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REVIEW **Open Access**

Rheology of the lower mantle: a review

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Abstract

We review our current understanding of the rheological properties of the lower mantle based both on materials science and geophysics points of view. We assume a simple model of the lower mantle that is made of only two minerals: bridgmanite (Br) (Mg,Fe)SiO_3 and ferropericlase (Fp) (Mg,Fe)O, and address a question of (i) which mineral is weaker (lower viscosity), (ii) how does lower mantle viscosity change with depth and location, and (iii) discuss implications for shear localization. We first review plausible mechanisms of deformation based on the deformation mechanism map on the normalized stress and temperature space. We conclude that likely mechanism of deformation in the lower mantle is either diffusion creep or power-law dislocation creep. Based on this review, we discuss recently proposed models by Cordier and his group (Cordier in Nature 481:177–181, 2012; Cordier in Nature 613:303–306 , 2023) where either asthermal creep (i.e., low-temperature plasticity) or pure climb creep (not power-law dislocation creep) would play an important role. We conclude that these models are not acceptable because (1) many aspects of their models are incompatible with experimental observations and theoretical models of deformation of most materials including oxides and metals and (2) these models are not consistent with the distribution of seismic anisotropy. Hence, we focus on power-law dislocation creep and diffusion creep. We review previously published results on deformation (by dislocation creep) and diffusion, we conclude that Fp is weaker than Br. The radial (depth) depth and lateral variation of viscosity is discussed based on the estimated activation volume and estimated variation of grain-size. Geophysical studies suggest only modest depth variation of viscosity that demands relatively small activation volume (V^* ($< 3 \times 10^{-6}$ m^3/mol)). Plausible models to explain small activation volume are discussed including the role of extrinsic diffusion. Grain-size also controls viscosity if deformation is by diffusion creep. Okamoto and Hiraga (J Geophys Res, 2024. 10.1029/2023JB027803), Solomatov et al. (Phys Earth Planet Inter 129:265–282, 2002) estimated the grain-size evolution in the lower mantle based on the kinetics of grain-growth and the role of a phase transformation. In contrast, there are other papers (e.g., Paul et al. in Prog Earth Planet Sci 11:64, 2024; Rozel in Geochem Geophys Geosyst, 2012. 10.1029/2012GC004282) where grain-size distribution is estimated assuming that grain-size is controlled by dynamic recrystallization. The validity of assumption is questionable because dynamic recrystallization occurs due to deformation by dislocation creep but not by diffusion creep and the absence of seismic anisotropy indicates that diffusion creep dominates in most of the lower mantle. Finally, we review the published models of shear localization that would explain the long-term preservation of geochemical reservoirs in the lower mantle. Accepting that two minerals (Fp and Br) in the lower mantle have largely different viscosity, Ballmer et al. (Nat Geosci 10:236–240, 2017) proposed that the presence of regions of compositional difference (difference in Fp/Br ratio) leads to localized deformation (deformation mainly in the weaker regions). However, in addition to the ad hoc nature of this model, there is no strong evidence for the presence of large variation in Fp/Br in the lower mantle that makes the validity of this model questionable. There are some papers where processes of shear localization are explored without invoking the presence of regions of large rheological contrast. Thielmann et al. (Geochem Geophys Geosyst, 2020. 10.1029/2019GC008688) presented the results of theoretical study of deformation of initially homogeneous

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two-phase mixture (Fp and Br) and showed that deformation causes the elongation of a weak Fp that promotes shear localization. In this model, the rheological contrast between Fp and Br was assumed to be independent of strain. However, Cho and Karato (J Geophys Res 2022. 10.1029/2021JB022673 ; Phys Earth Planet Inter, 2024. 10.1016/j.pepi.2024) showed that when deformation is by diffusion creep, the rheological contrast increases with strain due to the evolution of stress concentration caused by grain elongation. They showed that this will promote strain weakening particularly in simple shear that would lead to shear localization. Consequently, the tendency for shear localization is stronger in their model than a model where rheological contrast is assumed to be independent of strain.

1 Introduction

Chemical evolution and dynamics of Earth's interior depend on the nature of (partial) melting and resultant melt-solid segregation and on the rheological properties. Although the understanding of the nature of melting and the solid-melt separation have made a major progress during the last a few decades (e.g., Andrault et al. 2011; Dasgupta and Hirschmann 2006; Karato et al. 2020; Ohtani 1983, 1988), our understanding of its rheological properties in the deep mantle such as the lower mantle is still highly limited. This is largely due to the difficulties in conducting experimental studies on plastic deformation than conducting melting behavior under the deep Earth conditions. The goal of this paper is to review the current status of our understanding of the rheological properties of the lower mantle.

One of the important questions on the geochemistry of the lower mantle is to explain why the lower mantle has a large-scale geochemical heterogeneity for billions of years as evidenced by the trace element compositions of the ocean island basalt (OIB) as compared to the upper mantle that has relatively homogeneous composition as inferred from the composition of the mid-ocean ridge basalt (MORB) (Hofmann 1997). The long-term preservation of geochemically distinct regions in the lower mantle implies that mixing (stirring) by convection is inefficient in the lower mantle although mixing is efficient in the upper mantle.

Inefficient mixing of the lower mantle as compared to the efficient mixing of the upper mantle could be explained if (i) the lower mantle has much higher viscosity than the upper mantle, and/or (ii) if regions with different composition have largely different viscosity (e.g., Manga (1996), and/or (iii) the degree of localization and the time dependence of convection (e.g., Olson et al. 1984; van Keken et al. 2002). Regarding the magnitude of viscosity, not only the viscosity at a given pressure (P) and temperature (T), but we also need to understand the pressure dependence of viscosity because the pressure range in the lower mantle is large (~24–135 GPa). The degree of localization is closely related to the relative strength of co-existing minerals. If a volumetrically minor phase (ferropericlase (Fp)) is weaker than a major phase (bridgmanite (Br)), shear localization likely occurs that will limit the degree of mixing (e.g., Chapter 16 of (Karato 2008)).

Currently, our understanding of these three issues is highly limited. For example, regarding the relative strength, two contrasting models have been proposed: Fp is weaker than Br (e.g., Girard et al. 2016; Karato 1989b)) or Fp is stronger than Br (Cordier et al. 2023). Regarding the depth variation of viscosity, experimental studies are highly limited (e.g., Mei et al. 2008; Tsujino et al. 2022; Yamazaki and Irifune 2003)) and theoretical studies on diffusion show a wide variety of pressure dependence depending on the nature of diffusion (extrinsic versus intrinsic, diffusing species (O, Si or Mg(Fe)) (e.g., Cordier et al. 2023; Karato 1981)). Also, regarding the pressure dependence of viscosity, a broad range of activation volume (V*) has been reported ($2\text{–}10 \times 10^{-6}$ m^3/mol, e.g., (Ammann et al. 2010; Cordier et al. 2023; Karato 1981; Yamazaki and Irifune 2003)) that would lead to a difference in viscosity more than 10 orders of magnitude.

The goal of this review is to summarize the current status of our understanding of the rheological properties of the lower mantle from materials science and geophysical points of view. We consider a simple model of the lower mantle that is composed of ~60–70% bridgmanite (Br, $(Mg,Fe)SiO_3$) and ~20% ferropericlase (Fp, (Mg,Fe)O) (and ~10% other phases such as $CaSiO_3$ perovskite (davemaoite)), and will address a question of either Fp is weaker than Br, and also seek a mineral physics model to explain geophysically inferred only modest depth dependence of lower mantle viscosity. The ignorance of volumetrically minor phases can be justified because volumetrically minor phases would have an important contribution to the bulk strength of the material only when minor phases are much weaker than others and minor phases are interconnected. Although davemaoite is weaker than other phases (Immoor et al. 2022), its volume fraction is limited except for subducted oceanic crust, and it does not form a continuous film (e.g., (Kuwahara et al. 2018)). Consequently, its role in deformation of the lower mantle is limited.

The question of relative strength of two minerals is important because if volumetrically minor Fp is weaker than volumetrically major Br, then strain weakening and resultant shear localization is likely (e.g., (Handy 1994), see also Chapter 16 of (Karato 2008)) that will explain the long-term preservation of geochemical reservoirs (e.g., (Hofmann 1997)). However, as we will review below, largely contrasting models have been published. (Karato 1989b; Yamazaki and Karato 2001) proposed that Fp is weaker than Br, whereas (Cordier et al. 2023) proposed that Fp is stronger than Br. If the latter model were correct, deformation in the lower mantle will not be localized much and the explanation of long-term preservation of geochemical reservoirs would be difficult. Therefore, it is important to understand the reasons for these different conclusions. Also, the interpretation of a modest depth variation of lower mantle viscosity is a challenge and one would need to consider some details of microscopic mechanisms of deformation that control the pressure dependence of viscosity.

