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# Heat partitioning in metal-silicate plumes during Earth differentiation

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ABSTRACT

We present an analytical model for core formation that includes both metal diapirs and liquid conduits and produces a superheated core. We start from the standard model of Earth formation consisting of accretion by planetesimal or planetary embryo impacts large enough to create global or local magma oceans. Our model consists of two main fluid dynamical structures: 1) large liquid metal diapirs and 2) narrow conduits with liquid silicate and small liquid metal droplets. We assume that transport of metal from a near-surface magma ocean to the growing core occurs within these structures. The release of gravitational potential energy from the descent of the metal is converted to heat through viscous dissipation, which is broadly distributed around the large liquid metal diapirs but localized in the conduit around the smaller metal droplets. This model gives an average core temperature during accretion and a post-accretion average core temperature for a broad range of impact histories and chemical quilibration assumptions. We consider three types of Earth accretion histories with impacts that are evenly spaced over 30 Ma. These include: 1) equal-sized impacts, 2) equal-sized impacts with a final Mars-sized impact, and 3) exponentially decreasing impact sizes with a final Mars-sized impact. Our preferred model is based on core–mantle chemical equilibrium estimates that correspond to a fractionation, f, the ratio of metal in the diapir to the total metal content in each impact of $f=0.64$ and accretion history type 3. This model produces a post-accretion average core temperature that ranges from 5930 K to 5700 K for 8 to 90 impacts, respectively. The efficiency of the metal-silicate conduits, defined as the ratio of the heating of the core by the conduits to their potential energy released has an approximate value of 0.59 for our preferred model. The residual post-accretion structures include a basal magma ocean and a network of mantle conduits that may dictate the style of Hadean tectonics.

## 1. Introduction

The abundances present today in the mantle of short-lived decay products such as $^{27}$Mg and $^{182}$W relative to non-radiogenic isotopes suggest that Earth differentiated during accretion over 30–40 Ma time, with most of its mass accreting within about 10 Ma and fully accreted by 100 Ma (Halliday, 2008; Halliday and Wood, 2007; Walter and Tronnes, 2004; Wood et al., 2006). Based on SPH simulations, a large Mars-sized impact is thought to have occurred at the end of the accretion of the Earth resulting in the moon formation (Canup, 2004), a key process in this study. Although it is widely thought that core formation took place during accretion, rather than later on, there are relatively few constraints on how the core formed.

Consideration of the physical processes by which the Earth accreted offers one approach to this question. The current paradigm for Earth accretion emphasizes the importance of large impact events (Canup, 2004; Halliday, 2008; Rudge et al., 2010; Tonks and Melosh, 1993; Wood et al., 2006). It is likely that many of these impact events involved massive, already differentiated projectiles that released enough kinetic and potential energy to produce substantial melting of both the projectile and the target, creating localized and possibly global magma oceans (Albarede, 2009; Corgne et al., 2008; Halliday, 2008; Pahlevan and Stevenson, 2007; Rudge et al., 2010; Tonks and Melosh, 1993; Walter and Tronnes, 2004). Parts of the already existing planetesimal proto-cores become fragmented during impact while other portions of the proto-core stay coherent and sink to the bottom of the magma ocean quickly. Once a large mass of metal reaches the bottom of the magma ocean it sinks through the solid mantle in the form of a large liquid diapir.

In a magma ocean environment, both metal-silicate segregation and equilibration processes occur on much shorter time scales than normally found under sub solidus conditions (Halliday and Wood, 2007). In particular, the processes of chemical and thermal equilibration between metal that is disseminated by impacts and silicates are greatly accelerated if the metal phase is finely dispersed, on the scale of droplets for example (Höink et al., 2006). Theory (Dahl and Stevenson, 2010) predicts enormous Reynolds numbers and highly turbulent conditions during the initial decent of the dispersed metal in a magma ocean, and laboratory experiments predict partial although incomplete equilibration between the silicate and metal components (Righter et al., 1997). Heuristic considerations (Rubie et al., 2003) and calculations (Sasaki and Abe, 2007) indicate that large metal masses are likely to be unstable in a

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magma ocean and will tend to break up, forming smaller droplets during free fall. Once the metal free fall phase is over, further approach to chemical equilibrium becomes more problematic. A key unanswered question for chemical equilibration is what fraction of the metal becomes dispersed into small droplets and what fraction remains in the form of larger masses. We assume the equilibration process occurs solely in the magma ocean and therefore only the dispersed metal droplets and not the large metal masses become equilibrated. This is clearly an over simplification, but allows us to apply geochemical constraints to our model. We further assume that dispersed metal droplets gradually rain to the bottom of the magma ocean, entering the conduits that trail the diapirs formed by the larger masses of the remnant impactor proto-cores (Karato and Murthy, 1997; Ricard et al., 2009; Ziethe and Spohn, 2007).

Although it is possible that the entire silicate mantle could have been molten at times (Solomatov, 2000; Tonks and Melosh, 1993), evidence suggests it is unlikely that the metal now in the core equilibrated within a totally molten mantle. First, the solidification time scales of a totally molten mantle are likely very short. According to thermal history calculations, deep magma oceans would solidify in $10^3 - 10^5$ years, whereas a shallow magma ocean could persist in the upper mantle up to $10^8$ years, depending on the radiative thickness of its atmosphere (Solomatov, 2000). Second, the abundances of moderately siderophile elements in the present day mantle suggest equilibration pressures and temperatures that place the base of a magma ocean at mid-mantle or upper-mantle depths. Ni, Co, V, Cr, Si and Nb are all found in higher abundances in the mantle than predicted for near-surface equilibration, because these elements strongly partition into metal at low pressure (Rubie et al., 2007). Specifically equilibration between Fe–Ni–S-alloy and liquid silicate at present day depths near 1200 km or pressure lower than 30 GPa explains the Ni and Co abundances in the mantle today (Righter and Drake, 2001; Wood et al., 2006), and mantle V, Cr, Si, and Nb abundances constrain the temperature of equilibration to a temperature above 2400 K and below 4200 K (Li and Agee, 2001).

