Tectonophysics 481 (2010) 82–98

Contents lists available at ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

Rheology of the deep upper mantle and its implications for the preservation of the continental roots: A review

Shun-ichiro Karato

Yale University, Department of Geology and Geophysics, New Haven, CT 06520, USA

ARTICLE INFO

Article history:
Received 9 July 2008
Received in revised form 26 March 2009
Accepted 7 April 2009
Available online 17 April 2009

Keywords:
Rheology
Continental root
Dislocation creep
Water
Pressure
Partial melting
Mantle convection

ABSTRACT

The longevity of deep continental roots depends critically on the rheological properties of upper mantle minerals under deep upper mantle conditions. Geodynamic studies suggest that the rheological contrast between the deep continental and oceanic upper mantle is a key factor that controls the longevity of the continental roots. Current understanding of rheological properties of deep upper mantle is reviewed to examine how a large enough rheological contrast between the continental and oceanic upper mantle develops that leads to the longevity of the deep continental roots. Based on the microstructures of naturally deformed deep continental rocks as well as on the observations of seismic anisotropy, it is concluded that power-law dislocation creep dominates in most of the deep upper mantle. Deformation by power-law creep is sensitive to water content and therefore the removal of water by partial melting to form depleted continental roots is a likely mechanism to establish a large rheological contrast. The results of experimental studies on the influence of temperature, pressure and water content on plastic flow by power-law dislocation creep are reviewed. The degree of rheological contrast depends critically on the dependence of effective viscosity on water content under "wet" (water-rich) conditions but it is also sensitive to the effective viscosity under "dry" (water-free) conditions that depends critically on the influence of pressure on deformation. Based on the analysis of thermodynamics of defects and high-temperature creep, it is shown that a robust estimate of the influence of water and pressure can be made only by the combination of low-pressure ($<$0.5 GPa) and high-pressure ($>$5 GPa) studies. A wide range of flow laws has been reported, leading to nearly 10 orders of magnitude differences in estimated viscosities under the deep upper mantle conditions. However, based on the examination of several criteria, it is concluded that relatively robust experimental data are now available for power-law dislocation creep in olivine both under "dry" (water-free) and "wet" (water-saturated) conditions. These data show that the influence of water is large (a change in viscosity up to ~4 orders of magnitude for a constant stress) at the depth of ~200–400 km. I conclude that the conditions for survival of a deep root for a few billions of years can be satisfied when "dry" olivine rheology with a relatively large activation volume ($>(10\text{–}15)\times10^{-6}$ m^3/mol)) is used and the substantial water removal occurs to these depths. High degree of water removal requires a large degree of melting in the deep upper mantle that could have occurred in the Archean where geotherm was likely hotter than the current one by ~200 K presumably with the help of water.

1. Introduction

Understanding of the physical properties of the upper mantle is critical to the study of dynamics of this region. For example, the longevity of deep continental roots depends critically on the contrast in density and viscosity between the continental roots and surrounding convecting mantle (e.g., Doin et al. (1997); Lenardic and Moresi (1999); Shapiro et al. (1999a)). The density contrast can be estimated through the estimates of temperature profile and the change in major element chemistry and is relatively well understood. It is widely accepted that the continental lithosphere is nearly neutrally buoyant despite low temperatures due to the small difference in the major element chemistry (e.g., Jordan (1981); Poudjom Djomani et al. (2001); Kelly et al. (2003)). Consequently, the major factor that controls the stability of the continental roots is their rheological properties, particularly the rheological contrast between deep continental roots and surrounding (oceanic or subduction zone) upper mantle. However, the evaluation of the rheological contrast has been hampered by the lack of reliable quantitative studies on these properties under the deep upper mantle conditions.

In order to assess the role of rheological properties, I first review the current understanding of temperature distribution and chemical composition of the continental upper mantle and I conclude that the role of differences in temperature and/or major element composition (and the variation in grain-size) is unlikely to account for a large

---

E-mail address: shun-ichiro.karato@yale.edu.

0040-1951/$ – see front matter © 2009 Elsevier B.V. All rights reserved.
doi:10.1016/j.tecto.2009.04.011

rheological contrast required to preserve a continental lithosphere. A possible role of the lower water content of the continental lithosphere is highlighted, as first pointed out by Pollack (1986). However, the degree of rheological control of the stability of the deep continental roots is sensitive to the rheological properties under the deep upper mantle conditions that were not well understood until very recently. For example, the experimental data Pollack used, i.e., Carter (1976); Post (1977), have been shown to have major uncertainties particularly with regard to the influence of water, and the water weakening effect in olivine was in fact not demonstrated by these studies (e.g., Chopra and Paterson (1981)). During the subsequent years, high-resolution rheological data have been obtained for olivine at low pressures using a gas-medium deformation apparatus (<0.5 GPa; (Karato et al., 1986; Mei and Kohlstedt, 2000a,b)). However, as I will demonstrate in the following, the rheological results determined at pressures below ~0.5 GPa cannot be extrapolated to pressures exceeding a few GPa, because the pressure effect is large and non-monotopic. Consequently, although these low-pressure results are reliable and form the basis for the conceptual understanding of microscopic mechanisms of creep, these results cannot be used to discuss upper mantle rheology below ~20 km depth.

Many efforts during the last ~10 years have been made to obtain quantitative experimental data on (olivine) rheology under higher pressures. However, due to the technical difficulties in the experimental studies under high pressures (and to the complications in the plastic properties), it is not straightforward to obtain robust experimental results on plastic properties under high pressure conditions. In fact, published results on plastic deformation of olivine show a large scatter. The value of the activation volume, a key parameter to characterize the pressure effect, for example, ranges from ~0 to ~27×10^{-6} m^3/mol that corresponds to a more than 10 orders of magnitude difference in effective viscosity (at a constant stress). Although some intermediate values of activation volume were adopted in some earlier papers (e.g., Karato and Wu (1993); Hirth and Kohlstedt (1996, 2003)), there was no strong basis for such choices and widely different values were reported by recent experimental studies. Therefore it is essential to critically evaluate the published results on plastic deformation of upper mantle minerals and to estimate various key parameters (temperature, water content, grain-size) in the deep continental upper mantle in order to assess the mechanism by which the continental lithosphere has survived for ~3 Gyrs.

In the next section, I will first review geophysical and geological studies on the continental upper mantle, including its thermo-chemical structures and the models to explain the longevity of the deep continental roots. Next, I will examine flow laws of the upper mantle minerals (mostly those of olivine) and finally discuss the cause for the survival of the deep continental roots.

2. Geophysical and geological studies on the continental upper mantle and geodynamic studies on the longevity of the continental roots

2.1. Thickness and structure of the continental lithosphere (Fig. 1)

The thickness of the continental lithosphere has often been estimated through seismological studies. In these studies, the continental lithosphere is defined in two different ways. In most studies (e.g., Jordan (1975); Anderson (1979) see also Artemieva (2009)), the continental lithosphere is identified as a near surface region below the continents where the seismic velocities are faster (say by more than 0.5%: see Fig. 1a) than an average mantle. Some technical issues, particularly the issue of anisotropy have been discussed (e.g., Anderson (1979); Gung et al. (2003)), and these authors discussed that when the influence of anisotropy is included, the actual thickness of the Archean continents is ~200–250 km rather than ~300–400 km (however, from geothermal profiles, (Artemieva, 2006) showed that some young (~2.7–2.9 Gyrs) Archean continents have thicker roots, >250 km).

The significance of the "thickness" of the continental lithosphere by this method is not clear because the average seismic wave velocities do not directly correspond to the properties of the mantle. Also the choice of velocity anomaly of 0.5% is arbitrary. A more direct approach is to use the depth variation in seismic anisotropy. Anisotropy is caused mainly by lattice-preferred orientation of minerals (e.g., Wenk (1985); Chapter 21 of Karato (2008)), and hence it is directly related to plastic flow. For example, when a region of a continent is rheologically strong, that region will not be deforming and any seismic anisotropy there will be a frozen anisotropy. In contrast, when materials become soft enough to allow appreciable strain-rates, say $>10^{-15}$ s^{-1} (corresponding to the viscosity of 10^{20} (10^{22}) Pa s for the stress of 0.1 (10) MPa), then anisotropy there will reflect deformation in the recent past. Consequently the depth variation in anisotropy would be used to constrain rheological stratification (Fig. 1b). Based on the analyses of SKS waves, Vinnik et al. (1992) reported that the azimuthal anisotropy in the continental upper mantle follows fossil plate tectonic motion in the shallow part, but in the deep regions, the direction of the azimuthal anisotropy reflects the current (or near recent) plate motion (the transition depth is not well constrained in this study). Similar results were obtained by Deschamps et al. (2008) through the analyses of surface waves who showed a complicated depth varying azimuthal anisotropy. Similarly, using surface wave data, Gung et al. (2003) showed that the continental upper mantle below ~200–250 km has high radial anisotropy.

The age measurements of xenolith samples using Re–Os isotope systematics showed that the ages of xenolith samples from south African craton are 2.7±0.2 Gyrs in the depth range of ~30 to ~200 km that agrees with the age of the crust within the errors (Carlson et al., 1999b) (Fig. 1c). This indicates that the whole cratonic lithosphere in this region has been maintained from the time when the crustal materials were formed despite large lateral translation of continents relative to the surrounding mantle. However, some exceptions are also reported. For example, in the Slave craton (Canada; (Kopylova and Russell, 2000)) and in Wyoming (Carlson et al., 1999a) where deeper xenoliths show younger ages and less depleted chemical characteristics. Also, this technique can be applied only to the depth where samples are collected. Therefore no information is obtained below ~200 km from this technique.

Estimating temperature–depth profiles in the continental upper mantle are critical to the study of stability of the continents. There are two sources for constraining temperature–depth profile. First, one can use the chemical composition of xenoliths brought to the surface from <200 km depth if one assumes that the pressure–temperature conditions recorded in xenoliths reflects "steady-state" geotherm. Together with the observed surface heat flow and the estimated distribution of radiogenic heat production, temperature–depth profiles can be inferred (e.g., (Rudnick et al., 1998; Artemieva, 2006)). Alternatively, observed seismic wave velocity versus depth profiles can be inverted to temperature–depth profiles using a mineral physics-based inversion scheme (Röhm et al., 2000). This latter work provides a unique constraint on the geothermal profiles in the deep upper mantle (>200 km) that is not accessible by xenolith samples. They found that although seismic wave velocities vary a lot among different regions in the shallow regions in the continental upper mantle (<150 km), the variation in seismic wave velocities below ~250 km is small. When velocity difference is inverted for temperature difference, temperature variation in the shallow regions is up to ±300 K, whereas it is less than ±100 K in the deep part (~200–300 km) among various continents. In the deep upper mantle (>250 km), the temperature distribution in both oceanic and continental regions reaches to the similar values, i.e., an adiabat with a potential temperature of ~1520–1570 K.

In sum, both seismological and petrological studies suggest a thickness of the deep roots of ~200 km (with some variations) for old continents. Consequently, in the following, I will examine under which conditions the rheological contrast between the deep continental upper mantle and surrounding mantle becomes large enough to protect the continental roots for billions of years. In doing so I will assume a temperature difference of less than ±100 K. Some geodynamic considerations suggest that the upper mantle temperatures have decreased with geologic time, but the rate at which temperatures

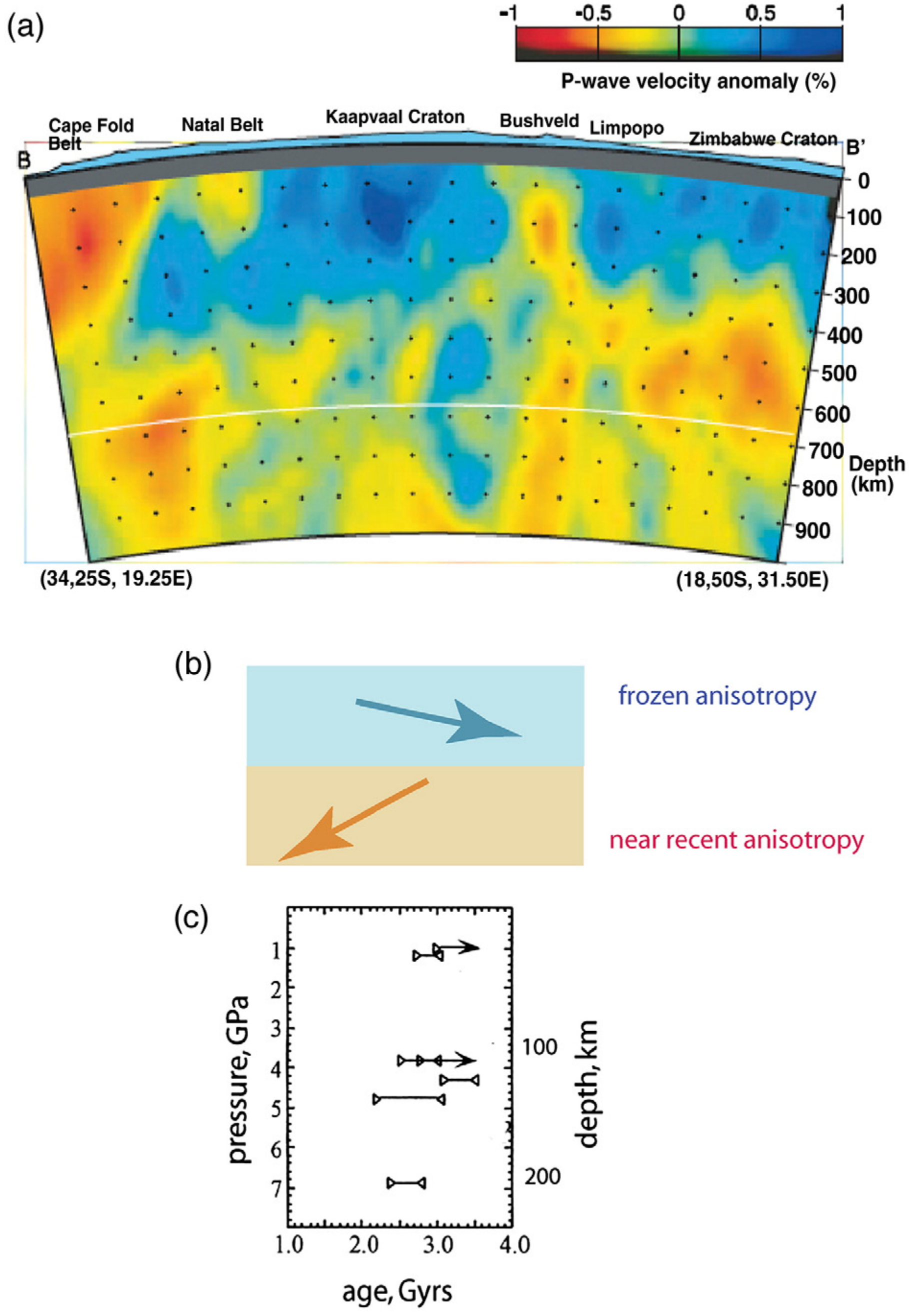


Fig. 1. The structures of the continental upper mantle. (a) Average velocity anomalies (data from James et al. (2001) (adopted from Carlson et al. (2005))): average seismic wave velocities in the shallow continental upper mantle is faster than the average Earth model such as PREM (Dziewonski and Anderson, 1981). This is likely due to the lower temperature (and different chemical composition). A high velocity region is often identified as the continental lithosphere (e.g., Jordan (1975); Anderson (1979)), but due to the influence of several factors and to the arbitrariness of the magnitude of velocity anomalies for this definition, this provides only a vague definition of the continental root (see e.g., Carlson et al. (2005); Artemieva (2009)). More direct evidence is the distribution of "ages" with depth (Carlson et al., 1999b, 2005) but the sampling is limited to ~200 km depth. (b) Stratification in seismic anisotropy: in the shallow region where viscosity is high, anisotropy reflects fossil deformation when that region was hot and had low viscosity, whereas in the deep hot region where viscosity is low enough, anisotropy reflects current on-going (or near recent) deformation. Therefore the transition from fossil to current (near recent) anisotropy (i.e., anisotropy caused by current or recent deformation) marks a depth above which no substantial deformation occurs in the geological time-scale. (c) Distribution of estimated ages of continental upper mantle (from Carlson et al. (1999b)) showing a narrow range of ages indicating that this portion of continental lithosphere has moved as a single unit after its formation for ~2.5 Gyrs.

have decreased is ~70 K/Gyrs (e.g., Abbott et al. (1994)). However, the long-term variation of geotherm is not well constrained. For example, based on the results of melting experiments, Takahashi (1990) suggested that the Archean geotherms away from mid-ocean ridges might be cooler than the current values. Similarly, using an unconventional scaling law for secular cooling, Korenaga (2006) proposed a much slower cooling rate (~30 K/Gyrs). In the following, I will use the geotherms shown in Fig. 2 to calculate the rheological properties and the influence of secular cooling will also be discussed.