In this paper, we first discuss the dominant mechanism of plastic deformation in the lower mantle based on a combination of results of mineral physics studies as compared to the results of the distribution of seismic anisotropy. Then, we examine the relative strength (creep strength) of Fp and Br is examined for two deformation mechanisms (diffusion creep, power-law dislocation creep), and we discuss the depth variation of the creep strength in the lower mantle with these mechanisms, and finally we discuss plausible processes for shear localization that would explain the long-term presence of geochemical reservoirs in the lower mantle. Figure 1 illustrates the mantle structure with the emphasis on the (likely) contrast in the viscosity of co-existing phases between the upper mantle and the lower mantle.

IWL is interconnected weak layer and LBF is load-bearing framework. This figure also includes a sketch of localized deformation suggested by the experimental results by (Girard et al. 2016) showing that ferropericlase (Fp) is substantially weaker than bridgmanite (Br). Two insets show the nature of mixing of two phases: in the upper mantle, a large volume is made of a weak mineral (olivine) in the hot asthenosphere (e.g., (Farla et al. 2013)), whereas in the lower mantle volumetrically smaller mineral, ferropericlase (Fp) is likely weaker than the major phase (bridgmanite, Br). However, either Fp is weaker than Br is controversial. Also, the depth variation of rheological properties is not well understood for the lower mantle where pressure changes from ~ 24 GPa (at 660 km) to ~ 135 GPa (at 2890 km).

2 Geophysical and geochemical constraints on the mantle composition and its implications for the rheological properties

The composition of the lower mantle can be inferred from the seismic wave velocities and densities determined by geophysical studies combined with the results of high-pressure experiments of phase transformations. These studies show that the lower mantle is composed of bridgmanite ((Mg,Fe)SiO_3), ferropericlase ((Mg,Fe)O) and other minor phases such as davemaoite (CaSiO3) (e.g., (Ringwood 1991)). There were some debate about the volume fraction of these phases in the lower mantle. If the mantle composition is independent of depth, then the lower mantle would have ~ 60–70% bridgmanite, ~ 15–25% ferropericlsae and ~ 5–10% other phases (e.g., davemaoite) (e.g., Murakami et al. 2012; Ringwood 1991)). However, because the difference in elastic properties between these minerals is so small that it is difficult to estimate the fraction of these phases from seismological observations alone. We use the results of seismic tomography suggesting the whole mantle convection (e.g., Bercovici and Karato 2003; Káreason and van der Hilst 2000)) to assume a simple model corresponding to depth-independent composition. We will also discuss that a pure bridgmanite model has a difficulty in explaining the lower mantle rheology inferred from geodynamic modeling.

3 Some basics of plastic deformation

3.1 Deformation mechanisms: a review

Unlike elastic deformation, plastic deformation occurs with a variety of mechanisms. Also, although elastic properties can be compared using a single property, elastic constant, plastic properties cannot be compared using a single property. Rather than a single property, plastic properties are characterized by a relation between stress and strain rate. For each mechanism, there is a certain formula to represent such a relationship.

Therefore, the first step in studying the rheological properties of the lower mantle is to identify plausible mechanism of deformation under the lower mantle conditions. This can be done using a deformation mechanism map (Frost and Ashby 1982). Figure 2a shows a deformation mechanism map of MgO, one of the important minerals in the lower mantle. This map is for 100 μ m grain-size on the parameter space of stress (σ) and temperature (T). Note that both stress and temperature are shown as normalized stress and temperature, i.e., $\sigma/\mu(P)$ and $T/T_m(P)$ where $\mu(P)$ is shear modulus at pressure P and $T_m(P)$ is melting temperature at pressure P. Deformation mechanism maps can also be constructed for different grain-size.

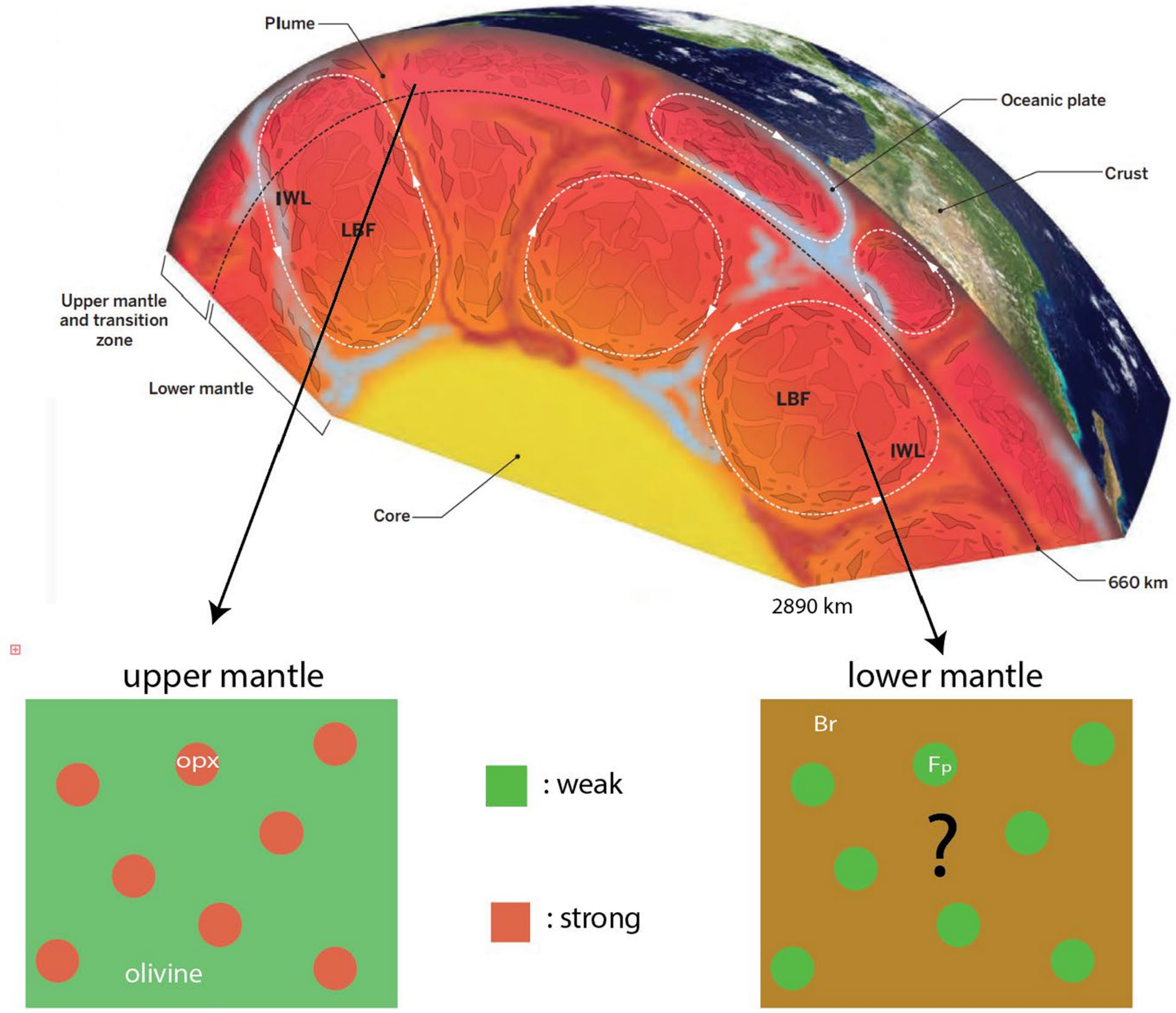

Fig. 1 A schematics of mantle structure (modified from (Chen 2016))

Three mechanisms may plan an important role: (i) Plasticity (Peierls mechanism), (ii) power-law (dislocation) creep and (iii) diffusion creep (diffusional flow). Plasticity is a deformation mechanism that occurs at low temperatures and/or high stress where deformation occurs by the glide motion of dislocations. In this regime, dislocation can move when there is high enough stress to overcome the barrier imposed by the chemical bonding and/or the barrier caused by impurity particles. In this regime, the strength of a matter (stress needed for deformation at a given strain-rate) is only weakly sensitive to temperature. In this regime, deformation can occur even at $T = 0$ K if stress exceeds a threshold value called eth Peierls stress. Power-law dislocation creep is deformation where strain is caused by dislocation glide whereas strain-rate is controlled by dislocation climb. Because dislocation climb is highly sensitive to temperature, the strength of a matter is highly sensitive to temperature, Deformation in this regime causes lattice-preferred orientation (LPO) that results in seismic anisotropy. Finally, in diffusion creep regime, deformation is due to diffusional mass transport across the grains. Consequently, deformation in this regime is sensitive to grain-size and to temperature. Because diffusional mass transport does not rotate the crystallographic orientations, there will be no LPO and hence no seismic anisotropy if deformation is by diffusion creep.

