Thermal evolution of the Martian core: Implications for an early dynamo

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ABSTRACT

Mars is thought to have possessed a dynamo that ceased ~0.5 b.y. after the formation of the planet. A possible, but ad hoc, explanation is an early episode of plate tectonics, which drove core convection by rapid cooling of the mantle. We present an alternative explanation: that the Martian core was initially hotter than the mantle after core formation, providing an initial high heat flux out of the core. A core initially 150 K hotter than the mantle can explain the early dynamo without requiring plate tectonics. Recent experimental results suggest that potassium is likely to partition into the Martian core, potentially providing an extra source of energy to power a dynamo. We find that the radioactive decay of ^{40}K cannot explain the inferred dynamo history without the presence of a hot core. Our results also suggest that core solidification is unlikely to have occurred, because this process would have generated a long-lived (>1 b.y.) dynamo. If, as we conclude, the core is entirely liquid, it must contain at least ~5 wt% sulfur. An initially hot core is consistent with geochemical evidence for rapid core formation and incomplete thermal equilibration with the mantle. Thus, the early history of planetary dynamos provides constraints on the processes of accretion and differentiation.

Keywords: Mars, core, thermal evolution, dynamos, potassium, plate tectonics.

INTRODUCTION

The discovery of strongly magnetized crust by the Mars Global Surveyor (MGS) magnetometer experiment indicates the presence of an earlier epoch in which Mars possessed a dynamo (Acuña et al., 1999; Connerney et al., 1999). The majority of the magnetic anomalies are confined to the ancient, heavily cratered southern highlands. Their absence in and around the large impact basins of Hellas and Argyre imply that the planet did not possess an intrinsic magnetic field at the time these impact events occurred (Acuña et al., 1999), ca. 4 Ga. Although it has been suggested that the dynamo onset postdated the large basins (Schubert et al., 2000), magnetic and textural studies of SNC meteorite ALH84001 (Weiss et al., 2002) suggest a dynamo was operating prior to 4 Ga. It is therefore likely that the Martian dynamo only operated before this time.

Planetary dynamos are thought to be driven by convection in a liquid, iron-rich core (e.g., Gubbins et al., 1979; Stevenson, 2001). A solidifying core will generate compositional convection in the liquid part of the core. In the absence of core solidification, the heat flux out of the core must exceed that of adiabatic cooling (Stevenson, 2001) to drive thermal convection. The heat flux out of the core is controlled by the ability of the overlying mantle to remove heat. Nimmo and Stevenson (2000) suggested plate tectonics (Sleep, 1994) as a means to drive an early, short-lived dynamo on Mars. They demonstrated that an early phase of plate tectonics would rapidly cool the mantle and thus generate a sufficient heat flux out of the core to drive core convection. They also showed that the cessation of plate tectonics would rapidly halt core convection. However this explanation is rather ad hoc, as strong evidence for an episode of plate tectonics is lacking (Pruis and Tanaka, 1995; Zuber, 2001; Halliday et al., 2001) and the timing of the interpreted transition to a stagnant lid regime is arbitrary and chosen to correlate with the apparent ~500 m.y. cessation of the dynamo. Further, Breuer and Spohn (2003) find the present-day crustal thickness and the inference of a monotonically declining crustal production rate to be easier to reconcile with a stagnant lid regime operating throughout the planet's history. They also suggest that a superheated core is required to explain an early dynamo in this case.

The gravitational energy of core formation was estimated by Solomon (1979) to raise the mean internal temperature of Mars by ~300 K. If thermal equilibration between the core and mantle is not complete, the core will be initially hotter than the mantle and the heat flux out of the core will be elevated early in the planet's history. Core convection would be sustained for a brief period as the hot mantle has a short dynamical time constant, providing a natural explanation for a brief, early dynamo without requiring ad hoc assumptions. In this paper we investigate the effect of an initially hot core on the dynamo history. Further, we explore the role of radiogenic heating in the core on the basis of experimental results that demonstrate that significant amounts of potassium (on the order of hundreds of parts per million) can be dissolved in Fe-S liquids at the relatively low pressure, low-temperature conditions relevant to formation of the Martian core (Gessmann and Wood, 2002; Lee and Jeanloz, 2002; Murthy et al., 2003). In this paper we show that the decay of ^{40}K alone cannot explain the cessation of the dynamo after ~500 m.y. and, at most, had only a secondary effect. We find that the dynamo would have been sustained for a few hundred million years if the core was initially 150–200 K hotter than the mantle, thus providing a consistent explanation for the dynamo behavior without invoking the speculative hypothesis of plate tectonics. Further, our results demonstrate that core solidification cannot explain a dynamo of the short duration required for Mars.

