Iron-Oxide Exsolution an Important Dynamo Power Source

Abstract

Earth has been observed to have a global magnetic field for at least the past 4.1 billion years [CITE], but recently proposed high thermal conductivity measurements [CITE] and young inner-core ages [CITE] make it difficult to sustain Earth's magnetic field by cooling and inner-core growth alone [CITE]. Exsolution of light elements such as magnesium (O'Rourke and Stevenson 2016) and silicon (Hirose et al. 2017) from Earth's core have been proposed as sources of energy to resolve this problem. We show that exsolution of iron-oxide is a powerful third source of energy and that the exsolution of all three species is controlled by interactions with the mantle. We use coupled whole-earth thermodynamic and chemical interaction model to estimate the entropy available to power a dynamo across Earth's history and examine signals and timing of inner-core nucleation.

Introduction

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Model

We model the thermal evolution of the Earth using a coupled core-mantle thermodynamic model following the methods of Stevenson (1983) for parameterized mantle convection and Nimmo (2015) for core thermodynamics. We add chemical reactions between the core and mantle to this system, modeled as an interaction layer at the base of the mantle that is allowed to chemically equilibriate with the core fluid. This governs the exsolution of light elements from the core into the mantle, which contribute to the heat and entropy budget of the core. This system is cast as a system of ODEs and solved forward in time from Earth's formation to the present day.

Thermodynamic Model

Core Thermodynamics

We model the thermodynamics of the core by modifying the method of Nimmo (2015). This model assumes the CMB heat flow is not significantly subadiabatic, which is satisfied by all of our valid runs. Nimmo tracks the contribution to the heat and entropy budgets of the core from all relevant sources. To his formulation, we add terms for latent heat release and gravitational energy release

for MgO, SiO2, and FeO exsolution from Earth's core. The gravitational energy and entropy released by exsolution is expressed as

$ Q_{g,i} = \int \rho \psi \alpha_i \frac{dc}{dt} dV $, $E_{g,i} = Q_{g,i}/T_c$

from Table 1 in that work. This requires an estimate of the compositional expansion α_i for each species. We use $\alpha_{MgO} \sim 0.84$ (O'Rourke, Korenaga 2016) and $\alpha_{SiO2} \sim 1.117$ (Hirose et al. 2017). For FeO, we perform a simple hard-sphere estimate to compute the change in density between a parcel of fluid with 99 wt% Fe, 1 wt% O and 100 wt% Fe to obtain $\alpha_{FeO} \sim 0.28$.

We compute latent heat release for each species through $Q_{L,i} = L_{H,i}\frac{dm_i}{dt}$, where $\frac{dm_i}{dt}$ is the change in the mass of the species with time and $L_{H,i}$ is the latent heat of exsolution for each species. We use $L_{H,SiO2} = 4300$ kJ/kg (Hirose et al., 2017), while $L_{H,MgO} = L_{H,FeO}=910$ kJ/kg represents a value intermediate between that for SiO2 and inner-core solidifcation. This value has little effect on the system, as the total heat released by exsolution of light elements is quite small in all of our model runs. In fact, we find <5% difference in model run outcomes when setting all latent heat of exsolution values to zero. Note also that unlike latent heat from inner-core growth, the latent heat from exsolution does not contribute to the entropy production in the core as exsolution occurs at the CMB. In fact, exsolution has a negative effect on the energy available to power a dynamo because it decreases the core cooling rate.

Mantle Thermodynamics

Chemical exchange between the core and the interaction layer at the base of the mantle cause only small changes to the physical properites of the bulk medium and do not significantly affect mantle dynamics. Therefore, we use a simple 1D parameterized whole-mantle convection model following the method of Stevenson (1983). We use an Arrhenius mantle viscosity with present day viscosities varying from $\mu_p = 10^{19.5}$ to 10^{21} Pa s. We use Stevenson's values for all parameters except for radiogenic heat production in the mantle, which we replace using the method from Korenaga (2005) which includes four individual radioactive species.