An inspection of Fig. 2a shows that deformation of the lower mantle $\left({}^{T}/{}_{T_m} = 0.5 - 0.8, \sigma/\mu < 10^{-3}\right)$

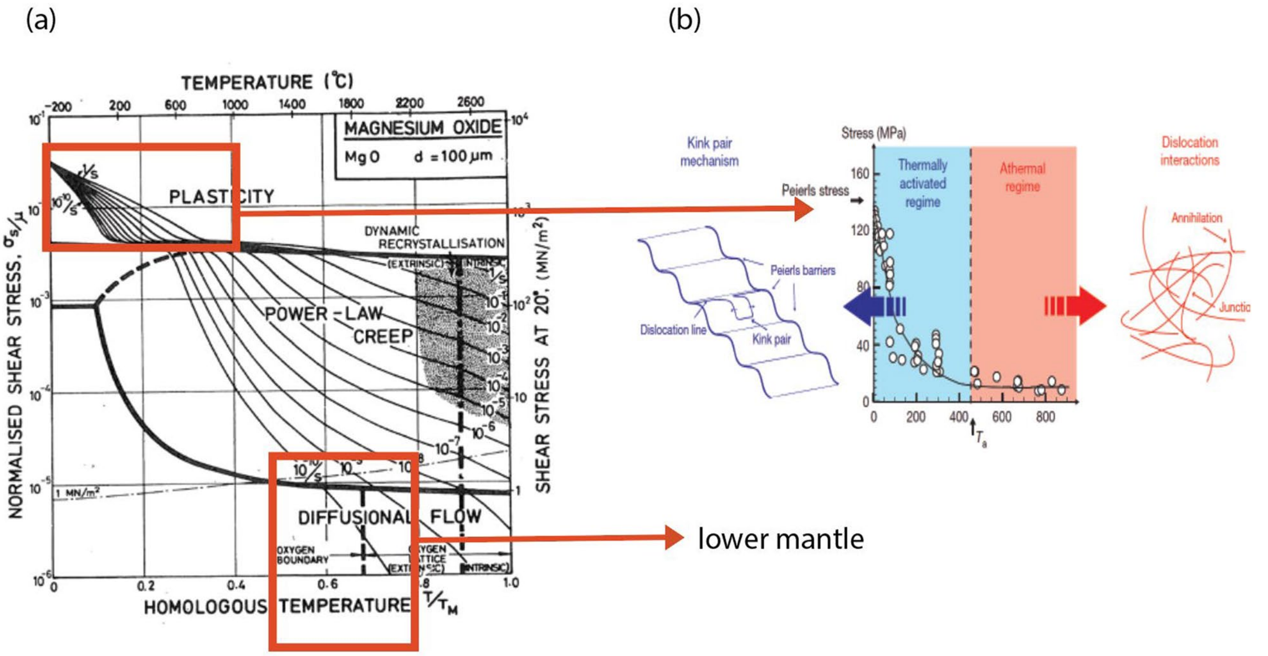

Fig. 2 The deformation mechanism map (stress-temperature relationships for a range of strain-rate) of MgO (d: grain-size). **a** A complete deformation mechanism map for a broad range of stress and temperature (from (Frost and Ashby 1982). Stress and temperature are shown as normalized fashion ($\sigma/\mu(P)$; μ : shear modulus), $T/T_m(P)$; T_m : melting temperature). Typical stress-temperature conditions in the lower mantle, and the stress-temperature relationship considered by Cordier et al. (2012) are shown by two boxes. **b** A stress-temperature relationship considered by Cordier et al. (2012) (from their Fig. 1)

occurs either by the power-law dislocation creep or by diffusion creep. This general trend applies to all other solids (Frost and Ashby 1982). Although there are less data available for materials with the perovskite structures, similar deformation mechanism map can be constructed and we obtain the same conclusion (e.g., (Karato 1998a, b)). Plasticity (the Peierls mechanism) operates only at low temperatures and high stress such as in the bending subducted slabs (e.g., (Goetze and Evans 1979)).

However, the use of deformation mechanism map to infer deformation mechanisms involves large uncertainties because of a large uncertainty in the laboratory data as well as in the estimated temperature, stress and grain-size. Seismic anisotropy provides a clue on the deformation mechanism. Seismic anisotropy is most likely caused by the non-random crystallographic orientation distribution in a rock (lattice-preferred orientation (LPO); see Chapter 14 and 21 of (Karato 2008)) that occurs when deformation is by dislocation glide but not for deformation by diffusion creep. The above conclusion, either diffusion creep or power-law dislocation creep dominates in the lower mantle, is consistent with the distribution of seismic anisotropy (see the next section).

4 Geophysical observations relevant to rheological properties of the lower mantle

Although materials science studies (either experimental or theoretical) play a key role in the study of mantle rheology, both experimental and theoretical studies are challenging and contain major limitations as will be discussed later in this paper. Therefore, it is important to examine constraints from geophysical studies such as the estimates of mantle viscosity from the analyses of time-dependent deformation and the distribution of seismic anisotropy and of viscosity.

4.1 Seismic anisotropy

Let us first discuss the observations on seismic anisotropy. We discuss this first because seismic anisotropy provides strong constraints on deformation mechanism. Any discussion on rheology such as the depth and lateral variation in viscosity can be made only when we know the relevant deformation mechanism(s).

Needs Seismic anisotropy in the mantle shows a marked depth dependence (Fig. 3a). Seismic anisotropy is absent in most of the lower mantle (Meade et al. 1995). (Karato et al. 1995) used this observation to suggest that most of the lower mantle deforms by diffusion creep (or superplasticity). However in the bottom layer

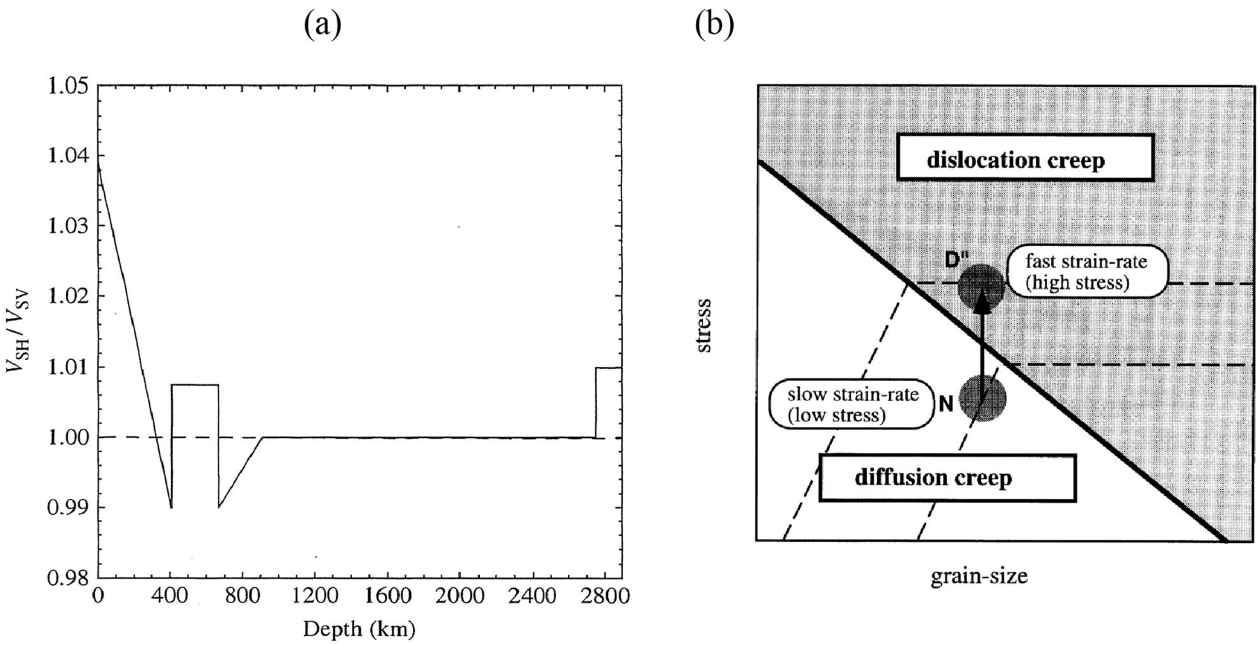

Fig. 3 **a** Depth variation of two seismic shear wave velocities (V_{SH} and V_{SV}) (based on the results by (Montagner and Kennett 1996) based on long wavelength normal mode seismology (reproduced from (Karato 2008))) In the most lower mantle, V_{SH}=V_{SV}, i.e., seismologically isotropic. But near the bottom of the lower mantle (the D″ layer), there is substantial anisotropy. Presence of anisotropic regions in the D″ layer is also shown by (Wolf et al. 2024). **b** A schematic model to explain the distribution of seismic anisotropy (from (Karato 1998b)) In the boundary layer (e.g., the D″ layer), stress is higher than in other regions (N: normal region) and hence LPO (lattice-preferred orientation) would develop

of the lower mantle (the D" layer), substantial anisotropy is detected (e.g., (Montagner and Kennett 1996) (similar results were reported by the body-wave seismology (Wolf et al. 2024)).

There are some reports on seismic anisotropy in the shallow lower mantle (e.g., Ferreira et al. 2019; Sturgeon et al. 2019)). Similar to seismic anisotropy in the D" layer, seismic anisotropy in the shallow lower mantle is likely caused by high stress caused by slab subduction.

