Icarus 218 (2012) 278–289

Contents lists available at SciVerse ScienceDirect

Icarus

journal homepage: www.elsevier.com/locate/icarus

Impact-induced mantle dynamics on Mars

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

[a] The Johns Hopkins University Applied Physics Laboratory, 11100 Johns Hopkins Road, Laurel, MD 20723-6099, United States
[b] Department of Physics, University of Toronto, 60 St. George St., Toronto, ON, Canada M5S 1A7

ARTICLE INFO

Article history:
Received 11 May 2011
Revised 18 October 2011
Accepted 30 November 2011
Available online 11 December 2011

Keywords:
Geophysics
Impact processes
Mars, interior
Thermal histories

ABSTRACT

At least 20 impact basins with diameters ranging from 1000 to 3380 km have been identified on Mars, with five exceeding 2500 km. The coincidental timing of the end of the sequence of impacts and the disappearance of the global magnetic field has led to investigations of impact heating crippling an early core dynamo. The rate of core cooling (and thus dynamo activity) is limited by that of the overlying mantle. Thus, the pre-existing thermal state of the mantle controls the extent to which a sequence of impacts may affect dynamo activity. Here, we examine the effects of the initial thermal structure of the core and mantle, and the location of an impact with respect to the pre-existing convective structure on the mantle dynamics and surface heat flux.

We find that the impacts that formed the five largest basins dominate the impact-driven effects on mantle dynamics. A single impact of this size can alter the entire flow field of the mantle. Such an impact promotes the formation of an upwelling beneath the impact site, resulting in long-lived single-plume convection. The interval between the largest impacts is shorter than the initial recovery time for a single impact. Hence, the change in convective pattern due to each impact sets up a long term change in the global heat flow. These long-term changes are cumulative, and multiple impacts have a synergistic effect.

1. Introduction

1.1. Observations

At least 20 exposed and buried "giant" impact basins with observed diameters, D_b exceeding 1000 km have been identified on Mars (Frey, 2008). Of these, five basins have $D_b > 2500$ km. The crater retention ages for these basins have been obtained by counting smaller craters, quasi-circular depressions (QCDs) in topography, and circular thin areas (CTAs) in crustal thickness models, with diameters greater than 300 km on the rims and interiors of the giant basins (Frey, 2008). The absolute model ages of these basins (Hartmann and Neukum, 2001) are largely clustered between 4.2 and 4.1 Ga. This spike in basin ages may be related to the Late Heavy Bombardment (Cohen et al., 2000; Gomes et al., 2005).

Observations of crustal magnetism (Acuña et al., 2001) provides strong evidence that such 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 (Arkani-Hamed et al., 2008; Roberts et al., 2009).

1.2. Previous work

The coincidental timing of the sequence of impacts and the disappearance of the magnetic field, suggests that former may have caused the latter (Arkani-Hamed et al., 2008; Roberts et al., 2009; Arkani-Hamed and Olson, 2010a,b). It is well known that large impacts can introduce substantial amounts of heating to planetary interiors (Reese et al., 2002; Monteux et al., 2007; Watters et al., 2009). Roberts et al. (2009) modeled the effects of impact heating on the mantle dynamics, in particular on the evolution of CMB heat flow. They found that the largest impacts heated the lower mantle resulting in a reduction in CMB heat flow that was unfavorable for a dynamo. Arkani-Hamed and Olson (2010a,b) on the other hand, suggested that direct impact heating of the core, with a lower specific heat than that of the mantle, may result in a substantial increase in core temperature and in the CMB heat flow. Moreover, the heating of the outer core results in stable stratification, which also serves to shut down dynamo activity. In both cases, a warm region, or "thermal blanket" prevents cooling of much of the core. In the first case, the thermal blanket is the lower mantle; in the second, it is the outermost core.

These previous studies have focused on the effects of basin-forming impacts in the deep interior and at the CMB. Here, we examine the effects of the initial thermal structure of the core and mantle, and the location of an impact with respect to the pre-existing convective structure on the mantle dynamics and surface heat flux.

---

* Corresponding author. Fax: +1 240 228 8939.
E-mail address: James.Roberts@jhuapl.edu (J.H. Roberts).

0019-1035/$ - see front matter © 2011 Elsevier Inc. All rights reserved.
doi:10.1016/j.icarus.2011.11.038

Table 1
Parameters for Noachian Mars.

Parameter	Symbol	Value	Reference
Planetary radius	R_0	3390 km	Bills and Ferrari (1978)
Core radius	R_c	1700 km	Yoder et al. (2003)
Mantle density	ρ_0	3500 kg m^{-3}	Sohl and Spohn (1997)
Gravitational acceleration	g	3.7 m s^{-2}	Esposito et al. (1992)
Coefficient of thermal expansion (at surface)	α_0	3×10^{-5} K^{-1}	Roberts and Zhong (2004)
Coefficient of thermal diffusion (at surface)	κ_0	1.29×10^{-6} m^2 s^{-1}	Schatz and Simmons (1972)
Specific heat	C_p	1200 J kg^{-1} K^{-1}	Roberts and Zhong (2004)
Surface temperature	T_0	220 K	Roberts and Zhong (2004)
Initial CMB temperature	T_{CMB}	2000 K	Nimmo and Stevenson (2000) and Hauck and Phillips (2001)

Table 2
Radioactive heating parameters for bulk silicate Mars.

