



Reconciling laboratory and observational models of mantle rheology in geodynamic modelling



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ABSTRACT

Experimental and geophysical observations constraining mantle rheology are reviewed with an emphasis on their impact on mantle geodynamic modelling. For olivine, the most studied and best-constrained mantle mineral, the tradeoffs associated with the uncertainties in the activation energy, activation volume, grain-size and water content allow the construction of upper mantle rheology models ranging from nearly uniform with depth to linearly increasing from the base of the lithosphere to the top of the transition zone. Radial rheology models derived from geophysical observations allow for either a weak upper mantle or a weak transition zone. Experimental constraints show that wadsleyite and ringwoodite are stronger than olivine at the top of the transition zone; however the uncertainty in the concentration of water in the transition zone precludes ruling out a weak transition zone. Both observational and experimental constraints allow for strong or weak slabs and the most promising constraints on slab rheology may come from comparing inferred slab geometry from seismic tomography with systematic studies of slab morphology from dynamic models. Experimental constraints on perovskite and ferropericlase strength are consistent with general feature of rheology models derived from geophysical observations and suggest that the increase in viscosity through the top of the upper mantle could be due to the increase in the strength of ferropericlase from 20–65 GPa. The decrease in viscosity in the bottom half of the lower mantle could be the result of approaching the melting temperature of perovskite. Both lines of research are consistent with a high-viscosity lithosphere, a low viscosity either in the upper mantle or transition zone, and high viscosity in the lower mantle, increasing through the upper half of the lower mantle and decreasing in the bottom half of the lower mantle, with a low viscosity above the core. Significant regions of the mantle, including high-stress regions of the lower mantle, may be in the dislocation creep (power-law) regime. Due to our limited knowledge of mantle grain size, the best hope to resolve the question of whether a region is in diffusion creep (Newtonian rheology) or dislocation or grain-boundary creep (power-law rheology), may be the presence or absence of seismic anisotropy, because there is no mechanism to rotate crystals in diffusion creep which would be necessary to develop anisotropy from lattice preferred orientation. While non-intuitive, the presence or absence of a weak region in the upper mantle has a profound effect on lower mantle flow. With an asthenosphere, the lower mantle organizes into a long-wavelength plan form with one or two (degree 1 or degree 2) large downwellings and updrafts, which may contain a cluster of plumes. The boundary between the long-wavelength lower mantle flow and upper region flow may be deeper, likely 800–1200 km, than the usually assumed base of the transition zone. There are competing hypotheses as to whether this change in flow pattern is caused by a change in rheology, composition, or phase.

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1. Introduction

The rheology of the mantle plays a major role in mantle convection (Davies, 1986; Davies and Richards, 1992; Bercovici, 2007, 2015, and references therein), plate motion (Bercovici, 2003, and

references therein), the evolution of subducted slabs (Billen and Gurnis, 2001, 2003; King, 2007; Liu et al., 2008; Billen, 2008; Becker and Faccenna, 2009; Schellart, 2011; Wada and King, 2015; and references therein), glacial isostatic adjustment (Cathles, 1975; O'Connell, 1976; Peltier and Andrews, 1976; Peltier, 1986; Mitrovica, 1996; Kaufmann and Lambeck, 2000; Tosi et al., 2005; Forte, 2007; Steffen and Wu, 2011, and references therein), the attenuation of seismic waves in the mantle (e.g., Gordon, 1967;

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Karato et al., 2008, and references therein), the rate of surface deformation after earthquakes (e.g., Pollitz et al., 2001; Freed et al., 2012) and plate-scale tectonics (e.g., Davies, 1977; Bunge et al., 1996, 1997; Tackley, 2000; Becker and O'Connell, 2001; Zhong et al., 2000; Lowman et al., 2004; Torsvik et al., 2006; van Heck and Tackley, 2008; Foley and Becker, 2009; Bower et al., 2015; Liu, 2015). Thus, rheological properties of the mantle have a first-order influence on many of the big picture questions in geodynamics including the recycling of subducted slab material, the mixing time of reservoirs within the mantle (e.g., van Keken and Ballentine, 1998), and the thermal evolution of the mantle (e.g., Christensen, 1984; Gurnis, 1989; Korenaga, 2009). As such, our understanding of the origin and evolution of the Earth is limited by our understanding, or lack thereof, of the rheological properties of the Earth's mantle. A detailed review including a historical perspective on the study of rheological properties of the Earth's mantle can be found in Karato (2010). In this review I will focus on the aspects of rheology that are of greatest interest to geodynamic modelling.

The terms viscosity and rheology are often used interchangeably in geodynamics, without undo confusion. Ranalli (1987) defines rheology as “the study of the flow of matter, either in a liquid state or a solid state under conditions in which it responds with plastic flow rather than deforming elastically in response to an applied force.” E.C. Bingham introduced the term rheology around 1920, inspired by the aphorism of Simplicius, *panta rhei*, which translates, “everything flows” (Beris and Giacomin, 2014). As such, the early work on the deformation of the interior of the Earth predates the naming of the field of rheology.