(Karato 1998a, b) proposed that deformation in most of the lower mantle is by diffusion creep whereas in the boundary layers (e.g., the D" layer) where stress is higher, deformation may occur by dislocation creep (Fig. 3b). A numerical study by (McNamara et al. 2002) on the stress distribution supports this notion. The possibility of LPO (lattice-preferred orientation) of post-perovskite is discussed as a cause for the anisotropy in the D" layer by Wolf et al. (2024), but almost nothing is known about the rheological properties of post-perovskite, and furthermore, elastic anisotropy of post-perovskite is substantially smaller than that of Fp (Yamazaki and Karato 2007). Therefore, we prefer a model to explain D" layer anisotropy by the LPO of Fp.

We note that there are some previous works (e.g., (Miyazaki et al. 2013)) where LPO development by diffusion creep was proposed. However, their conclusion is likely an artefact of their experimental approach as discussed by (Karato 2024).

4.2 Geophysical estimates of mantle viscosity from time-dependent deformation and gravity anomalies associated with convection

Time-dependent deformation such as the post-glacial isostatic adjustment (GIA) or post-seismic deformation provides some constraints on mantle rheology (e.g., Nakada and Lambeck 1987, 1989; Peltier 1998)). Also, the observations on gravity field as combined with the topography and the density distribution estimated from seismic tomography provide constraints on mantle viscosity (e.g., (Hager and Clayton 1989)). An example of one-dimensional model is shown in Fig. 4 (from (Lau et al. 2016)). Those models are subjected to large uncertainties, but a general feature is a modest increase in viscosity (by a factor of ~ 100) in the most part of the lower mantle (from 660 km to ~ 2000 km). The modest increase in viscosity implies that the pressure dependence of viscosity is small (in terms of activation volume (V*), it should be less than ~ 3×10^{-6} m^3/mol). This provides a strong constraint on the plausible atomistic processes of deformation in the lower mantle as we will show later in this paper. Somewhat different approaches were used to infer lower mantle viscosity. (Čížkova et al. 2012) used sinking

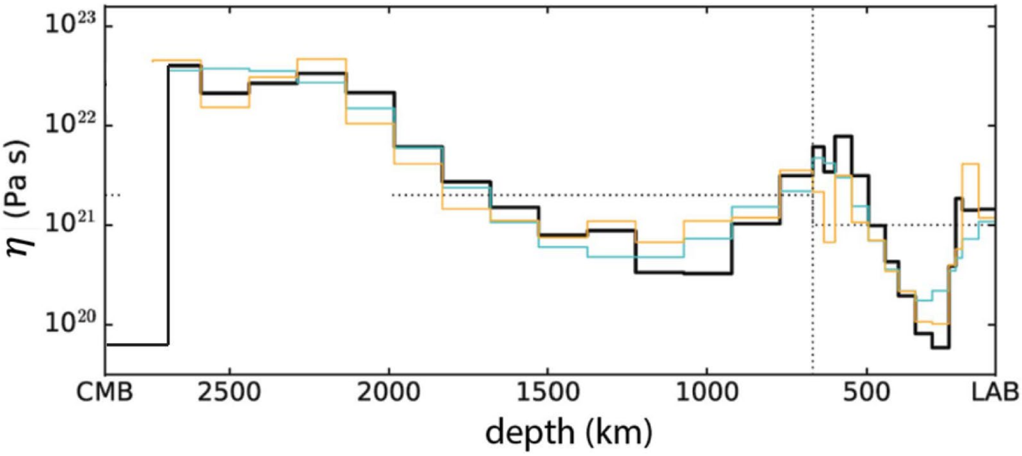

Fig. 4 A viscosity (η)-depth model based on geophysical observations (after (Lau et al. 2016) and (Nakada and Karato 2012))

velocity of subducted lithosphere in the lower mantle to infer the lower mantle viscosity of ~ 10^{22} Pa s that is not far from the viscosity model shown in Fig. 4**.**

We note, however, that using a geodynamic modeling approach, (Han et al. 2024) discussed a difficulty in reproducing the geoid and slab structure form the viscosity model similar to the one show in Fig. 4**.** One possible cause for this difficulty is the viscosity versus temperature and depth (pressure) relation that they used ($\eta(T,r) = \eta_r(r) \cdot exp[E \cdot (0.5 - T)]$ (E: normalized activation energy, T: normalized temperature, *r*: radius (pressure)). This equation implies that cold subducting slabs will have substantially higher viscosity than the surrounding mantle and hence they will not deform much in the deep mantle. This is inconsistent with the results of seismic tomography showing intensive deformation of slabs in the transition zone particularly in the west Pacific (e.g., Fukao et al. 2001; Kárason and van der Hilst 2000)). (Karato et al. 2001) suggested that a conventional model such as the relation (1) in (Han et al. 2024)) does not apply when a phase transformation results in grain-size reduction.

A modest increase in viscosity is supported also by the seismological model of depth variation of seismic wave attenuation (Q^{-1}). Attenuation is negatively correlated with viscosity ($Q^{-1} \propto \eta^{-\alpha} (\alpha \approx 0.3)$ (η : viscosity); Chapter 3 and 18 of (Karato 2008)). Geophysical analyses show that attenuation is less (by a factor of 3–5) in the lower mantle than the upper mantle (e.g., (Romanowicz and Mitchell 2015) indicating that the viscosity of the lower mantle is modestly higher (by a factor of 10–100) than in the upper mantle.

Note that there is no constraint on the viscosity of the lowermost lower mantle (~ 2700–2890 km) by a study of (Lau et al. 2016)) although the estimated viscosity of $(2\text{–}3) \times 10^{22}$ Pa s is shown in Fig. 4. The viscosity of this layer is important in discussing the topography of the core-mantle boundary (e.g., (Kanda and Stevenson 2006)) and the melt content in the D" layer (Hernlund and Jellinek 2010) (see also (Karato 2014)). The viscosity of the D" layer can be estimated using observations such as tidal deformation that have sensitivity to the viscosity of the D" layer. Using these observations, (Nakada and Karato 2012) showed a substantially lower viscosity (~ 10^{18}–10^{19} Pa s) in the D" layer than the viscosity shown in Fig. 4**.** This makes a big difference in our understanding the dynamics of this layer (e.g., (Karato 2014)) (see also (Peltier and Drummond 2010)).

5 Comments on Cordier's models

However, recently, several papers were published where different models of deformation mechanisms in the lower mantle were proposed. (Cordier et al. 2012) proposed that deformation of MgO in the lower mantle occurs mostly by the athermal mechanism where creep strength (viscosity) is insensitive to temperature. (Boioli et al. 2017) (see also (Carrez et al. 2024)) proposed that deformation in the lower mantle is by pure climb creep in which strain is due to dislocation climb. And importantly, these authors concluded that Fp is stronger (higher viscosity) than Br in the lower mantle (see also (Cordier et al. 2023)). This is in contrast to the results reported by (Girard et al. 2016; Yamazaki and Karato 2001), and these two models will predict different strain-weakening behavior and hence different tendency for shear localization. Also, these two models have different implications for the distribution of seismic anisotropy. Therefore, it is important to evaluate the validity of these models based both on materials science and geophysical points of view.

(Cordier et al. 2012) considered the stress-temperature relationship shown in Fig. 2b to propose a model of thermal-athermal transition in the mode of plastic deformation. First, the transition from high-temperature sensitivity to low-temperature sensitivity as temperature increases is not a trend that is observed in a broad range of temperatures. In a broader temperature range, deformation mechanism changes from plasticity regime (temperature insensitive regime) at low temperatures to power-law creep (or diffusion creep) (temperature sensitive regimes) at high temperatures (Fig. 2a). In other words, deformation mechanism changes from athermal regime at low temperatures to thermal regime at high temperature (this is opposite to (Cordier et al. 2012)'s model). Indeed, experimental studies on plastic deformation of MgO (and other materials) at high temperature show highly temperature sensitive flow law $\left(\dot{\varepsilon} \propto \exp\left(-\frac{H^*}{RT}\right)\right)$ (e.g., (Stretton et al. 2001)) and produce strong lattice-preferred orientation (LPO) (e.g., (Yamazaki and Karato 2002)). In contrast, (Cordier et al. 2012)'s model predicts that deformation at high temperature is athermal and hence the strength is insensitive to temperature and no LPO would develop.

The concept of thermal mechanism at low temperatures and athermal mechanism at high temperatures proposed by (Cordier et al. 2012) seems odd. To understand why they obtained such an unconventional view, we compare a complete deformation mechanism map for MgO (Fig. 2a, from (Frost and Ashby 1982)) with a stress-temperature relationship considered by (Cordier et al. 2012) (Fig. 2b). Note that the stress-temperature relationship considered by (Cordier et al. 2012) corresponds to the stress-temperature relationship in the "plasticity" regime (a low-temperature regime) in the complete deformation mechanism map. In this regime, the stress-temperature relation shows a gradual change in the slope (on the stress-temperature plane; Fig. 2b). In a standard model (e.g., Frost and Ashby 1982)), this is attributed to the transition from the Peierls mechanism (lattice friction-control) at low temperatures to obstacle-control at high temperatures (Frost and Ashby 1982) within the plasticity (athermal) regime (not from thermal mechanism at low temperatures to athermal mechanism at high temperatures as proposed by (Cordier et al. 2012)).

