removal of the impact-heated base of the mantle away from the core-mantle boundary. The core temperature of a model with impact is spherically symmetric, but not adiabatic. The heat flux at the core-mantle boundary is estimated from the cooling history of a stratified core (see below). On the other hand, the core temperature is adiabatic in a model with no impact.

[17] Figure 4a shows snapshots of the 2-D temperature distribution of model A with impact and illustrates details of mantle dynamics at every 1 Myr interval. The upwelling of the highly buoyant hot upper mantle as a giant plume dominates the postimpact mantle dynamics. It develops upward motion of the entire column of the mantle directly beneath the impact site down to the core-mantle boundary and results in appreciable lateral motion in the base of the mantle, which brings the surrounding colder material to the subimpact region. Moreover, the impact-heated and stratified core with spherically symmetric temperature (see below) heats the entire base of the mantle and creates a low-viscosity layer on the entire surface of the core, which further facilitates the lateral motion of the base of the mantle and prevents the impact-heated lowermost mantle to reduce the heat flux out of the core. The snapshots of the melt depletion field (Figure 4b) better illustrate the mantle dynamics. The impact-heated, hot plume of low-density solid residue ascends rapidly leaving almost no trace near the core-mantle boundary by 4 Myr. The plume remains in the subimpact hemisphere. The quiescence of the antipodal hemisphere characterizes the model with no impact, exhibiting negligible convection and rendering the temperature distribution almost spherically symmetric. For this reason, the model with no impact is not presented. Note that setting the preimpact velocity to zero immediately before the impact disentangles the effects of convection initiation from the effects of the impact-heating to a very good extent. The antipodal hemisphere exhibits negligible convection initiation. The postimpact mantle dynamics are almost entirely controlled by impact, at least for the 100 Myr investigated.

[18] Figure 5 shows the temperature profiles along the axis of symmetry directly beneath the impact site at every 2 Myr and the solidus and liquidus of dry peridotite adopted for the mantle. The upper parts of the mantle melt extensively. The melt in excess of 3% is extracted and put on the surface of Mars instantaneously, as mentioned before. The temperature at the core-mantle boundary gradually reduces toward the preimpact value as the stratified core cools. Also included in Figure 5 is the radially averaged, spherically symmetric profile of the preimpact temperature distribution immediately before the impact. It is clear that convection in the preimpact model was effective enough to create a well-defined thermal boundary layer near the core-mantle boundary.

4. Direct Impact Heating of the Core

[19] Shock waves propagate in the mantle as spherical waves centered on the isobaric sphere. Upon impinging the core-mantle boundary part of the waves is transmitted to the core. This section explains the direct heating of the core by the transmitted shock waves. I closely follow the technique developed by Arkani-Hamed and Olson [2010a]. Briefly, shock waves are traced using seismic ray theory [Aki and Richards, 2002] and the seismic velocities of the mantle and core listed in Table 1. The results obtained using the impedance match method [Watters et al., 2009] or the shock dynamic method [Han and Yin, 1993] are similar to that of the ray theory. Shock heating in the core is calculated for the material properties of liquid iron (Table 1). Figure 6 shows the resulting temperature increase in the core for the four models. The black regions of the core do not receive shock waves, because the shock wave velocity in the core is less than that in the mantle and Snell’s law prohibits the shock waves to enter these regions, as mentioned before. Figure 6 shows that the high velocity impact heats the shallower parts of the core in the subimpact region more than the low velocity impact, whereas the antipodal region of the core is less heated by the high velocity impact. This is because the effective shock pressure $P_b = \rho_v (P_s - P_m)$ used in the foundering shock heating model of Watters et al. [2009] increases with radius in the antipode region. Although the shock pressure, $P_v$, decreases with distance from the center.
of the isobaric sphere, the lithostatic pressure $P_o$ decreases even faster with radius (Figure 2a).