2.2. Compositions of the continental upper mantle

Mineralogical composition of the continental upper mantle is constrained by the composition of mantle xenoliths. Since I will discuss the rheological properties of deep continental roots, I will focus on the composition of the deep continental upper mantle. The dominant rock in this region is garnet lherzolite. The (cratonic) continental upper mantle is made of olivine-rich garnet peridotites (olivine: 60–80% mol fraction, orthopyroxene: 10–30%, garnet: 0–10%,

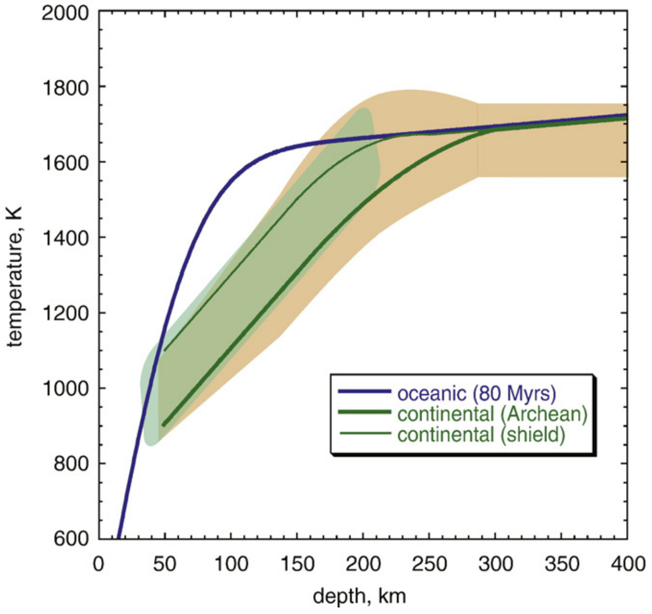


Fig. 2. The temperature-depth profiles for a typical old oceanic upper mantle (age of 80 Myrs) and for continents. For continents, two geotherms are shown corresponding to typical shield and Archean craton. Geotherms below 100 km are considered in the present study. The green region represents the temperature–depth data from mantle xenoliths, and the orange region corresponds to the temperature–depth profiles inferred from the inversion of seismic surface wave data from continental upper mantle.

clinopyroxene: 0–10%) (e.g., Walter (2005)). The depth variation in mineralogy is not large (e.g., Lee (2003); Carlson et al. (2005)). The volume fraction of garnet and clinopyroxene is so small that their effect on rheological properties can be ignored. Also the volume fraction of orthopyroxene is not very large, and the observations of a majority of garnet peridotites from the continental lithosphere show no major contrast in strain between olivine and orthopyroxene, or somewhat higher strength of orthopyroxene than olivine (e.g., (Mercier and Nicolas, 1975)). However, there are notable exceptions. For example, the "sheared lherzolites" i.e., highly deformed peridotites found from deep continental upper mantle (~160–200 km depth) (Boyd, 1973; Boullier and Gueguen, 1975) show the evidence of localized deformation in fine grained orthopyroxene. But deformation of sheared lherzolites occurs with a much shorter time scales than the time scale of mantle convection (e.g., Goetze (1975); Skemer and Karato (2008)). Also, recently Ohuchi and Karato (submitted for publication) showed that orthopyroxene is much weaker than olivine under low temperature conditions (see also Raleigh et al. (1971)). But at high temperatures, the rheological contrast between olivine and orthopyroxene becomes small. Therefore in the following, we consider that the rheological properties of the deep upper mantle as those of olivine. Also the presence of lattice-preferred orientation in most of xenoliths from continental upper mantle including ultra-deep xenoliths coming from the depth deeper than ~300 km (Jin, 1995) suggests that deformation is by dislocation creep.

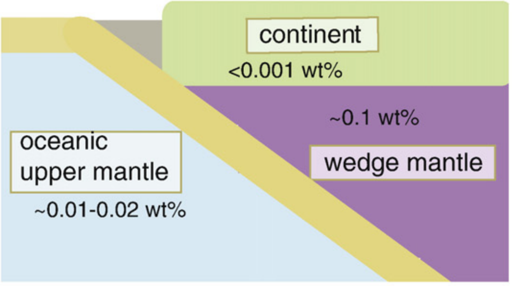
There are some differences in major element chemistry such as the Mg/(Mg + Fe) ratio (percentile of this ratio is known as *Mg#*) and the modal fraction of minerals. The *Mg#* for a primitive upper mantle (pyrolite) is estimated to be ~88, whereas depleted regions have higher values, 90–92 (e.g., Ringwood (1975)). *Mg#* of the typical oceanic upper mantle is ~90 whereas for the continental upper mantle, it can be higher (*Mg#* ~90–92). The modal fraction of minerals is different between depleted continents and surrounding mantle (less garnet and clinopyroxene, and more orthopyroxene in the continents), but such a difference does not have a large effect on rheological properties in the present approximation where the rheological properties are determined by those of olivine.

Water contents in the continental upper mantle are not well constrained. For example, in calculating the strength profile of the continental upper mantle, Kohlstedt et al. (1995) used the experimental results at "wet" (water-saturated) conditions, whereas the experimental data at "dry" (water-free) conditions were used for the oceanic upper mantle. However, as we will show, if the continental upper mantle is "wet" and the oceanic upper mantle is "dry", then the viscosity of the continental upper mantle would be lower than that of the oceanic upper mantle and the deep continental roots would not have survived for the geological time. Xenolith samples from kimberlite magmas show a range of water content, but in all cases much less than the saturation limit (less than ~0.015 wt.%; e.g., Bell and Rossman (1992a)).

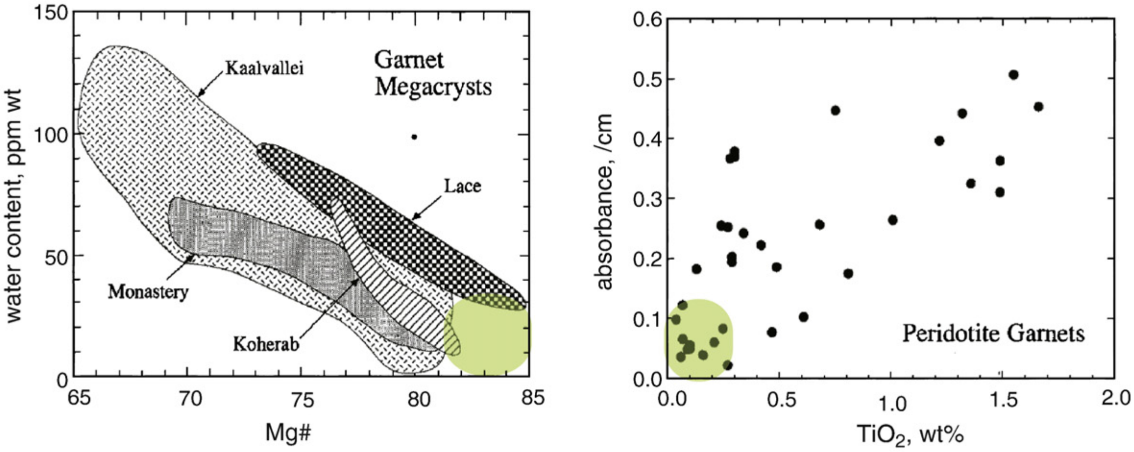
There are two major challenges in estimating the water content in the deep continental upper mantle from mantle xenoliths. First, because the diffusion of water in minerals is fast (this is particularly the case where hydrogen is dissolved by itself, charge compensated by diffusion of vacancies or electron holes), observed water contents in minerals rarely reflect their original water contents. In fact, for olivine in which the chemical diffusion of water is known to be fast (Kohlstedt and Mackwell, 1998), there is very little correlation between the water content and the concentration of other elements (Bell et al., 2004). In contrast, for garnet where dissolution of hydrogen is considered to occur as a combined substitution of Si with Al (Mookherjee and Karato, submitted for publication), chemical diffusion of water requires diffusion of Al that is slow. Therefore the water content in garnet more faithfully reflects the equilibrium water content in the mantle than olivine. Given the water content of garnet, the corresponding water content in olivine can be calculated from the known partition coefficient of water between these two minerals. Mookherjee and Karato (submitted for publication) showed that the partition coefficient is less than unity between olivine and garnet (at low water fugacity) indicating that the water content in garnet gives an upper limit for the water content in olivine.

There is a clear correlation between the water content in garnet and the concentration of elements that are enriched in the kimberlite magma (e.g., FeO, TiO_2; Bell et al. (2004)). In mantle xenoliths, sheared lherzolites are enriched with these elements compared to coarse-granular peridotites (Danchin, 1979). An implication of this observation is that the pristine continental upper mantle has smaller water contents than the values seen in mantle xenoliths from the kimberlite magma (Mookherjee and Karato, submitted for publication). In fact, there are a number of mantle xenoliths that show much smaller water contents (e.g., San Carlos olivine has almost no water; (Mackwell et al., 1985)). Many mantle xonoliths (from south Africa) showed olivine A-type fabrics (Ben Ismail et al., 2001; Skemer and Karato, 2008), which indicates that they were deformed under water-poor conditions (<0.001 wt.%) (for olivine fabrics see Karato et al. (2008)). Water-poor conditions are also consistent with a generally accepted view that the continental lithosphere is a residue of extensive partial melting (e.g., Jordan (1981); Carlson et al. (2005)). This point will be discussed in more detail when I examine the condition for the survival of the deep continental roots. An argument to support such a model was presented by Hirth et al. (2000) based on the inferred electrical conductivity. In contrast, a common view in the petrology/geochemistry is that the source region of mid-ocean ridge basalt (MORB) contains ~0.01–0.02 wt.% of water (e.g., Ito et al. (1983); Dixon et al. (2002); Hirschmann (2006)) and the water content is higher in the upper mantle in the subduction zone (Ito et al., 1983; Stolper and Newman, 1994; Shito et al., 2006): the maximum amount of water (in olivine) in the subduction zone will be close to the saturation limit, ~0.1 wt.%. In summary, the water content is up to ~0.1 wt.% in the subduction zone upper mantle, ~0.01–0.02 wt.% in the oceanic upper mantle (asthenosphere), and less than 0.001 wt.% in the continental upper mantle with local enrichment due to metasomatism (Fig. 3). Note that in a typical environment where

(a)



(b-1): continental mantle



(b-2): ultra-high pressure eclogite (subduction zone?)

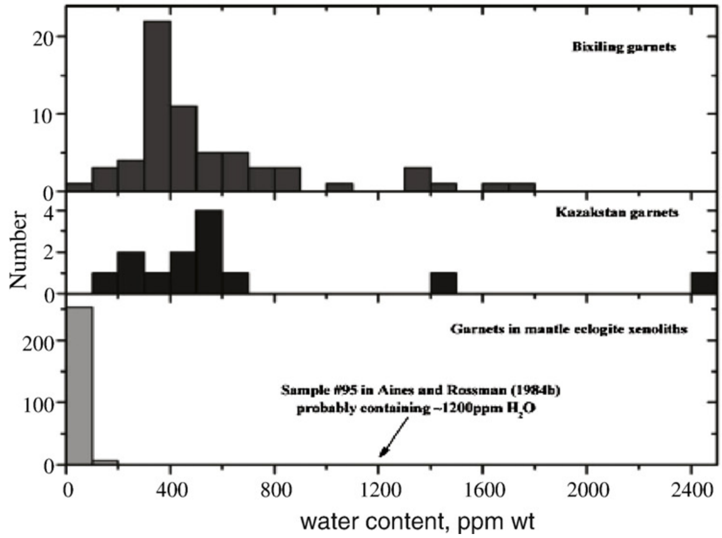


Fig. 3. (a) A cartoon showing the water distribution in the upper mantle. The water content in the oceanic asthenosphere is well constrained from the observations on MORB to be ~0.01–0.02 wt.% (Ito et al., 1983; Dixon et al., 2002; Hirschmann, 2006). Wedge mantle in the subduction zone contains a large amount of water, ~0.1 wt.% (Stolper and Newman, 1994; Ito et al., 1983; Shito et al., 2006). In contrast, the water content in the continental upper mantle is less constrained. A plausible model is that the majority of the continental lithosphere has low water content (less than 0.001 wt.%) due to a large degree of partial melting, whereas locally continental upper mantle is enriched with water due to the influence of volatile-rich magmatism such as kimberlite. Most samples from the continental upper mantle is from these regions. (b-1) Correlation of water content and Mg# or TiO_2 content in garnet from continental upper mantle (data from Bell and Rossman (1992a)) (absorbance is proportional to water content). Garnet is chosen because diffusional loss or gain of water is likely less important for garnet than olivine. The kimberlite xenolith from deep continental upper mantle show 0.001–0.01 wt.% of water (Bell and Rossman, 1992b; Bell et al., 2004) but these values likely over-estimate the water content in the pristine continental upper mantle. Garnet unaffected by metasomatism likely have high Mg# and low TiO_2 (Danchin, 1979) shown by green regions. (b-2) Water contents in garnet in some ultra-high pressure eclogites (data from Xia et al. (2005)). Ultra-high pressure eclogites are considered to have come from subduction zone environment. These garnets show high water contents nearly the saturation limit, ~0.05–0.1 wt.%.

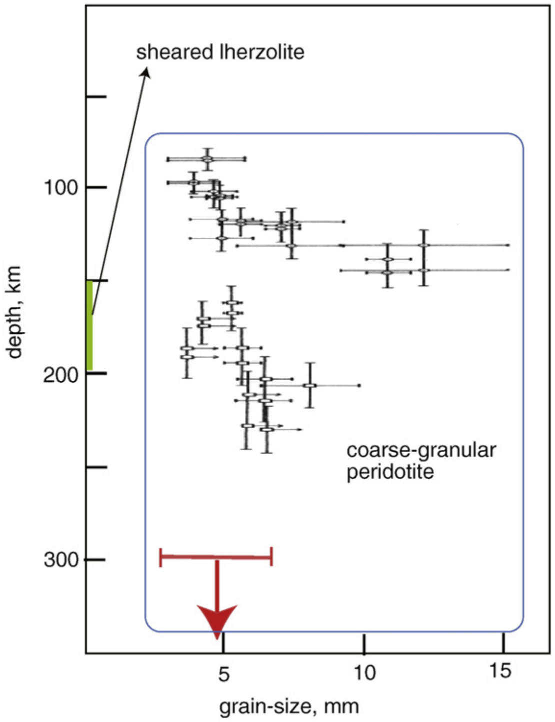


Fig. 4. Grain-size in the continental upper mantle from peridotite xenoliths (modified from Mercier (1980), the data for deep (>300 km) upper mantle is from Jin (1995), JAG-84-318).

deep continental roots are subjected to convectional erosion, two upper mantle materials are involved one depleted (water-poor) materials in the continental roots and another relatively water-rich materials in the convecting surrounding mantle (see also Fig. 5). Therefore it is possible that the rheological contrast in these materials is due partly to the difference in water content.