Furthermore, according to (Cordier et al. 2012)'s model, no lattice-preferred orientation (LPO) would develop by high-temperature dislocation creep because the development of LPO requires rotational deformation (see Chapter 14 of (Karato 2008)) but deformation in the high-temperature athermal regime in (Cordier et al. 2012)'s model is mostly by dislocation climb that produces irrotational deformation hence no LPO. However, experimental studies on deformation of (Mg,Fe)O aggregates in the dislocation creep regime (e.g., Long et al. 2005; Yamazaki and Karato 2002)) at high temperatures show strong LPO that is inconsistent with the model by (Cordier et al. 2012). Furthermore, (Cordier et al. 2012)'s model predicts seismic anisotropy does not exist anywhere in the lower mantle that is not consistent with seismological observations showing the presence of seismic anisotropy in an appreciable fraction of the D" layer (e.g., Montagner and Kennett 1996; Wolf et al. 2024)). In (Cordier et al. 2012)'s model, this thermal/athermal transition is due to the high crystal symmetry of Fp that makes Fp stronger than Br because of strong dislocation–dislocation interactions in Fp than Br. This discussion is incorrect because many metals have highly symmetric crystal structures and hence dislocations with different Burgers vectors can interact to form a network to cause work hardening. However, experimental and theoretical studies show that this work hardening is balanced by recovery leading to a steady-state creep where materials with simple crystal structures tend to show low strength compared to materials with more complicated crystal structures (e.g., (Karato 2013)). Because we believe that Cordier's model is not valid as we just showed, we do not accept their conclusion that Fp is stronger than Br.

Another issue of (Cordier et al. 2012)'s model is that they consider that in the effective stress exponent in the high-temperature athermal regime is infinite. In their model, the effective stress exponent in the low-temperature thermal regime decreases with temperature, but above a threshold temperature (Ta), i.e., in the athermal regime, it becomes infinite (their Fig. 3). This is physically incorrect. As shown in Supplementary Materials, the effective stress exponent indeed decreases with temperature (this point agrees with (Cordier et al. 2012)'s analysis on the low-temperature regime), but at higher temperatures, the stress exponent remains small ($n = 1$ for diffusion creep, $n \sim 3$ for power-law dislocation creep).

Second, (Boioli et al. 2017) estimated the mobility of dislocation glide and climb in Br at high pressures by theoretical calculations and concluded that under high pressures, glide velocity is reduced more than climb velocity and hence dislocation climb itself produces strain and controls the strain-rate at high pressures ("pure climb creep" model that was first proposed by (Nabarro 1967)). They discussed that the strain-rate for pure climb creep exceeds the strain-rate by diffusion creep for a plausible range of grain-size and proposed that this mechanism would dominate in the lower mantle (see also (Carrez et al. 2024)).

However, pure climb creep model contains a few fundamental problems both from materials science and geophysical points of view and cannot be accepted. Pure climb creep will produce no LPO and hence no seismic anisotropy. This is not consistent with the laboratory observations (e.g., (Tsujino et al. 2016)) nor with seismological observations showing the presence of seismic anisotropy in broad regions in the D" layer as summarized before.

Also pure climb creep has another major problem from materials science point of view. (Boioli et al. 2017) proposed that although the velocity of dislocation climb (υ_c) is lower than the velocity of dislocation glide (υ_g) at low pressures ($\upsilon_c < \upsilon_g$), climb velocity becomes higher than glide velocity ($\upsilon_c > \upsilon_g$) at high pressures and hence both strain and strain-rate are controlled by dislocation climb (pure climb creep) at high pressures. The conclusion of $\upsilon_c > \upsilon_g$ at high pressures is based on a theoretical estimate of $\upsilon_g(P,T)$ and $\upsilon_c(P.T)$ with an assumption that a dislocation is saturated with jogs. This assumption leads to higher activation energy and hence activation volume for dislocation glide than dislocation climb leading to $\upsilon_c > \upsilon_g$ at high pressures.

However, the validity of this conclusion is questionable from the materials science point of view. Let us first review a theoretical model of the pressure (and temperature) dependence of dislocation velocity (see Chapter 10 of (Karato 2008)). Generally, dislocation velocity depends on pressure (P), temperature (T) and stress (σ) (we ignore the influence of chemical environment (e.g., oxygen fugacity, hydrogen content) to simplify the discussion). At relatively high temperatures and low stress (this is a case for a majority of the lower mantle), stress dependence of dislocation velocity can be assumed to follow $\upsilon(P,T) = M(P,T)\cdot\sigma$ where $M(P,T)$ is dislocation mobility. Therefore, the velocity ratio $\xi_{g/c}(P,T) = \upsilon_g(P,T)/\upsilon_c(P,T)$ is independent of stress, i.e., $\xi_{g/c}(P,T) = M_g(P,T)/M_c(P,T)$ (we ignore the influence of chemical environment for simplicity).

The conclusion by (Boioli et al. 2017) that $\upsilon_g < \upsilon_c$ for bridgmanite comes from the stronger pressure dependence of glide velocity than climb velocity in their model. The pressure dependence of dislocation mobility can be given by the following relationship that is applicable to a broad range of thermally activated processes (see Chapter 10 of (Karato 2008)),

$$M(P,T) = M_o\exp(-\frac{G^*(P)}{RT}) \approx M_o\exp(-\frac{E^*+PV^*}{RT}) \tag{1}$$

with

$$V^* = \frac{E^*}{K}(2\gamma - \frac{2}{3}) \tag{2}$$

where K is bulk modulus, γ is the Grüneisen parameter (a non-dimensional constant) whose value is ~ 1.5 for most processes (dislocation glide, dislocation climb, diffusion etc.) (e.g., Chapter 4 of (Karato 2008)). Consequently, we have

$$\xi_{g/c}(P,T) = \frac{M_{go}}{M_{co}}\exp\left\{-\frac{\left[1+\frac{P}{K}\left(2\gamma-\frac{2}{3}\right)\right](E_g^*-E_c^*)}{RT}\right\} \tag{3}$$

and hence

$$\frac{\xi_{g/c}(P,T)}{\xi_{g/c}(0,T)} = \exp\left[-\frac{\frac{P}{K}\left(2\gamma-\frac{2}{3}\right)\left(E_g^*-E_c^*\right)}{RT}\right]. \tag{4}$$

This relation indicates that the pressure dependence of velocity contrast depends on the difference in activation energy ($\Delta E = E_g^* - E_c^*$). Some examples of the relation (4) are shown in Fig. 5.

$\xi_{g/c}(P,T) = \upsilon_g(P,T)/\upsilon_c(P,T)$ and $\upsilon_g \propto \exp\left(-\frac{E_g^*+PV_g^*}{RT}\right)$, $\upsilon_c \propto exp\left(-\frac{E_c^*+PV_c^*}{RT}\right)$ are glide and climb velocity, respectively, that has different activation energy (E_g^*, E_c^*). Pressure dependence is given by the activation volume $V_{g,c}^* = \frac{E_{g,c}^*}{K}\left(2\gamma-\frac{2}{3}\right)$ where γ is Grüneisen parameter (we assume $\gamma = 1.5$), $K = K_o + P\left(\frac{dK}{dP}\right)$ ($K_o =$

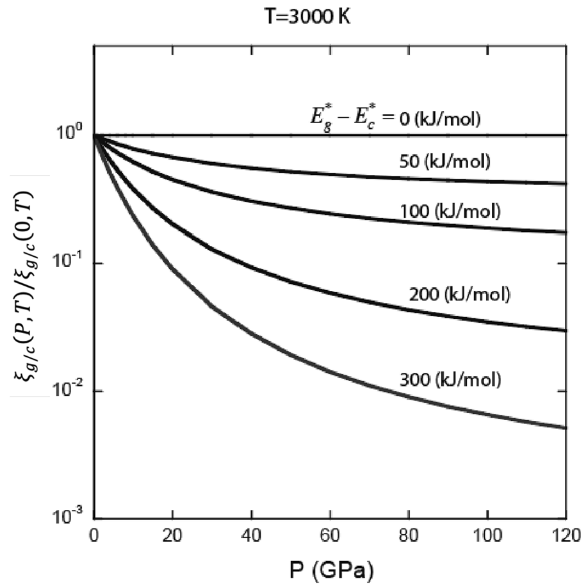

Fig. 5 Pressure dependence of dislocation velocity (mobility) ratio $\xi_{g/c}(P,T)/\xi_{g/c}(0,T)$

150 GPa, $\left(\frac{dK}{dP}\right) = 4$)). The ratio $\xi_{g/c}(P,T)\left[\equiv \frac{\upsilon_g(P.T)}{\upsilon_c(P.T)}\right]$ decreases with pressure but the degree to which this ratio decreases with pressure depends strongly on $\Delta E = E_g^* - E_c^*$. When a dislocation is saturated with jog, $\Delta E = E_g^* - E_c^*$ is large, whereas if a dislocation is not saturated with jog, $\Delta E = E_g^* - E_c^*$ is small.