MODEL

We have developed a planetary thermal evolution model, based on that of Nimmo and Stevenson (2000), to explore the effect of varying initial core temperatures and potassium core concentrations on the history of the Martian dynamo. Temperature changes in the core result from the heat flux into the mantle and radiogenic heating of ^{40}K. Heat fluxes across the conductive boundary layers of the core-mantle boundary (CMB) and the lithosphere are determined by using parameterized convection; a surface stagnant lid convection (Solomatov, 1995) is assumed throughout the planet's history. The thickness of the boundary layers and, hence, the heat flux across the boundaries depend on the temperature-dependent mantle viscosity. As a result, dynamo activity is intimately related to the thermal history of the mantle.

We calculate the rate of entropy production within the core as a function of time by using the methods of Gubbins et al. (1979) and Nimmo et al. (2003). The rate of entropy production is related to the power available to drive a dynamo. In the absence of core solidification, the heat flux due to thermal conduction is the maximum heat flux that can be extracted without triggering core convection and is given by (Nimmo and Stevenson, 2000):

$$F_{\text{cond}} = -k\left(\frac{dT}{dr}\right)_{\text{adiabat}} = \frac{k\alpha g(r)T_c}{C_p}, \quad (1)$$

where k is the thermal conductivity, dT/dr is the core adiabat, r is the radius, T_c is the temperature of the core at the CMB, C_p is the specific heat capacity, and α is the thermal expansivity, assumed to be constant. The acceleration due to gravity, $g(r)$, is calculated by using the method described in Nimmo et al. (2003) and is ~3.27 m·s^{-2} at the CMB. The thermodynamic values used are listed in Table DR1[1], and the values for the mantle are those of Nimmo and Stevenson (2000). For a constant density core, F_{cond} scales with core radius, as does the volume to surface area ratio. As a consequence, convective cessation is invariant with radius (Stevenson, 2001).

In the absence of core solidification, the rate of entropy production available to drive a dynamo is given by (Nimmo et al., 2003):

$$\Delta E = E_R + E_s - E_k, \quad (2)$$

where E_R is the entropy due to radioactive heating in the core and depends on the radioactive heating per unit mass, E_s is the specific heat term and depends on the rate at which the core cools, dT_c/dt, and E_k is the conductive contribution and is a function of the adiabat. The advantage of the entropy approach is that it allows core solidification to be addressed (Gubbins et al., 1979). If core solidification occurs, there are additional contributions from latent heat, compositional convection, and the release of gravitational energy (for further details, see Gubbins et al., 1979; Nimmo et al., 2003). The actual rate of entropy production required to drive the geodynamo is not known (Roberts et al., 2003); here we assume that it is small, i.e., any positive ΔE results in a dynamo.

PARAMETERS

Properties of the Martian core are constrained by analysis of Martian meteorites (Wänke and Dreibus, 1988; Longhi et al., 1992) and by the planet's mass and moment of inertia (Sohl and Spohn, 1997; Folkner et al., 1997). Uncertainties in the core density and size remain, as these quantities depend on the interior temperature profile and light element abundance (Schubert and Spohn, 1990). Further, it is unclear how the thermal expansivity varies with depth. Results by Yoder et al. (2003) from MGS radio-tracking data indicate that the core is at least partly liquid and has an inferred radius of between 1520 and 1840 km.

TABLE 1. VALUES FOR MELTING CURVE

Quantity	Value	Units
T_{m0}	1811	K
T_{m1}	13.35×10^{-12}	Pa^{-1}
T_{m2}	-13.94×10^{-23}	Pa^{-2}
θ	2.41	–
χ	0.142	–

Note: T_m is mantle temperature; χ is mass fraction of S; θ is depression of Fe melting temperature by S.

