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A re-evaluation of metal diapir breakup and equilibration in terrestrial magma oceans

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ABSTRACT

Due to mechanisms such as impact heating, early atmospheric thermal blanketing, and radioactive heating, the presence of at least one global magma ocean stage in the early histories of terrestrial planets seems unavoidable. In such a context, a key question to constrain the early thermo–chemical evolution of the Earth is how much iron diapirs provided by differentiated impactors emulsified during their sinking towards the bottom of an early magma ocean.

In the past years, several workers have focused on this question, using however various approaches and making different assumptions. While most studies favor rapid breakup and equilibration of iron bodies during their sinking through the magma ocean, recent work suggests that iron bodies of size comparable or greater than a few tens of kilometers may preserve most of their initial volume as they reach the bottom of a magma ocean, therefore leading to metal–silicate disequilibrium.

To clarify the discrepancies and the differences among studies I have conducted a series of numerical simulations and theoretical calculations to derive the conditions and the timing for the breakup of metal diapirs of any size, sinking through a silicate magma ocean, with a large range of plausible viscosity values. The obtained breakup criterion is used to derive stable diapir sizes and their ability to equilibrate with the surrounding silicates. I show that for plausible magma ocean viscosities, diapirs with initial radii smaller than the thickness of a magma ocean rapidly break up into stable diapir sizes smaller than 0.2 m, at which metal–silicate equilibration is rapidly achieved.

1. Introduction

The earliest stages of planetary evolution are punctuated by a great variety of events that have reshaped the Earth and other terrestrial planets and influenced their long term evolution. During the accretional stages of planetary evolution, planetesimals driven by gravitational interactions aggregate rapidly into growing planetary embryos (Kokubo and Ida, 1996; Kortenkamp et al., 2000; Weidenschilling, 1976). Throughout the duration of the accretional growth period, physical processes, such as radioactive decay, impact heating, and possibly thermal blanketing, significantly contribute to increasing the protoplanet's temperature. Once the radius of the growing terrestrial planet reaches a critical size of about 1000 km, the kinetic energy provided by incoming planetesimals becomes large enough to trigger local melting (Coradini et al., 1983; Davies, 1985; Safronov, 1978; Sasaki and Nakazawa, 1986; Senshu et al., 2002; Tonks and Melosh, 1992). As inferred from N-body simulations, during the late stages of planetary accretion (oligarchic growth), impacts become more energetic (Kokubo and Ida, 1996), leading to more frequent melting events of greater extents. Moreover, the latest stages of planetary accretion are also characterized by the occurrence of giant impacts (Benz et al., 1986; Canup, 2004), followed by isostatic readjustment, which are expected to trigger global scale melting as well as possible vaporization on terrestrial planets (Tonks and Melosh, 1993). The presence of a steamed impact-heated atmosphere would also favor the occurrence of large scale melting events (Abe, 1997). Therefore, with such an array of mechanisms favoring high temperatures, the occurrence of at least one global scale melting event during the early stages of terrestrial planet formation seems unavoidable (Tonks and Melosh, 1993).

In addition, radioactive heating produced by the disintegration of ^{26}Al and ^{60}Fe could also yield super-solidus temperatures, even in bodies of modest sizes such as planetesimals (Merk et al., 2002; Walter and Trϕnnes, 2004). Therefore, incoming impactors may have already been differentiated, with a small iron core surrounded by silicate material.

In such a context a key question is how much iron diapirs provided by differentiated impactors have emulsified during their sinking towards the bottom of an early magma ocean. Addressing this problem allows one to put strong constraints on metal–silicate equilibration processes (Dahl and Stevenson, 2010; Karato and Murthy, 1997; Rubie et al., 2003; Rubie et al., 2007; Wood et al., 2006). This is of prime importance for the interpretation of cosmochemical data such as Hf/W and U/Pb chronometers, allowing to put bounds on the timing of the accretion and differentiation of the Earth and other terrestrial bodies (Dahl and Stevenson, 2010; Kleine and Rudge, 2011; Kleine et al., 2004a; Kleine et al., 2004b; Rudge et al., 2010). In addition, the size of iron bodies sinking through the solid or molten silicate proto-mantle determines the heat distribution within a young terrestrial planet (Ichikawa et al., 2010; King and

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Olson, 2011; Lin et al., 2011; Monteux et al., 2009; Monteux et al., 2011; Ricard et al., 2009; Samuel et al., 2010; Senshu et al., 2002), which influences the subsequent long-term planetary thermal evolution.

Until recently, it was thought that even large iron diapirs sinking through a silicate magma ocean would rapidly break up into centimeter or millimeter-sized droplets (Ichikawa et al., 2010; Rubie et al., 2003; Stevenson, 1990). Such small droplets would allow metal-silicate equilibration to occur within sinking distances of just a few tens of meters (Karato and Murthy, 1997; Rubie et al., 2003). However, this scenario was recently questioned by (Dahl and Stevenson, 2010), who derived theoretical models to account for the turbulent erosion of iron diapirs via Kevin–Helmholtz and Rayleigh–Taylor mechanisms. The main result of their study is that large diapirs (*i.e.*, of radius >10 km) can survive complete erosion and preserve most of their initial volume as they reach the bottom of a ~1000 km thick magma ocean, therefore leading to metal–silicate disequilibrium.