The velocity ratio ($\xi_{g/c}(P,T)\left[\equiv \frac{\upsilon_g(P.T)}{\upsilon_c(P.T)}\right]$) depends on the difference in activation energy ($\Delta E = E_g^* - E_c^*$), but it also changes with pressure. Therefore, even if $\xi_{g/c}(P,T) > 1$ at low pressure, it could change to $\xi_{g/c}(P,T) < 1$ at high pressure. In (Boioli et al. 2017), they chose a parameter set corresponding to $\xi_{g/c}(P_o,T)\left[\equiv \frac{\upsilon_g(P_o.T)}{\upsilon_c(P_o.T)}\right] > 1$ at low P (P_o). They also assumed that dislocations are saturated with jogs, and hence $\Delta E = E_g^* - E_c^*$ is large (~200–300 kJ/mol). In such a case, this ratio can change to $\xi_{g/c}(P,T)\left[\equiv \frac{\upsilon_g(P,.T)}{\upsilon_c(P,T)}\right] < 1$ at high pressure as you can see from Fig. 5.

However, the validity of the assumption of jog saturation is questionable. To evaluate the validity of this assumption, let us recall some basics of dislocation climb. Climb motion of dislocation occurs by the migration of jogs through atomic diffusion. Consequently, climb velocity depends on jog density and diffusion coefficient as

$$\upsilon_c \propto D(P,T) \cdot C_j(P,T) \tag{5}$$

where D is diffusion coefficient ($D(P,T) = D_o \exp\left(-\frac{G_D^*}{RT}\right)$; G_D^* is the activation free energy of diffusion), C_j is jog density (e.g., (Hirth and Lothe 1982)). Generally jog density depends on temperature and pressure as $C_j \propto \exp\left(-\frac{G_j^*(P,T)}{RT}\right)$ (where $G_j^*(P,T)$ is the formation free energy of a jog (e.g., (Hirth and Lothe 1982)). Hence the activation free energy of dislocation climb is given by $G_c^* = G_D^* + G_j^*$. In a material with low dislocation energy (e.g., most metals), G_j^* is small and a dislocation line is saturated with jogs. In such a case, we can assume that C_j is constant ($C_j = \frac{1}{b}$ (b: the length of the Burgers vector)), and $G_c^* = G_D^*$. The validity of this assumption can be tested from the observed plastic anisotropy. In most minerals such as olivine, diffusion is nearly isotropic whereas plastic deformation by dislocation creep is highly anisotropic. This suggests that anisotropy comes from anisotropy of jog density, i.e., dislocations are not saturated with jogs (Karato 2010) ((Ammann et al. 2010) showed that diffusion is also nearly isotropic in bridgmanite. Given high plastic anisotropy of Br seen by the radial X-ray diffraction (Girard et al. 2016; Tsujino et al. 2022) and near isotropic diffusion, a dislocation in Br is also under-saturated with jogs similar to olivine).

Since jog density decreases with pressure, if a dislocation is under-saturated with jogs at low pressure (as shown for olivine), jog saturation at higher pressures (the lower mantle) is highly unlikely. If the assumption of jog saturation is removed, then E_c^* will be larger and $\Delta E = E_g^* - E_c^*$ is smaller. Consequently, pressure effects on $\xi_{g/c}(P,T)\left[\equiv \frac{\upsilon_g(P.T)}{\upsilon_c(P.T)}\right]$ is small and if glide velocity is higher than climb velocity at low pressure, it should also be the case at higher pressures (Fig. 5).

In Cordier et al. (2023), they argued that Br is weaker than Fp because pure climb occurs in Br. However, as we discussed above, we consider that the physical basis for the pure climb model is dubious. (Tsujino et al. 2016) showed that at P ~ 24–27 GPa, strong LPO is formed by the power-law dislocation creep implying that strain is by glide (not by climb) and hence pure climb creep is not operating in Br at least under the shallow lower mantle conditions.

In summary, we conclude that the validity of new models of plastic deformation in the lower mantle (athermal creep (Cordier et al. 2012)) and pure climb creep (Boioli et al. 2017)) is highly questionable. Consequently, when discussing the relative strength of Fp and Bp and the depth variation of viscosity (creep strength), we only consider power-law dislocation creep and diffusion creep regimes, and in case of dislocation creep.

6 Relative creep strength of ferropericlase (Fp) and bridgmanite (Br)

6.1 Dislocation creep regime

The relative creep strength of Fp and Br can be estimated by several ways. For the power-law dislocation creep regime, (Karato 1989b) reviewed then existing data on high-temperature deformation of oxides with various structures (including B1 (NaCl) structure and perovskite structure) to conclude that compared at the same $T/T_m(P)$ and $\sigma/\mu(P)$ and at a given strain-rate (he considered laboratory strain-rate, $\dot{\varepsilon} = 10^{-5}s^{-1}$), materials with the B1 structure (e.g., Fp) is substantially weaker than materials with perovskite structure (e.g., Br). Since the homologous temperature ($T/T_m(P)$) is similar for Fp and Br in the lower mantle (e.g., (Yao et al. 2021)), we expect that Fp is weaker than Br in the dislocation creep regime.

Experimental studies on plastic deformation of Fp and Br in the dislocation creep regime are summarized in Fig. 6 (we do not cite results from high-pressure, low-temperature experiments using a diamond anvil cell (DAC) because results from these experiments likely correspond to the Peierls regime (Fig. 2a) that are irrelevant to the lower mantle rheology). Among them, (Girard et al. 2016) provided the first experimental evidence

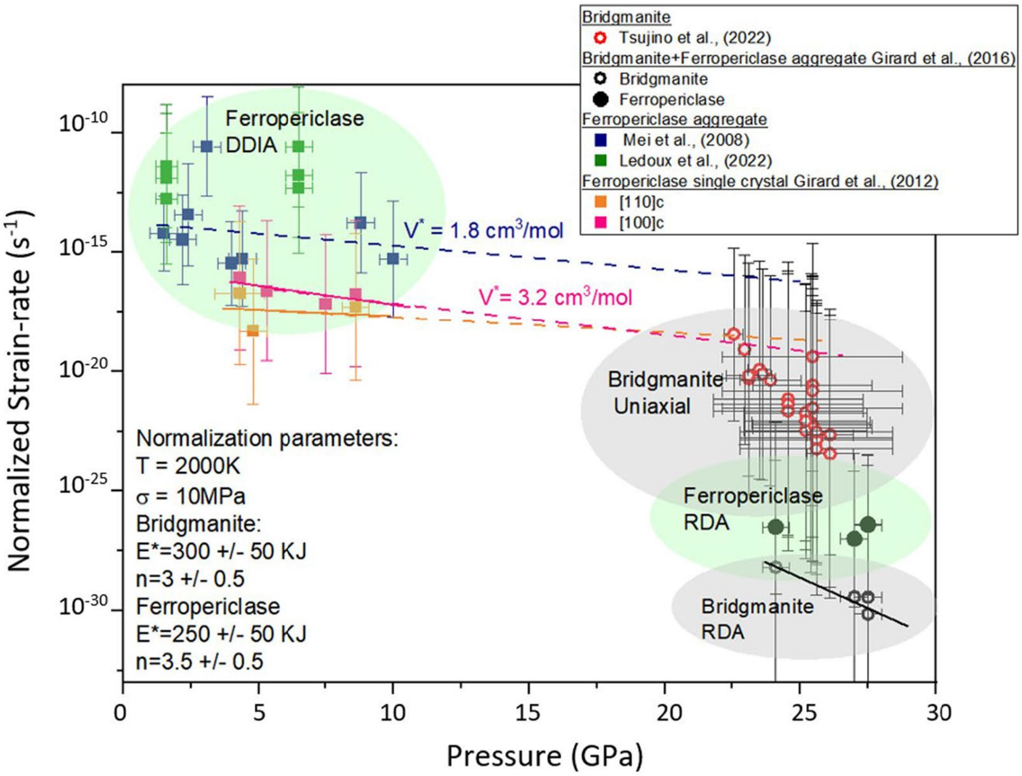

Fig. 6 Results of experimental studies on plastic deformation on Fp and Br

to support this notion, i.e., Fp is weaker than Br. They deformed a mixture of Fp and Br at the shallow lower mantle condition (P = 24–27 GPa, T ~ 2100 K) at the strain-rate of ~ 10^{-5} s^{-1} using the rotational Drickamer apparatus (RDA) and determined the creep strength by radial X-ray diffraction. The stress values estimated from different diffraction planes (hkl) show a wide variation indicating that deformation mechanism is dislocation creep (Karato 2009). Their results show that Br is substantially stronger than Fp under these conditions. This result is consistent with the results by (Karato 1989b) but the exact mechanism of deformation, i.e., either the power-law creep or the Peierls mechanism was unknown.