For nominal conditions we adopt a core S concentration suggested by Wänke and Dreibus (1988) of 14.2 wt%. The compressed core density is estimated from densities of Fe (face-centered cubic structure) (Boehler, 1986) and FeS (IV) (Fei et al., 1995) solid phases adjusted to typical Martian core pressures and temperatures by using a third-order Birch-Murnaghan equation and the assumed value of thermal expansivity. The volume change due to melting is obtained from the Clausius-Clapeyron equation to estimate the liquid density (Longhi et al., 1992). Utilizing estimates of the planetary mass and moment of inertia (Sohl and Spohn, 1997; Folkner et al., 1997), this approach yields a core radius of 1627 km (Table DR1; see footnote 1).

For the initial mantle temperature, we use 2500 K, the mantle solidus at the CMB (Boehler, 1996; Hirschmann, 2000; Herzberg et al., 2000). If the mantle were initially completely molten, it would rapidly cool to solidus temperatures, as a completely liquid mantle would convect so vigorously that its cooling time, ~10^4 yr, would be shorter than the time scale for growth from planetesimals (Melosh, 1990). This assumption is consistent with geochemical models that indicate that Mars had a magma ocean overlying a solid mantle when the core formed (Righter and Drake, 1996; Kong et al., 1999).

The core-melting curve is derived from a best fit to the empirical results of Boehler (1986, 1992):

$$T_m = T_{m0}(1 - \theta\chi) \times (1 + T_{m1}P + T_{m2}P^2), \quad (3)$$

and is a function of pressure, P, and mass fraction of S, χ, where T_{m0}, T_{m1}, and T_{m2} are constants and θ accounts for the depression of the Fe melting temperature by S (Table 1). The melting temperatures of pure Fe and the Fe + FeS eutectic of ~2200 K and ~1500 K, respectively, at Martian core pressures, imply that $\theta = 2.41$ (Fei et al., 2000).

The present-day temperature at the top of the core, T_c, is estimated as follows. The temperature at the top of the mantle is assumed to be similar to that in Earth, ~1300 °C, and the temperature at the bottom of the mantle is derived from an adiabat similar to equation 1 by using constant values appropriate for the mantle (Nimmo and Stevenson, 2000). The temperature drop across the bottom thermal boundary layer is controlled by the rheological temperature scale at the CMB and is ~50 K (Morris and Canright, 1984; Manga et al., 2001). We therefore obtain a present-day value for T_c of ~2000 K.

Comparison of the estimated core adiabat and empirically derived melting curves of Fe + FeS mixtures reveals that for 14.2 wt% S, the core is likely to be entirely liquid with the adiabat ~500 K above the melting curve (Fig. DR1; see footnote 1). Core solidification would require the S content to be less than a few percent. Because the adiabat and solidus are nearly parallel at Martian core pressures, if the core is at least partly molten, as the results of Yoder et al. (2003) suggest, then it is likely to be entirely liquid.

RESULTS

Figure 1 shows results from our thermal evolution model for a nominal case with core content of 14.2 wt% S, an initial mantle temperature of 2500 K, and a stagnant lid throughout the planet's history. We varied the initial core temperatures above the starting mantle temperature of 2500 K and achieved an early dynamo with initial core temperatures of >2650 K (black lines; Figs. 1A–1C) without potassium in the core. The gray curves are the results where the initial core temperatures are the same as the mantle temperature without potassium (dark gray) and with 170 ppm potassium (lighter gray) in the core.

It can be seen that stagnant-lid cooling alone is insufficient to drive a dynamo during the first ~500 m.y. unless the core starts ~150 K hotter than the mantle. This early high temperature provides an initial high heat flux out of the core that rapidly declines and levels off during the first 500 m.y. (Fig. 1B). The temperature difference between the core and mantle is greatest during this period (Fig. 1C), and the entropy production exceeds the conductive threshold, i.e., ΔE is positive (Fig. 1A). When the mantle-core starting temperatures are the same, the addition of 150 ppm potassium in the core is required to generate a dynamo. However, entropy production in this case becomes great enough to generate a dynamo only after ~500 m.y. and with 170 ppm potassium; as presented in Figure 1A, this persists for >500 m.y., making it difficult to reconcile with the inferred dynamo history.