In addition to temperature and depth, siderophile element abundances also constrain how much chemical equilibration the metal and silicate experienced prior to final core segregation. According to Rudge et al. (2010), about 36% of the metal equilibrated with the silicate mantle prior to the end of core formation, and according to Sasaki and Abe (2007), as much as 20% equilibrated during each large impact. These percentages could be interpreted as the mass fraction of metal dispersed into droplets by impact and mixing processes. What was the form of the remaining metal? Fluid dynamical models of core formation have shown that coherent masses of liquid metal tend to form nearly or partially spherical diapirs, through Rayleigh–Taylor instabilities and other mechanisms (Monteux et al., 2009; Olson and Weeraratne, 2008; Sasaki and Abe, 2007; Walter and Tronnes, 2004). There is limited possibility for such large metal diapirs to equilibrate in the deep mantle. Diapirs can increase the temperature in the surrounding silicate by viscous dissipation, possibly enough for melting, which would increase the chemical diffusivity and somewhat accelerate chemical equilibration (Samuel and Tackley, 2008). However, the degree of chemical equilibration has been shown to be minimal in large diapirs (Samuel et al., 2010) and is expected to be negligible in comparison with the equilibration between metal droplets and silicate melt in a magma ocean.

The thermal state of the present day Earth in the core–mantle boundary (CMB) region offers constraints on the degree of thermal equilibration between core-forming metals and mantle silicates during core formation. Today there is an intrinsic temperature difference of 1000–2000 K across the D" layer separating the adiabatic portion of the lower mantle from the outer core (Anderson, 2002; Buffett, 2003). Adiabatic extrapolation of the 4100 K temperature estimate for the outer core at the CMB to surface pressure indicates the core now has +1000 K of superheat (potential temperature relative to surface melting temperature), whereas the present day mantle has far less superheat (Nimmo, 2007). In other words, the present day thermal disequilibrium of the core and mantle is about as great as their chemical disequilibrium was during Earth formation, in spite of the fact that chemical diffusion is far slower than thermal diffusion. The present day superheat of the core could be a consequence of heat sources in the core. Possible heat sources for superheating the core after its formation include radioactive decay of $^{40}$K, ohmic and viscous heating. However, none of these is large enough to compensate for the secular cooling of the core over time. More plausibly, core superheat is a relic of the core formation process (Golabek et al., 2009; Ke and Solomatov, 2009). Energy sources available to heat the core during its formation include shock wave heating derived from the kinetic energy of impacts (Arkani-Hamed and Olson, 2010) and the change in potential energy to thermal energy via viscous dissipation as metal descends through the mantle (Monteux et al., 2009).

There are many proposed mechanisms for core formation, but only some of these preferentially heat the core. Metal percolation is possible in planetesimals and under high pressures due to small dihedral angles (Shannon and Agee, 1998; Yoshino et al., 2003), however, this mechanism does not increase the core's temperature preferentially, as the heat produced by percolation is dispersed throughout the mantle. Metal from the projectile could also directly merge with the proto-core if the projectiles were large enough to produce a magma ocean that extended to the CMB (Tonks and Melosh, 1993). Merging of cores through an entirely molten mantle is not considered here because the heat absorbed by merging cores is hard to quantify, the likelihood that the entire Earth was repeatedly liquified by impacts seems remote, and finally, this mechanism does not allow for any post-accretion equilibration of metal and silicate.

In this paper we argue that the same dynamical process of core formation that controls core–mantle chemical equilibrium also controls the core superheat. Specifically, we propose a segregation process that includes both metal diapirs and metal droplets in metal-silicate plumes allowing for partial chemical equilibration and producing a superheated post-accretion core. Consistent with the standard accretion model, we assume that large impacts create transient magma oceans on the growing Earth. Some of the metal from each impacting projectile sinks to the bottom of the magma ocean and through the solid mantle in large liquid diapirs, while the remainder is dispersed in smaller masses, assumed here to be droplets. The relative proportion of large metal masses to small metal droplets is a model parameter and is assumed to have the same value for all impacts.

Previous studies come to different conclusions about the heat supplied to the core by large diapirs (Monteux et al., 2009; Samuel et al., 2010), but the general consensus is that much of their viscous heating remains in the mantle due to the large viscosity difference between the liquid metal diapir and solid silicate mantle. Accordingly, the diapirs in this study are assumed to experience only adiabatic increases in temperature during their decent. We further assume that the dispersed metal descends from the magma ocean to the growing core within liquid metal-silicate conduits, consisting of superheated silicate melt and superheated metal droplets. Superheating in the conduits occurs as the potential energy of the droplets is converted to heat by viscous dissipation, which is localized within the narrow conduits. Only a portion of this localized viscous heating is lost by conduction into the solid silicate mantle, so the metal droplets are superheated as they enter the growing core. A similar model was proposed by Ke and Solomatov (2009) through which a low-viscosity iron diapir mostly dissipating heat in the mantle, whereas iron sinking through the mantle in a channel preferentially heats the iron. Like Ke and Solomatov (2009) we find that heat partitioning depends on the structure of the descending iron, and we also consider heat and mass exchange between metal-silicate conduit and the mantle.