THE PRESSURE DEPENDENCE OF CREEP IN OLIVINE: CONSEQUENCES FOR MANTLE FLOW

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Olivine is the most abundant and weakest phase in the uppermost mantle, hence its flow properties control the rheology of the region. While the flow properties of olivine have been studied extensively, previous experiments have been restricted to pressures below 1.6 GPa. Creep of olivine, like most materials at high temperature, is well-described by the power-law creep equation,

$$\dot{\varepsilon} = A \sigma^n \exp(-\Delta G^*/RT), \qquad (1)$$

where $\dot{\varepsilon}$ is the strain-rate, A is a constant, σ is the applied stress, n is a constant usually between 3 and 4, ΔG^* is the activation free energy, R is the gas constant and T is the absolute temperature.

The pressure dependence of creep is presumed to be reflected in the activation volume, ΔV^*, where $\Delta G^* = \Delta E^* + P\Delta V^* - T\Delta S^*$. Extrapolations of the low-pressure data to mantle conditions, when attempted, have generally relied on the assumption that the rate-controlling step in the deformation process is diffusion of oxygen, leading to an estimation that the pressure-dependence of creep is small ($\Delta V^* \sim 10 cm^3$/mole).

We have determined the pressure dependence of steady-state creep of an anhydrous synthetic harzburgite over the pressure range 1-3 GPa in a modified Griggs apparatus using a new liquid cell that greatly reduces uncertainties in friction. We find a strong pressure dependence of the flow stress (Fig. 1), corresponding to an apparent ΔV^* of 25 cm^3/mole. However, direct determination of ΔH^* as a function of pressure shows that ΔH^* decreases strongly with pressure, hence the pressure-dependence of our data cannot be interpreted in terms of ΔV^*. The positive pressure-dependence of the flow stress, therefore, is a consequence of the stronger decrease in ΔS^*.

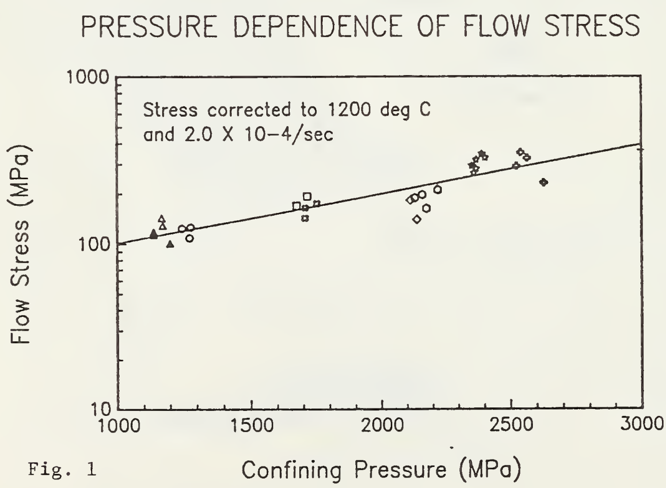

Fig. 1

Previously, a number of workers have suggested that diffusion kinetics and rheology should scale with the homologous temperature, T/Tm, where Tm is the temperature of melting, or the solidus for solid solutions (c.f. Sammis et al, 1981). Predictive equations from these theories fail to describe our data. We have developed, however, a generalized theory that recaptures both our data and those from simple systems. Our derivation shows that the pressure dependence of creep precisely follows the pressure dependence of melting. That is, at constant T/Tm, the creep strength is independent of pressure, The line in Fig. 1 is calculated from this relationship using only the data at 1.2GPa.

This discovery provides for the first time a reliable method for extrapolation to mantle conditions. To calculate the strain-rate for a given stress at depth, one needs only the solidus, the geotherm and rheological data at one pressure. The only modification of (1) that is necessary is substitution of the "equivalent temperature", T_{eq}, where

$$T_{eq} = {}^{p}\left(\frac{T}{T_m}\right){}^{o}T_m \qquad (2)$$

The superscripts p and o denote, respectively, the pressure of interest and the pressure at which the temperature dependence of creep has been determined. ΔH^* in (1) similarly becomes $^{o}\Delta H^*$.

Our extrapolation procedure predicts that reduction of the solidus due to trace quantities of volatiles should be reflected in a similar reduction in the strength of the solid. The predicted water-weakening for olivine is comparable to that observed by Chopra and Paterson (1984), and the predicted CO_2-induced weakening at pressures above 2.8GPa has been confirmed in our laboratory.

Extrapolation of our experimental results to the base of the olivine-bearing mantle using the dry peridotite solidus of Herzberg (1983) and various published suboceanic geotherms, produces an approximately isoviscous mantle between 200 and 400km depth. Our preferred geotherm has, at shallow depths, a slope appropriate for "average" suboceanic mantle and is tied to the olivine-spinel transition at 1725K, 1.3GPa (Lees, et. al., 1983). This geotherm yields at 200-400Km, for a stress of 0.1MPa, an effective viscosity of 10^{21}Pa sec and for a stress of 1.0MPa, yields 10^{18}Pa sec (Fig. 2). Of particular interest is the lack of any low-viscosity channel beneath the lithosphere. Similar calculations for H_2O-bearing and CO_2-bearing mantles yield effective viscosities two to four orders of magnitude lower.

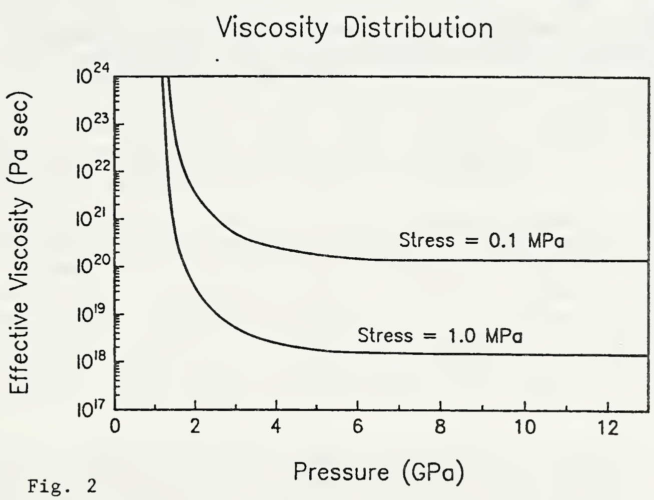

Fig. 2

The ubiquitous association of CO_2 emanations with volcanism (Barnes et al., 1978), the presence of CO_2-filled fluid inclusions in mantle xenoliths (Roedder, 1965) and the presence of exsolved fluid bubbles in mantle olivine (Green and Gueguen, 1983) indicate that at least those regions of the upper mantle associated with volcanic eruptions should reflect the weakening implied by the CO_2-related solidus reduction. It is possible, therefore, that the asthenosphere may be weakened by introduction of carbon. Since our results indicate that carbon distributed throughout the upper mantle would result in a viscosity too low to be compatible with a variety of observations, we suggest that the carbon may be transported directly from the deep mantle or core-mantle boundary in narrow plumes of the sort envisioned by Morgan (1971) and Loper (1985), among others. Previous plume models are initiated by thermal instabilities produced by cooling of the outer core. However, our results imply that exsolution of carbon from the core and dissolution into the silicates also could produce reduction in viscosity and probably an increase in buoyancy.

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