For reasonable ranges of S content (~5–30

[1]GSA Data Repository item 2004014, Table DR1 (parameter values for Mars thermal evolution model) and Figure DR1 (melting curve and adiabat for Martian core), is available online at www.geosociety.org/pubs/ft2004.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

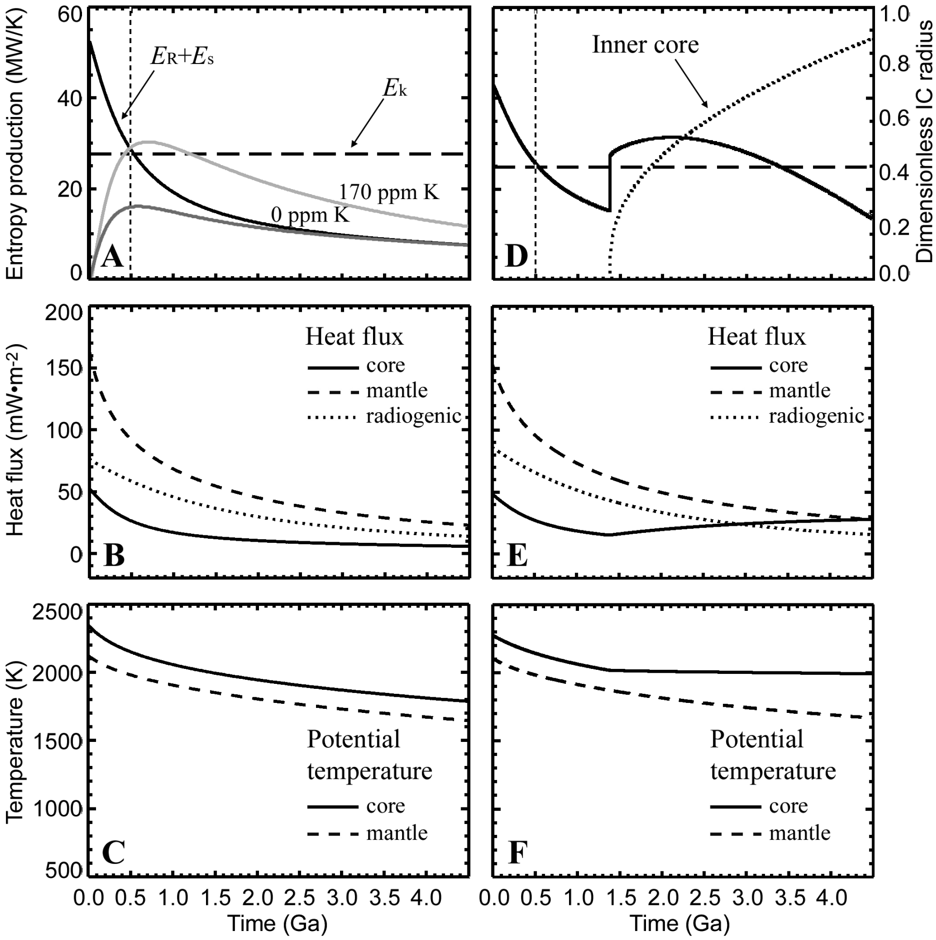

Figure 1. Results of thermal evolution model. A: Entropy production within core as function of time. Black solid line represents entropy production for core initially 200 K hotter than mantle. Dynamo occurs when $E_R + E_s$ exceeds dashed line E_k (see equation 2). Dark gray line is entropy production when core and mantle start at same temperature. Light gray line is same as dark gray line but with 170 ppm K in core. Vertical dashed line indicates approximate time at which Martian dynamo stopped. B: Heat-flux evolution as function of time for core initially 200 K hotter than mantle. C: Potential temperature evolution of core and mantle. D: As for A but with melting curve changed such that core solidification occurs. Dotted line indicates ratio of inner-core radius to core radius as function of time (scale on right). E: As for B but with core solidification occurring. F: As for C but with core solidification occurring.

wt% S), we find that the core remains liquid throughout the planet's history. However, to demonstrate the effect of core solidification on a dynamo, we utilize a pure Fe core and select a low thermal-expansivity value for Fe (5.8×10^{-5}) and a lower value for T_{m2} in the solidus (-1.394×10^{-23}) to ensure the intersection of solidus and adiabat. We do not include potassium in the core for this example and start with a hot core, as in Figures 1A–1C. In this run the core begins to solidify after ~1.3 b.y. (Fig. 1D). The core never completely solidifies, and the rate of solidification diminishes as the inner core grows. The S remains in the outer core, and the increasing concentration of S impedes solidification.