What about the grain-size? Grain-size can have an important effect on rheological properties under some conditions (e.g., Karato (2008), Chapter 8 and 19). However, most of deep mantle rocks show a small range of grain-size, a few mm (e.g., Mercier (1979, 1980); Fig. 4) except for "sheared lherzolites". Sheared lherzolites are upper mantle xenoliths from the depth of ~150 km to ~200 km (Boyd, 1973; Goetze, 1975)) showing a smaller grain-size (on the order of ~0.1 mm), and they are potentially important for the discussion of deep continental rheology. However, the structural and compositional analyses of these rocks suggest that these rocks were deformed by some short-term events (on the time scale of ~10^{3-4} years (Skemer and Karato, 2008)) and their microstructure is not representative of a long-term deformation. I conclude that the grain-size of a typical continental upper mantle is several mm and the corresponding stress level is ~0.1–1 MPa if one accepts that grain-size is controlled by dynamic rercystallization (e.g., Karato et al. (1980); Mercier (1980); Urai et al. (1986); Drury and Urai (1990); van der Wal et al. (1993)). These results will be used to discuss the possible mechanisms of deformation in the deep upper mantle.

In summary, the differences in temperature and major element chemistry between the continental and surrounding oceanic upper mantle, or between the continental lithosphere and asthenosphere are small. In the following, we will first examine to what extent the differences in these factors could affect the stability of the deep continental roots, and second, we will explore the influence of another factor, the water content.

2.3. Preservation of the continental lithosphere

Geochemical and geophysical observations indicate that old continental lithosphere has survived for 2–3 Gyrs. There are two different processes that could have destroyed the continental lithosphere. One is the gravitational instability caused by density contrast (the Rayleigh–Taylor instability, e.g., Houseman and Molnar (1997)) and the convective erosion (e.g., Lenardic and Moresi (1999)) (Fig. 5).

The continental lithosphere is cooler than the surrounding mantle, and therefore it is likely to be denser. If it were denser, then gravitational instability would have destabilized the continental lithosphere. However, the analyses of gravity signals and the composition of continental xenoliths show that the continental upper mantle is chemically different from the oceanic upper mantle, and the chemical buoyancy nearly compensates for thermal buoyancy (e.g., Jordan (1975); Shapiro et al. (1999b)). Let us, for now, assume that the density contrast is small and the continental lithosphere is neutrally buoyant. In such a case, the important process that could destroy the continental lithosphere is convective erosion, i.e., the removal of continental materials due to the viscous force exerted by the convection current in the nearby regions. Previous geodynamic studies showed that a viscosity contrast of ~10^3 is required to maintain the integrity of the continental lithosphere for ~3 Gyrs against convective erosion (e.g., Lenardic and Moresi (1999); Shapiro et al. (1999a)). The viscosity contrast of two materials with non-linear rheology depends on either stress or strain-rate is kept constant (Chapter 19 of Karato (2008)). Because the convective erosion occurs when two materials are in contact without any constraints on strain-rate, this viscosity contrast must be the viscosity contrast at the same stress and not at the same strain-rate. In the following I will examine how this condition can be met based on the analysis of experimental data on rheological properties of the upper mantle materials.

3. Some fundamentals of rheological properties

Although the average viscosity of Earth mantle is well constrained from some geodynamic studies (e.g., Mitrovica (1996), Chapter 18 of Karato (2008)), these studies provide only weak constraints on the lateral or small-scale variation in rheological properties such as the rheological

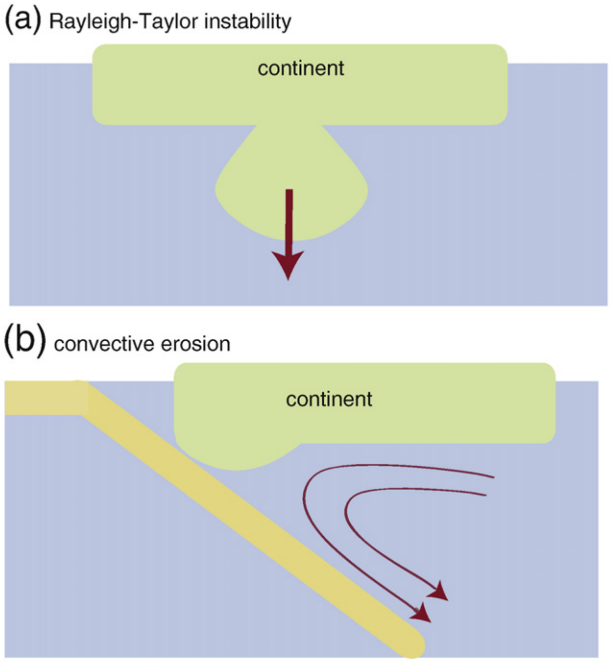


Fig. 5. Schematic drawings showing two mechanisms of destruction of the deep continental roots. (a) Rayleigh–Taylor instability caused by density inversion. (b) convective erosion. When density contrast is small as suggested by the gravity observations, the second mechanism (convective erosion) dominates.

contrast between the continental and the oceanic upper mantle. Some efforts have been made to infer the lateral variation in rheological properties from geodynamic studies (e.g., Zhang and Christensen (1993); Cadek and Fleitout (2003)), but the spatial resolution in these studies is limited. Therefore, it seems most appropriate to use laboratory data on rheological properties combined with geophysical, geochemical inferences on some key parameters (such as temperature, water content) to infer the degree of the rheological contrast across the continental–oceanic upper mantle boundary.

However, experimental approach to mantle rheology is challenging for the following reasons. First there is a vast difference in the time scale of deformation between laboratory studies and deformation in Earth, and any results from experimental studies need to be extrapolated in order to apply them to deformation in Earth. Second, it is difficult to conduct quantitative deformation experiments under deep Earth conditions. Based on the coordinated efforts during the last several years, the latter difficulties have been overcome and quantitative deformation experiments can now be conducted to pressures exceeding ~17 GPa at temperatures to ~2200 K (Karato and Weidner, 2008). However the large difference in the time scale (strain-rate) always exists that implies that one needs to extrapolate laboratory data to estimate the rheological properties of Earth's interior. In order to justify such an extrapolation, one needs to demonstrate that the microscopic mechanisms operating in the given set of laboratory studies are the same as those operating in Earth. Also the mathematical formula ("constitutive relationship") must be consistent with the microscopic physics of deformation.

There are a few different approaches to accomplish these challenges. A common procedure is to conduct well-defined laboratory deformation experiments (under limited conditions), to analyze the data to find an appropriate constitutive relationship, and to examine the deformation microstructures of laboratory samples to place constraints on the microscopic mechanisms of deformation. Plastic deformation in solid materials occurs by the motion of crystalline defects. Upon applying a deviatoric stress, concentration of defects will be modified and after certain strain, steady-state distribution of defects is established leading to steady-state deformation. For steady-state deformation, a unique relationship between strain-rate ($\dot{\varepsilon}$) and other variables (stress (σ), temperature (T), pressure (P), water fugacity (f_{H2O}) (or water content (C_{W})), grain-size (L)) can be defined. Such a relationship (constitutive relationship) holds for a particular mechanism (i-th mechanism) and in general written as (e.g., Kocks et al. (1975); Karato (2008))

$$\dot{\varepsilon}_i = f_i(\sigma, T, P; f_{\mathrm{H2O}}, L) \tag{1a}$$

and the corresponding effective viscosity is given by $\eta_i = \sigma / 2\dot{\varepsilon}_i$. When water content is fixed (a closed system behavior), then instead of water fugacity, one can use water content, C_{W}.

$$\dot{\varepsilon}_i = f_i'(\sigma, T, P; C_{\mathrm{W}}, L). \tag{1b}$$

Important deformation mechanisms in Earth's interior and in experimental studies include (i) (power-law) dislocation creep, (ii) diffusion creep and (iii) exponential creep (the Peierls mechanism). Each mechanism has its own constitutive relationship and the above three mechanisms operate independently. Fig. 6 illustrates how the dependence of effective viscosity on some variables changes when deformation mechanism changes. In the experimental study of plastic deformation, effective viscosity (strength at a given strain-rate) is measured in a limited parameter space and the results are extrapolated to estimate the effective viscosity of Earth's interior. As seen in Fig. 6, extrapolation of experimental data is valid only when extrapolation is within the same deformation mechanism where the same constitutive relationship determined in the lab would apply.

When several mechanisms of deformation operate, the use of a deformation mechanism map is a convenient way to understand which mechanism is dominant in a certain parameter space (Frost and Ashby, 1982). Fig. 7 shows a deformation mechanism map of olivine on a stress versus grain-size space at a given temperature, pressure and water content. In this map, I chose temperature and pressure conditions appropriate for the deep upper mantle (water content is assumed to be low, i.e., "dry"). Using the estimated grain-size and stress in the deep upper mantle, one can conclude that likely mechanism that may operate in Earth is either power-law dislocation creep or diffusion creep for low strain-rate deformation that occurs in Earth's interior (I will provide a detailed discussion to sort out which one is more important in the deep continental upper mantle). However, in laboratory deformation experiments where strain-rates are higher and grain-size is smaller (in most cases), likely mechanisms of deformation are exponential creep (the Peierls mechanism), power-law dislocation creep or diffusion creep depending on the exact conditions of deformation experiments. In particular, the exponential creep (the Peierls mechanism) tends to dominate under high-pressure conditions because a material becomes strong at high pressures. Results from such experiments cannot be extrapolated to Earth's deep interior.

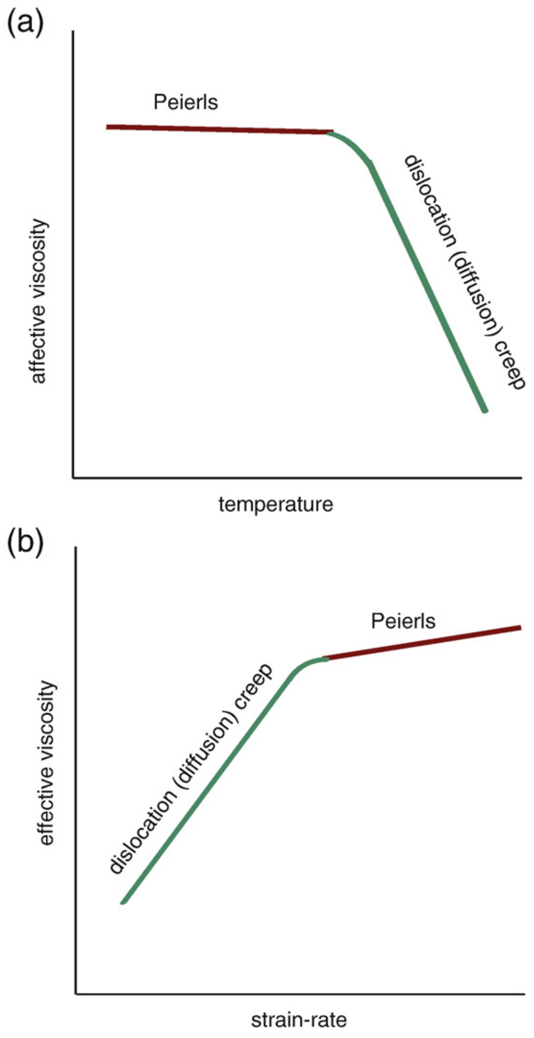


Fig. 6. Schematic diagrams showing the influence of temperature (a) and strain-rate (b) on the effective viscosity for the Peierls mechanism and power-law creep. At low temperatures or high strain-rates, the Peierls mechanism (exponential creep) dominates for which $\eta_{\mathrm{eff}} \propto \sqrt{T \cdot \log\dot{\varepsilon}}$ whereas at high temperatures and/or slow strain-rates, the power-law creep dominates for which $\eta_{\mathrm{eff}} \propto \dot{\varepsilon}^{\frac{1}{n}} \exp\left(\frac{H^*}{nRT}\right)$. Results from one regime cannot be extrapolated to another regime.

The reliability of the conclusions from the deformation mechanism map depends on the reliability of the data used to construct such a map and also on the estimated values of parameters such as the stress.

Both of them have large uncertainties. An alternative approach that is free from these uncertainties is to use microstructural observations to get some insights into the dominant deformation mechanisms in the deep upper mantle. Almost all mantle xenoliths show strong lattice-preferred orientation (e.g., Ben Ismail and Mainprice (1998)), and they also show evidence of dynamic recrystallization (e.g., Drury and Urai (1990)) suggesting the operation of dislocation creep. This is also true for an ultra-deep xenolith from deep continental upper mantle (deeper than 300 km) of South Africa (Jin, 1995). In addition, recent seismological observations show evidence for strong seismic anisotropy in the deep upper mantle (e.g., Gung et al. (2003)). Taken together, I conclude that there is strong case for dislocation creep as a dominant deformation mechanism in the deep upper mantle although grain-size sensitive creep such as diffusion creep may operate under limited conditions (particularly low temperature shear zones: Handy (1989); Jin et al. (1998); Warren and Hirth (2006)).

Now let us discuss some details of the flow law for dislocation creep. Plastic deformation at high temperature and low stress occurs by thermally activated motion of crystalline defects and therefore the strain-rate depends exponentially on temperature. Defect concentration and/or mobility also depend on water fugacity, and in some cases, the deformation involves grain-boundary processes. So a generic equation to describe the dependence of strain-rate on some thermodynamic and structural parameters is (e.g., Karato (2008))

$$\dot{\varepsilon} = A \cdot f_{H2O}^{r}(P,T) \cdot L^{-m} \cdot \sigma^{n} \cdot \exp\left(-\frac{E^{*} + PV^{*}}{RT}\right) \tag{2a}$$

where $\dot{\varepsilon}$ is strain-rate, A is a pre-exponential factor that is insensitive to thermo-chemical conditions, $f_{H2O}(P,T)$ is the fugacity of water that depends on pressure (P) and temperature (T), L is grain-size, σ is deviatoric stress, E^* is activation energy, V^* is activation volume, and R is the gas constant (r, m and n are non-dimensional constants that depend on the mechanisms of deformation). If we use a relation $C_W \propto f_{H2O}$, the water fugacity can be replaced with the water content in the flow law equation,

$$\dot{\varepsilon} = A' \cdot C_W \cdot L^{-m} \cdot \sigma^{n} \cdot \exp\left(-\frac{E'^{*} + PV'^{*}}{RT}\right). \tag{2b}$$

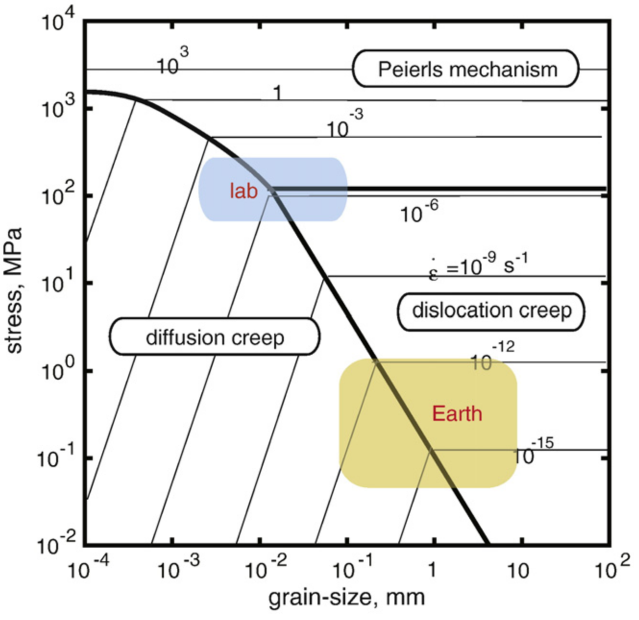


Fig. 7. A deformation mechanism map for olivine under "dry" (water-free) conditions at ~200 km conditions ($P = 7$ GPa, $T = 1700$ K) on the grain-size and stress space. The power-law creep constitutive relation established by Kawazoe et al. (2009) and the diffusion creep constitutive relationship by Mei and Kohlstedt (2000a) with the activation volume of 6×10^{-6} m^3/mol are used. The exact location of the boundaries has some uncertainties, but the general conclusions discussed in the text remain valid. Plausible ranges of grain-size and stress in the deep upper mantle and in laboratory experiments are shown by shaded regions.

