Icarus 226 (2013) 20–32

Contents lists available at SciVerse ScienceDirect

Icarus

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

Dynamics of core merging after a mega-impact with applications to Mars' early dynamo

J. Monteux [a,b,*], A.M. Jellinek [b], C.L. Johnson [b,c]

[a] Laboratoire de Planétologie et de Géodynamique de Nantes, UMR CNRS, Université de Nantes, France
[b] Department of Earth, Ocean and Atmospheric Sciences, University of British Columbia, Canada
[c] Planetary Science Institute, Tucson, AZ, United States

ARTICLE INFO

Article history:
Received 3 October 2012
Revised 23 April 2013
Accepted 9 May 2013
Available online 18 May 2013

Keywords:
Mars, Interior
Magnetic fields
Impact processes
Planetary dynamics

ABSTRACT

A giant impact occurring within the first 500 Myr of martian history may have been responsible for the dichotomy between the northern lowlands and the southern highlands and may have influenced the initiation or cessation of early and short-lived core dynamo. We hypothesize that a significant volume of metallic iron from a differentiated impactor merged with a preexisting martian core. We investigate the dynamics and thermal effects of this core merging, assuming that the impactor's core sank as a single metallic diapir through a solid mantle. We explore the consequences of this process for dynamo action and for Mars' magnetic field history. For large impacts (with radii larger than 100 km) and plausible mantle viscosities, merging is expected to occur in less than 1 Myr. Depending on the temperature-dependence of the mantle viscosity, viscous dissipation within the diapir may be very large. Where thermal mixing of the hot diapir into a preexisting core is complete, merging can increase the temperature gradient to the surrounding mantle and consequently drive a dynamo until this additional heat is transferred to the mantle, which takes on the order of 100 Myr. If merging leads to strong thermal stratification in the core, however, dynamo action may be inhibited.

© 2013 Elsevier Inc. All rights reserved.

1. Introduction

Mega impacts (with an impactor/target radius ratio between one tenth and 1) probably played an important role in the late history of terrestrial planetary accretion. An impact between the proto-Earth and a Mars-size protoplanet can, for example, explain the Earth–Moon system (Hartmann and Davis, 1975) and a large impact removing part of the silicate mantle is a hypothesis for the high iron/silicate ratio on Mercury (Smith, 1979; Benz et al., 1988). Among the hypotheses for the origin of the martian dichotomy including an endogenic origin (Elkins-Tanton et al., 2003; Roberts and Zhong, 2006; Citron and Zhong, 2012) or a plate-tectonics feature (Sleep, 1994), an exogenic origin by a mega impact that displaced crustal material from the northern to the southern hemisphere seems to be the most plausible candidate (Wilhelms and Squyres, 1984; Nimmo et al., 2008; Andrews-Hanna et al., 2008; Marinova et al., 2008, 2011). Models for the martian impact suggest that the impactor was 800–1300 km in radius and hit the planet with a speed comparable to or larger than the martian escape velocity (i.e. $v_{imp} > 5\ \mathrm{km\ s^{-1}}$) within the first 500 Myr of martian history (Frey, 2006). Such a dramatic event can generate a debris disk around Mars that could have re-accreted and formed the martian moons (Rosenblatt, 2011).

In addition to a history of large impacts, Earth, Mercury and Mars have, or have had, an internally generated magnetic field. Mercury and the Earth have active core dynamos, whereas widespread crustal magnetism strongly suggest that Mars had an early internally-generated magnetic field (Acuña et al., 1999; Hood et al., 2003; Lillis et al., 2008a) that ceased by around 4.0 Ga (Acuña et al., 1999; Johnson and Phillips, 2005; Lillis et al., 2008b). The timing of the initiation of the martian dynamo is difficult to constrain and strongly depends on the differentiation processes that occurred during the first million years of martian history (Monteux et al., 2011). The cause of the cessation of the martian dynamo is also still currently debated. Recent models explore the effects of large impacts on the dynamo generation process, and in particular on the cessation of dynamo action as a result of a reduction of the core–mantle boundary (CMB) heat-flux (Roberts et al., 2009; Watters et al., 2009; Roberts and Arkani-Hamed, 2012). Shock heating within the core can also increase the CMB heat flow and create a thermal stratification that prevents heat loss from the inner part of the core, inhibiting core convection (Arkani-Hamed and Olson, 2010). Alternatively, other models show that the thermal anomaly induced by a large impact and the formation of a hot molten iron layer from the impactor's core at the CMB can favor dynamo generation (Reese and Solomatov, 2010).

