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Thermal evolution of Earth with magnesium precipitation in the core

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ABSTRACT

Vigorous convection in Earth's core powers our global magnetic field, which has survived for over three billion years. In this study, we calculate the rate of entropy production available to drive the dynamo throughout geologic time using one-dimensional parameterizations of the evolution of Earth's core and mantle. To prevent a thermal catastrophe in models with realistic Urey ratios, we avoid the conventional scaling for plate tectonics in favor of one featuring reduced convective vigor for hotter mantle. We present multiple simulations that capture the effects of uncertainties in key parameters like the rheology of the lower mantle and the overall thermal budget. Simple scaling laws imply that the heat flow across the core/mantle boundary was elevated by less than a factor of two in the past relative to the present. Another process like the precipitation of magnesium-bearing minerals is therefore required to sustain convection prior to the nucleation of the inner core roughly one billion years ago, especially given the recent, upward revision to the thermal conductivity of the core. Simulations that include precipitation lack a dramatic increase in entropy production associated with the formation of the inner core, complicating attempts to determine its age using paleomagnetic measurements of field intensity. Because mantle dynamics impose strict limits on the amount of heat extracted from the core, we find that the addition of radioactive isotopes like potassium-40 implies less entropy production today and in the past. On terrestrial planets like Venus with more sluggish mantle convection, even precipitation of elements like magnesium may not sustain a dynamo if cooling rates are too slow.

1. Introduction

The dynamo created in Earth's liquid outer core has survived for billions of years. Paleomagnetic studies of unmetamorphosed rocks with ages near 3.45 Gyr unambiguously show that the strength of Earth's global magnetic field at that time was at least half its present-day value (e.g., Tarduno et al., 2010; Biggin et al., 2011). No rocks of sufficiently low metamorphic grade have been found from earlier epochs, so the question of whether our magnetic field is even older remains unanswered. Recently, detrital zircon crystals found in the Jack Hills of Western Australia were proposed to record field intensities of modern magnitudes (Tarduno et al., 2015). These data are controversial, however, because zircon-bearing rocks in the Jack Hills may have suffered pervasive remagnetization related to the emplacement of a nearby igneous province (e.g., Weiss et al., 2015). In any case, how to power convection in the core and thus a dynamo for the vast majority of Earth's history remains one of the most pressing puzzles in geophysics.

Thermal convection in the core is possible if the heat flow across the core/mantle boundary (CMB) exceeds the rate at which heat is conducted along an adiabatic temperature gradient (e.g., Stevenson, 2003). Over the past few years, some theoretical calculations (e.g., de Koker et al., 2012; Pozzo et al., 2012) and diamond-anvil cell experiments (e.g., Gomi et al., 2013; Seagle et al., 2013; Ohta et al., 2016) have indicated that the thermal conductivity of the core's iron-rich alloy is a factor of two to three larger than prior estimates. The conductive heat flux is ∼10–15 TW at present according to these new values. However, countervailing evidence from high-pressure experiments that the previous, low values are actually correct has also been presented recently, so debate over this issue will likely continue (Konôpková et al., 2016).

Cooling rates approaching twice the conductive heat flux have been suggested as the minimum required to compensate for Ohmic dissipation (e.g., Stelzer and Jackson, 2013). But this dissipation mainly occurs at high harmonic degree and its scaling with dipole field strength is uncertain. Since the dissipation due to the low harmonics alone is far less than the actual heat flow, maintaining the observed field with a heat flow only mildly in excess of conduction along the adiabat is possible in principle. In any case, the actual CMB heat flow of ∼5–15 TW estimated from seis-

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mology and mineral physics (e.g., Lay et al., 2008) may be only marginally sufficient to sustain the dynamo by thermal convection alone. Fortunately, the dynamic chemistry of the core yields additional sources of energy.

The exclusion of light elements from the solidifying inner core provides enough compositional buoyancy to drive convection today. Once compositional buoyancy is present, the heat flow out of the core need not exceed conduction along the adiabat (i.e., convection can even carry heat downwards). In practice, models with a growing inner core also benefit from the significant release of latent heat and accordingly require less rapid cooling. Conventional calculations have indicated that the inner core nucleated roughly one billion years ago (e.g., Labrosse et al., 2001). The age of the inner core is several hundred million years less in models with increased CMB heat flow and thus faster cooling/freezing to accommodate the revised values for thermal conductivity (e.g., Nimmo, 2015; Labrosse, 2015).

The energy available for dissipation in dynamo generation dramatically increases once the inner core forms, which might imply a larger magnetic field according to scaling laws where the buoyancy flux determines the global field strength (e.g., Christensen, 2010). In some canonical models, the inner core thus prevents the dynamo from turning off (e.g., Stevenson et al., 1983), but these models do not explain the current total heat flow of Earth. Biggin et al. (2015) claimed to observe an increase in Earth's dipole moment associated with the formation of the inner core in the Mesoproterozic. Given the relevant experimental and statistical uncertainties, however, the available data are arguably consistent with roughly constant field intensities throughout the Precambrian (e.g., Smirnov et al., 2016).

