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The thermal evolution of Mercury's Fe–Si core

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ABSTRACT

We have studied the thermal and magnetic field evolution of planet Mercury with a core of Fe–Si alloy to assess whether an Fe–Si core matches its present-day partially molten state, Mercury's magnetic field strength, and the observed ancient crustal magnetization. The main advantages of an Fe–Si core, opposed to a previously assumed Fe–S core, are that a Si-bearing core is consistent with the highly reduced nature of Mercury and that no compositional convection is generated upon core solidification, in agreement with magnetic field indications of a stable layer at the top of Mercury's core. This study also present the first implementation of a conductive temperature profile in the core where heat fluxes are sub-adiabatic in a global thermal evolution model.

We show that heat migrates from the deep core to the outer part of the core as soon as heat fluxes at the outer core become sub-adiabatic. As a result, the deep core cools throughout Mercury's evolution independent of the temperature evolution at the core-mantle boundary, causing an early start of inner core solidification and magnetic field generation. The conductive layer at the outer core suppresses the rate of core growth after temperature differences between the deep and shallow core are relaxed, such that a magnetic field can be generated until the present. Also, the outer core and mantle operate at higher temperatures than previously thought, which prolongs mantle melting and mantle convection. The results indicate that S is not a necessary ingredient of Mercury's core, bringing bulk compositional models of Mercury more in line with reduced meteorite analogues.

1. Introduction

Observations of planet Mercury provide enigmatic constraints on its thermal evolution. The ancient crust of Mercury (e.g., Marchi et al., 2013) and its partially liquid present-day iron (Fe) – rich core (Margot et al., 2007) are indicative of a low planetary cooling rate. Measurements by NASA's MErcury Surface, Space Environment, Geochemistry, and Ranging (MESSENGER) spacecraft indicate that Mercury's magnetic field is actively generated by core convection today (Anderson et al., 2012), which requires a superadiabatic core cooling rate to be thermally driven. Previous thermal evolution studies have assumed a substantial sulfur (S) abundance in Mercury's core to reconcile these observations. S significantly reduces the core solidification temperature, such that the partially liquid present-day state of the core can be matched by a higher core cooling rate, and the S fractionation in the liquid upon core solidification provides compositional buoyancy that helps to generate and sustain core convection (e.g., Stevenson et al., 1983; Schubert et al., 1988; Hauck et al., 2004; Grott et al., 2011; Tosi et al., 2013). The high elemental S/Si ratio of Mercury's surface confirms the presence of substantial S abundances on Mercury (Nittler et al., 2011).

The low intensity and large scale geometry of Mercury's magnetic field (Anderson et al., 2012) has been reconciled by a thin outer shell core dynamo (Stanley et al., 2005), or by a deep core dynamo covered by a conductive liquid core layer that attenuates small scale rapidly varying magnetic signatures that propagate toward the surface (Christensen, 2006; Christensen and Wicht, 2008). The deep core dynamo scenario is supported by the consistent result among thermal evolution studies that the core-mantle boundary (CMB) heat flux becomes sub-adiabatic in the first billion years (Gyr) of Mercury's evolution (Stevenson et al., 1983; Schubert et al., 1988; Hauck et al., 2004; Grott et al., 2011; Tosi et al., 2013), thermally stratifying an outer core layer. However, these thermal evolution studies always fixed the core's internal energy distribution by the adiabat, which is inconsistent with heat fluxes becoming sub-adiabatic in the outer core.

A key problem for the deep core dynamo scenario with an Fe–S core is that the compositional convection, related to S fractionation in the liquid core, penetrates the outer core's ther-

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mally stable layer and produces a magnetic field that is inconsistent with observations (Manglik et al., 2010). A scenario where Fe precipitates in shallow core regions and ‘snows’ down to be re-molten in the deeper core has been proposed, which stratifies an outer core layer compositionally (Chen et al., 2008; Vilim et al., 2010; Cao et al., 2014). This requires a S-core concentration above ∼4 wt% to sufficiently flatten the Fe–S melting curve, such that it intersects the temperature profile in a shallow core region and Fe-snow is produced (Chen et al., 2008; Dumberry and Rivoldini, 2015).

These S-rich core compositions for Mercury are difficult to reconcile with other key characteristics of the planet. The large proportion of metallic Fe (60–80 wt%) in Mercury’s bulk (Hauck et al., 2013) and the planet’s low FeO surface content (<∼2wt%) (Nittler et al., 2011) indicate that Mercury is relatively poor in oxygen (O). At the corresponding low O fugacity conditions (between −4.5 and −7.3 log units below the Fe–FeO buffer Zolotov et al., 2012), S dominantly fractionates into the mantle whereas the Si fractionation in the core increases (McCoy et al., 1999; Berthet et al., 2009; Malavergne et al., 2010; Boujibar et al., 2014; Cartier et al., 2014; Chabot et al., 2014). If Si is also a substantial component of Mercury’s core, the low core density of an Fe–S–Si core alloy with >4 wt% S is difficult to reconcile with the moment of inertia of Mercury’s core, mantle and bulk planet (Rivoldini and Van Hoolst, 2013). Additionally, the theory that Mercury’s reduced characteristic is related to an accretion of early condensates that separated from the solar nebular at high temperatures near the Sun predicts a low S content for Mercury relative to other terrestrial planets (Lewis, 1972). In combination with the S-poor olivines and pyroxenes that may form the lower mantle in the mantle crystallization sequence (Brown and Elkins-Tanton, 2009; Cartier et al., 2014), the high S/Si ratio of Mercury’s surface may represent an upward migration of S during the planet’s differentiation and may be indicative of a S-deficient deep interior and core in particular.