* Corresponding author at: Laboratoire de Planétologie et de Géodynamique de Nantes, UMR CNRS, Université de Nantes, France.
E-mail address: julien.monteux@univ-nantes.fr (J. Monteux).

0019-1035/$ - see front matter © 2013 Elsevier Inc. All rights reserved.
http://dx.doi.org/10.1016/j.icarus.2013.05.008

At the time of the proposed giant impact, both Mars and the impactor were probably differentiated (Yoshino et al., 2003). Models suggest that although some material was ejected far from Mars, the majority of the mass of the impactor's core was retained within the planet and merged with the pre-impact martian core (Canup, 2004; Ćuk and Stewart, 2012). The aim of this study is to characterize the dynamics of core merging as a result of a diapiric descent of molten iron (Monteux et al., 2009) on a Mars-size planet. In addition, we investigate how the processes of impact and core merging might influence the thermal regime of Mars' core and, in turn, magnetic field generation (cf., Monteux et al., 2011).

2. Thermo-chemical state before the martian mega-impact

2.1. Pre-impact interior of Mars

The initial structure and thermal state of a growing planet is determined by the characteristics of its accretion from chondritic material (Safronov, 1978; Kaula, 1979; Agee, 1997). During accretion, heating driven by a combination of the dissipation of impact energy and the decay of short lived radionuclides such as ^{26}Al and/or ^{60}Fe (Yoshino et al., 2003; Monteux et al., 2007) increases the mean internal temperature and gives rise to a radial temperature gradient that depends on the accretion rate relative to the rate of radiative cooling to space (Kaula, 1979; Senshu et al., 2002). If the growth rate is very high in comparison to surface cooling, this heating can ultimately cause partial or complete melting of the chondritic material (Yoshino et al., 2003) and lead to extensive metal/silicate separation (Tonks and Melosh, 1992; Senshu et al., 2002; Monteux et al., 2009).

Hf/W chronology suggests that core formation happened during the first 10–30 myr of Mars' history (Lee and Halliday, 1997; Nimmo and Kleine, 2007). Such a rapid process involves extensive melting potentially enhanced by radiogenic heating as a result of the decay of short-lived radionuclides (Yoshino et al., 2003), impact heating (Tonks and Melosh, 1992; Senshu et al., 2002; Monteux et al., 2009) and gravitational energy conversion during metal/silicate separation (Stevenson, 1989; Ricard et al., 2009). Metal/silicate separation can occur via a wide range of phenomena such as percolation (Shannon and Agee, 1996), the sedimentation of metallic rain through a magma ocean (Rubie et al., 2003; Höink et al., 2005) or a large diapir sinking through a solid mantle after an impact (Tonks and Melosh, 1992; Monteux et al., 2009). Whatever the mechanism, Mars' internal structure characterized by a ~1700 km diameter Fe-core was mostly established within ~10 Myr of the planet's formation (Yoder et al., 2003) (cf., Fig. 1a).

The gravitational heat released during martian core formation was partitioned between the planet's core and mantle. The fraction of gravitational heat taken up by the metal or the silicate fraction depends strongly on the rheology of the planet and on the segregation mechanisms (Samuel and Tackley, 2008; Monteux et al., 2009; Ke and Solomatov, 2009). The combined processes leading to core formation yield a wide range of possible early thermal states, depending on the nature and timescale of the core formation process and the heat transfer properties of Mars' early mantle. In particular, the core could initially have had a temperature close to the deep mantle temperature if thermal equilibration was efficient. Alternatively, it could have been hotter than the mantle if the gravitational potential energy released during core formation was largely retained within the core itself, a situation which would lead to potentially strong cooling to the mantle (Fig. 1a).