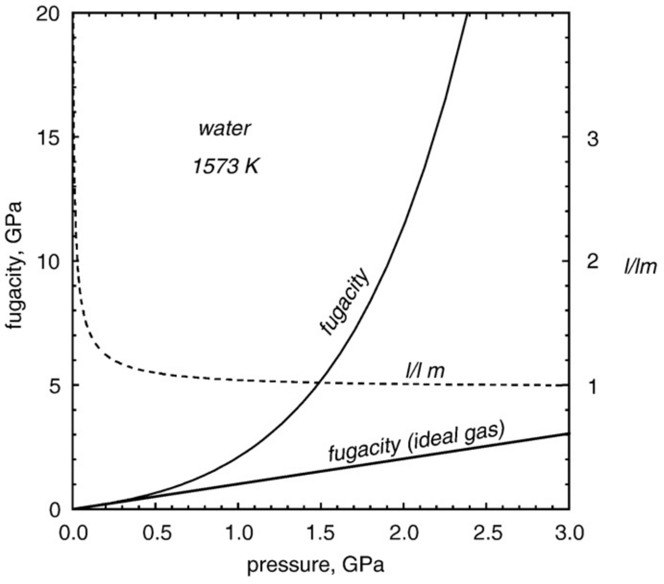


Fig. 8. The fugacity of water as a function of pressure (at $T = 1573$ K) (from Karato (2008)). Shown together is the ratio of mean molecular distance, l, to the molecular size l_m. Non-ideal gas behavior becomes important when l/l_m becomes close to 1.

For dislocation creep (at modest stress level), usually $m = 0$ and $n > 1$.

The water fugacity strongly depends on both pressure (P) and temperature (T) and the functional form to describe the pressure and temperature dependence of water fugacity changes with pressure (Fig. 8). At low pressures and high temperatures, water behaves like an ideal gas, whereas at high pressures, water molecules become close together and water behaves like a highly incompressible, non-ideal gas. Consequently, at low pressures, $f_{H2O}(P,T) \propto P$ whereas at high pressures, $f_{H2O}(P,T) \propto \exp\left(\frac{PV_{H2O}}{RT}\right)$ where V_{H2O} is the molar volume of water and this transition occurs at ~0.5 GPa. Consequently, it is essential to obtain experimental data for a pressure range including both below and above ~0.5 GPa to determine the flow law parameters under water-rich ("wet") conditions.

The defect species responsible for plastic deformation likely changes when water is added. For example, when there is no water, diffusion may occur using empty vacancies but under water-rich conditions diffusion can occur using hydrogen-filled vacancies (e.g., Hier-Majumder et al. (2005)). In these cases, one needs to consider at least two mechanisms of deformation, one deformation under "dry" (water-free) conditions and another deformation under "wet" (water-rich) conditions with different activation energies and volumes. In these cases, there will be two different flow laws, and one may distinguish flow law under "wet" condition ($\dot{\varepsilon}_{wet} = f_{wet}(T,P,\sigma,f_{H2O};L)$ or $\dot{\varepsilon}_{wet} = f_{wet}(T,P,\sigma,C_W;L)$) from that under "dry" conditions ($\dot{\varepsilon}_{dry} = f_{dry}(T,P,\sigma,L)$). Thus,

$$\dot{\varepsilon}_{dry} = A_{dry} \cdot \sigma^{n_{dry}} \cdot \exp\left(-\frac{H_{dry}^{*}}{RT}\right) \tag{3}$$

and

$$\dot{\varepsilon}_{wet} = A_{wet} \cdot f_{H2O}^{r}(P,T) \cdot \sigma^{n_{wet}} \cdot \exp\left(-\frac{H_{wet}^{*}}{RT}\right) \tag{4a}$$

$$= A'_{wet} \cdot C_W^{r} \cdot \sigma^{n_{wet}} \cdot \exp\left(-\frac{H'^{*}_{wet}}{RT}\right) \tag{4b}$$

Table 1
Flow law parameters of power-law dislocation creep of olivine.

Reference	$\log_{10}A$	n	r	E^* (kJ/mol)	V^* (10^{-6} m^3/mol)	Starting material	P (GPa)	T (K)	Water content[j] (ppm H/Si)	Apparatus
"Dry"										
Kawazoe et al. (2009)	5.04[e]	3.5[e]	–	530[e]	15–20	Hot-pressed aggregate made from San Carlos olivine	4.9–9.6	1300–1870	<50	RDA
Raterron et al. (2007)[a]	-8.2	2.6 ± 0.3[f]	–	112 ± 40	3.0 ± 0.5	Single crystal of synthetic Mg_2SiO_4 olivine	2.1–7.5	1373–1677	<90	D-DIA
Raterron et al. (2007)[b]	-9.2	2.7 ± 0.3[f]	–	104 ± 40	1.0 ± 0.5	Single crystal of synthetic Mg_2SiO_4 olivine	2.1–7.5	1373–1677	<90	D-DIA
Li et al. (2006)	3.66[g]	3.0[g]	–	470[g]	0 ± 5	Powder made from San Carlos olivine	3.5–9.6	1073–1473	570–3800	D-DIA
Karato and Jung (2003)	6.1 ± 0.2	3.0 ± 0.1[g]	–	510 ± 30[g]	14 ± 2	Hot-pressed aggregate made from San Carlos olivine	2.0	1573	<100	Griggs (solid medium)
Karato and Rubie (1997)	5.38[h]	3	–	540[h]	14 ± 1	Hot-pressed aggregate made from San Carlos or synthetic olivine	11–15	1300–1900	–	Kawai
Bussod et al. (1993)	5.05[i]	3.5[i]	–	540[i]	5	Single crystal of San Carlos olivine	6.0–13.5	1720–1870	–	Kawai
Borch and Green (1989)	8.00	3.3	–	505	28	Hot-pressed aggregate made from Balsam Gap dunite	0.6–2.1	1250–1700	[l]	Griggs (liquid medium)
"Wet"										
Karato and Jung (2003)[c]	2.9 ± 0.1	3.0 ± 0.1[g]	1.20 ± 0.05	470 ± 40[g]	24 ± 3	Hot-pressed aggregate made from San Carlos olivine	1.0–2.0	1473	800–1500	Griggs (solid medium)
Karato and Jung (2003)[d]	0.56 ± 0.02	3.0 ± 0.1[g]	1.20 ± 0.05	410 ± 40	11 ± 3	Hot-pressed aggregate made from San Carlos olivine	1.0–2.0	1473	800–1500	Griggs (solid medium)
Mei and Kohlstedt, (2000a,b)	3.66	3.0 ± 0.1	0.69[k]	470 ± 40[k]	0	Hot-pressed aggregate made from San Carlos olivine	0.10–0.45	1473–1573	~110–380	Paterson
	3.18	3.0 ± 0.1	0.98[k]	470 ± 40[k]	20					
	2.70	3.0 ± 0.1	1.25[k]	470 ± 40[k]	38					

Unit of A is $s^{-1}(MPa)^{-n-r}$.

[a] For [100](010) slip system.

[b] For [001](010) slip system.

[c] Open system.

[d] Closed system.

[e] From Hirth and Kohlstedt (2003).

[f] Stress exponent is not determined but assumed.

[g] From Mei and Kohlstedt (2000b).

[h] From Karato et al. (1986).

[i] From Bai et al. (1991). Value of A is recalculated as that under oxygen fugacity of $10^{-2.9}$ bar.

[j] Based on Paterson's (1982) calibration.

[k] The values of r and V^* are not constrained independently by this study because of the strong trade-off (see text).

[l] A later study (Young et al., 1993) showed a substantial amount of water in their sample.

respectively, where $\dot{\varepsilon}$ is strain-rate, $A_{\mathrm{dry,wet}}$ is a pre-exponential factor, $n_{\mathrm{dry,wet}}$ is the stress exponent, $f_{\mathrm{H2O}}(P,T)$ is the fugacity of water, r is a constant, P is pressure, T is temperature, σ is differential stress, $H^*_{\mathrm{dry,wet}}(= E^*_{\mathrm{dry,wet}} + PV^*_{\mathrm{dry,wet}})$ is activation enthalpy ($E^*_{\mathrm{dry,wet}}$: activation energy, $V^*_{\mathrm{dry,wet}}$: activation volume) and R is the gas constant. Note that the water fugacity in Eqs. (1a) or (4) is well defined only under water-saturated conditions. As we will discuss later, in many experimental studies the water fugacity (content) was not clearly defined that leads to a major uncertainty in the interpretation of the data.

From Eqs. (2a), (2b) it is clear that strain-rate (or the effective viscosity) is very sensitive to temperature and pressure, but also to water fugacity and in some cases grain-size. Therefore one needs to determine the sensitivity of rheological properties to these variables in laboratory studies, and also the values of these variables in Earth need to be estimated.

4. Experimental constraints on the rheological properties of olivine under the deep upper mantle conditions

There have been several reviews on the rheological properties of upper mantle materials (mostly for olivine) (Karato and Wu, 1993; Kohlstedt et al., 1995; Hirth and Kohlstedt, 2003). However, rheological properties of the upper mantle were not well constrained in these studies mainly because of the lack of quantitative experimental results at high pressures.

Several new results have been published on olivine rheology since these reviews were published and Table 1 contains a list of published results on olivine rheology for power-law dislocation creep (for *Mg#* ~90). The viscosity corresponding to each flow law is plotted for a typical mantle geotherm for $\sigma = 0.1$ MPa in Fig. 9. In this calculation, I assumed that the activation volume is constant when the experimental results need to be extrapolated in pressure. Borch and Green (1987) proposed that the pressure dependence of high-temperature creep follows the homologous temperature scaling, i.e., $\dot{\varepsilon} \propto \exp\left(-\frac{\beta \cdot T_m(P)}{T}\right) \propto \exp\left(-\frac{E^* + PV^*}{RT}\right)$ where T_m is solidus (β is a non-dimensional constant). The solidus-pressure curve of dry peridotite has a strong curvature (Takahashi, 1986) and it would imply that the activation volume is reduced appreciably with pressure. However, this model is inconsistent with the observation that the rheological properties of pure olivine are very similar to those of peridotite (Zimmerman and Kohlstedt, 2004). The solidus of olivine is higher than that of peridotite by several hundreds of degrees and if such a model were valid, then the rheological properties of peridotite would be much softer than those of pure olivine aggregates.

Fig. 9 shows a large scatter of published results on olivine rheology even under "dry" conditions particularly when the rheological properties under high pressures are compared. In order to understand the possible reasons for such a large scatter, it is important to review the nature of experimental studies of plastic deformation. Currently, undisputable high quality data on plastic deformation are obtained only by using a gas-medium deformation apparatus with an internal load-cell (Paterson, 1990). Experimental results using this apparatus under carefully controlled chemical environment and microstructures have been published and formed a basis for the modern experimental study of plastic deformation of minerals (e.g., Karato et al. (1986); Mei and Kohlstedt (2000a,b); Rybacki and Dresen (2004)). However, this apparatus can be operated only at low pressures (<0.5 GPa corresponding to a depth of <15 km), which results in a major limitation for the application of results from this apparatus to understand plastic deformation in Earth's interior. For example, the influence of pressure (activation volume) cannot be constrained well for mantle minerals because the influence of pressure at low pressures is very small compared to the uncertainties. This is particularly true if the grain-size sensitive flow law for grain-boundary sliding accommodated dislocation creep proposed by Hirth and Kohlstedt (2003) is used (Appendix A). Consequently, the rheological properties of most of the upper mantle (plastic deformation occurs mostly in the deep upper mantle) are essentially unconstrained by the data from low-pressure studies.

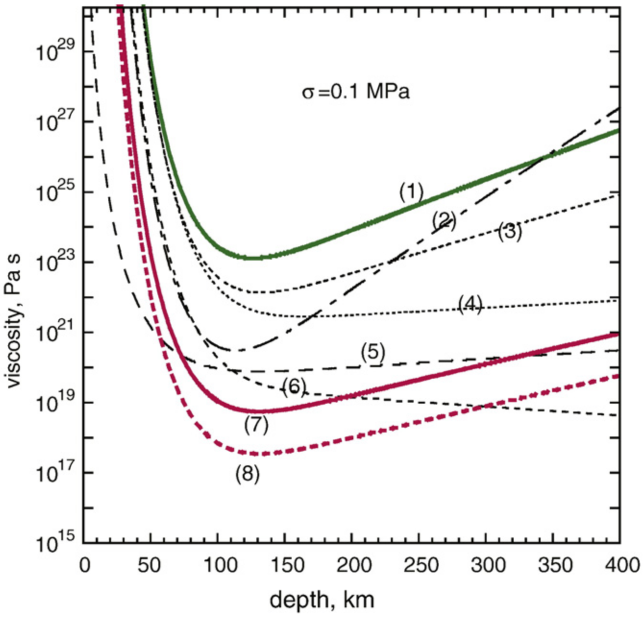


Fig. 9. Viscosity versus depth relation for the upper mantle with a typical oceanic geotherm calculated from the power-law creep constitutive relationship for olivine. Except for (7) and (8) all the results are reported to correspond to nominally "dry" conditions. A reference stress of 0.1 MPa is used. The absolute value of viscosity depends on the stress as $\eta \propto \sigma^{1-n}$. For the explanation for the discrepancy among different results, see text. Data source: (1): (Kawazoe et al., 2009) (an activation volume of 17×10^{-6} $\mathrm{m^3/mol}$ is used). (2): (Borch and Green, 1987) (activation volume is assumed to be constant; see text). (3): (Karato and Rubie, 1997). (4): (Bussod et al., 1993). (5): (Raterron et al., 2007). (6): (Li et al., 2006). (7): (Karato and Jung, 2003) ($C_{\mathrm{w}} = 1000$ ppm H/Si (~0.007 wt.%)). (8): (Karato and Jung, 2003) ($C_{\mathrm{w}} = 10{,}000$ ppm H/Si (~0.07 wt.%)).

The situation is more complicated when the influence of water is to be evaluated. When plastic deformation is influenced by the presence of water, the rate of deformation depends on the amount of water dissolved in a mineral. The amount of water dissolved in a mineral depends on the fugacity of water that increases with pressure, leading to the softening with pressure. At the same time, pressure reduces the defect mobility that increases the viscosity with pressure. Fig. 10 illustrates how these two effects compete. At low pressures, fugacity effects dominate and the viscosity decreases with pressure. However, at high pressures, the activation volume effects become important and viscosity increases with pressure if the activation volume is large. The degree to which activation volume effects change the viscosity can be large but the activation volume cannot be well constrained by low-pressure experiments (see Appendix A). This leads to large uncertainties in extrapolating low-pressure data on "wet" rheology to high pressures. However, data above ~1–2 GPa are sensitive to the activation volume effect, so that if one combines low-pressure (<0.5 GPa) and high-pressure (>1 GPa) data, one obtains a tight constraint on the flow law that can be extrapolated to the deep upper mantle if the stress under these conditions is determined with sufficient precision (see also Karato (2006) and Appendix A).