[20] The differentially heated core becomes quite unstable and an energetic thermal adjustment takes place within a very short time and redistributes the shock-heated fluid onto spherically symmetric isothermal surfaces with increasing temperature as a function of distance from the center, resulting in a stable thermal stratification. The evolution of the temperature distribution during stratification inside a core with characteristics similar to those of the model core considered here [Arkani-Hamed and Olson, 2010b, Figure 3] indicates that the core becomes stratified within a short period of 1–2 kyr, which is also assumed in the present study. The stratified temperature is calculated by determining the volume fraction of the core versus the shock-heated temperature increase. Assuming that the core was initially close to a well-mixed state the stable thermal stratification is then calculated by adding the stratified impact temperature increase to the preimpact adiabatic temperature and refilling the core according to temperature. Figure 7 shows the spherically symmetric temperature distributions in the stratified core for the four models thus calculated. The preimpact adiabatic temperature is assumed 2000 K at the core-mantle boundary. The impact-heated part of the core is concentrated in the upper ~300 km. Note that the shock-induced temperature perturbations are far larger than the temperature perturbations associated with convection in iron-rich planetary cores, estimated to be less than 1 K [e.g., Stevenson, 1987; Christensen and Wicht, 2007]. The direct impact heating of the core and subsequent thermal stratification can effectively suppress a preexisting thermally driven core convection for a limited time [Arkani-Hamed and Olson, 2010b].

5. Cooling of the Stratified Core

[21] The upward concentration of the impact heated, high-temperature part in the stratified core suppresses core convection shortly after the impact. However, because of the very low viscosity of the liquid iron, convection develops in the outer parts and heat is transferred to the mantle by convection. This section presents the poststratification thermal evolution of the core calculated following the procedure adopted by Arkani-Hamed and Olson [2010a, Appendix A].

[22] Figure 1 shows that about 20% of the core-mantle boundary receives direct shock wave and major shock heating actually occurs only on ~10%. The base of the mantle overlying the remaining 80% of the boundary is not impact heated at all and it dominates subsequent cooling of the core which has a spherically symmetric temperature. Moreover, the fast upwelling of the buoyant upper mantle directly beneath the basin pulls up the impact-heated part of

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**Figure 5.** Temperature profiles of model A (Table 2) along the axis of symmetry beneath the impact site at certain times after the impact. The numbers on the curves denote time after the impact in Myr. Curves S and L are the mantle solidus and liquidus. The “before” curve is the spherically symmetric temperature distribution immediately before impact. It shows that a pronounced thermal boundary layer was formed at the core-mantle boundary before the impact. Regions with temperatures higher than solidus are partially molten. Note the rapid upwelling of the impact-heated regions and the temperature decrease in the lower parts of the mantle to almost its preimpact condition, due to the lateral motion of the mantle to fill the space drained by the upwelling. Note that the temperature at the core-mantle boundary gradually reduces as the core cools.
CT/C14 is temperature, \( \alpha \) is the thermal diffusivity. The convective outer layer is determined at the beginning of each time step \( \Delta t \) which assures that heat does not diffuse by more than \( \delta/2 \) during that interval; \( \kappa_c \) denotes the thermal diffusivity of the core. Note that the time steps are variable because of the time varying thickness of the boundary layer (see below). Figure 8 shows the temperature profiles in the core and the overlying mantle layer at certain times after the core stratification for model A. The other models have similar temperature profiles but with different values. I use kinematic viscosity of 1 m\(^2\)/s for the outer core, which is higher than the viscosity values suggested for the outer core of the Earth based on geodetic, geomagnetic, and seismic studies [Secco, 1995, Tables 1–3]. I also examined viscosity values of \( 10^{-5} \) and \( 10^{-4} \) m\(^2\)/s for short time periods due to the huge computer time demand. The results were almost identical, emphasizing that the thermal evolution of the stratified core is mainly controlled by the overlying mantle layer. Following the stratification there is a hot layer near the surface of the core, specified by curve 0 in Figure 8, which makes the temperature gradient positive and suppresses core convection (the adiabatic temperature gradient in a convecting core is negative). Note that while the impact heat is being exhausted to the mantle through convection in the outer hot layer, the middle part of the core between 1000 and 1650 km radii is actually heated up, largely by the downward heat conduction from the outer hot layer and partly by upward heat conduction along the adiabat from the central part. This increases the temperature gradient and results in a sub-adiabatic core below the outer hot layer.