Chemical Reactions

Chemical Interaction Layer

We model chemical interactions between the core and mantle as a system of equilibrium reactions with the base of the lower mantle. First-principles calculations and seismic observations argue for a pyrolitic lower mantle, necessitating the convesion of minerals such as olivine and ferropericlase transported from the upper mantle into Bridgemanite at depth. While the exact mechanisms of this process are not well-understood, it requires interaction between mineral species at relatively large distances over timescales of millions of years. Thus, at long

enough timescales the chemical state of the lower mantle should be able to be modeled as an equilibrium reaction system.

A similar argument can be made about core-mantle interactions. Material from the lower mantle can both diffuse into and exolve out of the core, setting up a system of equlibrium reactions. To model this, we assume a typical lengthscale for chemical interactions in the mantle and keep track of the moles of each mantle mineral in this layer. There is considerable uncertainty concerning the mechanisms by which Earth's core and mantle interact, and therefore the appropriate lengthscale is uncertain. If chemical interaction is dominated by chemical diffusion, lengthscales of H~1-10m over 10Myr may be appropriate [CITE]. If grain-boundary diffusion occurs at the CMB, core material could interact H~100m [CITE], while up to H~1km could be appropriate if core material physically intrudes into the lower mantle [CITE].

Exsolved material from the core builds up in this layer over time, altering the composition of the interaction layer and therefore affecting the exsolution of core species. During chemical interactions, this layer moves with the background mantle convection along the free surface of the CMB, setting up a "conveyer belt" system as in figure 1. In this system, the layer is removed by mantle upwellings or plumes and replaced with fresh background mantle due to downwellings or sinking slabs on a timescale τ set by the background mantle convection. Analytical stability analyses and 2D numerical mantle convection codes show that exsolved material in our simulations never reaches the required thickness and buoyancy contrast to form Rayleigh-Taylor instabilities, in contrast with other recent results (Helffrich 2018).

To model this "conveyer belt" layer overturn, we include a layer erosion term in our system of differential equations that pushes each species towards the background mantle composition value. Assuming a pyrolitic background mantle composition and specifying a layer thickness sets the total number of moles $M_{b,i}$ of each species i in the layer. Our equations are cast in terms of absolute number of moles, so the erosion term $\left(\frac{dM_i}{dt}\right)_{erosion}$ for each species takes is a function only of the difference $M_i - M_{b,i}$ and the timescale of layer overturn τ (see supplement).
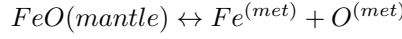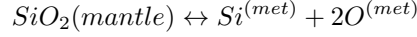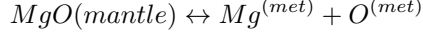
In our model, we assume a present day τ_p between 200 and 800 Myr. The early earth likely had more vigorous convection, so we set the timescale at early earth $\tau_0 = \tau_p/8$ and set τ at intermediate times using an exponential relationship with a timescale of 1 Byr.

We assume a pyrolitic background lower mantle based on seismic and first-principles estimates. This consists of 83% Bridgemanite, 16% ferroperilase, and 1% free SiO2, with a magnesium number of 0.8 [CITE].

Core Chemistry

We include four components in the core in our model: Fe, Mg, Si, O. We track

the absolute number of moles of each species M_i in the core, and calculate the mole fractions as $X_i = M_i/M_c$ where M_c is the total number of moles in the core. These species exchange with the mantle through the equilibrium reactions

$$MgO(mantle) \leftrightarrow Mg^{(met)} + O^{(met)}$$

$$SiO_2(mantle) \leftrightarrow Si^{(met)} + 2O^{(met)}$$

$$FeO(mantle) \leftrightarrow Fe^{(met)} + O^{(met)}$$

as can be seen in figure 1. Each of these reactions is governed by their temperature dependent equilibrium constant. We use the values reported by Hirose et al. (2017) for SiO2 and FeO and Badro et al. (2015) for MgO. For each species we calculate the equilibrium constant as