In summary, high-resolution data from the gas-medium apparatus at low pressures (<0.5 GPa) have major limitations in estimating the rheological properties of most of the upper mantle. Therefore the critical development in recent years is to expand the pressure range of quantitative rheological studies to higher pressures under well-characterized chemical environment particularly under well-controlled water content. Obtaining quantitative rheological data above ~0.5 GPa is not trivial because there is no commercially available

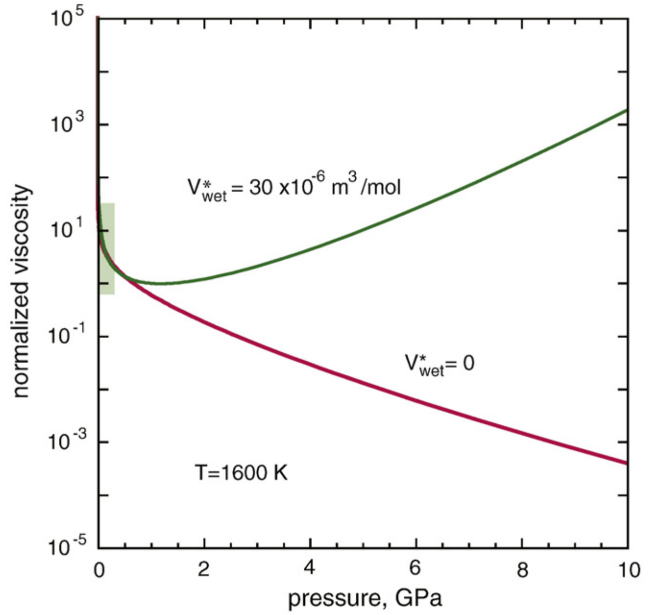


Fig. 10. A diagram illustrating the range of uncertainties if only low-pressure data are used to estimate the viscosity of the deep upper mantle under "wet" conditions. The flow law of a form Eq. (4a) is used and the pressure and temperature dependence of water fugacity calculated from the equation of state was included in this calculation. At low pressures (<0.5 GPa), the fugacity effect dominates and viscosity is reduced with pressure. At high pressures (>1 GPa), the activation volume term can also be important for a reasonably large activation volume. Both effects need to be characterized, but the activation volume effect can be determined only from data at pressures exceeding ~1 GPa. If only data below ~0.5 GPa are used, activation volume is unconstrained that leads to large uncertainties in estimating the viscosity in the deep upper mantle.

apparatus with a guaranteed resolution on mechanical data. A commonly used high-pressure deformation apparatus above 1 GPa is the solid-medium Griggs apparatus (e.g., Tullis and Tullis (1986)). With this apparatus, deformation experiments can be performed to ~3 GPa. However, stress is usually measured by an external load cell that leads to very large uncertainties in the stress estimates (e.g., Gleason and Tullis (1995)). With an improved sample assembly to reduce the influence of friction on the load cell reading, some quantitative results on flow law can be obtained (e.g., Green and Borch (1987); Borch and Green (1987, 1989)). Alternatively, the dislocation density–stress relationship may also be used to determine the stress (Karato and Jung, 2003). Therefore if other key factors such as the water content are well controlled, some quantitative rheological data can be obtained to ~3 GPa using this apparatus. However, the pressure effects (without water) are still small compared to the errors in the stress estimates and the determination of pressure effects with this apparatus is difficult.

During the last several years, important progress has been made in determining the flow laws in upper mantle minerals (mostly olivine) through laboratory studies under high pressures (to ~10 GPa) to characterize the influence of pressure and water. This is a result of development of new types of high-pressure deformation apparatus combined with the use of synchrotron X-ray radiation facility (Karato and Weidner, 2008). However the reported flow laws using these apparatus show a large scatter (Fig. 9). One should recall that in order for any experimental data on plastic deformation to be applied to Earth's deep interior, all of the following points need to be satisfied: (i) A dense sample, either a single crystal or a dense polycrystalline aggregate must be used as a sample. When a powder sample is used, then the influence of compaction and/or grain crushing will complicate the interpretation of the results. (ii) The collected data must correspond to those at steady-state creep. (iii) The flow law operating in the lab must be power-law creep (including a linear rheology) rather than the exponential flow law such the one for Peierls mechanism. (iv) The water content (water fugacity) during an experiment must be well defined (water content must be known and should not change too much during a single run). (v) Stress must be measured accurately enough. (vi) The measurements must be made in a broad enough pressure range in order to determine the flow law that can be extrapolated to the deep upper mantle conditions. As we will discuss, in many of the published results on high-temperature rheology (of olivine), some of these conditions are not met that prevents us from applying them to the upper mantle of Earth.

Some of the early works on high-pressure deformation used powder samples (e.g., Chen et al. (1998); Li et al. (2003, 2004, 2006)), and no detailed discussions will be made on these results except (Li et al., 2006) where rheological data were obtained using a newly developed DDIA apparatus. The importance of points (ii), (iii), and (v) is obvious, but the points (iv) and (vi) may require some discussions. The point (iv) is regarding the water content. In early 1980s, Paterson and his colleagues noted that when a natural rock is used in a deformation experiment at high pressure and temperature, dehydration of hydrous minerals (that are present almost always in natural samples on grain-boundaries) provides a large amount of water to the sample (Chopra and Paterson, 1981, 1984). When a natural sample is used at high temperatures, partial melting could also occur that will reduce the water fugacity if the system is under-saturated with water (Karato, 1986). Also, although no water is intentionally added, a significant amount of water may go into a sample from the pressure medium in a high-pressure experiments (see Huang et al. (2005); Nishihara et al. (2006)). Therefore water content of a sample must be determined both before and after each deformation experiment. However, such measurements were made in very few cases. For example, the experimental study by Green and Borch (1987) (see also Borch and Green (1987, 1989)) reported a large activation volume for nominally "dry" olivine, but a later study showed a substantial amount of water in their samples (Young et al., 1993). Because the water content during the experiments is uncertain, these data cannot be applied to Earth's interior.

High-pressure data (to ~15 GPa) by Karato and Rubie (1997) on "dry" olivine provide some constraints on the activation volume ($V^* = (14–18) \times 10^{-6}$ m^3/mol), but the flow law was only roughly constrained in that study because the stress values were not determined precisely. It is critical to obtain quantitative data for the power-law creep regime at pressures higher than a few GPa to place tight constraints on the rheological properties under deep upper mantle. Above ~4 GPa, quantitative rheological measurements can only be made by the use of newly developed apparatuses DDIA (deformation DIA, (Wang et al., 2003)) or RDA (rotational Drickamer apparatus, (Yamazaki and Karato, 2001; Xu et al., 2005)). However, the use of these high-pressure apparatus (Griggs-apparatus, DDIA, RDA) is not trivial and great care needs to be taken in order to obtain results that can be applied to Earth's interior. There are only a few studies that satisfy the criteria discussed above. For instance, although newly developed DDIA was used in the study by Li et al. (2006) and Raterron et al. (2007), applicability of these results to deep upper mantle is questionable for the following reasons. In the study by Li et al. (2006), where powder samples were used, water content was not well characterized (water content varies from one run to another) and strain in each segment of deformation experiment was small and steady-state was not confirmed. Raterron et al. (2007) reported a small activation energy for "dry" olivine single crystals (~100 kJ/mol compared to the well-known value of ~400–500 kJ/mol for olivine) that suggests the operation of the Peierls mechanism rather than power-law creep (the stress exponent was not determined in their study but rather the stress exponent was assumed in their analysis). In a recent study by Kawazoe et al. (2009), they conducted a series of large-strain deformation experiments to determine the steady-state flow law of olivine aggregates in both power-law creep and the Peierls mechanism regimes to ~10 GPa and ~1900 K (at $\dot{\varepsilon} \approx 10^{-5}$ s^{-1}). The water content in the samples was determined after annealing (before deformation) as well as after each deformation experiment, and shown to be less than ~4 ppm wt ("dry" conditions). They obtained an activation volume of $V^* = (15–20)\times10^{-6}$ m^3/mol for power-law creep (inferred from deformation microstructures and temperature sensitivity of strength). Karato and Jung (2003) determined the water content in each sample and obtained the strength corresponding to nearly steady-state power-law creep and applied the flow law formula of Eq. (3) to determine the flow law parameters, particularly r and V^* through the simultaneous inversion. The flow law parameters such as r and V^* reported in these studies have sound theoretical basis as discussed by Karato and Jung (2003) (see also Chapter 10 of Karato (2008)). Consequently, I consider that the results by Karato and Jung (2003) and by Kawazoe et al. (2009) are the best available data for olivine rheology for the power-law creep regime under "wet" and "dry" conditions respectively. However, results from some other studies are also included in the present analysis to illustrate the importance of flow laws in evaluating the longevity of the deep continental roots.

Note that the results by Kawazoe et al. (2009) are obtained at conditions to ~10 GPa and ~1900 K, and therefore there is no need for extrapolation in pressure (and temperature) when these results are applied to deformation of the deep upper mantle (~200–300 km depth). Only extrapolation is extrapolation in stress, but the stress exponent in the dislocation creep in olivine is well established ($n = 3.0–3.5$; (Bai et al., 1991; Mei and Kohlstedt, 2000b)).

5. Possible explanation of the preservation mechanisms of the continental lithosphere

Let us first evaluate the possible roles of differences in temperature and major element chemistry on the viscosity contrast between deep continental roots and the surrounding mantle. The influence of temperature depends on activation enthalpy, H^*. The value of activation enthalpy depends on activation volume and therefore I included a range of activation volume (from 0 to 30×10^{-6} m^3/mol in this calculation (our preferred value is $(15–20)\times10^{-6}$ m^3/mol). The results are shown in Fig. 11. The results depend somewhat on the assumed activation volume but the conclusion is not very sensitive to the activation volume.

The influence of difference in *Mg#* can be estimated using an empirical relation that the influence of *Mg#* comes mainly through its influence on activation enthalpy, $H^* = H_0^*\left(1-\alpha\cdot\left(1-\frac{Mg}{100}\right)\right)$ where H_0^* is the activation enthalpy for a reference state (*Mg#* = 100) and α (= 0.093) is a constant (Zhao et al., 2009). Again the influence of the difference in *Mg#* is small (less than a factor of 5 for the variation of *Mg#* of less than 5).

I conclude that within the plausible range of parameters, the influence of difference in temperature and major element composition such as *Mg#* is too small to cause a large enough viscosity difference ($>10^3$) to stabilize the deep continental roots against convective erosion. And consequently, the influence of water content is the most likely factor that controls the viscosity contrast. The viscosity contrast that is relevant for erosion of continent is the viscosity contrast between sub-continental mantle and the continental roots. Based on various geological and geochemical observations, sub-continental upper mantle, particularly the one near subducting slabs likely contains a large amount of water (to ~0.1 wt.%). In contrast, the continental root is likely to have undergone a high degree of partial melting and hence depleted with water.

Let us calculate how the given de-watering will increase the viscosity. Given the flow law under "wet" conditions, the degree to which water affects the viscosity (strain-rate) of a material depends on the degree of water depletion (de-watering), but it also depends on the flow law under "dry" conditions. Note that when the water content is reduced, then the contribution to strain-rate from processes similar to deformation under "dry" conditions will also make an important contribution. Therefore the actual strain-rate for a "wet" sample with any arbitrary water content will be given by $\dot\varepsilon \approx \dot\varepsilon_{wet}(C_W) + \dot\varepsilon_{dry}$. In other words, $\dot\varepsilon_{dry}$ gives the minimum value of strain-rate (the maximum value of viscosity). This point was not incorporated in the calculation of the effect of de-watering by Hirth and Kohlstedt (1996), and consequently, the degree of hardening due to de-watering calculated by them is the upper limit (the degree of hardening would be infinite, in their model, if all water is removed).

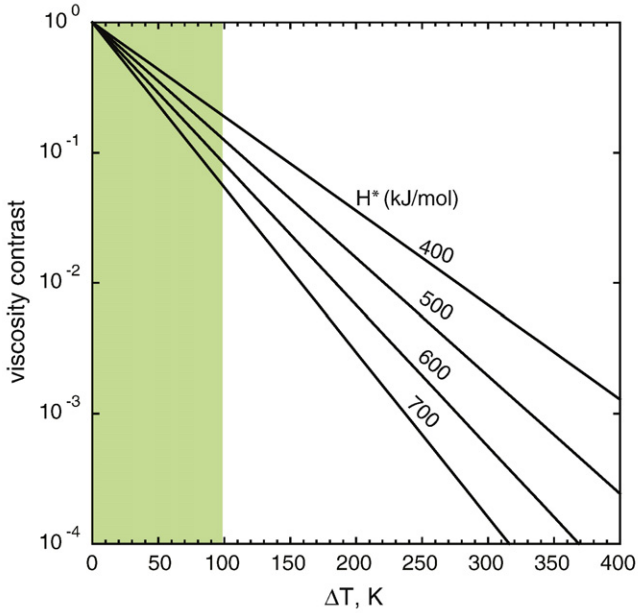


Fig. 11. A diagram showing the influence of temperature difference (ΔT) on the viscosity contrast. The viscosity contrast depends on the activation enthalpy ($H^* = E^* + PV^*$). For a plausible range of activation energy and volume, the values of H^* range from ~400 kJ/mol to ~700 kJ/mol. The viscosity contrast is less than a factor of ~10 for the temperature difference of ~100 K.

Let us consider a case where the water content is reduced by partial melting. As the water content becomes small, strain-rate will be smaller until it becomes comparable to the strain-rate at "dry" (water-free) conditions, $\dot\varepsilon_{dry}$. Therefore the degree of hardening due to water loss is determined by

$$\xi = \frac{\eta(\text{final})}{\eta(\text{initial})} = \frac{\dot\varepsilon\left(C_W^{\text{cont}}\right)}{\dot\varepsilon\left(C_W^0\right)} \tag{5}$$

where viscosity is calculated at a fixed stress, C_W^0 is the water content of a material before partial melting and C_W^{cont} is the water content in the continent after de-watering due to partial melting. Given the flow law parameters for "wet" conditions, we need to know (i) the initial water content, C_W^0, (ii) the degree of water depletion, $\frac{C_W^{\text{cont}}}{C_W^0} = \frac{C_W^{\text{final}}}{C_W^{\text{initial}}}$ as well as (iii) the strain-rate under "dry" conditions $\dot\varepsilon_{dry}$ that defines the base line.

The initial water content can be estimated from petrological observations as discussed earlier (~0.01–0.1 wt.%). The degree of water depletion by partial melting $\frac{C_W^{\text{cont}}}{C_W^0}$ depends on the degree of melting and the mode of melt removal. Using the degree of melting inferred from geochemical observations (30–50%; (Carlson et al., 2005)), the minimum degree of de-watering can be calculated if one assumes "batch melting" (i.e., total equilibrium melting), leading $\frac{C_W^{\text{cont}}}{C_W^0} = \frac{C_W^{\text{final}}}{C_W^{\text{initial}}} \sim 10^{-3}–10^{-2}$. If "fractional melting" (local equilibrium melting) is assumed then much larger degree of water removal can be predicted. Geochemical observations on MORB suggest fractional melting (Johnson et al., 1990) with respect to Ti, Zr and other rare earth elements. However, because diffusion of water (hydrogen) is much faster than these elements, melting might occur as batch melting with respect to water (hydrogen). Therefore I consider that the water depletion by a factor of ~10^{-3} is a conservative estimate for the continental roots.

