

Role of Water and Melts on Upper Mantle Viscosity and Strength

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Melt and water are two of the most important elements influencing the viscosity of the rocks in regions of Earth's upper mantle such as in the mantle wedge above a subducting plate. Over the past five years, laboratory deformation experiments under controlled thermodynamic conditions have yielded quantitative relationships describing the dependence of strain rate, $\dot{\epsilon}$, and, thus, viscosity, η , on melt fraction, ϕ , and hydrogen or hydroxyl concentration, C_{OH} , as well as on differential stress, σ , grain size, d , temperature, T , and pressure, P . These constitutive equations provide a critical part of the framework necessary for modeling processes such as convective flow in and melt extraction from partially molten mantle environments.

Constitutive Equations

A number of experimental studies have been carried out to measure the effects of temperature, differential stress, grain size, melt composition, melt fraction and water content on the viscosity of partially molten olivine-rich aggregates (Cooper and Kohlstedt 1984, 1986; Borch and Green 1990; Bussod and Christie 1991; Jin et al. 1994; Hirth and Kohlstedt 1995a,b; Mei and Kohlstedt 2000a,b). The results of these studies have recently been reviewed by Kohlstedt (1992), Kohlstedt and Zimmerman (1996), and Kohlstedt et al. (2000). Results from deformation experiments on plastic flow of rocks and minerals are frequently analyzed in terms of a power-law creep equation of the form

$$\dot{\epsilon} = A_1 \frac{\sigma^n}{d^p} \exp\left(-\frac{Q + PV}{RT}\right) \quad (1)$$

in order to describe the dependence of strain rate, $\dot{\epsilon}$, on differential stress, σ , grain size, d , pressure, P , and temperature, T . In Eq. 1, A_1 is a materials parameter, Q is the activation energy for creep, V is the activation volume for creep, and R is the gas constant. For olivine-rich rocks with and without melt under anhydrous and hydrous conditions, at low stresses, $n = 1$ and $p \approx 3$, indicative of creep rate controlled grain boundary diffusivity. At higher stresses, $n \approx 3.5$ and $p = 0$, characteristic of creep rate governed by dislocation processes.

To take into account the influence of melt on creep rate, Eq. 1 must be modified. In their analysis of the role of melt on diffusion creep of olivine-basalt aggregates, Cooper and Kohlstedt (1986; 1989) emphasized the importance of (i) local stress enhancement due to replacement of part of each grain with melt and (ii) enhanced transport kinetics resulting from rapid or 'short-circuit' diffusion through the melt. These authors concluded that deformation rate in the diffusion creep regime was limited by ionic diffusion through melt-free grain boundaries though enhanced by the two factors noted above. Their model, however, underestimates the effect of melt on strain rate by an amount that becomes significant at melt fractions, ϕ , greater than 0.03 to 0.05.

In a paper with Keleman and co-workers (1997), Hirth noted that the data in both the dislocation creep regime and the diffusion creep regime could be well-described by modifying the power-law relation given in Eq. 1 to

$$\dot{\epsilon} = A_{2a} \frac{\sigma^n}{d^p} \exp(\alpha \phi) \exp\left(-\frac{Q + PV}{RT}\right) \quad (2a)$$

with $\alpha = 45$ for anhydrous experiments. Recent work in our laboratory suggests that the effect of melt on strain rate might more meaningfully be written as (Mei and Kohlstedt 2000c)

$$\dot{\epsilon} = A_{2b} \frac{\sigma^n}{d^p} \left(\frac{1}{1-\phi}\right)^\beta \exp(\alpha \phi) \exp\left(-\frac{Q + PV}{RT}\right) \quad (2b)$$

where the term $(1-\phi)^{-\beta}$ arises from local stress enhancement effects and the term $\exp(-\alpha \phi)$ enters due to enhanced grain boundary diffusion kinetics resulting from the changes in grain boundary composition that occur in melt-bearing samples. For values of $\beta = (3n + 1)/2$ (Ashby 1983; Chen 1985) of 2 and 6 in the diffusion and dislocation creep regimes, respectively, $\alpha \approx 26$ for water-saturated olivine-basalt samples (Mei and Kohlstedt 2000c).

If water is added to a melt-free sample, Eq. 1 must be again be modified. To quantify the effect of water on viscosity or strain rate, experiments were recently carried out under water-saturated conditions (Bai and Green 1998; Mei and Kohlstedt 2000a,b). Water fugacity was varied by deforming samples at different confining pressures, following the approach used by Mackwell and Kohlstedt (1990) and Bai and Kohlstedt (1992) in their investigations of diffusivity and solubility of water in olivine. The resulting flow law is of the form

$$\dot{\epsilon} = A_{3a} \frac{\sigma^n}{d^p} f_{\text{H}_2\text{O}}^s \exp\left(-\frac{Q + PV}{RT}\right) \quad (3a)$$

where the water fugacity exponent s is 0.7-0.8 in both the diffusion and the dislocation creep regime. If the dependence of solubility on water fugacity (Kohlstedt et al. 1996) is now incorporated into Eq. 3a, then the dependence of creep rate can be expressed directly in terms of the water, that is, OH concentration, C_{OH} , as

$$\dot{\epsilon} = A_{3b} \frac{\sigma^n}{d^p} C_{\text{OH}}^1 \exp\left(-\frac{Q + PV}{RT}\right) \quad (3b)$$

Finally, a general form for the flow law can be written to include the effects of stress, grain size, melt fraction, water concentration, pressure and temperature as

$$\dot{\epsilon} = A_{4a} \frac{\sigma^n}{d^p} C_{\text{OH}}^1 \left(\frac{1}{1-\phi} \right)^\beta \exp(\alpha \phi) \exp\left(-\frac{Q+PV}{RT} \right) \quad (4a)$$