[25] The thermal boundary layer thickness \( \delta \) of the convecting outer core is determined at the beginning of each time step following King et al. [2009], as described by Arkani-Hamed and Olson [2010a]. The boundary remains thinner than 20 m for a major part of the thermal evolution of the stratified core (Figure 9a).

[26] The convection in the outer core is penetrative with gradual thickening of the convecting layer, as seen from the time variations of its thickness (Figure 9b). However, shortly after the exhaustion of the major part of the impact
heat the convection zone penetrates very rapidly to the center and global convection of the core resumes. This rapid penetration is due to the fact that the temperature in the deeper parts of the core remains close to the adiabatic temperature as seen in Figure 8. The rapid penetration increases the characteristic length of convection by a factor of \(\sim 3\) and the thermal Rayleigh number by a factor of \(\sim 27\), two orders of magnitude higher than that required for regeneration of dynamo as indicated by the numerical models of Kuang et al. [2008]. Figure 9b shows that depending on the size and the velocity of the impactor it may take about 35–85 Myr for the convective layer to exceed the critical thickness and the core become capable of regenerating a strong core dynamo.

Figure 10 shows the heat flux out of the core of the four models. The heat flux starts at a very high value because the core stratification places the most impact-heated part of the core directly beneath the core-mantle boundary. However, this substantially heated part is quite thin and cools rapidly allowing the heat flux to reduce appreciably. It is worth remarking that the heat flux shown in Figure 10 is used as the cooling boundary condition at the core-mantle boundary in the calculations of the mantle dynamics explained above.

[27] Figure 10 shows the heat flux out of the core of the four models. The heat flux starts at a very high value because the core stratification places the most impact-heated part of the core directly beneath the core-mantle boundary. However, this substantially heated part is quite thin and cools rapidly allowing the heat flux to reduce appreciably. It is worth reminding that the heat flux shown in Figure 10 is used as the cooling boundary condition at the core-mantle boundary in the calculations of the mantle dynamics explained above.

[28] The hemispheric dichotomy of the magnetization of the Martian crust was recently related to a single hemispheric dynamo in the south [Stanley et al., 2008]. It was suggested that a 1° mantle convection or a giant Borealis impact raised the temperature at the core-mantle boundary in the northern hemisphere and created a hemispheric temperature distribution on the core. The colder core-mantle boundary in the south allowed appreciable heat loss from the core which powered a single hemispheric dynamo. The magnetic source bodies in the Martian crust reside in the upper 45–50 km [e.g., Voorhies et al., 2002] and they require 50–100 Myr to cool through their magnetic blocking temperatures and acquire strong magnetization enough to give rise to the magnetic anomalies of Mars, which are an order of magnitude stronger than the magnetic anomalies of the Earth. Figures 4 and 5 emphasize that a single hemispheric dynamo which possibly resulted from the giant Borealis impact does not last more than 5 Myr. It is worth noting that although direct impact heating of the base of the mantle is limited to less than 20% of the core’s surface, the thermally stratified core with a spherically symmetric temperature substantially heats the entire base of the mantle by over 200 K within the first 5 Myr (see Figure 8). This significantly reduces the viscosity of the mantle immediately above the core-mantle boundary and creates a globe encircling, low-viscosity layer overlying the entire core. The low-viscosity layer facilitates lateral motion of the lowermost mantle and the upwelling of the impact-heated buoyant material, and further diminishes the lateral variations of temperature at the base of the mantle. The time estimated to suppress the lateral variations of temperature in the base of the mantle is much shorter than the time required for a hemispheric dynamo to stay active and magnetize the crust in the southern hemisphere.