The degree to which viscosity changes due to de-watering (for a fixed temperature) is shown in Fig. 12. It is shown that the degree of hardening depends both on the degree of water removal and also on the activation volume for "dry" olivine. If the degree of de-watering is minimum (~10^{-3}), then a large activation volume (high viscosity) for "dry" olivine is needed to achieve a factor of ~10^3 change in viscosity without any difference in temperature. I conclude that if temperature is the same between the oceanic and continental deep mantle, it is marginal to establish a large enough viscosity contrast to preserve the continental roots for a few billion years.

I have also evaluated the combined effect of differences in temperature and water content. Combinations of water depletion factor ($C_\mathrm{W}^\mathrm{cont}/C_\mathrm{W}^0 = C_\mathrm{W}^\mathrm{final}/C_\mathrm{W}^\mathrm{initial}$) and temperature difference between the surrounding mantle and the continental roots ($\Delta T = T_\mathrm{surrounding} - T_\mathrm{continent}$) that yield a viscosity contrast of 10^3 are calculated for a range of activation volume for "dry" olivine (Fig. 13). The ($C_\mathrm{W}^\mathrm{cont}/C_\mathrm{W}^0$,$\Delta T$) values that fall to the south east of each curve satisfy the conditions for the preservation of the deep continental roots. For small activation volumes (less than ~5×10^{-6} m^3/mol), a relatively large temperature difference between the continental roots and the surrounding mantle is needed to preserve the continental roots. However, for activation volumes larger than 10×10^{-6} m^3/mol, a broad range of combination of water depletion factor ($C_\mathrm{W}^\mathrm{cont}/C_\mathrm{W}^0$) and temperature difference between the surrounding mantle and the continental roots ($\Delta T = T_\mathrm{surrounding} - T_\mathrm{continent}$) can satisfy the conditions for the preservation of the continental roots. Although water effect alone may result in a large enough strengthening as shown by Fig. 12, a combined temperature and water content effect helps explain the longevity of the deep continental roots. However, a large temperature contrast is unlikely particularly near the continent–ocean boundary where convective erosion occurs because of thermal diffusion.

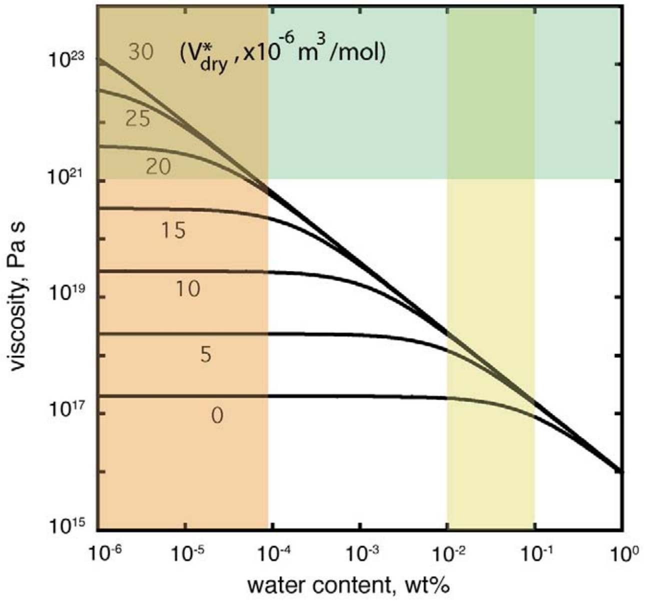


Fig. 12. The influence of water depletion on viscosity (at ~200 km depth: $P = 7$ GPa, $T = 1700$ K, $\sigma = 0.1$ MPa). The degree of strengthening by de-watering depends on (i) the initial water content, (ii) the amount of de-watering and (iii) on the "dry" rheology that defines a base line. The range of initial water content is shown by a yellow region that corresponds to the inferred water contents in the oceanic asthenosphere or in the upper mantle in the subduction zone (0.01 to 0.1 wt.%). The partial melting will reduce the water content (to a region shown by an orange region) and increases the viscosity. The degree of viscosity increase is calculated using the flow law of "wet" olivine by (Karato and Jung, 2003) for a range of activation volume for "dry" olivine. In order to have more than a factor of 10^3 increase in viscosity by de-watering, activation volume larger than ~17×10^{-6} m^3/mol is needed for "dry" olivine if there is no temperature difference between continent and oceanic mantle.

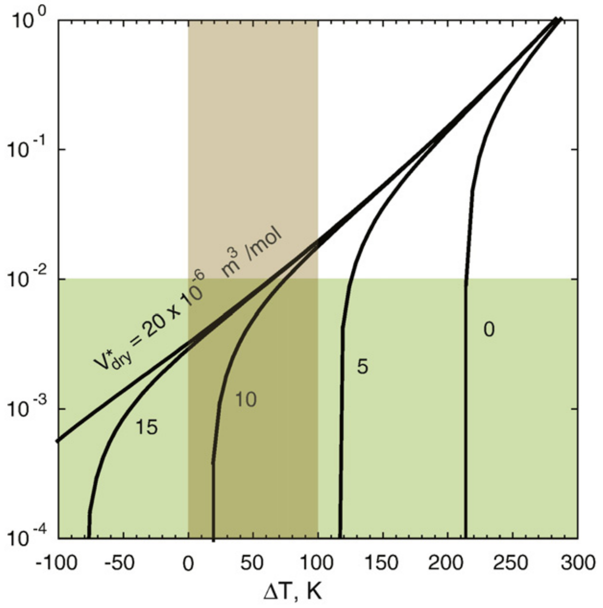


Fig. 13. Combinations of temperature difference ($\Delta T = T_\mathrm{surrounding} - T_\mathrm{continent}$) and water content contrast ($C_\mathrm{W}^\mathrm{cont}/C_\mathrm{W}^0 = C_\mathrm{W}^\mathrm{final}/C_\mathrm{W}^\mathrm{initial}$) that yield the viscosity contrast of 10^3 (at ~200 km depth: $P = 7$ GPa, $T_\mathrm{surrounding}$ (temperature of the surrounding mantle) = 1700 K, $\sigma = 0.1$ MPa) that is needed for the preservation of the continental roots for a range of activation volume for "dry" olivine (V^*_dry). A combination of ($C_\mathrm{W}^\mathrm{cont}/C_\mathrm{W}^0$,$\Delta T$) below ("south–east" of) each curve satisfies the condition for preservation of the deep continental roots. For an activation volume for "dry" olivine of less than 5×10^{-6} m^3/mol, a substantial temperature difference (more than 100 K) is needed to stabilize the continental roots. For a range of activation volume of (10–20×10^{-6} m^3/mol), a broad range of combination of ($C_\mathrm{W}^\mathrm{cont}/C_\mathrm{W}^0$,$\Delta T$) can satisfy the conditions for the preservation of the deep continental roots.

It should be noted that the proposed mechanism of stabilization of continental roots implies that the thickness of the continent correlates with the depth to which extensive partial melting occurred. The thickness of the continents correlated positively with the age (e.g., Artemieva (2006)), and therefore this implies that the melting in the geological past likely occurred at a deeper depth.

6. Discussion

6.1. Rheology of the deep upper mantle

I have shown that the rheological properties of olivine in the power-law creep regime are reasonably well characterized based on the combination of high-resolution low-pressure (<0.5 GPa) data using the gas-medium deformation apparatus and the results of high-pressure studies (>1 GPa). The keys to establish a robust flow law are (i) the robust determination of stress under pressures exceeding ~1 GPa, (ii) careful characterization of water contents in the samples, (iii) the assessment of steady-state deformation and (iv) the examination of flow law regime. Only when all of these issues are carefully examined, a reasonably well-defined flow law can be obtained that can be applied to the upper mantle. These results provide a plausible explanation for the longevity of the deep continental roots.

However, there remain some uncertainties in the flow law of the upper mantle. Main remaining uncertainties are (i) the pressure dependence of diffusion creep in both under "wet" and "dry" conditions and (ii) the role of orthopyroxene. The possible role of diffusion creep remains unclear. Karato and Wu (1993) suggested that diffusion creep might be important under deep upper mantle conditions because the activation volume for diffusion creep is likely smaller than that of dislocation creep. Now the activation volume (and energy) of dislocation creep is well constrained. However, the activation volume for diffusion creep in olivine is not yet constrained well. If one accepts the reported value of activation volume for diffusion creep (15×10^{-6} m^3/mol under "dry" conditions) by Mei and Kohlstedt (2000a), and if one assumes a commonly observed grain-size of 3–10 mm (e.g., Mercier (1980)), then one will find that dislocation creep is always dominant. On the other hand if one uses the reported value of activation volume by Béjina et al. (1999) (~0 m^3/mol), then for the same grain-size, diffusion creep will almost always be dominant in the deep upper mantle, and this is inconsistent with the microstructural observations on mantle rocks. The cause for the difference between these two studies is unknown, but possible causes for the difference are (i) Mei and Kohlstedt's (2000a) results are for Coble creep and therefore the activation volume is that of grain-boundary diffusion whereas Béjina et al. 's (1999) results are for bulk diffusion of Si, and/or (ii) uncertainties in the activation volume in both studies are so large that the discrepancy may be within the uncertainties. In any case, based on the observation that a deep xenolith (>300 km) shows strong LPO (Jin, 1995) (see also Ben Ismail and Mainprice (1998); Ben Ismail et al. (2001)) and the evidence for strong seismic anisotropy in the deep continents (Gung et al., 2003), I consider that the hypothesis of transition from dislocation creep in the shallow upper mantle to diffusion creep in the deep upper mantle does not have strong support.

6.2. Does grain-boundary sliding affect the rheological properties in the dislocation creep regime?

Hirth and Kohlstedt (2003) emphasized that there is some influence of grain-size even in the so-called power-law dislocation creep regime. They argued that the published results are best fitted by a constitutive relationship that includes grain-size, $\dot{\varepsilon}\propto\sigma^{3.5}/L^2$ and Warren and Hirth (2006) proposed that such a deformation mechanism should be added to the deformation mechanism map using the above cited constitutive relationship. Such a constitutive relationship is known to work for creep involving both grain-boundary sliding and dislocation motion at a fixed grain-size (e.g., (Langdon, 1994)). Although such an intermediate regime may indeed play some roles in Earth, its significance is limited and there are some theoretical issues on the use of constitutive relationship proposed by Hirth and Kohlstedt (2003). First and the most important is the fact that the magnitude of such an effect is minor compared to the influence of pressure and water. The conditions under which this type of flow law dominates are limited to a narrow range of the parameter space. For example, Hirth and Kohlstedt's (2003) analysis suggested that this mechanism might be important in a limited grain-size range that modifies the strain-rate by a factor of ~10 (see also a discussion by De Bresser et al. (2001)). This is much smaller than the influence of water (and pressure) discussed earlier.

Second, the formulation of grain-size sensitive creep adopted by Hirth and Kohlstedt (2003) is not applicable to deformation involving dynamic recrystallization. A large number of observations on naturally deformed peridotite show that dynamic recrystallization plays an important role in natural deformation (e.g., Poirier and Nicolas (1975); Urai et al. (1986); Drury and Urai (1990)). When dynamic recrystallization occurs, then grain-size is not an independent variable but becomes a function of stress, and therefore the constitutive relationship needs to be modified.

Finally, if indeed the influence of grain-size is large, then this leads to a serious question as to the validity of using any low-pressure data to infer the pressure dependence of creep. As discussed in detail in the Appendix A, extremely high-resolution measurements of mechanical properties are needed in order to determine the pressure dependence of creep by low-pressure experiments. Grain-size is a parameter that cannot be determined precisely due to its statistical distribution. Consequently, any appreciable influence of grain-size will lead to a serious limitation for the determination of pressure effects on creep from low-pressure experiments. In summary, I conclude that the influence of grain-size in the dislocation creep regime is at most modest and much less than the possible effects of water and pressure.

6.3. Is the continental lithosphere really dry?

The above analysis based on olivine power-law creep rheology shows that if de-watering due to deep partial melting occurs in a material that contains a substantial water (>100 ppm wt), then the viscosity of the depleted residual, i.e., the continental roots, will increase more than by a factor of ~10^3. Consequently, I consider that the hypothesis proposed by Pollack (1986) is now supported by the experimental results on rock deformation. There are additional observations to support the notion of water-depleted continental lithosphere. First is, as I discussed before, the observation of olivine fabric showing the dominance of A-type fabric (e.g., Jin (1995); Ben Ismail and Mainprice (1998); Ben Ismail et al. (2001)) that develops only at water-poor conditions (Karato et al., 2008). Second, the mineralogy of the peridotites from the continental upper mantle shows evidence of a large degree of "depletion" by partial melting, i.e., a large *Mg#* and small Al and Ca content compared to pyrolite (e.g., Carlson et al. (2005)).

A complication, however, is the influence of metasomatism that would add incompatible components including hydrogen (water). It should be noted that melting in the deep upper mantle likely occurs with a help of addition of incompatible components such as water (e.g., Grove et al. (2002)). Consequently, water depletion due to partial melting is likely caused by water addition and therefore it is natural to expect heterogeneous structure of regions that have been affected by deep mantle melting. Extensive metasomatism would introduce a large amount of water (refertilization) that would reduce the strength of the mantle. Evidence of refertilization is provided by Griffin et al. (2003). Such refertilization events would reduce the viscosity in these regions causing the removal of the continental lithosphere as evidenced by Gao et al. (2002) (see also Li et al. (2008)). It is important to reanalyze the water contents in mantle xenoliths to sort out the influence of metasomatism to infer the water content of the "pristine" continental upper mantle.

Another complication is the influence of other volatiles such as CO_2. The upper mantle of Earth contains both CO_2 than H_2O (and compounds containing both carbon and hydrogen such as CH_4) (e.g., Pasteris (1984); Canil and Scarfe (1990); Wood et al. (1996)). The presence of CO_2 affects the fugacity of water and hence indirectly influences the rheological properties of olivine. Also, the dissolution of carbon in olivine might directly affect its rheological properties. Currently nothing is known about the influence of carbon on olivine rheology except for preliminary results on carbon solubility at relatively low pressures (Tingle et al., 1988; Keppler et al., 2003; Shcheka et al., 2006).

7. Concluding remarks

I have reviewed the current status of geological, geochemical and geophysical observations on the continental upper mantle as well as the results of experimental studies on plastic deformation of olivine. I conclude that the viscosity contrast between the deep continental roots and surrounding mantle has the most important control on the longevity of the continental roots. I show that a robust estimate of rheological properties of the upper mantle can only be made by the combination of low-pressure (<0.5 GPa) and high-pressure (>1 GPa) data. However, there have been large discrepancies among the published results on olivine rheology above ~1 GPa. A detailed discussion was provided to resolve this issue. I conclude that the main cause for the discrepancy is the incomplete characterization of water content in many previous studies, and among a large number of studies so far published, only a few of them satisfy key requirements and can be used to infer the deep upper mantle viscosity. These data can explain the longevity of the deep continental roots due mainly to hardening by de-watering as originally proposed by Pollack (1986).

However, there are several issues that need to be investigated in more detail. First, the flow law of olivine under "wet" conditions was investigated only to a low pressure (<2 GPa). Extension of such a study to higher pressures is important. There is a large scatter in the pressure dependence of diffusion and diffusion creep in olivine. Careful studies on the pressure effects on diffusion and diffusion creep under controlled chemical environment need to be conducted. Also, the role of secondary phases such as orthopyroxene needs to be evaluated in more detail. No robust rheological data is currently available for orthopyroxene that is stable only above ~0.5 GPa where a high-resolution gas-medium deformation apparatus cannot be used (for a new study at higher pressures, see Ohuchi and Karato (submitted for publication)).