While the extrapolation of this equation from melt-bearing to melt-free conditions poses no mathematical difficulty, the extrapolation from hydrous to anhydrous conditions is clearly not possible as it would yield $\dot{\epsilon} = 0$. In the case of melt, Eq. 4a describes the rheology of samples with $\phi > 0$ and of samples with $\phi = 0$ provided that the addition of melt to a rock does not induce a change in the mechanism of deformation, such as in the model of Cooper and Kohlstedt (1986, 1989). In the case of water, two separate flow laws are required, one for anhydrous conditions and one for hydrous conditions. It is not possible to write one flow law to describe flow in both dry and wet environments because the charge neutrality condition (i.e., point defect chemistry) is not the same when water is present as the charge neutrality condition that operates when water is absent (Kohlstedt and Mackwell 1998; Mei and Kohlstedt 2000a,b). For samples deformed under anhydrous conditions, the flow law given in Eq. 4a must be rewritten without the hydroxyl concentration term as

$$\dot{\epsilon} = A_{4b} \frac{\sigma^n}{d^p} \left(\frac{1}{1-\phi} \right)^\beta \exp(\alpha \phi) \exp\left(-\frac{Q+PV}{RT} \right) \quad (4b)$$

To date, experiments on melt-free and melt-bearing samples under hydrous and anhydrous conditions indicate that Eqs. 4a and 4b adequately describe flow in both the diffusion and the dislocation creep regime. That is, while separate equations are required to describe deformation in anhydrous and hydrous environments (i.e., Eqs. 4b and 4a, respectively), additional equations are not required to characterize deformation of melt-free and melt-bearing rocks. Again, the reader is reminded that not only the parameter A but also the parameters n , p , Q and V will differ from one creep regime to the next (e.g., from dislocation creep to diffusion creep) and from anhydrous to hydrous deformation conditions.

Implication for Upper-Mantle Rheology

Mei and Kohlstedt (2000c) recently investigated the effect of water and melt on diffusion and dislocation creep of olivine-basalt aggregates under water-saturated conditions. For samples deformed under anhydrous conditions, the flow law describing diffusion creep can be written as

$$\dot{\epsilon} = A_{\text{dry}}^{\text{disl}} \frac{\sigma^1}{d^3} \exp(28\phi) \exp\left(-\frac{Q_{\text{dry}}^{\text{disl}}}{RT} \right) \quad (5a)$$

For samples deformed under hydrous conditions, the flow law in the diffusion creep regime is

$$\dot{\epsilon} = A_{\text{wet}}^{\text{diff}} \frac{\sigma^1}{d^3} C_{\text{OH}} \exp(28\phi) \exp\left(-\frac{Q_{\text{wet}}^{\text{diff}}}{RT}\right) \quad (5b)$$

Similar flow laws have been determined in the dislocation creep regime with $n = 3.5$ and $p = 0$. With increasing water concentration, the transition from the flow law for anhydrous conditions (Eq. 5a) to that for hydrous conditions (Eq. 5b) takes place at a water concentration of about 50 H/10⁶Si. The experimental results were also analyzed using an equation with a dependence of strain rate on melt fraction of the form given in Eq. 4; however, for the purpose of extrapolating from laboratory to mantle conditions, Eq. 5 suffices.

The constitutive equations for anhydrous and for hydrous environments, formed by summing the appropriate flow laws for diffusion creep and dislocation creep, provide the framework necessary to model geodynamical processes in water-rich and water-depleted regions of the upper mantle both with and without melt present. Laboratory creep experiments have demonstrated that a small amount of melt does not dramatically weaken polycrystalline olivine in either the dislocation or the diffusion creep regime; addition of 3 vol% melt increases the creep rate (reduces the viscosity) by only a factor of about 3. Since the melt fraction in most partially molten regions in the upper mantle beneath a mid-ocean ridge is expected to be small (<0.03) (e.g., Salters and Hart 1989; Johnson et al. 1990), the effect of partial melting on the rheology of upper mantle appears in general to be modest. In addition, since the reduction in strain rate due to the presence of a small amount of melt is similar for both diffusion and dislocation creep, partial melting in the upper mantle should not promote a transition in deformation mechanism between dislocation creep and diffusion creep.

For melt directly to have a significant effect on mantle viscosity, the melt must be trapped, possibly by cold lithosphere above the region of partial melting, so that the melt can accumulate. If the melt fraction reaches 0.07-0.08, the viscosity will decrease relative to the melt-free value by a factor of ~10. However, melt can indirectly impact the viscosity of partially molten rocks through its influence on water content, as originally suggested by Karato (1986) and subsequently analyzed in detail by Hirth and Kohlstedt (1996). In their analysis of the effect of partial melting on the rheology of lithospheric mantle, these researchers argued that upwelling mantle beneath a mid-ocean ridge will be effectively dried out at a depth of ~65 km due to partitioning of water from the minerals into the melt during pressure-release melting (Hirth and Kohlstedt 1996). As a result, the viscosity of the mantle in and below the source region for mid-ocean ridge basalt (a depth of ≥100 km) will be a factor of >100 smaller than the viscosity of the water-depleted rocks above. This analysis suggests that the base of the oceanic plate corresponds to a compositional boundary layer defined by the extraction of water from mantle minerals during partial melting. In the mantle wedge, viscosity will depend critically on both water content and melt fraction and, hence, on the mechanism of transport of water from the subducting slab into the overlying mantle.

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