6. Regeneration of the Core Dynamo

[29] The thermal stratification quenches the core convection within a few thousand years and the preexisting
magnetic field decreases by more than an order of magnitude within one dipole decay time, \( \sim 10 \text{ kyr} \) \citep{Arkani-Hamed2010b}. However, the convecting outer core is capable of generating a dynamo. Here I make a rough estimate of the dipole magnetic field intensity arising from the mean magnetic field that is expected to be generated in the convecting outer core. Using the scaling equation (48) of \citet{Christensen2006}, the mean magnetic strength inside the convecting outer core is determined by

\[
B = \frac{0.9 \mu_0^{1/2} \rho_{\text{i}}^{1/6} (\alpha_c g_R c d_C)}{(C_{\text{pc}} R_i)^{1/3}},
\]

where \( \mu_0 = 4\pi \times 10^{-7} \text{ H/m} \) is the magnetic permeability, \( \rho_i \) (7500 kg/m\(^3\)) and \( g_c \) (3.56 m/s\(^2\)) are the core density and the gravitational acceleration at the core-mantle boundary, respectively, \( \alpha_c = (1 \times 10^{-5} / \text{K}) \) and \( C_{pc} = 600 \text{ J/kg/K} \) are the thermal expansion coefficient and the specific heat of the core, respectively, and \( R_i \) (1700 km), \( R_s \), and \( d = (R_s - R_i) \) are the outer radius (the core radius), the inner radius, and the thickness of the layer, respectively. Note that \( R_i \) and \( d \) are time-dependent because of the penetrative nature of the convection. The buoyancy flow over the entire surface \( Q_B \) in equation (48) of \citet{Christensen2006} is expressed in equation (8) in terms of the advective heat flux per unit area \( q_{\text{adv}} \). In a thermally driven dynamo \( Q_B = 4 \pi R_c^2 \alpha_c q_{\text{adv}} / C_{pc} \). The advective heat flux in the core is the heat flux at the core-mantle boundary \( q \), shown in Figure 10, minus the conductive heat flux, \( q_{\text{cond}} \), along the adiabat. In a convective layer of highly conductive iron, the heat transfer by conduction is smaller than but comparable to the heat transfer by advection. I replace \( q_{\text{adv}} \) by \( q \) to estimate an upper limit for \( B \). Figure 11 shows the mean magnetic field intensity generated inside the convecting outer core. The field intensity is mainly controlled by the heat flux \( q \), which decreases in time, and by the thickness of the convecting layer \( d \), which increases in time. The radius to the bottom of the convecting layer \( R_i \) also decreases in time but relatively slowly. The rapid decrease of the field intensity in the first few million years is due to the fast decrease of the heat flux, while the convecting layer remains thin (see Figure 9b). For a given core model, and except for the rapid decrease in the first few million years, the magnetic field intensity increases gradually with time until the convecting layer penetrates to the center of the core. The field intensity remains lower than \( 4 - 6 \times 10^5 \text{ nT} \) before global convection resumes. Note that the dipole component of the magnetic field is probably \( 3 - 10 \) times weaker than the mean field as estimated by \citet{Christensen2006}. Also the dipole field decays by a factor of \((R_i/R_s)^3 \sim 0.13\) as it reaches the surface of Mars at radius \( R_i \). Therefore, the magnetic dipole field of the convecting outer core expected at the surface of Mars is at least two orders of magnitude weaker than those shown in Figure 11. Higher-degree harmonics decay even more rapidly. Estimates of the paleointensity of the magnetic field at the surface of Mars at around 4.2 Ga, based on the oldest Martian meteorite ALH84001 \citep[e.g.,][]{Weiss2008} and on the magnetostrophic force balance in its liquid core \citep{Arkani-Hamed2005b} indicate that the dipole core field intensity at the surface of Mars was comparable to that on present-day Earth. This indicates that the dipole field generated by the convecting outer core at the surface of Mars is at least two orders of magnitude weaker than that required to magnetize the magnetic source bodies in the

