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On the thermal and magnetic histories of Earth and Venus: Influences of melting, radioactivity, and conductivity

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ABSTRACT

The study of the thermal evolution of Earth's interior is uncertain and controversial in many respects, from the interpretation of petrologic observations used to infer the temperature and dynamics of the interior, to the physics and material properties governing heat transport. The thermal history of Venus is even more uncertain, but the lack of a dynamo at present in an otherwise similar planet may provide additional constraints on terrestrial planet evolution. In this paper a one dimensional thermal history model is derived that includes heat loss due to mantle melt eruption at the surface to explore its influence on the thermal and magnetic history of Earth and Venus. We show that the thermal catastrophe of Earth's mantle, which occurs for a present day Urey ratio of 1/3 and convective heat loss exponent of $\beta = 1/3$, can be avoided by assuming a rather high core heat flow of ~15 TW. This core heat flow also avoids the new core paradox by allowing for the geodynamo to be thermally powered prior to inner core growth for core thermal conductivities as high as 130 $\mathrm{W m^{-1} K^{-1}}$. Dynamo regime diagrams demonstrate that the mantle melt eruption rate has a minor effect on the history of mobile lid planets due to the efficiency of plate tectonic convective heat loss. However, if Earth were in a stagnant lid regime prior to 2.5 Ga, as has been proposed, then at least ~5% of mantle melt is required to erupt in order to thermally power the paleodynamo at that time. Dynamo regime diagrams for stagnant lid Venus models indicate that more than half of the melt generated in the mantle is required to erupt in order to overcome the insulation imposed by the stagnant lid and drive a dynamo. This implies that with an Earth-like mantle radioactivity the Venusian dynamo shut down ~0.3 Ga for an eruption efficiency of 50%, and ~3 Ga for an eruption efficiency of zero. Consequently, a stagnant lid alone does not prevent a core dynamo if melting of the upper mantle provides a substantial mantle heat sink.

1. Introduction

One of the most curious features of the Solar System is that the Earth is so different from Venus, its nearest planetary twin. The first order similarities between the two planets are striking: Venus' mass and bulk density are 82% and 95% of Earth's, respectively, and their moments of inertia are indistinguishable beyond 0.33 (Bills et al., 1987; Kaula, 1990). They therefore share a common bulk composition and are similarly differentiated. Despite these similarities, their dichotomies are at least as numerous. Earth has plate tectonics, while Venus has a stagnant lid (Solomon et al., 1992). Earth has a strong magnetic field maintained by convective dynamo action in its core, while Venus has no detectable magnetic field (Phillips and Russell, 1987). Earth's atmosphere, moderate in pressure and temperature, sustains complex precipitation-evaporation and weathering cycles, while Venus is covered by a massive runaway greenhouse atmosphere with surface temperatures high enough to preclude the precipitation of any major volatiles species (Driscoll and Bercovici, 2013). These dramatic differences today imply a divergence in their evolution from surface to core. In this sense, the Earth–Venus dichotomy represents a fundamental puzzle that must be explained by any generalized theory of terrestrial planet evolution.

In this paper, the thermal and magnetic evolution of Earth and Venus are revisited using a one dimensional thermal history model. This modeling approach has a long and sometimes controversial history (e.g. Schubert et al., 1979; Stevenson et al., 1983; Christensen, 1985; Richter, 1985; Spohn, 1991; Korenaga, 2006; Davies, 2007; Labrosse and Jaupart, 2007). Uncertainties in Earth's present day energy budget, which are critical to inferring its

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history, persist due to the inherent difficulty in inferring the properties of the deep interior. Specifically, there is considerable uncertainty in the present day mantle and core secular cooling rates (Herzberg et al., 2010; Lay et al., 2008) and radiogenic concentrations (Gessmann and Wood, 2002; Murthy et al., 2003), although geoneutrino detectors promise new constraints in the near future (Araki et al., 2005; Dye, 2012). In addition, the dynamics of mantle heat loss, which is intimately connected to surface tectonics, is the subject of considerable debate (e.g. Conrad and Hager, 1999; Korenaga, 2003; van Hunen and van den Berg, 2008; Davies et al., 2009; Leng and Zhong, 2010; Buffett and Becker, 2012). Rather than propose a new solution to these difficulties we adopt a generalized thermal history formulation so that a range of models can be explored. The one substantial addition to the thermal history formulation developed here is the addition of heat loss due to mantle melt eruption (e.g. Moore and Webb, 2013).

The thermal history of the Earth is better constrained than that of Venus due to geologic, paleomagnetic, and geochemical signals preserved in its crust. These observations indicate that Earth has maintained plate tectonics and a strong dynamo generated magnetic field since at least the end of the Archean (Herzberg et al., 2010; Biggin et al., 2012), and possibly earlier (Condie and Kröner, 2008; Tarduno et al., 2010). Despite the increasing number of observations and advances in geophysical modeling, the actual cooling rates of Earth's mantle and core remain uncertain. This uncertainty is manifest in the range of thermal evolution models and cooling histories proposed (e.g. Stevenson et al., 1983; Richter, 1985; Christensen, 1985; Schubert et al., 2001; Labrosse and Macouin, 2003; Korenaga, 2006, 2008; Nimmo et al., 2004; Nimmo, 2007; Davies, 2009; Aubert et al., 2009).