2.2. Interior structure of the impactor

In this study, we consider impactors with a radius in the range 200–800 km. The lower bound for our range is motivated by the impactor size needed to create large impact basins such as Hellas or Utopia, and the upper bound is motivated by the minimum impactor radius needed in exogenic models for the dichotomy boundary (Marinova et al., 2008). Assuming that both the impactor and the target body had chondritic compositions, their volumetric metal fractions, f_0, should be similar (we consider that the impactor has the same metal content as Mars and we use f_0 = 12.5% (Stevenson, 2001)). Hence, for 200–1300 km diameter impactors, an additional volume of core material with a radius between 100 km and 700 km merges with the preexisting core (Fig. 1).

3. Thermo-chemical state after a mega-impact

3.1. Mantle heating and melting

Kinetic energy of the impactor is dissipated as a result of the irreversible work done by shock waves in damaging crustal rocks (Senshu et al., 2002; Monteux et al., 2011) as well as heating and melting the target material. This dissipation process is a complex mechanism that is still poorly constrained specially for giant impact events. In our models, we consider that post-impact heating and melting mostly occurs within a spherical region with a volume V_{ic} (and a radius R_{ic}) that is typically taken to be 3 times larger than the volume V_{imp} of the impactor itself (O'Keefe and Ahrens, 1977; Croft, 1982; Pierazzo et al., 1997). The energy available to heat and melt the target planet is $\Delta E = \gamma m_{imp} v_{imp}^2/2$, where γ is the fraction of the kinetic energy of the impactor ultimately dissipated to heat up the mantle (O'Keefe and Ahrens, 1977), m_{imp} is the impactor mass and v_{imp} is the impact velocity. The energy needed to melt a silicate volume V_{ic} is $\Delta E_{m,Si} = \rho_{Si} V_{ic} L_{Si}$, where ρ_{Si} and L_{Si} are the density and latent heat of the silicate material. The energy needed to melt the impactor core is $\Delta E_{m,Fe} = \rho_{Fe} f_0 V_{imp} L_{Fe}$ with L_{Fe} and ρ_{Fe} are the density and latent heat of the impactor's core.

In the heated region the temperature increases uniformly from an initial value T_0 by an amount ΔT_0 (Fig. 1c). The excess temperature ΔT_0 decreases rapidly and smoothly with distance r from the boundary of the isothermal anomaly as approximately $\Delta T_0 (R_{ic}/r)^m$ (Fig. 1c). Following Senshu et al. (2002), and fitting the decay of peak pressure with distance away from the edge of the isobaric core $m \approx 4.4$ (Monteux et al., 2007). The energy needed to increase the temperature inside and outside the isobaric core is $\Delta E_{th} = h_m \rho_{Si} V_{ic} C_{p,Si} \Delta T_0$, where $C_{p,Si}$ is the specific heat of the silicate material and h_m is a geometric parameter representing the amount of heat that is used to increase the temperature inside and outside the isobaric core relative to the amount of heat used to increase the temperature by ΔT_0 within the isobaric core (Senshu et al., 2002; Monteux et al., 2011) and

$$h_m = 1 + \frac{3(2m-5)}{2(m-3)(m-2)} \approx 2.7 \tag{1}$$

Hence, from the following energy balance:

$$\Delta E = \Delta E_{th} + \Delta E_{m,Si} + \Delta E_{m,Fe} \tag{2}$$

and using $m_{imp} = \rho_0 V_{imp} = \rho_0 V_{ic}/3$, we obtain:

$$\Delta T_0 = \frac{1/6 \gamma \rho_0 v_{imp}^2 - L_{Si}\rho_{Si} - f_0/3 L_{Fe}\rho_{Fe}}{h_m \rho_{Si} C_{p,Si}} \tag{3}$$

Assuming that the impact velocity is equal to the escape velocity of the planet we obtain a minimum estimate for the kinetic energy where $v_{imp} = v_{esc} = \sqrt{2gR}$ with $g = 4/3\pi G \rho_0 R$, G the gravitational constant and R the radius of the target planet. After some algebra:

$$\Delta T_0 = \frac{4/9\pi\gamma\rho_0^2 G R^2 - L_{Si}\rho_{Si} - \frac{f_0}{3} L_{Fe}\rho_{Fe}}{h_m \rho_{Si} C_{p,Si}}, \tag{4}$$

Download English Version:

https://daneshyari.com/en/article/10701279

Download Persian Version:

https://daneshyari.com/article/10701279

Daneshyari.com