From the observational side, the "thickness" of the continental upper mantle needs to be studied in more detail. A change in seismic velocity alone provides a loose definition of the thickness of the continent. The depth variation of seismic anisotropy (e.g., Vinnik et al. (1992); Gung et al. (2003); Deschamps et al. (2008)) will provide an important tool to define the thickness of the continents. Distribution of electrical conductivity can also provide some constraints on the water content (e.g., Hirth et al. (2000); Wang et al. (2006)). More detailed works in these areas will be helpful to understand the rheological structure of the deep continents. Progress in both observational and experimental studies is critical to improve our understanding of evolution and dynamics of the deep continental roots.

Acknowledgments

This study was supported by the NSF and COMPRES. Takaaki Kazoe prepared Table 1 and Fig. 9. I thank Eiji Ohtani for the discussion on deep upper mantle melting, Dave Kohlstedt for sending the pre-print of their work, Frédéric Deschamps for the discussion of seismic anisotropy and Kazuhiko Otsuka for helpful comments. Comments by an anonymous reviewer and the guest editor, Irina Artemieva, were helpful in clarifying the presentation.

Appendix A. Needs for a broad pressure range for constraining rheological properties of minerals in the upper mantle

Consider a power-law creep formula,

$$\dot{\varepsilon} = A \cdot \sigma^n \cdot \exp\left(-\frac{E^* + PV^*}{RT}\right). \qquad (A-1)$$

Because the pressure dependence is exponential, it is weak at low pressures but becomes strong at high pressures. The magnitude of pressure effects increases exponentially with pressure. Experimental errors also increase with pressure but the errors increase much less than exponentially. Therefore the effect of larger signature of pressure effects over-compensates for the effect of increased error and high-pressure experiments have major advantage over low-pressure experiments in determining the pressure effects on rheological properties. To see this, let us estimate the errors in the estimation of activation volume. The activation volume can be determined by the measurements of strain-rate (or stress) at two (or more) different pressures,

$$V^* = -\frac{RT}{P_1 - P_2}\log\frac{\dot{\varepsilon}\,(P_1)}{\dot{\varepsilon}\,(P_2)} \text{ for constant stress tests} \qquad (A-2a)$$

or by

$$V^* = \frac{nRT}{P_1 - P_2}\log\frac{\sigma(P_1)}{\sigma(P_2)} \text{ for constant strain}-\text{rate tests} \qquad (A-2b)$$

where $P_{1,2}$ are pressures at which measurements of strain-rate (or stress) are made. Therefore if errors are due solely to the errors in strain-rate (or stress) measurements and if errors follow the Gaussian distribution, the standard deviation in the estimated V^*, $\delta(V^*)$, is related to the standard deviation in the measured stress or strain-rate as

$$\frac{\delta(V^*)}{V^*} \approx \frac{2RT}{(P_1 - P_2)V^*}\delta(\log\dot{e}\ ) \qquad (A-3a)$$

or

$$\frac{\delta(V^*)}{V^*} \approx \frac{2nRT}{(P_1 - P_2)V^*}\delta(\log\sigma) \qquad (A-3b)$$

respectively where $\delta(X)$ is the standard deviation of a quantity X. Fig. A1 shows a plot of $\delta(V^*)$ versus P_1–P_2 for various values of $\delta(log\sigma)$ or $\delta(log\dot{\varepsilon})$. It is seen that the error in the activation volume is reduced significantly with the increase in the span of pressure. For a small pressure span such as 0.3 GPa, precise determination of pressure effect is difficult particularly when strain-rate is sensitive to grain-size (note that Hirth and Kohlstedt (2003) argue that in both dislocation creep and diffusion creep regimes, strain-rate is sensitive to grain-size. In such a case the uncertainties in grain-size measurements (and the uncertainties in the variation of grain-size during a run) cause a large error in the estimate of pressure dependence of deformation).

The nature of errors is somewhat different for "wet" rheology. This comes from the fact that the rheological properties under "wet" conditions are sensitive to two factors, $f_{H2O}(P,T)$ and $\exp(-\frac{PV^*}{RT})$, and the pressure dependence of these two factors is different. At low pressures (say < 0.5 GPa), the variation of $\exp(-\frac{PV^*}{RT})$ is small but the water fugacity increases almost linearly with pressure, $f_{H2O}(P,T) \propto P$. Consequently, strain-rate for a given stress increases with pressure when pressure is increased under water-saturated conditions (see the lower pressure portion of Fig. 10). Activation volume is essentially determined by the difference between the strain-rate (stress) extrapolated from low pressures following $f_{H2O}(P,T) \propto P$ (with $V^* = 0$) and the actual strain-rate (stress), and therefore activation volume is constrained by a large difference in strain-rate (or stress) as illustrated in Fig. 10 (see a pressure region above ~1 GPa).

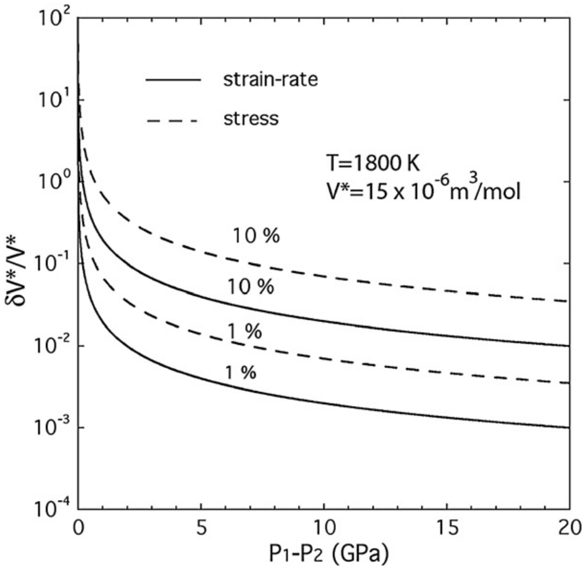


Fig. A1. Errors in the estimated activation volume δV^* determined by a measurement of creep strength (stress) or strain-rate for a pressure range of P_1–P_2 with errors of stress or strain-rate of 10% or 1%. Solid curves correspond to those from stress measurements (at constant strain-rate), and broken lines to those from strain-rate measurements (at constant stress). Relative errors, $\delta V^*/V^*$, depends on the activation volume, δV^*, and the results for $\delta V^* = 15 \times 10^{-6}$ m^3/mol are shown.

Consequently, even from relatively low-pressure range (say $P<2$ GPa), activation volume can be determined with reasonable precision under "wet" conditions as shown by (Karato and Jung, 2003).

References

Abbott, D., Burgess, L., Longhi, J., 1994. An empirical thermal history of the Earth's upper mantle. Journal of Geophysical Research 99, 13835–13850.

Anderson, D.L., 1979. The deep structure of continents. Journal of Geophysical Research 84, 7555–7560.

Artemieva, I.M., 2006. Global 1° × 1° thermal model TC1 for the continental lithosphere: implications for lithosphere secular evolution. Tectonophysics 416, 245–277.

Artemieva, I.M., 2009. The continental lithosphere: reconciling thermal, seismic, and petrologic data. Lithos 109, 23–46.

Bai, Q., Mackwell, S.J., Kohlstedt, D.L., 1991. High temperature creep of olivine single crystals 1. Mechanical results for buffered samples. Journal of Geophysical Research 96, 2441–2463.

Béjina, F., Jaoul, O., Liebermann, R.C., 1999. Activation volume of Si diffusion in San Carlos olivine: implications for upper mantle rheology. Journal of Geophysical Research 104, 25529–25542.

Bell, D.R., Rossman, G.R., 1992a. The distribution of hydroxyl in garnets from the subcontinental mantle of southern Africa. Contributions to Mineralogy and Petrology 111, 161–178.

Bell, D.R., Rossman, G.R., 1992b. Water in Earth's mantle: the role of nominally anhydrous minerals. Science 255, 1391–1397.

Bell, D.R., Rossman, G.R., Moore, R.O., 2004. Abundance and partitioning of OH in a high-pressure magmatic system: megacrysts from the Monastery Kimberlite, South Africa. Journal of Petrology 45, 1539–1564.

Ben Ismail, W., Mainprice, D., 1998. An olivine fabric database: an overview of upper mantle fabrics and seismic anisotropy. Tectonophysics 296, 145–157.

Ben Ismail, W., Barroul, G., Mainprice, D., 2001. The Kaapvaal craton seismic anisotropy: petrological analyses of upper mantle kimberlite nodules. Geophysical Research Letters 28, 2497–2500.

Borch, R.S., Green II, H.W., 1987. Dependence of creep in olivine on homologous temperature and its implication for flow in the mantle. Nature 330, 345–348.

Borch, R.S., Green II., H.W., 1989. Deformation of peridotite at high pressure in a new molten cell: comparison of traditional and homologous temperature treatments. Physics of Earth and Planetary Interiors 55, 269–276.

Boullier, A.M., Gueguen, Y., 1975. SP-mylonites: origin of some mylonites by superplastic flow. Contributions to Mineralogy and Petrology 50, 93–104.

Boyd, F.R., 1973. A pyroxene geothem. Geochemica and Cosmochemica Acta 37, 2533–2546.

Bussod, G.Y., Katsura, T., Rubie, D.C., 1993. The large volume multi-anvil press as a high P–T deformation apparatus. Pure and Applied Geophysics 141, 579–599.

Cadek, O., Fleitout, L., 2003. Effect of lateral viscosity variation in the top 300 km on the geoid and dynamic topography. Geophysical Journal International 152, 566–580.

Canil, D., Scarfe, C.M., 1990. Phase relations in peridotite + CO_2 systems to 12 GPa: implications for the origin of kimberlite and carbonatite stability in the Earth's upper mantle. Journal of Geophysical Research 95, 15805–15816.

Carlson, R.W., Irving, A.J., Hearn Jr., B.C., 1999a. Chemical and isotopic systematics of peridotite xenoliths from the Williams Kimberlite, Montana: clues to processes of lithosphere formation, modification and destruction. In: Gurney, J.J., Gurney, J.L., D.P. M, S.H., Richardson (Eds.), Proceedings of the 7th International Kimberlite Conference. Red Roof Design, Cape Town, pp. 90–98.

Carlson, R.W., et al., 1999b. Re–Os systematics of lithospheric peridotites: implications for lithosphere formation and preservation. In: Gurney, J.J., Gurney, J.L., D.P. M, S.H., Richardson (Eds.), Proceedings of the VIIth International Kimberlite Conference. Red Roof Design, Cape Town, pp. 99–108.

Carlson, R.W., Pearson, D.G., James, D.E., 2005. Physical, chemical, and chronological characteristics of continental mantle. Review of Geophysics 43. doi:10.1029/2004RG000156.

Carter, N.L., 1976. Steady state flow of rocks. Review of Geophysics and Space Physics 14, 301–360.

Chen, J., Inoue, T., Weidner, D.J., Wu, Y., Vaughan, M.T., 1998. Strength and water weakening of mantle minerals, olivine, wadsleyite and ringwoodite. Geophysical Research Letters 25, 575–578.

Chopra, P.N., Paterson, M.S., 1981. The experimental deformation of dunite. Tectonophysics 78, 453–573.

Chopra, P.N., Paterson, M.S., 1984. The role of water in the deformation of dunite. Journal of Geophysical Research 89, 7861–7876.

Danchin, R.V., 1979. Minerals and bulk chemistry of garnet lherzolite and garnet harzburgite xenoliths from the Premier Mine, South Africa. In: Boyd, F.R., Meyer, H.O.A. (Eds.), The Mantle Samples: Inclusions in Kimelrlites and Other Volcanics. American Geophysical Union, Washington DC, pp. 104–126.

De Bresser, J.H.P., ter Heege, J.H., Spiers, C.J., 2001. Grain size reduction by dynamic recrystallization: can it result in major rheological weakening? International Journal of Earth Sciences 90, 28–45.

Deschamps, F., Lebedev, S., Meier, T., Trampert, J., 2008. Azimuthal anisotropy of Rayleigh-wave phase velocities in the east-central United States. Geophysical Journal International 173, 827–843.

Dixon, J.E., Leist, L., Langmuir, J., Schiling, J.G., 2002. Recycled dehydrated lithosphere observed in plume-influenced mid-ocean-ridge basalt. Nature 420, 385–389.

Doin, M., Fleitout, L., Christensen, U.R., 1997. Mantle convection and stability of depleted and undepleted continental lithosphere. Journal of Geophysical Research 102, 2771–2787.

Drury, M.R., Urai, J., 1990. Deformation-related recrystallization processes. Tectonophysics 172, 235–253.

Dziewonski, A.M., Anderson, D.L., 1981. Preliminary reference Earth model. Physics of Earth and Planetary Interiors 25, 297–356.

Frost, H.J., Ashby, M.F., 1982. Deformation Mechanism Maps. Pergamon Press, Oxford. 168 pp.

Gao, S., Rudnick, R.L., Carlson, R.W., McDonough, W.F., Liu, Y.-S., 2002. Re–Os evidence of ancient mantle lithosphere beneath the north China craton. Earth and Planetary Science Letters 198, 307–322.

Gleason, G.C., Tullis, J., 1995. A flow law for dislocation creep of quartz aggregates determined with the molten slat cell. Tectonophysics 247, 1–23.

Goetze, C., 1975. Sheared lherzolites: from the point of view of rock mechanics. Geology 172–173.

Green II., H.W., Borch, R.S., 1987. The pressure dependence of creep. Acta Metallurgica 35, 1301–1305.

Griffin, W.L., O'Reilly, S.Y., Natapov, L.M., Ryan, C.G., 2003. The evolution of lithospheric mantle beneath the Kalahari craton and its margins. Lithos 71, 215–241.

Grove, T.L., Parman, S.W., Bowring, S.A., Price, R.C., Baker, M.B., 2002. The role of an H_2O-rich fluid component in the generation of primitive basaltic andesites and andesites from the Mt. Shasta region, N. California. Contributions to Mineralogy and Petrology 142, 375–396.

Gung, Y., Romanowicz, B., Panning, M., 2003. Global anisotropy and the thickness of continents. Nature 422, 707–711.

Handy, M.R., 1989. Deformation regimes and the rheological evolution of fault zones in the lithosphere: the effects of pressure, temperature, grain size and time. Tectonophysics 163, 119–152.

Hier-Majumder, S., Anderson, I.M., Kohlstedt, D.L., 2005. Influence of protons on Fe–Mg interdiffusion in olivine. Journal of Geophysical Research 110. doi:10.1029/2004JB003292.

Hirschmann, M.M., 2006. Water, melting, and the deep Earth H_2O cycle. Annual Review of Earth and Planetary Sciences 34, 629–653.

Hirth, G., Kohlstedt, D.L., 1996. Water in the oceanic upper mantle—implications for rheology, melt extraction and the evolution of the lithosphere. Earth and Planetary Science Letters 144, 93–108.

Hirth, G., Kohlstedt, D.L., 2003. Rheology of the upper mantle and the mantle wedge: a view from the experimentalists. In: Eiler, J.E. (Ed.), Inside the Subduction Factory. American Geophysical Union, Washington DC, pp. 83–105.

Hirth, G., Evans, R.L., Chave, A.D., 2000. Comparison of continental and oceanic mantle electrical conductivity: is Archean lithosphere dry? Geochemistry, Geophysics, Geosystems 1. doi:10.1029/2000GC000048.

Houseman, G.A., Molnar, P., 1997. Gravitational (Rayleigh–Taylor) instability of a layer with non-linear viscosity and convective thinning of continental lithosphere. Geophysical Journal International 128, 125–150.

Huang, X., Xu, Y., Karato, S., 2005. Water content of the mantle transition zone from the electrical conductivity of wadsleyite and ringwoodite. Nature 434, 746–749.

