Earth and Planetary Science Letters 478 (2017) 192–202

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

www.elsevier.com/locate/epsl

Effects of basin-forming impacts on the thermal evolution and magnetic field of Mars

J.H. Roberts [a,*], J. Arkani-Hamed [b]

[a] Johns Hopkins University Applied Physics Laboratory, 11100 Johns Hopkins Rd., Laurel, MD, 20723, United States
[b] Dept. of Physics University of Toronto, Toronto, ON, M5S 1A7, Canada

ARTICLE INFO

Article history:
Received 7 October 2016
Received in revised form 25 July 2017
Accepted 25 August 2017
Available online xxxx
Editor: C. Sotin

Keywords:
geophysics
impact processes
Mars
thermal history
magnetic field
interiors

ABSTRACT

The youngest of the giant impact basins on Mars are either weakly magnetized or completely demagnetized, indicating that a global magnetic field was not present at the time those basins formed. Eight basins are sufficiently large that the impact heating associated with their formation could have penetrated below the core–mantle boundary (CMB). Here we investigate the thermal evolution of the martian interior and the fate of the global magnetic field using 3D mantle convection models coupled to a parameterized 1D core thermal evolution model. We find that the survival of the impact-induced temperature anomalies in the upper mantle is strongly controlled by the mantle viscosity. Impact heating from subsequent impacts can accumulate in stiffer mantles faster than it can be advected away, resulting in a thermal blanket that insulates an entire hemisphere. The impact heating in the core will halt dynamo activity, at least temporarily. If the mantle is initially cold, and the core initially superheated, dynamo activity may resume as quickly as a few Myr after each impact. However unless the lower mantle has either a low viscosity or a high thermal conductivity, this restored dynamo will last for only a few hundred Myr after the end of the sequence of impacts. Thus, we find that the longevity of the magnetic field is more strongly controlled by the lower mantle properties and relatively insensitive to the impact-induced temperature anomalies in the upper mantle.

1. Introduction

The strong magnetic anomalies detected on Mars by the Mars Global Surveyor (Acuña et al., 1999), and the lack of a global core field at present suggest that martian crust carries remanent magnetization acquired during the active period of a core dynamo in the past. The remanent magnetization of the oldest martian meteorite (ALH84001) indicates that a strong magnetic field existed at the martian surface at ∼4.1 Ga (e.g., Collinson, 1986, 1997; Kirschvink et al., 1997; Weiss et al., 2002; Antretter et al., 2003), but provides no information as to the longevity of the core dynamo.

The characteristics of the magnetic field over the 20 giant impact basins created between 4.2 to 3.8 Ga imply that the core field decayed rapidly within less than 100 Myr at around 4.1 Ga (Lillis et al., 2008), in good agreement with the absence of magnetic field inside Hellas, Argyre and Isidis basins (e.g., Acuña et al., 1999; Mohit and Arkani-Hamed, 2001; Hood et al., 2003). The weak magnetic anomalies over the Tharsis bulge emphasize the possibility that the core dynamo was at the final stage of decaying when the major part of the bulge was forming (e.g., Johnson and Phillips, 2005), or that the anomalies are remnants of originally stronger anomalies due to deep seated source bodies which have been partially demagnetized by the overlying volcanic layers of Tharsis. The absence of expected edge effects due to the rupture of the Tharsis bulge that has created Valles Marineris implies that the upper ∼10 km of Tharsis is not appreciably magnetized (Arkani-Hamed, 2004). Moreover, the lack of magnetic signatures of large volcanic structures such as Olympus, Ascraeus, Arsia, and Pavonis Montes, and the Elysium Rise indicates that no core field existed at the time these structures were formed.

Observations of crustal magnetism (Acuña et al., 2001) provides strong evidence that a global magnetic field existed early on, but vanished in the mid- to late Noachian. The ages of the basins correlate strongly with their magnetization strengths (Lillis et al., 2008), resulting in speculation that there may have been a causal relationship between the impacts, which created the basins and the disappearance of the magnetic field (Roberts et al., 2009).

Previous modeling of the impact heating of the core shows that the 7 largest impacts could cripple the core dynamo, but that the dynamo is resilient and reinitiates shortly after each impact

* Corresponding author.
E-mail address: James.Roberts@jhuapl.edu (J.H. Roberts).

http://dx.doi.org/10.1016/j.epsl.2017.08.031

(Arkani-Hamed, 2012). However, the resilience of the core dynamo may be partly an artifact. The thermal state of the mantle on the core dynamo was not taken into account in the modeling, and a fixed temperature boundary condition was imposed at the top of the lower thermal boundary layer of the mantle, which may artificially promote core cooling and dynamo activity. On Mars, the stagnant lid reduces heat flux out of the mantle, which is expected to be appreciably heated by radioactive elements in the early (first ~500 Myr) history, resulting in significant reduction of heat flux out of the core and hampering core cooling (Breuer and Spohn, 2003; Arkani-Hamed, 2005), and hence re-initiation of the core dynamo.

In the present study we investigate the thermal evolution of Mars in the first 1 Gyr on the basis of 3D mantle convection modeling while taking into account the effects of the 8 largest impacts (chronologically, Daedalia, Ares, Amazonis, Chryse, Scopolus, Acidalia, Utopia, and Hellas) as well as the temperature- and pressure-dependent mantle viscosity. By directly coupling the mantle convection with the core dynamics we investigate the effects of the above factors on the heat loss from the core, and show that the mantle convection was sluggish during the impact times. Successively battering the core dynamo by the 8 largest basin-forming impacts could have diminished the dynamo strength.