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{Core_Cooling.pdf}
\caption{The thermal evolution of model A (Table 2). The numbers on the curves denote the time after the impact in Myr. Only the upper 200 km of the core is displayed for clarity. The temperature in the deeper part is close to the initial adiabat. Note the region deeper than the 0 Myr curve is heated up and the temperature gradient is positive.}
\end{figure}
Figure 9. Time variations of the thickness of (a) the thermal boundary layer and (b) the convecting outer core of models A–D (Table 2). The dashed lines are for the case where the impactor iron layer is placed on the core. Note that the thickness of the boundary layer is smaller than the grid interval adopted (see text for details). The convecting layer gradually propagates to deeper parts of the core until the impact heat is exhausted and then it penetrates very rapidly.
Martian crust before the rapid penetration of convective layer to the center. The present calculations suggest that Borealis impact, if it occurred, was capable of crippling the core dynamo of Mars for about 35–85 Myr.

7. Effects of an Impactor Iron Layer

The giant Borealis impact is capable of increasing the temperature of the mantle by about 3000 degrees down to about 1000 km depth as seen in Figures 1 and 2. The temperature increase is enough to melt not only the upper mantle directly beneath the impact site but also the impactor. The upper mantle was already depleted in iron during early core formation prior, but the impactor may have an appreciable amount of iron which upon melting sinks due to its higher density compared to the low density of the surrounding silicate mantle. Reese and Solomatov [2010] studied several sinking mechanisms of the impactor iron and concluded that the major part of the iron could sink to the core-mantle boundary and spread over the core directly beneath the solid mantle, creating a super-heated iron layer on the core. This is an interesting possibility and deserves further study.

Here I investigate the effects of the super-heated impactor iron layer on the cooling of the stratified core discussed in the previous section, and determine the magnetic field possibly generated in the layer. I consider model A with an impactor of 1600 km diameter (see Table 2). By taking the particle velocity to be one half the impact velocity, Reese and Solomatov [2010] implicitly assumed the impactor and the target to be made of the similar material, as also adopted here. Accordingly, I assume that the iron content comprises 1/8 of the total volume of the impactor, and let its entire iron sink and spread on the core. This amounts to an impactor iron layer ~8 km thick. Depending on the sinking mechanism the impactor iron may take several days to several thousand years to descend and cover the core, and it may cool or further heat up while sinking [Reese and Solomatov, 2010]. I assume that the time it takes for the total impactor iron to reach the core is comparable with the time it takes for the core to be stratified (a few thousand years), a longer descent time would hamper core cooling by gradually replenishing hot impactor iron to the surface of the core. I also assume no appreciable changes in the temperature of the sinking iron due to friction with the surrounding mantle and set an initial temperature of 4270 K for the impactor iron layer placed on top of the stratified core. This consists of the impact temperature increase in the isobaric sphere (Figure 2) plus the preimpact temperature in the upper mantle. It also takes into account the adiabatic temperature increase from the upper mantle to the core-mantle boundary. The resulting enlarged core is allowed to cool following the procedure described in the previous section. Figure 12 shows the thermal evolution of the enlarged core. The super-heated layer cools very rapidly. It largely loses its heat to the mantle but also appreciably to the middle part of the core, which further delays exhaustion of the impact heat of the core. Included in Figure 11 is the mean magnetic field intensity generated inside the convecting outer core of the enlarged core. The impactor iron layer enhances the magnetic field intensity for about 2 Myr, but does not generate an appreciable core dynamo, while it delays the regeneration of a strong core dynamo by an additional ~40 Myr.