Ito, E., Harris, D.M., Anderson, A.T., 1983. Alteration of oceanic crust and geologic cycling of chlorine and water. Geochemica and Cosmochemica Acta 47, 1613–1624.

James, D.E., Fouch, M.J., VanDecar, J.C., Lee, v.f., Group, K.S.S., 2001. Tectospheric structure beneath southern Africa. Geophysical Research Letters 28, 2485–2488.

Jin, D., 1995. Deformation Microstructures of Some Ultramafic Rocks. M Sc Thesis, University of Minnesota, Minneapolis, 115 pp.

Jin, D., Karato, S., Obata, M., 1998. Mechanisms of shear localization in the continental lithosphere: inference from the deformation microstructures of peridotites from the Ivrea zone, northern Italy. Journal of Structural Geology 20, 195–209.

Johnson, K.T.M., Dick, H.J.B., Shimizu, N., 1990. Melting in the oceanic upper mantle: an ion microprobe study of diopsides in abyssal peridotites. Journal of Geophysical Research 95, 2661–2678.

Jordan, T.H., 1975. The continental tectosphere. Review of Geophysics and Space Physics 13, 1–12.

Jordan, T.H., 1981. Continents as a chemical boundary layer. Philosophical Transactions of the Royal Society of London. A 301, 359–373.

Karato, S., 1986. Does partial melting reduce the creep strength of the upper mantle? Nature 319, 309–310.

Karato, S., 2006. Influence of hydrogen-related defects on the electrical conductivity and plastic deformation of mantle minerals: a critical review. In: Jacobsen, S.D., van der Lee, S. (Eds.), Earth's Deep Water Cycle. American Geophysical Union, Washington DC, pp. 113–129.

Karato, S., 2008. Deformation of Earth Materials: Introduction to the Rheology of the Solid Earth. Cambridge University Press, Cambridge. 463 pp.

Karato, S., Jung, H., 2003. Effects of pressure on high-temperature dislocation creep in olivine polycrystals. Philosophical Magazine. A. 83, 401–414.

Karato, S., Rubie, D.C., 1997. Toward experimental study of plastic deformation under deep mantle conditions: a new multianvil sample assembly for deformation experiments under high pressures and temperatures. Journal of Geophysical Research 102, 20111–20122.

Karato, S., Weidner, D.J., 2008. Laboratory studies of rheological properties of minerals under deep mantle conditions. Elements 4, 191–196.

Karato, S., Wu, P., 1993. Rheology of the upper mantle: a synthesis. Science 260, 771–778.

Karato, S., Toriumi, M., Fujii, T., 1980. Dynamic recrystallization of olivine single crystals during high temperature creep. Geophysical Research Letters 7, 649–652.

Karato, S., Paterson, M.S., Fitz Gerald, J.D., 1986. Rheology of synthetic olivine aggregates: influence of grain-size and water. Journal of Geophysical Research 91, 8151–8176.

Karato, S., Jung, H., Katayama, I., Skemer, P.A., 2008. Geodynamic significance of seismic anisotropy of the upper mantle: new insights from laboratory studies. Annual Review of Earth and Planetary Sciences 36, 59–95.

Kawazoe, T., Karato, S., Otsuka, K., Jing, Z., Mookherjee, M., 2009. Shear deformation of dry polycrystalline olivine under deep upper mantle conditions using a rotational Drickamer apparatus (RDA). Physics of the Earth and Planetary Interiors 174, 128–137.

Kelly, R.K., Kelemen, P.B., Jull, M., 2003. Buoyancy of the continental upper mantle. Geochemistry, Geophysics, Geosystems 4. doi:10.1029/2002GC000399.

Keppler, H., Wiedenbeck, M., Shcheka, S.S., 2003. Carbon solubility in olivine and the mode of carbon storage in the Earth's mantle. Nature 424, 414–416.

Kocks, U.F., Argon, A.S., Ashby, M.F., 1975. Thermodynamics and kinetics of slip. Progress in Materials Sciences 19, 1–288.

Kohlstedt, D.L., Mackwell, S.J., 1998. Diffusion of hydrogen and intrinsic point defects in olivine. Zeitschrift für Phisikalische Chemie 207, 147–162.

Kohlstedt, D.L., Evans, B., Mackwell, S.J., 1995. Strength of the lithosphere: constraints imposed by laboratory measurements. Journal of Geophysical Research 100, 17587–17602.

Kopylova, M.G., Russell, J.K., 2000. Chemical stratification of cratonic lithosphere: constraints from the northern Slave craton, Canada. Earth and Planetary Science Letters 181, 71–87.

Korenaga, J., 2006. Archean geodynamics and the thermal evolution of the Earth. In: Benn, K., Mareschal, J.C., Condie, K.C. (Eds.), Archean Geodynamic Processes. American Geophysical Union, Washington DC, pp. 7–32.

Langdon, T.G., 1994. A unified approach to grain boundary sliding in creep and superplasticity. Acta Metallurgica et Materials 42, 2437–2443.

Lee, C.-T.A., 2003. Compositional variation of density and seismic velocities in natural peridotites at STP conditions: implications for seismic imaging of compositional heterogeneities in the upper mantle. Journal of Geophysical Research 108. doi:10.1029/2003JB0002413.

Lenardic, A., Moresi, L.N., 1999. Some thoughts on the stability of cratonic lithosphere: effects of buoyancy and viscosity. Journal of Geophysical Research 104, 12747–12759.

Li, L., Raterron, P., Weidner, D.J., Chen, J., 2003. Olivine flow mechanisms at 8 GPa. Physics of Earth and Planetary Interiors 138, 113–129.

Li, L., Weidner, D.J., Raterron, P., Chen, J., Vaughan, M.T., 2004. Stress measurements of deforming olivine at high pressure. Physics of Earth and Planetary Interiors 143/144, 357–367.

Li, L., et al., 2006. Deformation of olivine at mantle pressure using the D-DIA. European Journal of Mineralogy 18, 7–19.

Li, Z.-X.A., Lee, C.-T.A., Peslier, A.H., Lenardic, A., Mackwell, S.J., 2008. Water contents in mantle xenoliths from the Colorado Plateau and vicinity: implications for the mantle rheology and hydration-induced thinning of continental lithosphere. Journal of Geophysical Research 113. doi:10.1029/2007JB005540.

Mackwell, S.J., Kohlstedt, D.L., Paterson, M.S., 1985. The role of water in the deformation of olivine single crystals. Journal of Geophysical Research 90, 11319–11333.

Mei, S., Kohlstedt, D.L., 2000a. Influence of water on plastic deformation of olivine aggregates, 1. Diffusion creep regime. Journal of Geophysical Research 105, 21457–21469.

Mei, S., Kohlstedt, D.L., 2000b. Influence of water on plastic deformation of olivine aggregates, 2. Dislocation creep regime. Journal of Geophysical Research 105, 21471–21481.

Mercier, J.-C.C., 1979. Peridotite xenoliths and dynamics of kimberlite intrusion. In: Boyd, F.R., Meyer, H.O.A. (Eds.), The Mantle Sample: Inclusions in Kimberlites and Other Volcanics. American Geophysical Union, Washington DC.

Mercier, J.-C.C., 1980. Magnitude of the continental lithospheric stresses inferred from rheomorphic petrology. Journal of Geophysical Research 85, 6293–6303.

Mercier, J.-C.C., Nicolas, A., 1975. Textures and fabrics of upper mantle peridotites as illustrated by xenoliths from basalts. Journal of Petrology 16, 454–487.

Mitrovica, J.X., 1996. Haskell [1935] revisited. Journal of Geophysical Research 101, 555–569.

Mookherjee, M. and Karato, S., submitted for publication. Solubility of water in pyrope-rich garnet at high pressure and temperature. Chemical Geology.

Nishihara, Y., Shinmei, T., Karato, S., 2006. Grain-growth kinetics in wadsleyite: effects of chemical environment. Physics of Earth and Planetary Interiors 154, 30–43.

Ohuchi, T. and Karato, S., submitted for publication. Plastic deformation of orthopyroxene under the lithospheric conditions. Nature Geoscience.

Pasteris, J.D., 1984. Kimberlites: complex mantle melts. Annual Review of Earth and Planetary Sciences 12, 133–153.

Paterson, M.S., 1982. The determination of hydroxyl by infrared absorption in quartz, silicate glass and similar materials. Bulletin Mineralogie 105, 20–29.

Paterson, M.S., 1990. Rock deformation experimentation. In: Duba, A.G., Durham, W.B., Handin, J.W., Wang, H.F. (Eds.), The Brittle–Ductile Transition in Rocks: The Heard Volume. American Geophysical Union, Washington DC, pp. 187–194.

Poirier, J.-P., Nicolas, A., 1975. Deformation induced recrystallization due to progressive misorientation of subgrains, with special reference to mantle peridotites. Journal of Geology 83, 707–720.

Pollack, H.K., 1986. Cratonization and the thermal evolution of the mantle. Earth and Planetary Science Letters 80, 175–182.

Post, R.L., 1977. High-temperature creep of Mt. Burnett dunite. Tectonophysics 42, 75–110.

Poudjom Djomani, Y.H., O'Reilly, S.Y., Griffin, W.L., Morgan, P., 2001. The density structure of subcontinental lithosphere through time. Earth and Planetary Science Letters 184, 605–621.

Raleigh, C.B., Kirby, S.H., Carter, N.L., Avé Lallemant, H.G., 1971. Slip and the clinoenstatite transformation as competing processes in enstatite. Journal of Geophysical Research 76, 4011–4022.

Raterron, P., Chen, J., Li, L., Weidner, D.J., Cordier, P., 2007. Pressure-induced slip system transition in forsterite: single-crystal rheological properties at mantle pressure and temperature. American Mineralogist 92, 1436–1445.

Ringwood, A.E., 1975. Composition and Structure of the Earth's Mantle. McGraw-Hill, New York. 618 pp.

Röhm, A.H.E., Snieder, R., Goes, S., Trampert, J., 2000. Thermal structure of continental upper mantle inferred from S-wave velocity and surface heat flow. Earth and Planetary Science Letters 181, 395–407.

Rudnick, R.L., McDonough, W.F., O'Connell, R.J., 1998. Thermal structure, thickness and composition of continental lithosphere. Chemical Geology 145, 395–411.

Rybacki, E., Dresen, G., 2004. Deformation mechanism maps for feldspar rocks. Tectonophysics 382, 173–187.

Shapiro, N., Hager, B.H., Jordan, T.H., 1999a. Stability and dynamics of the continental tectosphere. Lithos 48, 135–152.

Shapiro, S.S., Hager, B.H., Jordan, T.H., 1999b. The continental tectosphere and Earth's long-wavelength gravity field. Lithos 48, 115–133.

Shcheka, S.S., Wiedenbeck, M., Frost, D.J., Keppler, H., 2006. Carbon solubility in mantle minerals. Earth and Planetary Science Letters 245, 730–742.

Shito, A., Karato, S., Matsukage, K.N., Nishihara, Y., 2006. Toward mapping water content, temperature and major element chemistry in Earth's upper mantle from seismic tomography. In: Jacobsen, S.D., Lee, S.v.d. (Eds.), Earth's Deep Water Cycle. American Geophysical Union, Washington DC, pp. 225–236.

Skemer, P.A., Karato, S., 2008. Sheared lherzolite revisited. Journal of Geophysical Research 113. doi:10.1029/2007JB005286.

Stolper, E.M., Newman, S., 1994. The role of water in the petrogenesis of Mariana trough magmas. Earth and Planetary Science Letters 121, 293–325.

Takahashi, E., 1986. Melting of a dry peridotite KLB-1 up to14 GPa: implications on the origin of peridotitic upper mantle. Journal of Geophysical Research 91, 9367–9382.

Takahashi, E., 1990. Speculations on the Archean mantle: missing link between komatiite and depleted garnet peridotite. Journal of Geophysical Research 95, 15941–15945.

Tingle, T.N., Green II., H.W., Finnerty, A.A., 1988. Experiments and observations bearing on the solubility and diffusivity of carbon in olivine. Journal of Geophysical Research 93, 15289–15304.

Tullis, T.E., Tullis, J., 1986. Experimental rock deformation. In: Hobbs, B.E., Heard, H.C. (Eds.), Mineral and Rock Deformation. American Geophysical Union, Washington DC, pp. 297–324.

Urai, J.L., Means, W.D., Lister, G.S., 1986. Dynamic recrystallization in minerals. In: Hobbs, B.E., Heard, H.C. (Eds.), Mineral and Rock Deformation: Laboratory Studies. American Geophysical Union, Washington DC, pp. 166–199.

van der Wal, D., Chopra, P.N., Drury, M., Fitz Gerald, J.D., 1993. Relationships between dynamically recrystallized grain size and deformation conditions in experimentally deformed olivine rocks. Geophysical Research Letters 20, 1479–1482.

Vinnik, L., Makayeva, L.I., Milev, A., Usenko, A.Y., 1992. Global patterns of azimuthal anisotropy and deformations in the continental mantle. Geophysical Journal International 111, 433–447.

Walter, M.J., 2005. Melt extraction and compositional variability in mantle lithosphere. In: Carlson, R.W. (Ed.), Treatise on Geochemistry. Elsevier, Amsterdam, pp. 363–394.

Wang, Y., Durham, W.B., Getting, I.C., Weidner, D.J., 2003. The deformation-DIA: A new apparatus for high temperature triaxial deformation to pressures up to 15 GPa. Review of Scientific Instruments 74, 3002–3011.

Wang, D., Mookherjee, M., Xu, Y., Karato, S., 2006. The effect of water on the electrical conductivity in olivine. Nature 443, 977–980.

Warren, J.M., Hirth, G., 2006. Grain size sensitive deformation mechanisms in naturally deformed peridotites. Earth and Planetary Science Letters 248, 438–450.

Wenk, H.-R., 1985. Preferred Orientation in Deformed Metals and Rocks: An Introduction to Modern Texture Analysis. Academic Press Inc., Orland. 610 pp.

Wood, B.J., Pawley, A.R., Frost, D.R., 1996. Water and carbon in the Earth's mantle. Philosophical Transactions of the Royal Society of London 354, 1495–1511.

Xia, Q.K., Sheng, Y.M., Yang, X.Z., Yu, H.M., 2005. Heterogeneity of water in garnet from UHP eclogites, eastern Dabieshan, China. Chemical Geology 224, 237–246.

Xu, Y., Nishihara, Y., Karato, S., 2005. Development of a rotational Drickamer apparatus for large-strain deformation experiments under deep Earth conditions. In: Chen, J., Wang, Y., Duffy, T.S., Shen, G., Dobrzhinetskaya, L.F. (Eds.), Frontiers in High-Pressure Research: Applications to Geophysics. Elsevier, Amsterdam, pp. 167–182.

Yamazaki, D., Karato, S., 2001. High pressure rotational deformation apparatus to 15 GPa. Review of Scientific Instruments 72, 4207–4211.

Young, T.E., Green II., H.W., Hofmeister, A.M., Walker, D., 1993. Infrared spectroscopic investigation of hydroxyl in β-$(Mg,Fe)_2SiO_4$ and coexisting olivine: implications for mantle evolution and dynamics. Physics and Chemistry of Minerals 19, 409–422.

Zhang, S., Christensen, U.R., 1993. Some effects of lateral viscosity variations on geoid and surface velocities induced by density anomalies in the mantle. Geophysical Journal International 114, 531–547.

Zhao, Y.-H., Zimmerman, M.E., Kohlstedt, D.L., 2009. Effect of iron content on the creep behavior of olivine: 1. Anhydrous conditions. Earth and Planetary Science Letters 287, 229–240.

Zimmerman, M.E., Kohlstedt, D.L., 2004. Rheological properties of partially molten lherzolite. Journal of Petrology 45, 275–298.