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Review article

Density of magmas at depth

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ABSTRACT

Knowing the density of silicate liquids at high pressure is essential to answer questions relevant to the presence of magmas at depth, whether that be in the present Earth or in its earliest times, during differentiation of the planet. Melts have unique physical and chemical properties, which vary as a function pressure, and chemical composition. The focus here will be on in situ measurements of the density of magmas, with a presentation of the available methods and of the main results obtained so far, including why some magmas may be trapped at depth. Understanding the macroscopical physical properties of magmas requires an accurate microscopic structural description. Structural descriptions of compressed magmas are becoming more widely available, from experiments and from theoretical calculations. These structural inputs are used to understand the compression mechanisms at stake in the densification of magmas, e.g. the collapse of voids, coordination increase for the major cations, and bond compressibility. These densification processes profoundly affect not only the physical properties of the melt, but also its chemical properties, i.e. the way element partition between the magma and a metallic melt or between the magma and crystals.

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1. Introduction

Magmas are essentially high pressure (*P*) objects, being formed at depth. Magmas that erupt on the surface were formed mostly below mid-oceanic ridges between 30 and 50 km depths (Lee et al., 2009), between 100 and 150 km depths at subduction ones (Syracuse and Abers, 2006; Schmidt and Poli, 2014), and down to 220 km for komatiite melts (Lee et al., 2009). In addition, seismology has probed discrete zones at depth where the presence of magmas is suspected: at the lithosphere–asthenosphere boundary (Schmerr, 2012), atop the 410 km (Tauzin et al., 2010) and 660 km (Schmandt et al., 2014) discontinuities, and in the ultra-low velocity zones (ULVZ) atop the core–mantle boundary (Williams and Garnero, 1996). The discrete depths at which magmas are detected raise the question of why they are trapped and/or produced only there. For the magmas that reach the surface, knowing their density vs *P* is essential to model their

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behaviour from segregation from the parent rock and consequent control on melt composition, to their ascent to the surface (Stolper et al., 1981). Magmas are ubiquitous objects in the solar system, both currently on the surface of terrestrial planets and some satellites (e.g. Io, the Moon), and in the past as planets were born molten (Harper, 1996). The structure of planets as we know them now has thus been inherited from this molten era, the magma ocean stage, although the extent in space and time of the magma ocean is a matter of debate (see (Carlson et al., 2014; Elkins-Tanton, 2012) for reviews on the topic). Matter and heat were transferred through the magma ocean either to the surface or to the core of planets, density exerting a first order control on these differentiation processes.

The compressibility of magmas has been measured extensively at ambient pressure using ultrasonic techniques (cf. (Rivers and Carmichael, 1987; Lange and Carmichael, 1987; Ai and Lange, 2008) and references therein), resulting in a comprehensive data-set. Pioneering high *P* density measurements on magmas was obtained with shock-wave techniques up to 23 GPa (Rigden et al., 1984), verifying the prediction of a density cross-over between melts and co-existing crystals (Stolper et al., 1981). The existence of melts/crystals cross-over(s) implies that magmas may be trapped at depth, eventually leading to the formation of hidden reservoirs enriched in incompatible elements. This major result has been confirmed by static high *P*–*T* sink–float measurements (Agee, 1998). The sink–float method has been successfully applied to different basic and ultrabasic melts (see (Ohtani, 2009) for a review of the method). The highest achievable pressures with this method have been reached using diamond markers (Suzuki et al., 1995; Sakamaki et al., 2006) and have revealed that diamond may float in the transition zone (Suzuki et al., 1995). The sink–float method is restricted to low viscosity melts in order for the sphere to fall, i.e. basic and ultrabasic melts. Different markers may be used: olivine, garnet and diamond, so that up to 4 to 5 points can be collected along the equation of state up to a maximum of 24 GPa. Based on these high *P* data and the ultrasonic ambient *P* data, thermodynamic models have been built to calculate the density of any magma under *P* from the partial molar volumes of the constituent oxide components (Ghiorso, 2004).