Metal-silicate partitioning experiments at reducing conditions yield Fe-rich metal with substantial Si concentrations and devoid of S (e.g., Cartier et al., 2014; Chabot et al., 2014). Because Si negligibly fractionates between solid and liquid metal (Kuwayama and Hirose, 2004), the solidification of an Fe–Si inner core does not produce significant compositional buoyancy, maintaining the stability of a thermally stable upper core layer. Although Si lowers the core melting temperature by only ∼15 K/(wt%Si), less efficient compared to the decrease of melting temperature with S of ∼50 K/(wt%S) (Kuwayama and Hirose, 2004; Morard et al., 2011), the core can accommodate higher Si than S contents within the moment of inertia constraints (Knibbe and van Westrenen, 2015). Additionally, Si increases the latent heat release upon solidification (Desai, 1985) and lowers the thermal conductivity of the metal (De Koker et al., 2012; Konopkova et al., 2016; Secco, in press).

Here, we present thermal evolution calculations with Si as Mercury’s dominant core alloying element, along with the first implementation of an outer core conductive layer in a global thermal evolution model. The main purpose is to examine the evolution and present day state of Mercury’s core in this scenario and assess the implications for Mercury’s magnetic field. Additionally, we examine the possibility of generating a magnetic field for a long period in the planet’s evolution in light of the recent detection of magnetized ancient crust (Johnson et al., 2015) and discuss trends in model outcomes with relevance to the recent estimates for Mercury’s crustal thickness (Padovan et al., 2015) and the evolution of Mercury’s mantle.

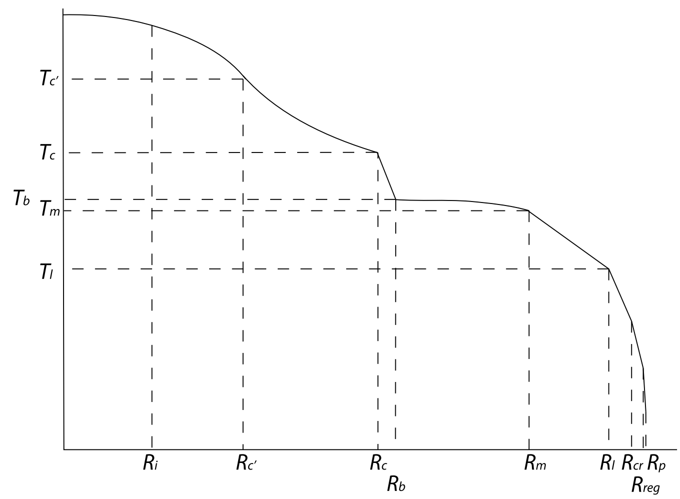

Fig. 1. A sketched interior configuration and temperature profile of Mercury. R_i refers to the inner core radius, $R_{c'}$ and $T_{c'}$ the radius of and temperature at the boundary of convective and conductive molten outer core, R_c and T_c the radius of and temperature at the CMB, R_b and T_b the radius of and temperature at the lower boundary of the mantle convection cell, R_m and T_m the radius of and temperature at the upper boundary of the mantle convective cell, R_l and T_l the radius of and temperature at the bottom of the lithosphere, R_{cr} and R_{reg} the bottom radii of the crust and regolith respectively and R_p the planet’s radius.

2. Methods

2.1. The thermal evolution model

Gravitational energy released during core-mantle differentiation is sufficient to heat Mercury to a molten and convective interior state (Schubert et al., 1988). Because Mercury’s surface does not show geological features of the planetary expansion related to this heating event, the age of the primary crust of ∼4.2 Ga (e.g. Marchi et al., 2013) sets a minimum age for core-mantle differentiation. If accretional energy has been sufficiently retained by the planet, differentiation occurred in tandem with planetary formation (Schubert et al., 1988). If accretional energy was largely radiated away, it may have taken several hundreds of millions of years after formation to reach sufficiently high temperatures by radiogenic heating for differentiation to start (Schubert et al., 1988; Noyelles et al., 2014). The starting point of the thermal evolution examined in this study is set at 4.3 Ga (just after core mantle differentiation) with a hot planet.

Mercury is modeled by concentric shells of a (partially) liquid Fe–Si metallic core, silicate mantle and silicate lithosphere. The solid part of the core grows from the planet’s center. A conductive layer forms at the top of the outer core when the CMB heat flux drops below the adiabatic heat flux. The mantle is characterized by a convective region in between conductive thermal boundary layers. We represent the upper part of the lithosphere by a low-density and low-conductivity crust and a small regolith layer with further reduced thermal conductivity (as in Grott et al., 2011). A schematic layout of the planet is given in Fig. 1.

2.2. The evolution of mantle and crust

The rate of change of the temperature T_m at the upper convective mantle boundary (R_m) is given by energy balance equation (e.g., Morschhauser et al., 2011)

$$V_m \rho_m c_m (1+St)\frac{dT_m}{dt} = -A_l\left(F_m + \left(\rho_{cr}L_{cr} + \rho_{cr}c_{cr}(T_m - T_l)\right)\frac{dD_{cr}}{dt}\right)$$