Inner-core formation results in increased levels of heat flux out of the core after ~1.3 b.y. (Fig. 1E) and the generation of additional entropy through latent heat release and compositional convection. As a result, the entropy production remains above the conductive threshold for ~2.5 b.y. These results are similar to those of Schubert et al. (1992), but are inconsistent with the inferred dynamo history.

The initial thermal state of the mantle and core is uncertain and depends on processes such as accretion, core formation, and giant impacts (Hauck and Phillips, 2002). Stevenson (2001) presented an initial very hot Mars as a scenario for an early dynamo. If the core and mantle start at the same temperature, we find that a temperature >2700 K is required to generate excess entropy during the first ~500 m.y. This seems unreasonably high as the entire planet would be molten and the mantle would rapidly cool to solidus temperatures.

Lowering the mantle viscosity increases the mantle's efficiency at removing heat from the core. A value of 1×10^{19} Pa·s for the reference viscosity at $T = 1500$ K will extract enough heat from the core to drive a dynamo. The viscosity of the Martian mantle is unknown, but this value is low compared to Earth's bulk upper mantle, which varies from 10^{20} to 10^{21} Pa·s (Mitrovica and Forte, 1997). The thermal expansivity, conductivity, and specific heat of the core also require excessively large deviations from the nominal values to generate excess entropy in the absence of a hot core. Further, the dynamo in these cases persists beyond 500 m.y.

DISCUSSION

The magnetized crust of the ancient highlands indicates that Mars possessed a dynamo during the first ~500 m.y. after the planet formed. Stagnant-lid cooling alone cannot explain this observation unless the core was initially hotter than the mantle by >150 K. Plate tectonics provides a possible explanation, but is speculative; no strong geologic evidence has been identified to support this hypothesis (Pruis and Tanaka, 1995; Zuber, 2001; Halliday et al., 2001).

An initial core temperature ~150 K hotter than the mantle is reasonable, as core formation is estimated to raise the mean temperature of the planet by 300 K (Solomon, 1979). Martian core formation occurred within 13 m.y. of solar system formation (Kleine et al., 2002), suggesting that iron segregation was a relatively rapid process. Geochemical evidence also suggests that the iron last equilibrated with the mantle at high temperatures and low pressures (Righter and Drake, 1996; Kong et al., 1999). The subsequent transport of iron to the core was probably sufficiently rapid that neither thermal nor chemical equilibration took place, in agreement with theoretical arguments (Stevenson, 1990). Under these circumstances, a core hotter than the mantle is a likely outcome.

The addition of potassium to the Martian core provides an additional source of power to generate a dynamo. Our model demonstrates that the heat derived from ~150 ppm ^{40}K is capable of generating the power required to drive a dynamo in a liquid Martian core without an initially hot core, but the dynamo occurs after an ~500 m.y. delay. As a result, ^{40}K could not have been the primary mechanism driving the Martian dynamo, but may have had a secondary effect.

Our models demonstrate that core solidification does not offer a viable explanation for the brief, early dynamo. If core solidification did take place on Mars, it likely would have generated a very long-lived dynamo (Schubert et al., 1992). The fact that no such dynamo is inferred suggests that the Martian core is entirely liquid. This places a lower bound on the core S content of ~5 wt% (Boehler, 1986, 1992).

It is clear from this study that the early history of a dynamo is dominated by the initial conditions. These initial conditions, in turn, are controlled by the rate at which planetary accretion and core differentiation proceed (Stevenson, 1990). Observations of ancient magnetization elsewhere in the solar system (e.g., the Moon, asteroids, and perhaps Mercury) may thus help to constrain the processes by which these bodies formed and differentiated.

ACKNOWLEDGMENTS

We thank Norm Sleep and Oded Aharonson for their thoughtful reviews. This research was supported by National Aeronautics and Space Administration Mars Data Analysis Program and the Royal Society.

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Manuscript received 25 June 2003
Revised manuscript received 14 October 2003
Manuscript accepted 15 October 2003

Printed in USA