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Experiments on fragmentation and thermo-chemical exchanges during planetary core formation

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ABSTRACT

The initial thermo-chemical state of telluric planets was largely controlled by mixing following the collision of differentiated proto-planets. Up to now, most models of planet formation simply assume that the iron core of the impactors immediately broke up to form an "iron rain" within a large-scale magma ocean, leading to the rapid equilibration of the whole metal with the whole mantle. Only recent studies have focused on resolving the fluid mechanics of the problem, with the aim to define more relevant diffusion–advection models of thermal and chemical exchanges within and between the two fluids. Furthermore, the influence of the viscosity ratio on this dynamical process is generally neglected, whilst it is known to play a role in the breakup of the initial iron diapirs and in the shape of the resulting droplets. Here we report the results of analog laboratory experiments matching the dynamical regime of the geophysical configuration. High speed video recording allows us to describe and characterize the fluid dynamics of the system, and temperature measurements allow us to quantify the diffusive exchanges integrated during the fall of the liquid metal. We find that the early representation of this flow as an iron rain is far from the experimental results. The equilibration coefficient at a given depth depends both on the initial size of the metal diapir and on the viscosity of the ambient fluid, whereas the falling speed is only controlled by the initial size. Various scalings for the diffusive exchanges coming from the literature are tested. We find good agreement with the turbulent thermal model developed by Deguen et al. (2014).

1. Introduction

The aggregation time scale of moon-size planets in a protoplanetary disk is on the order of 10–100 Myr (Chambers, 2004; O'brien et al., 2006). It is comparable to the typical time scale of differentiation of terrestrial bodies inferred from meteorite analysis (Lee and Halliday, 1996; Kleine et al., 2004) and numerical simulations (Neumann et al., 2012). This implies that Earth and other terrestrial bodies most likely formed from the collision of already differentiated proto-planets (Yoshino et al., 2003). These events were by themselves energetic enough to cause a local melting of the mantle (Safronov, 1978; Kaula, 1979; Reese and Solomatov, 2006; Monteux et al., 2007). In addition, there was enough ^{26}Al for the decay heat to elevate the mantle's temperature above solidus across depths of several hundreds of kilometers (Merk et al., 2002), and thus produce a deep global magma ocean (Tonks and Melosh, 1992).

After the impact, the liquid iron from the impactors' core sank through the molten silicate mantle allowing efficient equilibration, the details of which being a matter of fluid dynamics. Once formed, the core and mantle of terrestrial planets are weakly coupled regarding heat and element exchanges, because the surface of exchange is relatively small and because transfers can only take place through diffusion and very slow advection in the solid mantle. Consequently, the internal dynamics of these reservoirs strongly depend on their initial thermochemical states (Samuel et al., 2010; King and Olson, 2011). Also, our understanding of the event chronology taking place during planetary accretion relies on radio-nucleid: loose constraints on metal/silicate partitioning at the very end of accretion don't allow to discriminate between a wide range of possible scenarii (Kleine et al., 2004; Allègre et al., 2008).

Past studies provide us with a partial story on the flow occurring after the impact. Provided that the initial mass of liquid metal can be seen as a unique source referred to as a "diapir", it falls as a turbulent thermal (i.e. an isolated buoyant mass of fluid, in which gravitational potential energy is converted to turbulent motion, causing mixing: see Deguen et al., 2011), breaks-up within a depth of a few diapir's initial radii to capillary-sized droplets (Ichikawa et al., 2010; Samuel, 2012; Wacheul et al., 2014), except for the largest diapirs (Dahl and Stevenson, 2010) or if planetesimals cores

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undergo massive mixing during their impact (Kendall and Melosh, 2016). Turbulent thermals are dominated by collective effects between droplets, until they transition to an "iron rain" regime when diluted enough (Bush et al., 2003; Deguen et al., 2011). The size and speed of droplets in this last regime allow droplets to fully equilibrate with the ambient silicate well before reaching the bottom of the magma ocean (Stevenson, 1990; Rubie et al., 2003; Ichikawa et al., 2010; Ulvrová et al., 2011; Samuel, 2012). However, the typical length scale required for significant equilibration is very important because it sets the depth, *i.e.* the pressure of equilibration between the metal and silicate. Recent studies (Deguen et al., 2014) suggest that the smallest scales of the induced turbulence, even before fragmentation, could drastically reduce this equilibration length scale. It is then crucial to precisely understand how the fluid mechanics of this flow governs diffusion.

We report in the present article the results of laboratory experiments on a fluid system analog to the liquid silicate/ liquid iron system, where we performed measurements of both the fluid dynamics and the thermal diffusive exchanges over the depth of our experimental domain. The set-up and relevant parameters are described in Sections 2 and 3 respectively. The fluid dynamics of sedimentation and fragmentation are addressed in Section 4. Equilibration results are analysed in Section 5 in comparison with various models from the literature. Finally, conclusions and future works are shortly reviewed in Section 6.