The mean field to dipole field ratio of 3–10 estimated by Christensen and Aubert [2006] is probably conservative if applied to the magnetic field possibly generated in the impactor iron layer. Stanley et al. [2005] numerically
simulated the magnetic field generated inside a thin spherical shell overlying a conductive sphere, for the inner radius to outer radius ratios of up to 0.9, and concluded that the magnetic field in the thin shells is dominated by much higher-degree harmonics compared to the dipole field. For the case of the impactor layer this ratio is about 0.995, and the height of the tangent cylinder is only 330 km, too small to generate any appreciable dipole field at the surface of Mars.

8. Magnetic Signature of the Lowland

The giant Borealis impact excavates the entire crust at the impact site and creates a very deep transient cavity, which is largely filled by the fall back, the collapse of the basin rim, the isostatic upwelling of the impact-heated mantle, and partly by the superheated basaltic lava at temperatures well above the solidus of basalt which creates a lava lake covering the northern lowland. The basaltic lava would have acquired strong magnetization upon subsequent cooling below the magnetic blocking temperatures of its magnetic minerals if a strong core field existed.

To investigate this possibility, I determine the cooling of the lava lake and estimate the downward propagation of the Curie temperature inside the lake. Volcanism usually stacks thin layers of lava in the process of building a thick volcanic construct. This allows a given lava layer to cool efficiently from the top before being covered by the next lava layer. Here an extreme case scenario is considered, by assuming that the entire crust of ~30 km thickness beneath the lowland [Zuber et al., 2000] is the depth of the lava lake. For the lowland model of 8000 km diameter this amounts to about $15 \times 10^6 \text{ km}^3$ of basaltic lava. It is also assumed that all of the lava is emplaced at once immediately after the impact. This elongates the magnetization time of the lava lake. I take the initial temperature of the lava to be 2270 K, about 750 K higher than its solidus, and the underlying lithosphere is at solidus of 1350 K. It is assumed that the lava cools by heat conduction alone, at a much lower rate than cooling by convection. Note that these extreme measures lead to an over estimate of the cooling time of the lava lake. As the lava cools it crosses the melting temperature and acquires latent heat of melting, set to $4 \times 10^5 \text{ J/kg}$. I determine the thermal evolution of the upper 100 km of Mars consisting of the 30 km thick lava and 70 km thick underlying lithosphere. Except for the initial temperatures of the lava and the lithosphere other physical parameters are assumed to be identical for the two sublayers; for the solid and liquid phases a density of $\rho_b = 3000 \text{ kg/m}^3$, a specific heat of $C_{pb} = 1200 \text{ J/kg K}$, and a temperature-dependent thermal conductivity $K_b$ following Schatz and Simmons [1972] in the experimental model, equations (4a) and (4b). Approximating the lake by a laterally uniform horizontal layer, because the lake is much thinner than the radius of Mars, the one-dimensional heat conduction equation

$$\rho_b C_{pb} \partial T / \partial t = \partial (K_b \partial T / \partial z) / \partial z,$$  \hspace{1cm} (9)

is solved subject to the temperature boundary conditions of 230 K at the surface ($z = 0 \text{ km}$) and 1523 K at the base ($z = 100 \text{ km}$). No radioactive heating is taken into account because of the short period of thermal evolution considered. Figure 13 shows the temperature profiles at 2 Myr intervals. Also included in Figure 13 is the range of the magnetic blocking temperature of magnetite, which is likely the main magnetic carrier in the Martian crust [e.g., Dunlop and Arkani-Hamed, 2005], spanning from the Curie temperature of 840 K down to 740 K. The lava lake cools from top and bottom, but largely from the top. The sharp temperature difference initially existing between the lake and the

Figure 11. The intensity of the mean magnetic field generated in the convecting outer core of models A–D (Table 2). The dashed curve is for the case where the impactor iron layer is placed on the core.
underlying region diminishes within about 10 Myr. The entire lake cools through the magnetic blocking temperature range within 45–80 Myr. Figure 14 shows the depth to the Curie temperature, indicating that the rapid cooling of the lake from the top can result in precipitation of singledomain or pseudosingle-domain magnetic particles, which can acquire significant thermoremanent magnetization in the presence of a core dynamo. Note that possible minor volcanism occurred at later times would create thin layers of lava which cool rapidly to the atmosphere without thermally demagnetizing the underlying magnetized magma ocean significantly.