For Br (bridgmanite), later studies by Tsujino and his group ((Tsujino et al. 2016) and (Tsujino et al. 2022)) provided more details. First, (Tsujino et al. 2016) reported strong LPO indicating that deformation is dislocation creep where strain is (mostly) by dislocation glide (not by climb). Second, (Tsujino et al. 2022) determined the flow law of Br showing the power-law creep with $n = 3$–3.5, H* = 450–500 kJ/mol. These results show that deformation of bridgmanite under these conditions is by commonly observed power-law dislocation creep where strain is by glide but strain-rate is controlled by dislocation climb. (Tsujino et al. 2022) also compared their results with the results on MgO to show that under these conditions, Br is substantially stronger than Fp (the same conclusion as (Girard et al. 2016)).

For Fp (ferropericlase), (Girard et al. 2012) extended the study of plastic deformation of MgO to pressure from 4 to 9 GPa (T = 1273 to 1473 K) at the strain-rate of $(1\text{–}3)\times10^{-5}\ s^{-1}$. They reported that the pressure dependence of strength corresponding to $\frac{1}{2}\langle110\rangle\{110\}$ and $\frac{1}{2}\langle110\rangle\{100\}$ slip-systems is different and suggested that at ~ 23 GPa, the dominant (easy) slip system changes from $\frac{1}{2}\langle110\rangle\{110\}$to $\frac{1}{2}\langle110\rangle\{100\}$ confirming an earlier suggestion by (Karato 1998b) based on the change in the nature of chemical bonding and its influence on the choice of slip system with pressure. (Amodeo et al. 2012) also concluded this pressure-induced slip system change based on the computational approach.

All published results conducted at high temperature and pressures (T/Tm > 0.5) are compiled.

Extrapolation of these results to geological strain rate $(\dot{\varepsilon} \approx 10^{-15}s^{-1})$ is challenging, but the key is the stress exponent n. Figure 7 shows the effect of difference in stress exponent on the strength contrast at a geological strain-rate. Considering the error of Δn ~±0.5, we conclude that it is likely that the strength contrast observed

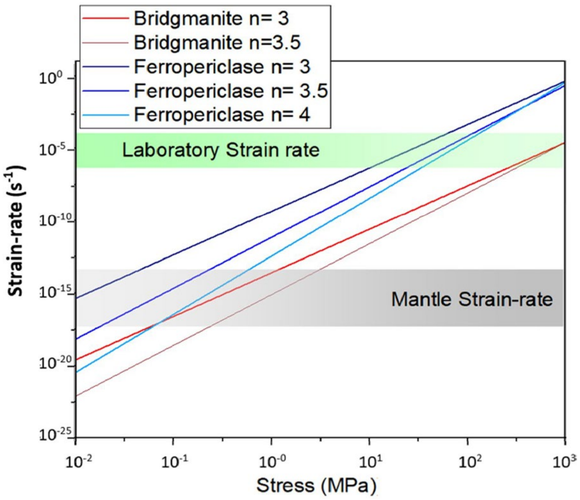

Fig. 7 Strength contrast between ferropericlase (Fp) and bridgmanite (Br) at various strain-rates in the power-law dislocation creep regime for various stress exponents (n) n: stress exponent. Fp is weaker than Br even at a geological strain-rate

at a laboratory strain-rate will persist at a geological strain-rate.

Experimental data for the laboratory strain-rate ($\dot{\varepsilon} = 10^{-5}s^{-1}$) are extrapolated to the geological strain-rate ($\dot{\varepsilon} = 10^{-15}s^{-1}$) assuming a range of different values of difference in stress exponent ($\Delta n = n_{\mathrm{Br}} - n_{\mathrm{Fp}}$). ($n_{\mathrm{Br}} =< spanclass =' convertEndash' > 3 - 3.5 < /span >$, $n_{\mathrm{Fp}} =< spanclass =' convertEndash' > 3 - 4 < /span >$)

6.2 Diffusion creep regime

How about the relative strength in the diffusion creep regime? Experimental studies on diffusion creep are available for MgO at room pressures, but no experimental results are available at high pressures. No experimental studies were made on diffusion creep in Br. However, there have been some reports on diffusion coefficients in Fp ((Mg,Fe)O) and Br ($MgSiO_3$) (e.g., Van Orman et al. 2003; Yamazaki and Irifune 2003; Yamazaki et al. 2000)). Given diffusion coefficients, the relative strength (viscosity) of Fp and Br can be estimated if grain-size is known. (Yamazaki and Karato 2001) presented the results of such a study based on the experimental results on self-diffusion coefficients. They concluded that (i) Fp is substantially weaker than Br and (ii) the depth variation in the strength is modest (1–2 orders of magnitude).

However, applications of those results are subject to a few uncertainties. First, experimental studies on diffusion under high pressures are challenging because at high pressures, diffusion coefficients tend to be under-estimated due to the stress development and/or the influence of water. In fact, in case of olivine where both diffusion (Béjina et al. 2003) and diffusion creep (Silber et al. 2024) are investigated to pressure range, activation volume for diffusion is substantially less than that for diffusion creep. Second, the relationship between diffusion coefficients and diffusion creep is not straightforward because of the complications caused by the role of grain-boundary diffusion (e.g., (Gordon 1973b)). Third, even if the results of diffusion coefficients are applicable to diffusion creep, the data so far obtained were to ~ 35 GPa and the extrapolation to the entire mantle (to ~ 135 GPa) is subject to a large uncertainty.

Therefore, in the following, we discuss the diffusion coefficients and their implications for the creep strength relevant to diffusion creep (or power-law dislocation creep) of Fp- and Br-based both experimental and theoretical studies. Let us first review some basics of diffusion coefficients in (ionic) crystal. Diffusion of atoms in a crystal occurs by random motion of atoms via point defects (e.g., (Karato 1981) and Chapter 8 of (Karato 2008)). Consequently, diffusion coefficient depends on the concentration and mobility of relevant point defect as $D \propto \Gamma_f.\Gamma_m = C_{\mathrm{d}} \cdot \Gamma_m$ where Γ_f is the probability that a neighboring cite is occupied by a defect ($\Gamma_f = C_{\mathrm{d}}$; C_{d} is the concentration of a point defect), and Γ_m is the probability that a defect jumps to the neighboring crystalline site. A few issues on diffusion coefficients need to be reminded. (1) Diffusion coefficients in a solid depend strongly on pressure (P) and temperature (T) because diffusion of atoms (or ions) in a solid occurs only via point defects (vacancies, interstitial atoms) and both their concentration and mobility depend strongly on T and P. (2) When C_{d} is controlled by the concentration of impurities, diffusion is called *extrinsic* diffusion. When C_{d} is controlled only by temperature and pressure but not by the impurities, it is called *intrinsic* diffusion and pressure effect is different between the two. (3) For a compound such as MgO, we need to consider diffusion coefficients of Mg and O and develop a model to understand how diffusion coefficients of various species interact in a given process.

Let us consider a case of MO ((Mg,Fe)O, Fp) (a similar consideration will apply to $(Mg,Fe)SiO_3$ (Br) with some modifications). The intrinsic process to form vacancies in MO ((Mg,Fe)O) is the formation of a Schottky pair, i.e.,

$$\mathrm{null} = V_M'' + V_O^{\cdot\cdot} \tag{6}$$

where V_M'' is a vacancy at M-site, $V_O^{\cdot\cdot}$ is a vacancy at O-site, (we use the Kröger-Vink notation). At equilibrium, we can get the concentrations of vacancies from the law of mass action, viz.,

$$\left[V_M''\right]\left[V_O^{\cdot\cdot}\right] = K_6(P,T) \propto \exp\left(-\frac{G_S^*}{RT}\right) \tag{7}$$

where $K_6(P,T)$ is the equilibrium constant for the reaction (7) and $G_S^*\left(=E_S^*+PV_S^*-TS_S^*\right)$ is the formation free energy of a Schottky pair, V_S^* is the volume of a Schottky pair (~ molar volume of MO). When there are no other charged defects, then the charge balance demands $\left[V_M''\right]_{\mathrm{int}}=\left[V_O^{\cdot\cdot}\right]_{\mathrm{int}}$ and we have,

$$\left[V_M''\right]_{\mathrm{int}}=\left[V_O^{\cdot\cdot}\right]_{\mathrm{int}}=K_6^{1/2}(P,T)\propto\exp\left(-\frac{G_S^*}{2RT}\right)\propto\exp\left(-\frac{PV_S^*}{2RT}\right) \tag{8}$$

These are the concentrations of vacancies in the intrinsic regime. Because the formation free energy of a Schottky pair is large for (Mg,Fe)O (~ 350–400 kJ/mol (Karato 1981)), the concentration of intrinsic vacancies is small and therefore even a small amount of impurities play a key role in diffusion to promote extrinsic diffusion.