With the advancement of X-ray synchrotron sources at the beginning of the century, measuring the properties of compressed non-crystalline materials such as melts became possible using different methods presented below. A main advantage of in situ methods is the unequivocal melting criterium with the concomitant disappearance of crystalline Bragg peaks and the appearance of a diffuse scattered signal on X-ray diffraction data. These techniques thus avoid relying on the interpretation of quench textures to assess melting that can be ambiguous for ultrabasic compositions that do not quench as glasses. Additionally, in situ methods may combine several approaches within the same experimental set-up, such as viscosity, ultrasonic, and structural measurements (Kono et al., 2014).

The outline of this paper is to first present the different in situ X-ray synchrotron based methods to measure the density of magmas at high *P*, their *P*–*T* domains of application, and their pros and cons compared to other methods. Next, the compression mechanisms at stake in magmas at depth will be discussed as a function of the pressure range at which they occur. This discussion will be closely based on the structural evolution of silicate melts, with inputs from both experiments and theory. Lastly, planetary implications will be discussed. Densification of magmas may imply their segregation at depth for some particular compositions and in some particular planetary contexts. Densification may also affect the way element partition between crystals and silicate melts, or between metallic and silicate melts.

2. How to measure the density of magmas at depth

This section highlights the pros and cons of the different in situ methods that have been developed to measure the density of magmas at high pressures, along with their domains of application in terms of pressure, temperature and chemical compositions.

2.1. In situ X-ray absorption

The X-ray absorption was initially developed to measure the density of relatively heavy metallic melts such as indium using large volume presses (Katayama et al., 1996, 1998), but was successfully applied later to liquid phosphorous (Katayama et al., 2004). The X-ray absorption method (Fig. 1) consists in measuring the intensity of the X-ray beam that is absorbed by the sample while scanning the press perpendicularly to the X-ray beam. The beam intensity is recorded using photodiodes or ionization chambers, typically within 20–30 min to have an adequate signal vs noise ratio. Each absorption scan is simulated with Relation (1) leading to values of $(\mu\rho)_{liq}$.

$$\frac{I}{I_o} = \int_{x,y} \exp\left(-(\mu\rho d)_{liq} - (\mu\rho d)_{env}\right) dxdy \qquad (1)$$

where μ is the mass absorption coefficient, ρ the density, d the X-ray path length and y the vertical size of the X-ray beam (typically 50 μm width × 50 μm height). The subscript *env* stands for the environment surrounding the sample. The sample absorption coefficient, $\mu(E)$ (E is the energy of the X-ray beam), must be either known or ideally measured for instance on the crystalline sample which density can be measured by in situ X-ray diffraction. This method allows to cover a fine *P*–*T* mesh, it is not restrictive in terms of chemical compositions,

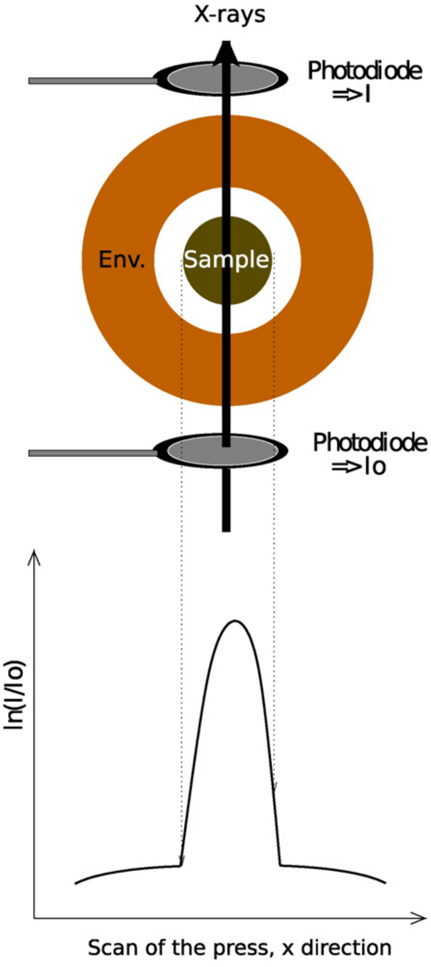

Fig. 1. Schematic of the X-ray absorption method.