2. Experimental set-up

As shown in Fig. 1, our set-up consists in a 45 × 45 × 100 cm polycarbonate tank filled with a mixture of water and UCON™ oil of density 1050 kg·m^{-3}, that is used here to mimic the molten silicate. Its viscosity can be adjusted over 5 orders of magnitude depending on the mass fraction of water, in order to reproduce the range of viscosity ratios of the liquid iron/silicate magma system at low pressure (Rubie et al., 2003). To mimic the impacting iron core, we use a balloon filled with galinstan, a gallium alloy that is liquid at room temperature. Its surface energy is 0.718 J·m^{-2}, and its density 6440 kg·m^{-3}. Before each experimental run, the mixture water/UCON™ oil is vigorously mixed in order to homogenize the temperature. A balloon is filled with a given mass of galinstan and precisely weighted. A thermocouple is inserted in the balloon before closing it. The balloon is then placed in a hot container in order to heat the galinstan. The viscosity of the mixture water/UCON™ decreases with temperature, therefore the temperature of the galinstan is elevated only by about ten degrees Celsius to avoid lubrication due to higher temperature in the thermal boundary layer. Eventually, the balloon is fixed at the top of the tank where it is immersed in the mixture water/UCON™ oil and popped immediately. The balloon takes only a few thousandths of seconds to retract and the galinstan then starts falling through the viscous fluid. At the bottom of the tank (70 cm below the surface) sits a square funnel made of polystyrene collecting the liquid metal and channeling it towards an adiabatic chamber. When the galinstan is collected, its residual speed mixes completely the temperature and the whole mass of metal reaches an almost constant mean temperature in a few seconds. A thermocouple is placed a few millimeters above the bottom of the adiabatic chamber in order to measure this final temperature. In addition, a third thermocouple is fixed in the middle of the tank one centimeter away from the wall in order to measure the temperature in the bulk of the ambient fluid.

For each experimental run, two events are clearly identifiable in the time series of the thermocouples, as seen in Fig. 2: the release of the galinstan and the moment it reaches the adiabatic chamber at the bottom of the tank. From theses signals we compute the temperature drop of the galinstan with respect to ambient fluid during its fall. Each run of the experiment is also recorded by a high speed camera PHOTRON Fastcam at 1000 frames per second with a resolution of 800 × 1240 pixels.

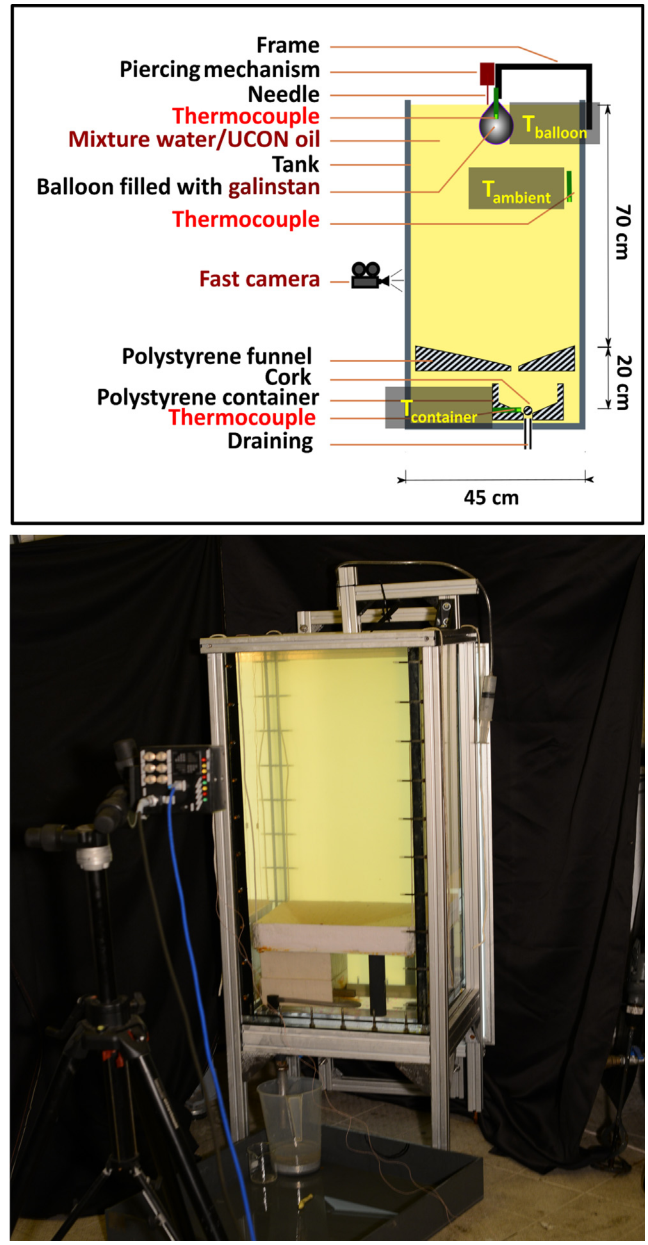

Fig. 1. Schematic and picture of the experimental set-up.

3. Parameters and non dimensional numbers

It is still unclear how the initial size of impacting planetesimal converts into the typical size of liquid iron structures at the surface of the magma ocean just after the impact (Canup, 2004). However, it is very likely that this typical size depends on the angle of the impact, since stretching and dilution of the core become very important when the impact is far from the vertical direction (Kendall and Melosh, 2016). In our approach, we consider this typical size as a free parameter, and correspondingly we vary the initial radius of our balloon of galinstan. The other free parameter is the viscosity of the ambient fluid, because the magma viscosity is

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