1.1. Theoretical considerations

Like many problems in physics, the quantitative study of deformation in Earth's interior can be traced back to Newton. Newton (1687) calculated the flattening of rotating Earth assuming a constant-density, rotating sphere in hydrostatic equilibrium. Based on the measured values of gravity and the rate of rotation, Newton estimated the flattening of the Earth, the difference between the equatorial, a_e , and polar, a_p , radii, divided by the equatorial radius, $f = (a_e - a_p)/a_e$ was 230. His value is close to the currently accepted value of 1/298.257, which is quite an achievement given the knowledge regarding the Earth's interior at that time. The difference is primarily due to the variation of density with depth within the Earth. An interesting point to note here is that the rheology of the Earth only plays a role in the change in the flattening with time in response to a change in rotation rate (Lambeck, 1980).

1.1.1. Newton's second law

Starting from the continuum-scale theoretical viewpoint, one can derive the equation of motion of a highly viscous fluid by balancing the internal body forces acting on a given volume of material with the surface forces acting on the boundary of the volume of the material. The details of such derivations can be found in numerous sources; here I follow Ricard (2007). The equation balancing body forces acting on a volume of material with surface tractions can be written

$$0 = \nabla \cdot \underline{\underline{\sigma}} + \vec{F} \quad (1)$$

where the tensor $\underline{\underline{\sigma}}$ is the total stress applied to the surface and \vec{F} is a vector containing the sum of the body forces (i.e., buoyancy). The total stress tensor can then be divided into the thermodynamic pressure, $-p\mathbf{I}$ where \mathbf{I} is the identity matrix, and a deviatoric stress tensor, $\underline{\underline{\tau}}$, which depends on velocities,

$$0 = \nabla \cdot \underline{\underline{\tau}} - \vec{\nabla} p + \vec{F}. \quad (2)$$

To study geodynamic problems, this equation is coupled with a conservation of mass equation,

$$0 = \vec{\nabla} \cdot (\rho \vec{u}), \quad (3)$$

and a conservation of energy equation,

$$\rho c_p \left(\frac{\partial T}{\partial t} + \vec{u} \cdot \vec{\nabla} T \right) = -\vec{\nabla} \cdot \vec{q} + \rho H \quad (4)$$

where \vec{u} is the velocity of the fluid, ρ is the density, c_p is the specific heat at constant pressure, T is the temperature, \vec{q} is the heat flux, t is time, and H is the rate of internal heating per unit volume.

1.1.2. Constitutive theory

There are too many unknown variables (\vec{u} , $\underline{\underline{\tau}}$, p , T , and \vec{q}) to solve Eqs. (2)–(4) and additional relationships between the unknown variables are required to close the problem. The deviatoric stress can be eliminated using

$$\underline{\underline{\tau}} = 2\eta \left(\underline{\underline{\dot{\epsilon}}} - \frac{1}{3} \nabla \cdot \vec{u} \right), \quad (5)$$

where η is the viscosity and $\underline{\underline{\dot{\epsilon}}}$ is the infinitesimal strain-rate tensor, defined by

$$\underline{\underline{\dot{\epsilon}}} = \frac{1}{2} (|\nabla \vec{u}| + |\nabla \vec{u}|^t). \quad (6)$$

When Eqs. (5) and (6) are substituted back into Eq. (2), the deviatoric stress is eliminated from the equations. Eq. (5) is a constitutive equation. Constitutive equations are relationships between two physical quantities and are specific to the material being studied and they approximate the response of the material to external forces or fields. The relationship between heat flux, \vec{q} , and temperature, T , given by

$$\vec{q} = -k \vec{\nabla} T, \quad (7)$$

where, k is the thermal conductivity is another example of a constitutive equation, and is commonly referred to as Fourier's law of heat conduction (e.g., Turcotte and Schubert, 2002). Eq. (7) can be used in order to remove the heat flow, \vec{q} , from Eq. (3).

Constitutive equations are not a part of the fundamental conservation laws. Eq. (2) is derived explicitly from Newton's second law (e.g., Ricard, 2007) and is not dependent on the material properties of the fluid. Eq. (3) is a direct result of the conservation of mass and Eq. (4) is derived from conservation of energy. Constitutive laws are properties of the materials being studied and are generally determined by using theoretical principles of material science with laboratory measurements (Karato, 2008) and progress is being made to understand them from first principles (e.g., Wentzcovitch et al., 2004).

1.2. Material science approach

Silicate minerals deform in response to a constant applied stress; this behavior is called high-temperature creep. Creep is a function of applied differential stress, pressure, and temperature. A material deforms, or creeps, more readily as it approaches its melting point. One can understand the macroscopic behavior of creep deformation in terms of defects in a crystal (e.g., point defects, dislocations, and grain boundaries). Examples of point defects include: sites in the crystal lattice where atoms are missing (vacancies), atoms that are found at sites that are normally not occupied (interstitial atoms), ions with different charge states (e.g., Fe^{2+} vs. Fe^{3+}), and impurities (c.f. Karato, 2008). Dislocations are linear or planar regions within a crystal where the atoms are out of position with

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