$K_1 = a_{Mg}a_O/a_{MgO} = X_{Mg}X_O/X_{MgO}$

assuming activities are all equal to one and other species present have no influence. In our model, we only allow Mg, Si, and O to leave the core, but not enter. An undersaturated core would allow light elements from the mantle to dissolve into the core fluid at the CMB. However, this would likely form a thin stratified layer at the CMB heavily enriched in light elements, which would prevent the reaction from continuing and would not mix into the bulk of the core. Thus, the bulk of the core composition would remain unchanged.

We model the mantle interaction layer as including five species: MgO, SiO_2, FeO, MgSiO_3, and FeSiO_3. The total number of moles of each species are tracked and governed by the reactions

$ MgSiO_3 \leftrightarrow MgO + SiO_2 $$FeSiO\_3 \leftrightarrow FeO + SiO\_2$$FeSiO_3 + MgO \leftrightarrow FeO + MgSiO_3$
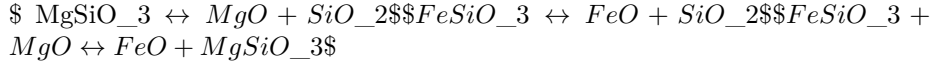
where the third reaction maintains stoichiometry in the layer and has an equilibrium constant defined as 1. This constrains K_4 to equal K_5. The lower mantle is expected to be mostly Bridgemanite, so that K_4 = K_5 << 1.

System of Equations

To determine the evolution of these reaction equations over time, we take the derivative of each equilibrium reaction with respect to temperature. These equations are tightly coupled due to the dependence of each mol fraction on both the number of moles of that species and the total number of moles in the core or mantle. We use the open source sympy linear equation solver to convert the equations into a system of first-order nonlinear ODEs in order to integrate them into the thermodynamic model.

For the fully coupled thermo-chemical model, we explicitly track the temperatures of the upper mantle and core-mantle boundary, as well as the absolute number

of moles of each of the four species in the core and five in the interaction layer. We specify the initial state by setting $T_{CMB,0}$ and $T_{UM,0}$ as well as the initial composition of the core $M_{Fe}, M_{Mg}, M_{Si}, M_O$. This sets the initial interaction layer composition by requiring it be in equilibrium with the core at formation. We find that many initial core compositions require initial interaction layer compositions very different from what we expect at the present day. In particular, many initial compositions require a layer highly enriched in ferropericlase. Thus, we adjust the equlibrium constant governing the Bridgeanite to ferropericlase ratio towards the expected present-day value on the same timescale τ as mantle overturn. Finally, we solve the complete system of equations using scipy's built-in ODEint solver.

Results

Using this model, we compute the evolution of Earth for a range of plausible parameters and keep solutions that result in inner-core sizes within 10% of the present-day value. We show the evolution of Earth for one set of parameters in figure 2. In this run, entropy production is initially dominated by secular cooling before FeO begins to exsolve out of the core and becomes the dominant source of entropy production. Then, as the core cools, both SiO2 and MgO begin to exsolve at distinct times, causing small spikes in total entropy available but still contributing less to the overall entropy budget than either FeO or secular cooling. For the ~2 Byrs before inner-core nucleation, secular cooling is the largest single contributor to the entropy budget, but exsolution of light elements stil produces a significant amount of entropy. Inner-core nucleation brings a powerful new source of entropy production which causes a spike in total available entropy for the dynamo and depresses the core cooling rate. This causes entropy production from cooling and exsolution to decline and allows inner-core entropy production to dominate all other sources to the present day.

The mantle interaction layer is significantly enriched in iron in the early days, with a Mg# ~ 0.25 (see fig. 2d). As exsolution slows near the present day, Mg is incorporated from background mantle convection and the Mg# increases to ~0.6. The layer is still enriched in iron at the present day compared to the background mantle, which has a Mg# ~0.9. This has possible implications for the origin of LLSVPs and ULVZs, as many authors have proposed iron enrichment as an origin of these features [TODO:CITE]. This plot represents a "typical" run of our model, but is far from the only possible Earth history.