Isotope	Present-day concentration	Half-life (Gyr)	Heating rate (W kg^{-1})
U	1.7×10^{-8}		
U-235		0.704	5.69×10^{-4}
U-238		4.47	9.46×10^{-5}
K	3.4×10^{-5}		
K-40		1.25	2.92×10^{-5}
Th	6.8×10^{-8}	14.0	2.64×10^{-5}

1.3. Motivating questions

The thermal evolution of the interior of Mars depends on the ability of the mantle to remove heat. Core cooling is limited by the vigor of convection in the much more viscous mantle above. Here, we suggest that this efficiency may depend on the initial pre-impact conditions of Mars. In particular, we examine the effects of the pre-existing thermal state of the mantle and frequency of large impacts on the impact-induced mantle dynamics.

In the following section we describe our modeling efforts, including shock heating of the mantle due to impacts and the subsequent thermal evolution of the mantle by thermal convection. Next we present the results of the convection models with particular attention to the evolution of the thermal structure. Finally, we discuss the implications of these results on the long-term thermal evolution of Mars, and attempt to quantify the effects of uncertainties in observations of basin size and age, scaling laws, and models of shock pressure and heating.

2. Modeling

2.1. Mantle convection

We model mantle dynamics using CitcomS (Zhong et al., 2000; Tan et al., 2006), a 3D finite-element model of convection in a spherical shell. Our model contains 1.8 million elements, each with an average thickness of 26 km, and spanning 4.2° in each horizontal direction. The mantle is assumed to be viscous and incompressible, with an infinite Prandtl number, and subject to the extended Boussinesq approximation (Christensen and Yuen, 1985). Thermal convection is governed by the equations of conservation of mass, momentum and energy:

$$\begin{aligned}
&\nabla \cdot \vec{u} = 0 \\
&-\nabla P + \nabla \cdot [\eta(\nabla \vec{u} + \nabla^T \vec{u})] + \delta\rho g \vec{e}_r = 0 \qquad (1) \\
&\frac{\partial T}{\partial t} + \vec{u} \cdot \nabla T = \nabla \cdot (\kappa \nabla T) + H
\end{aligned}$$

where $\vec{u}$ is velocity, P is pressure, η is dynamic viscosity, g is the gravitational acceleration, $\vec{e}_r$ is a unit vector in the radial direction, T is temperature, t is time, κ is the thermal diffusivity, and H is the internal heating, which includes adiabatic heating, frictional dissipation, and radioactive heat generation (Christensen and Yuen, 1985). The density varies due to temperature as $\delta\rho = \rho_0\alpha(T - T_0)$, where ρ_0 is the reference density, α is the thermal expansivity, and T_0 is the reference temperature.

We nondimensionalize Eqs. (1)–(3) using the following conversions:

$$x'_i = \frac{x_i}{R_0}, \quad u'_i = \frac{u_i}{\kappa_0/R'_0}, \quad t'_i = \frac{t}{R'_0/\kappa_0}, \quad \eta' = \frac{\eta}{\eta_0}, \quad P' = P\frac{R_0^2}{\eta_0\kappa_0},$$
$$T' = \frac{T - T_0}{\Delta T}, \quad T_S = \frac{T_0}{\Delta T'}, \quad H' = H\frac{R_0^2}{\kappa_0\rho_0 C_p \Delta T'}, \quad \kappa' = \frac{\kappa}{\kappa_0} \qquad (2)$$

where x_i represents a spatial coordinate, R_0 is the planetary radius, ΔT is the initial temperature difference across the mantle and C_P is the specific heat. All primed quantities are nondimensional. All '0' subscripts refer to the reference values of their respective quantities, typically chosen at the surface (though η_0 is the viscosity at the CMB). See Table 1 for the reference parameter values used in this study. Using these nondimensionalizations and dropping the primes for clarity, the momentum equation becomes:

$$-\nabla P + \nabla \cdot [\eta(\nabla \vec{u} + \nabla^T \vec{u})] + RaT\vec{e}_r = 0 \qquad (3)$$

where Ra, the Rayleigh number is defined as:

$$Ra = \frac{\rho_0 g \alpha_0 \Delta T R_0^3}{\kappa_0 \eta_0} \qquad (4)$$

The nondimensional mass and energy equations are formally identical to the dimensional versions. α varies linearly with depth, decreasing by a factor of two from the surface to the CMB. The conductivity, $\kappa = \rho C_p \kappa$, varies with temperature according to an empirical relationship (Schatz and Simmons, 1972).

$$k = \begin{cases} \frac{414.8 \text{ W m}^{-2}}{30.6 \text{ K} + 0.21\,T} & T < 500 \text{ K} \\ \frac{414.8 \text{ W m}^{-2}}{30.6 \text{ K} + 0.21\,T} + 0.0023(T - 500 \text{ K}) & T > 500 \text{ K} \end{cases} \qquad (5)$$

The mantle is cooled from above (by radiation into space), and heated from below (by secular cooling of the core) and from within by radioactive decay. The surface temperature is held constant. The CMB temperature is laterally homogeneous but varies with time as the core is cooled by the mantle. The time-dependent heating rate (Turcotte and Schubert, 2002) is given by:

$$\begin{aligned}
H = {} & 0.9928 C_0^U H^{U238} \exp\frac{t \ln 2}{\tau_{1/2}^{U238}} + 0.0071 C_0^U H^{U235} \exp\frac{t \ln 2}{\tau_{1/2}^{U235}} \\
& + C_0^{Th} H^{Th} \exp\frac{t \ln 2}{\tau_{1/2}^{Th}} + 1.19 \times 10^{-4} C_0^K H^{K40} \exp\frac{t \ln 2}{\tau_{1/2}^{K40}} \qquad (6)
\end{aligned}$$

where C_0^X is the present-day concentration of element X, H^{X1} is the heating rate of isotope X1, and $\tau_{1/2}^{X1}$ is the half-life of that isotope. The bulk concentrations (Wanke and Dreibus, 1994), half-lives, and heating rates (Turcotte and Schubert, 2002) for the long-lived