A spherical shell with a constant thickness and uniformly distributed magnetic particles which is magnetized by an internal magnetic field produces no magnetic field outside [e.g., Runcorn, 1975]. However, being punctured by subsequent impacts it can create appreciable edge effects strong enough to be observable at satellite altitudes of 100–400 km. There have been many large impact basins such as Utopia (diameter 3380 km) and Isidis (diameter 1350 km) created at around 4 Ga [Frey, 2008] that most likely punctured the entire thickness of the solidified lava lake about 400 Myr after Borealis impact. However, no edge effects of these punctured regions have been detected by Mars Global Surveyor [e.g., Acuña et al., 1999; Lillis et al., 2008], emphasizing that the magma ocean is not magnetized and probably no strong core field existed when the lake was cooling below its magnetic blocking temperature range. It is worth noting that Hellas basin which has punctured the magnetized crust in the southern hemisphere shows distinct edge effects at satellite altitudes [e.g., Mohit and Arkani-Hamed, 2004].

Borealis impact constrains the period when the strong core dynamo of Mars magnetized the crust. Assuming that the impact occurred at 4.5 Ga, the core dynamo of Mars must have resumed no earlier than around 4.4 Ga. On the other hand, the lack of magnetic signatures associated with giant impact basins such as Utopia and Hellas that were created at around 4.1 Ga implies that the core dynamo must have ceased no later than 4.1 Ga [Lillis et al., 2008]. Therefore, the core dynamo was likely active for about 100 Myr prior to Borealis impact and for about 300 Myr once the crippled core became active again. The strong magnetic anomalies of Mars require the source bodies to carry thermoremanent magnetization [Arkani-Hamed, 2003; Dunlop and Arkani-Hamed, 2005], which is acquired as they cooled through their magnetic blocking temperatures in the presence of a strong core field. Whether the source bodies are intrusive [e.g., Hood and Richmond, 2002] or are formed by stacking successive volcanic layers [Arkani-Hamed, 2005b], it may take ~100 Myr for them to acquire magnetization in the presence of a strong core field. This indicates that each active period of the core is long enough to magnetize source bodies of the large and extensive magnetic anomalies in the southern hemisphere with horizontal dimensions of over 1000 km and estimated thicknesses of over 30 km [e.g., Connerney et al., 1999; Nimmo and Gilmore, 2001; Arkani-Hamed, 2002; Voorhies et al., 2002; Langlais et al., 2004]. The magnetic anomalies of the southern hemisphere could be due to magnetization acquired partly prior to Borealis impact and partly after 100 Myr following the impact.

The large impacts that have created Hellas, Isidis, and Argyre basins have completely demagnetized the preexisting crust underlying about 80% of the basins [Mohit and Arkani-
It is expected that a much larger Borealis impact excavated a major part of the preexisting crust and demagnetized the remaining part, which explains the lack of strong magnetic signature over the entire lowland. The very weak localized magnetic anomalies observed over the lowland may arise from the bottom topography of the lava lake [Zuber et al., 2000, Figure 2], which was created by the collapse of the transient cavity and subsequent isostatic uplift.

**Figure 13.** The temperature profiles inside the magma ocean and the underlying lithosphere. The numbers on the curves are time after emplacement of the magma in Myr. The horizontal dash-dotted line shows the lower boundary of the ocean. The two vertical dashed lines specify the magnetic blocking temperature range of magnetite.

**Figure 14.** The depth to Curie temperature of magnetite (840 K) in the first 80 Myr after emplacement of the magma ocean. The horizontal dashed line is the bottom of the magma ocean.
of the mantle. Deeper parts of the lake must have cooled at later times and very slowly (see Figure 13), allowing magnetic particles to grow and become multidomain with weaker magnetic properties. It is possible that deeper parts of the lake cooled below the magnetic blocking temperatures and acquired weak magnetization at a later time and in the presence of the regenerated core dynamo.