Table 1 | Parameter | Description | Minimum | Maximum | | ————————- | ———— | ———— | ———— | | T_{CMB_0} | initial CMB temperature | 5000K | 6250K | | ν_p | present-day mantle viscosity | $10^{19.5}$ | 10^{21} | | $|w_{Mg,0}|$ initial Mg wt% in core | TODO | | $|w_{Si,0}|$ initial Si wt% in core | | | $|w_{O,0}|$ initial O wt% in core | | | | τ_p | present-day mantle overturn timescale | 200 Myrs | 800 Myrs | | H | interaction layer thickness | 30m | 1000m |

Entropy Production Over Time

We run a suite of parameters, varying the values listed in table 1 and keeping runs that have result in a present-day inner-core size within 10% of the real value. We show entropy histories of all valid runs in figure 3. Figure 3a shows the entropy available to power a dynamo, demonstrating a huge variety of possible histories, but with consistent trends between histories. There is generally a large amount of entropy production in early Earth which declines to a minimum in the billion years before inner-core nucleation before increasing suddenly as the inner-core begins to grow. Some valid histories have only thermal convection before inner-core nucleation, with no light-element exsolution at all. However, most valid runs include a significant amount of light-element exsolution.

FeO exsolution out of the core is a significant source of entropy production in all computed histories that undergo light-element exsolution. FeO contributes 50-100% of the compostional entropy production in the early earth, and 40-100% just before inner-core nucleation (figure 3b). In some histories MgO or SiO2 breifly produce more entropy that FeO, but in almost all histories FeO is the largest single source of compositional entropy production through the majority of Earth's history. However, after the inner core begins to grow, FeO exsolution becomes much less important, contributing less than 10% of the compositional entropy budget at the present day in most histories.

It is possible for cooling to either dominante or only produce a small fraction of the available entropy before inner-core growth takes over. However, general trends can be seen in figure 3c. Many histories show that entropy production is initially dominated by cooling before one or more elements begin to exsolve from the core and become the dominant source of entropy. Ove time, cooling produces a steadily greater fraction of total entropy until it contributes 30-75% of total entropy just before inner-core nucleation. Finally, after inner-core nucleation takes over, it produces ~15% of total entropy at the present day.

Inner-Core Nucleation

We use our suite of histories to examine the timinig of inner core nucleation and the change in entropy production with inner-core nucleation. In figure 4c we plot a histogram of inner-core nucleation timing, showing that the median inner-core nucleation is ~550 Mya in our model, with nucleation ages varying from ~350Mya to ~780Mya. These results are very similar to those of Nimmo (2015), as might be expected since we use a modified form of his core parameterization.

Our model always shows an increase in available entropy with inner-core nucleation, with a median increase of ~100% from the entropy available 1Myr before to 100Myr after nucleation. We show in figure 3b the histogram of entropy changes ranging from ~5% to ~500%. The largest entropy increases arise due to histories with very low available entropy before inner-core nucleation. Most

entropy chanages in our model are much higher than other recent estimates of the change in availble entropy with inner-core nucleation (e.g. Aubert 2017, CITE), but are again consistent with Nimmo (2015). A doubling of availble entropy might indicate that an increase in field stregth should be able to be observed in the paleomagnetic record (CITE Driscol dynamo scaling). However, available entropy might not correspond directly to a change in observable field strength. Inner-core nucleation changes the location and mechanism of the dominant source of entropy production, which could change both the scaling of field strength with etropy and the mophology of the generated field (CITE Nick's paper).

Conclusions

FeO exsolution is an important source of entropy to power Earth's dynamo. Exsolution of light elements from Earth's core is governed by the composition of the lowermost mantle. TODO