A comparison between these different studies is not necessarily straightforward because the underlying assumptions, the governing parameters, their ranges of variation, and to some extent the geometry and modeling approaches may be very different. For instance in Rubie et al. (2003) the authors derive a conceptual/parameterized model where the criterion used to determine the diapir stable sizes was based only on the value of the Webber number, implying that viscous effects are negligible. In Ichikawa et al. (2010) surface tension and viscous effects are taken into account, however, due to computational limitations of numerical experiments, the explored range of Reynolds numbers was restricted to low values and the effect of surface tension forces on diapir breakup and stable sizes was not investigated systematically, since the Webber number was fixed to a constant value close to unity. In addition, most of their exploration was performed in a 2D Cartesian geometry, also due to limitations imposed by computational run time. On the other hand, using theoretical turbulent modeling Dahl and Stevenson (2010) have focused only on diapirs where Reynolds and Webber numbers are both very large. Although they account for moderate diapir deformation, they made the assumption that iron diapirs do not breakup during their descent through the silicate magma ocean. In a recent study, Deguen et al. (2011) have investigated experimentally the mixing of dense bodies during their sinking. They find that large bodies are not fully mixed by turbulent instabilities, which confirms qualitatively the theoretical results of Dahl and Stevenson (2010). However, in their work sinking iron bodies are modeled as a cloud of dense particles/flakes instead of a continuous body. While collective behavior within particle clouds can occur at high Reynolds numbers, it is not clear whether the dynamics and the fragmentation processes are identical for particle clouds and for initially continuous bodies.

To clarify these discrepancies, I have conducted a series of numerical simulations and theoretical calculations to derive the conditions and the timing for the breakup of axisymmetric metal diapirs of any size, sinking through a silicate magma ocean with a large range of plausible viscosity values. The corresponding range of governing parameters covers more than 16 orders of magnitude. The obtained breakup criterion is used to derive stable diapir sizes and their ability to equilibrate chemically with the surrounding silicates.

The paper is organized as follows: the next section, Section 2 introduces the fluid dynamic problem, the corresponding governing parameters and their plausible ranges. Section 3 presents a simple analytical model to derive the general kinematics of iron diapirs sinking through a magma ocean in the absence of diapir breakup. This is followed by Section 4 devoted to the dynamics of diapir fragmentation, where general criteria for diapir breakup and the corresponding timing for breakup is derived using numerical experiments, scaling analysis and analytical theory. In the last section, Section 5, preceding the conclusion, the fluid dynamics results are applied to evaluate the ability of iron diapirs to equilibrate with the surrounding silicate magma ocean during their descent.

2. Governing parameters, parameter ranges and dynamic regimes

A main focus of this study is to constrain the size of iron bodies of density ρ_m sinking through a liquid silicate magma ocean of density ρ_s and viscosity η at a given velocity v, as sketched in Fig. 1. The iron diapirs are assumed to be initially spherical with a radius R_0, and are subject to surface tension forces acting at the interface to preserve a constant curvature κ. In principle, one should expect the diapir viscosity to be smaller than the surrounding silicate, however it is assumed here that there are no viscosity differences between the diapir and the silicates. This simplification should not affect the results and conclusions significantly.

With such a configuration the dynamics is entirely governed by two dimensionless numbers. The Reynolds number, which expresses the importance of inertia over viscous effects:

$$Re = \frac{\rho_s v R_0}{\eta}, \tag{1}$$

and the Webber number, which measures the importance of inertia over surface tension forces acting on the diapir surface:

$$We = \frac{\rho_s v^2 R_0}{\sigma}, \tag{2}$$

where σ is the coefficient of surface tension.

The value of Re defines three dynamical regimes for the sinking diapir (*e.g.*, (Crowe et al., 1997; Lamb, 1932) and references therein):

1. $Re<1$ corresponds to the Stokes regime of creeping flow where inertia is negligible.
2. $Re=1-500$ corresponds to the intermediate regime where the influences of both inertial and viscous effects are important.
3. For larger values of $Re(>500)$ viscous forces are negligible compared to inertia. This is called the Newton regime. Within this regime the flow around the diapir transits from laminar to turbulent at about $Re \sim 10^5$ (Fig. 2).

In the Stokes regime the diapir instantaneously reaches its terminal velocity v_s (regardless of the value of its initial velocity v_0), which is for a sphere:

$$v_s = \beta \frac{(\rho_m - \rho_s) g R_0^2}{\eta}, \tag{3}$$

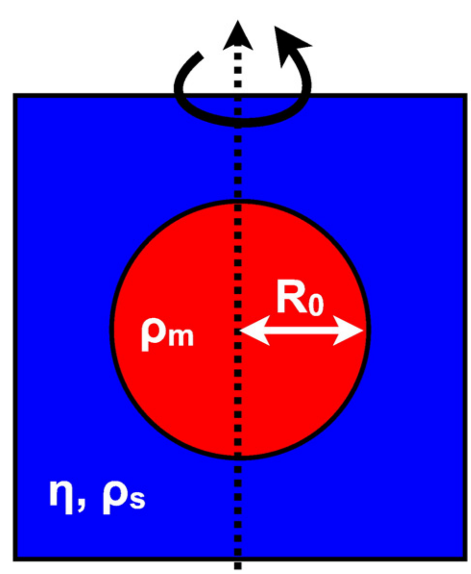

Fig. 1. Schematic representation of the problem. See text for further details.