2. Modeling

An impact generates a strong shock wave that propagates through the interior and heats the mantle and the core, with strongest heating directly beneath the impact site, and rapidly decaying with distance away from that point. A differentially heated, rotating, low viscosity core stratifies, creating a radially increasing temperature distribution in the core relative to an adiabat, and cripples thermally driven core convection. It takes of order 1 kyr after the impact event for thermal convection in the martian core to diminish, and another about 10 kyr for the existing core dynamo-driven magnetic field to decay (Arkani-Hamed and Olson, 2010a). As time passes, the core cools into the overlying mantle and convection resumes in a thin uppermost layer of the core, which increases in thickness in time and becomes capable of regenerating a new core dynamo (Arkani-Hamed and Olson, 2010b; Roberts and Arkani-Hamed, 2014).

2.1. Shock heating by basin-forming impacts

Here we present the method by which we model the heating of the martian interior by the sequence of impacts that created the eight largest basins on Mars during the heavy bombardment in the Solar System. An impactor diameter, D_{imp}, is determined from the diameter of the corresponding basin, D_b, using the crater scaling laws (Schmidt and Housen, 1987; Melosh, 1989; Holsapple, 1993),

$$D_{imp} = 0.69 D_{tr}^{1.28} V_{imp}^{-0.56} g^{0.28} \tag{1}$$

$$D_{tr} = 0.76\, D_b^{0.921} D_*^{0.079} \tag{2}$$

where D_* is the transition diameter from simple to complex crater structure (=7 km for Mars), D_{tr} is the transient crater diameter, V_{imp} is the impact velocity, and g is the gravitational acceleration at the surface of Mars. We adopt an impact velocity of 10 km/s, which is the average of the impact velocities on Mars (Neukum and Wise, 1976). We show in Table 1 the space–time coordinates and diameters of the eight largest impact basins reported by Frey (2008), along with estimates of the projectile diameters.

A giant basin-forming impact not only excavates and heats a large portion of the upper mantle of Mars, but also generates a strong shock wave that propagates through the interior and differentially heats the mantle and the core, with the most intense heating directly beneath the impact site. An impact creates a nearly

Table 1
Impact basin parameters.

Basin name	Lat (°N)	Long (°E)	D_b (km)	Age (Gyr)	D_{imp} (km)
Daedalia	−26.5	228.3	2639	4.199	434
Ares	4.0	343.9	3300	4.160	565
Amazonis	27.1	187.9	2873	4.154	480
Chryse	25.0	318.0	1725	4.140	222
Scopolus	6.9	81.8	2250	4.133	355
Acidalia	59.8	342.7	3087	4.132	510
Utopia	45.0	115.5	3380	4.111	580
Hellas	−42.3	66.4	2070	4.065	330

uniform shock pressure P_{iso} inside the so-called isobaric core of radius r_{iso} where the pressure is estimated (Melosh, 1989),

$$P_{iso} = \rho(C + S u_p) u_p; \quad r < r_{iso} \tag{3}$$

where ρ and C are the pre-shocked density and acoustic velocity, u_p is the particle velocity in the isobaric sphere ($u_p = \frac{1}{2} V_{imp}$, assuming similar target and impacting materials), S is a constant, r is the distance from the impact site at the surface, and $r_{iso} \sim 0.5 D_{imp}$. We list in Table 2 the physical parameter values used in this study.

Several different models have been proposed for the shock pressure distribution outside the isobaric region. In this study, we adopt the average model of Pierazzo et al. (1997) for the shock pressure in the mantle,

$$P_s = P_{iso}(r/r_{iso})^n; \quad r < r_{iso}; \ n = 1.84 - 2.61 \log(V_{imp}) \tag{4}$$

where V_{imp} is in km/s. A shock wave propagates in the mantle as a spherical wave. While crossing the core mantle boundary the pressure jumps suddenly (Arkani-Hamed and Ivanov, 2014). The shock wave then travels in the core while the pressure decays following a similar power law with the decay factor of 1.175 in the liquid iron core (Rae, 1970; Dienes and Walsh, 1970). The shock wave emerging out of the core in the antipodal hemisphere has already decayed, and the shock pressure has reduced below the Hugoniot elastic pressure. At this stage, the wave is probably reduced to an elastic wave and does not heat the mantle.

In Fig. 1a we show the impact-induced temperature increase ΔT in Mars caused by the Daedalia impactor, calculated using the foundering model of Watters et al. (2009), and before taking melting into account. This temperature increase is calculated as:

$$\Delta T = P_s \frac{(1 - 1/f)}{(2\rho S)} - \left(\frac{C}{S}\right)^2 (f - \ln f - 1) \tag{5}$$

where

$$f = \frac{-P_s}{[\beta(1 - \sqrt{((2P/\beta) + 1)})]} \tag{6}$$

and

$$\beta = \frac{(C^2 \rho)}{(2S)} \tag{7}$$

Fig. 1a shows that the impact induced shock wave heats Mars differentially. While the sub-impact region is appreciably heated, major part of the core receives no impact heating. About 20% of the core–mantle boundary receives direct shock wave and major shock heating actually occurs only on ~10% of the boundary.

Fig. 1b shows the impact-induced temperature increase (before taking melting into account) along the axis of symmetry passing through the impact site and the center of Mars for the 8 impactors. The impact velocity for all impacts is 10 km/s; consequently the pressures and the temperature increases created by the impacts inside their isobaric regions are identical, but the sizes