O'Rourke and Stevenson (2016) proposed the precipitation of magnesium-bearing minerals as an alternative power source. One or two weight percent of magnesium can partition into the core in the high-temperature aftermath of giant impacts during Earth's accretion according to earlier calculations (Wahl and Militzer, 2015) and subsequent diamond-anvil cell experiments (Badro et al., 2016). Because its solubility in iron alloy is strongly-temperature dependent, subsequent cooling quickly saturates the core in magnesium. Elements like aluminum and calcium may have similar thermodynamic properties (Badro et al., 2016), but their abundances are relatively small. Transporting magnesium-rich oxide or silicate across the CMB provides an order-of-magnitude more gravitational energy than freezing an equivalent mass of the inner core. Precipitation drives vigorous, compositional convection before the nucleation of the inner core, even without vastly higher CMB heat flow than today. O'Rourke and Stevenson (2016), however, only calculated the CMB heat flow implied by a constant rate of entropy production for the dynamo. In reality, mantle dynamics control CMB heat flow, so entropy production should vary over time.

The purpose of this paper is to describe simple models of Earth's thermal evolution that are consistent with the observed longevity of the dynamo. First, we describe how we couple a one-dimensional model of the core to simple scaling laws for mantle dynamics. We next identify which parameters control the amount of power available for the dynamo throughout geologic time. Specifically, we focus on the rheology of the boundary layer at the base of the mantle and the abundance of radioactive isotopes like potassium-40 in the core. After presenting representative simulations, we discuss the limitations of our model for early Earth history and the implications for other planets.

2. Theoretical formulation

In this section, we present a parametrized model for the coupled evolution of Earth's core and mantle. Fig. 1 shows the simplified structure with which we calculate thermal histories. Key model parameters are listed in Table 1. As in nearly all models of core history for the past fifty years, we assume that the core is sufficiently low viscosity that the convective state is extremely close to an isentropic and homogeneous state, except in thin boundary layers (e.g., Stevenson, 1987). Although most previous studies (e.g., Stevenson et al., 1983; Buffett, 2002) only consider a thermal boundary layer at the base of the mantle, we allow for the existence of a stagnant layer that may not participate in convection because it is compositionally dense (Hernlund and McNamara, 2015), possibly the solidified remnant of a basal magma ocean (e.g., Labrosse et al., 2007). The existence of this distinct chemical layer could explain why the thermal excess associated with mantle plumes may be less than half the total temperature contrast across the CMB (e.g., Farnetani, 1997). Because our primary focus is how mantle dynamics affect the evolution of the core, we do not model the dynamics of the crust and lithosphere in detail. Finally, we present simulations that demonstrate the effects of varying key parameters.

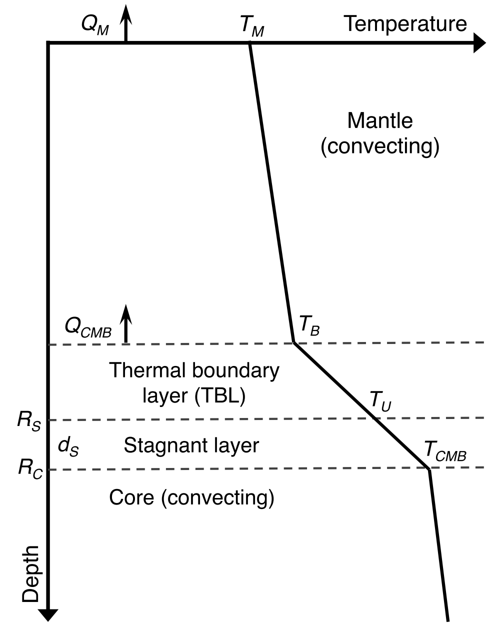

Fig. 1. Cartoon showing the assumed thermal structure of Earth and the key parameters tracked during simulations of Earth's evolution. The temperature gradients and vertical dimensions of each layer are not to scale.

Table 1
List of key parameters tracked during simulations of Earth's thermal evolution and their definitions.

Term	Definition
Q_M	Heat flow from the mantle
Q_{CMB}	Heat flow from the core
Q_R	Radiogenic heating in the core
Q_S	Secular cooling of the core
Q_P	Gravitational energy release from precipitation
Q_G	Gravitational energy release from the inner core
Q_L	Latent heat associated with the inner core
H_M	Radiogenic heating in the convecting mantle
T_M	Potential temperature of the mantle
T_B	Basal temperature of the convecting mantle
T_U	Temperature at the top of the stagnant layer
T_{CMB}	Temperature of the uppermost core
T_I	Temperature at the inner core boundary
R_I	Radius of the inner core
E_K	Entropy production associated with conduction
E_ϕ	Entropy production available for the dynamo