Planetary cooling models rely on a fluid dynamic description of heat transfer through a convecting fluid, where heat flow depends on fluid temperature to some exponent β, to compute the rate of heat loss from the mantle (Schubert et al., 2001). Although this method has been borne out by a variety of thermal convection experiments with variable viscosity fluids (e.g. Davaille and Jaupart, 1993; Giannandrea and Christensen, 1993; Manga and Weeraratne, 1999; Davaille and Limare, 2007), it fails to produce reasonable heat fluxes and temperature histories for the Earth with the low Urey ratios, the ratio of radiogenic to total heat loss, inferred geochemically. In fact, given the present day mantle heat flow, temperature, and radiogenic abundance, simple mantle thermal history calculations predict a thermal catastrophe only 1–2 Ga (Korenaga, 2006). Proposed solutions to this paradox include using a homologous temperature dependence for lithospheric viscosity (Christensen, 1985), changing the convective heat flow law to one controlled by lithospheric stiffness (Conrad and Hager, 1999; Korenaga, 2006), or substantial melting of the deep mantle throughout much of Earth history (Labrosse et al., 2007). These alternative models modify the mode of convective cooling so as to avoid a rapid increase in the mantle cooling rate going back in time, and produce significantly different thermal histories for the Earth that cannot be easily discriminated with present observations (e.g. Davies, 2009).

In the present energy budget of the Earth, heat loss due to melting of the mantle and subsequent cooling of that melt at the surface is a minor component of the total planetary heat output. This is expected from a simple estimate of the melt heat transport: $Q_{melt} = \dot{M}_{melt}(L + c_p\Delta T_{melt})$, where $\dot{M}_{melt}$ is the melt mass production rate, L is latent heat of melting, and $c_p\Delta T_{melt}$ is the internal heat transported by the melt adiabatically to the surface. Using typical values for melting at mid-ocean ridges of $\dot{M}_{melt} = 1.3\times 10^6\,\mathrm{kg\,s^{-1}}, L = 320\,\mathrm{kJ\,kg^{-1}}$, and $c_p\Delta T_{melt} = 1.5\times10^6\,\mathrm{J\,kg^{-1}}$, the melt heat loss $Q_{melt} = 2.4$ TW (Nakagawa and Tackley, 2012) is small compared to the total surface heat loss of 46 TW. However, a hotter mantle is expected to melt more of the mantle and produce hotter melt, and, therefore, amplify the importance of melt heat loss to the cooling of the planet (Richter, 1985). Moreover, mantle melting should be magnified in a stagnant lid planet like Venus where normal surface convective heat transfer is muted by conduction through a thick crust and a hotter mantle is expected. Melt heat loss has been referred to as a "heat pipe" in the context of Io (O'Reilly and Davies, 1981; Moore, 2003) and Venus (Turcotte, 1989), where eruptions tend to occur in the form of hot spots rather than ridges. The heat pipe mechanism has also been proposed as a major heat sink on the early Earth prior to the onset of plate tectonics (Moore and Webb, 2013). Using a 2-D numerical mantle convection model, Nakagawa and Tackley, 2012 demonstrated that melt heat loss can be a significant component of Earth's heat budget in an earlier hotter mantle. Similarly, Armann and Tackley, 2012 applied the same model to Venus and demonstrated that melt heat loss can play a much larger role in a stagnant lid planet. These studies indicate that heat loss due to melting, which has been under appreciated in previous thermal histories of Earth and Venus, is a crucial component of the interior energy balance and lifetime of the dynamo.

Planetary magnetic fields maintained by convective dynamo action offer a unique view into the energetic state of the core. Present day observations and theoretical considerations indicate that the geodynamo is powered by a combination of thermal and compositional buoyancy (Nimmo, 2007), both directly related to the core cooling rate. However, recent measurements of the electrical conductivity of core materials at the relevant pressure and temperature conditions indicate that the thermal conductivity of the core may be 2–3 times larger than previous thought (Pozzo et al., 2012; de Koker et al., 2012; Gomi et al., 2013), which implies that the heat required to keep the core adiabatically well mixed is 2–3 times higher as well. This strains the core heat budget so much that the traditional estimates of the core cooling rate of $\sim$10 TW would predict that the core today is not fully thermally convective, and that a stably stratified layer may exist in the outermost outer core (Pozzo et al., 2012; Gomi et al., 2013). If the geodynamo is driven mainly by compositional convection associated with light element release at the inner core boundary as it grows, then a high core conductivity raises the question as to what maintained the paleogeodynamo prior to inner core nucleation $\sim$1 Ga when the paleomagnetic field (Tauxe and Yamazaki, 2007) is known to have been quite strong? This is the so called "new core paradox" (Olson, 2013).

In this paper a 1-D thermal history model is developed and used to explore the influence of mantle melt heat loss efficiency, radiogenic heat production rates, and core conductivity on the thermal and magnetic evolution of Earth and Venus. Section 2 describes the basic thermal history model, including a discussion in §2.2 on the efficiency of convective mantle cooling of a mobile versus stagnant lid planet. Section 2.3 develops a parameterized model for the rate of heat loss due to melting of upwelling mantle and includes this new heat loss term in the thermal history equations. Section 3 describes the core thermal and magnetic evolution model. Simple models computed backwards in time, referred to as "histories", are presented in §4, and a solution to the thermal catastrophe of the mantle is proposed. Section 5 demonstrates how tectonic style leads to the divergence of the thermal and magnetic evolutions for Earth and Venus in forward "evolution" models starting from the same initial temperatures. Magnetic field regime diagrams in §5.3 are used to infer the range of parameters that are consistent with the observations of both planets. The implications of these results are discussed in §6.

2. Thermal history model

The thermal history of a terrestrial planetary interior, which is differentiated into a silicate mantle overlying an iron-rich core,