9. Conclusions

[38] I investigated the heating of the Martian interior by the putative Borealis impact using Pierazzo et al.’s [1997] shock pressure distribution model and the foundering shock heating model of Watters et al. [2009]. The northern lowland is modeled by a circular basin, and the effects of the impact velocity and the size of the impactor were estimated by using impact velocities of 6 and 10 km/s and model basin diameters of 7000 and 8000 km. The investigation spans four different time scales ranging from a few minutes, when the shock waves propagate and heat the mantle and core, to a few thousand years that it takes for the differentially heated core to stratify, to a few tens of thousand years for the diminishing of a possible preexisting core dynamo, and, finally, to several tens of millions of years for the stratified core to cool and reestablish a global convective and possibly regenerate a strong core dynamo. The model calculations lead to the conclusions that (1) impact heating in the upper mantle can dominate postimpact mantle dynamics; (2) impact heating of the deeper parts of the mantle may not suppress a preexisting core dynamo; (3) direct impact heating causes the core to become stratified, resulting in an upward increasing temperature distribution in the upper ∼300 km, which suppresses the thermal convection in the core and diminishes a possible preimpact thermally driven core dynamo in a few tens of thousand years; (4) during subsequent cooling of the stratified core a downward-propagating convection develops in the outer core which efficiently transfers heat to the mantle; (5) the convection zone thickens gradually while the impact heat is being exhausted to the mantle, but very rapidly once major part of the heat is exhausted; (6) depending on the size and velocity of the impactor, it takes on the order of 35–85 Myr for the core to exhaust the impact heat and reestablish a global thermal convection and likely regenerate a strong core dynamo; (7) a plausible superheated iron layer produced on the core by the sinking iron content of the impacting body is found to be too thin to generate a significant dynamo and create appreciable magnetic field at the surface of Mars (the layer, however, delays the cooling of the core and regeneration of a strong dynamo for an additional ∼40 Myr); and (8) Borealis impact causes considerable melting in the mantle and results in pervasive volcanism, filling the basin and creating a thick lava lake. The entire lake, ∼30 km thickness, cools below the Curie temperature within about 40–70 Myr, during which the core dynamo is crippled. It does not acquire thermoremanent magnetization in the absence of a core dynamo. However, because of appreciable topographic relief at the bottom of the lake, deeper parts of the lake may cool slowly and be magnetized by the regenerated core dynamo and create the localized weak magnetic anomalies observed over the lowland.

[39] The above conclusions must be regarded indicators, rather than rigorous. This is even more so for the models with impact velocity of 6 km/s, because the Pierazzo et al. [1997] pressure distribution model may be less accurate at impact velocities less than 10 km/s. The physical parameters entering the model calculations are poorly constrained, especially during the first 100–200 Myr of the history of Mars considered in this study. Whether the huge Borealis impact actually occurred and whether it occurred at 4.5 Ga is not clear. The initial temperature of the Martian interior immediately after the accretion is not well constrained. The mantle viscosity which controls mantle dynamics both before and after the impact is not known accurately. It may differ from the model used in this study by a few orders of magnitude. The 70% partitioning of the radioactive elements to the mantle has appreciable effect on the radioactive heating of the mantle, which in turn controls mantle rheology. The acoustic velocities adopted are based on the interpolation of the seismic velocities of Earth. The core viscosity of 1 m²/s used in the calculations seems much higher than expected which required considerable amount of computer time if adopted, a common shortcoming of all core dynamo models of terrestrial planets. However, the conclusion that Borealis impact, if it occurred, could have crippled the core dynamo of Mars for about 100 Myr seems plausible. The absence of a sizable core dynamo when the basaltic crust of the northern lowland was cooling is in good agreement with the lack of magnetic signature of the lowland as established by the Mars Global Surveyor data.

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