An important cause of the extrinsic diffusion in (Mg,Fe)O is the presence of ferric Fe (F^{3+}). A majority of Fe in (Mg,Fe)O is ferrous Fe (F^{2+}). A small amount of ferric Fe (F^{3+}) is produced by oxidation, i.e.,

$$\frac{1}{2}O_2+2Fe_M^{\times}=2Fe_M^{\cdot}+O_O^{\times}+V_M'' \tag{9}$$

where $Fe_M^{\times}$ is F^{2+} at M-site (site that is occupied by Mg (or Fe)), $Fe_M^{\cdot}$ is F^{3+} at M-site, and $O_O^{\times}$ is O^{2-} at O-site. At high pressures where oxygen is a highly non-ideal gas, volumes of both sides of relation (9) are nearly the same, and the concentration of V_M'' formed by reaction (9) is (nearly) independent of pressure (and is proportional to $f_{O_2}^{1/6}$). Evidence for extrinsic diffusion of Mg-Fe in (Mg,Fe)O is presented by (Otsuka and Karato 2015; Otsuka et al. 2013).

An impurity such as F^{3+} also affects the vacancy concentration at O-site (and hence diffusion of oxygen). From the relation (8), oxygen vacancy concentration can be related to $\left[V_M''\right]_{\mathrm{ext}}$,

$$\left[V_O^{\cdot\cdot}\right]_{\mathrm{ext}}=\frac{K_6(P,T)}{\left[V_M''\right]_{\mathrm{ext}}} \tag{10}$$

Therefore,

$$\frac{\left[V_O^{\cdot\cdot}\right]_{\mathrm{ext}}}{\left[V_O^{\cdot\cdot}\right]_{\mathrm{int}}}=\frac{K_6^{1/2}(P,T)}{\left[V_M''\right]_{\mathrm{ext}}}(<<1). \tag{11}$$

We conclude that when F^{3+} affects the M-site vacancy concentration, it will also affect the O-site vacancy concentration in such a way that $\frac{\left[V_O^{\cdot\cdot}\right]_{\mathrm{ext}}}{\left[V_O^{\cdot\cdot}\right]_{\mathrm{int}}}<<1$ and its pressure dependence is strong $\left(\left[V_O^{\cdot\cdot}\right]_{\mathrm{ext}}=\frac{K_6(P,T)}{\left[V_M''\right]_{\mathrm{ext}}}\propto\exp\left(-\frac{PV_S^*}{RT}\right)\right)$.

Figure 8 is a schematic diagram where the role of extrinsic process to control the vacancy concentration is shown as a function of pressure. Important points are:

(i) Pressure dependence of vacancies formed by the intrinsic process (Schottky pair formation) is large.
(ii) Concentration of M-site vacancy formed by ferric Fe is (nearly) independent of pressure (but depends on oxygen fugacity). Therefore, extrinsic diffusion coefficient of Mg (Fe) depends only weakly on pressure (depth).

Ferric Fe also modifies the concentration of O-site vacancy. In contrast to M-site vacancy formed by ferric Fe $\left(\left[V_M''\right]_{\mathrm{ext}}\right)$, concentration of extrinsic oxygen vacancy depends strongly on pressure (decreases strongly with pressure). Therefore, extrinsic diffusion coefficient of oxygen is highly sensitive to pressure (depth) (decreases strongly with pressure).

From these, we conclude that if oxygen diffusion controls the rate of deformation, then viscosity increases with depth too much (in the intrinsic regime, viscosity would increase by ~ 10 orders of magnitude or more (Fig. 9 (Karato 1981); similar results are reported by (Cordier et al. 2023)). Figure 9 shows the diffusion coefficients in (Mg,Fe)O in the lower mantle calculated based on a model of point defects (Karato 1981). A few points should be noted:

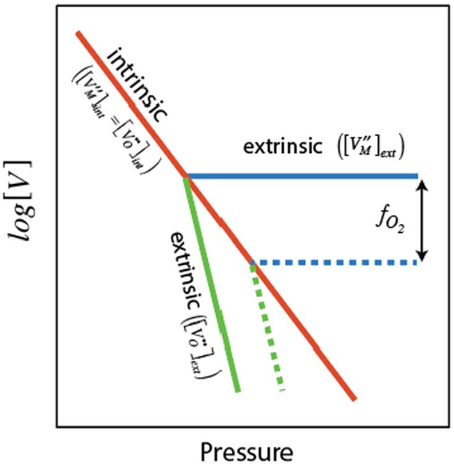

Fig. 8 A schematic diagram showing the pressure dependence of vacancy concentration [*V*] at M-site and O-site in (Mg,Fe)O (Fp) for both intrinsic and extrinsic regime (extrinsic regime is a regime where vacancy concentration at M-site is controlled by ferric Fe)

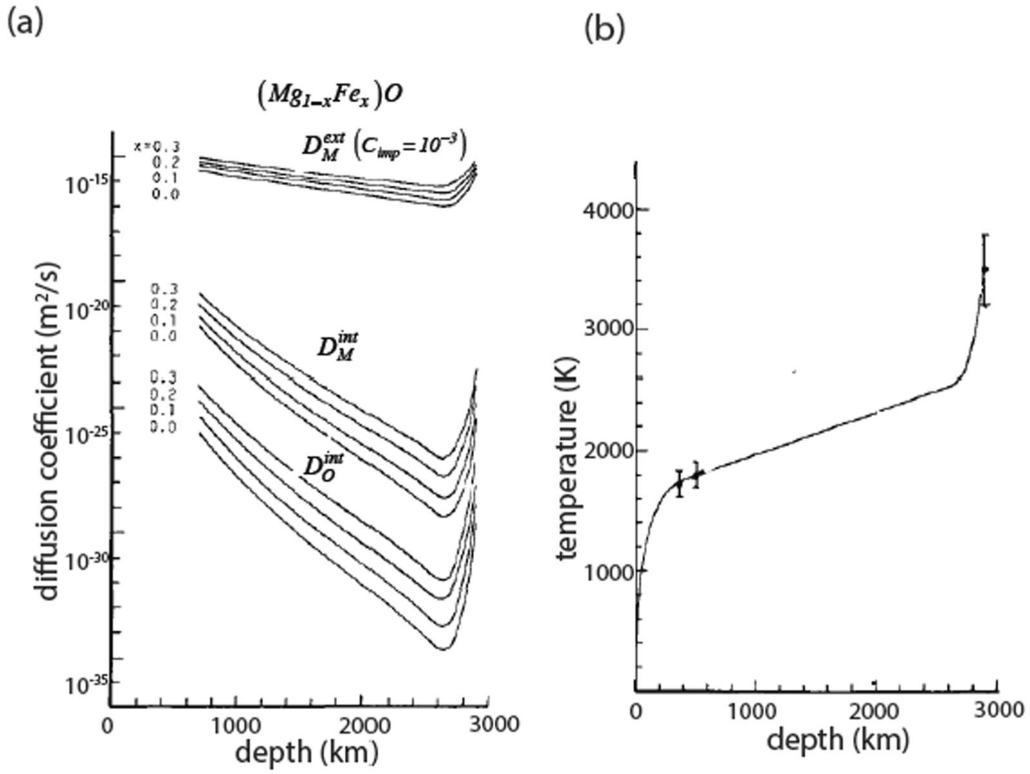

Fig. 9 Theoretically calculated diffusion coefficients in (Mg,Fe)O crystal in the lower mantle (after (Karato 1981)) **a** Diffusion coefficients of Mg (D_M) and O (D_O) (through the bulk crystal) D_M^{ext} : diffusion coefficient of M (Mg) in the extrinsic regime (C_{imp} is the impurity concentration (= vacancy concentration) and is assumed to be independent of pressure and temperature) D_M^{int}: diffusion coefficient of M in the intrinsic regime D_O^{int}: diffusion coefficient of O in the intrinsic regime x = Fe/(Fe + Mg) **b** A model of depth variation of temperature in the lower mantle

(1) Diffusion coefficients depend strongly on depth through their dependence on pressure (and temperature).
(2) Diffusion coefficients of Mg (Fe) (D_M) differ substantially from diffusion coefficients of O (D_O).
(3) For each species, the depth variation of diffusion coefficients differs substantially depending on either it is "intrinsic" (independent of impurities) or "extrinsic" (impurity-controlled).
(4) If vacancy mechanism operates, extrinsic diffusion is caused by $Fe^{2+} \leftrightarrow Fe^{3+}$ reaction and it enhances M (Mg,Fe) diffusion.

However, (Karato 1981) did not consider extrinsic diffusion of oxygen discussed above (Fig. 9). (Ammann et al. 2010) considered extrinsic diffusion of both Mg (Fe) and O where they assumed that the activation enthalpy for diffusion is activation enthalpy for defect migration (i.e., defect concentration is independent of pressure (and temperature)). The validity of this assumption for oxygen diffusion is questionable as discussed above.

To explain the modest depth dependence of viscosity in the lower mantle inferred from geodynamic studies, we need to assume that extrinsic diffusion of Mg (Fe) is a rate-controlling process of deformation in the lower mantle. If extrinsic diffusion of Mg (Fe) controls the diffusion creep rate, then, viscosity increases in the lower mantle will be ~ 1–2 orders of magnitude (Karato 1981; Yamazaki and Irifune 2003) that is consistent with geophysically inferred depth variation of viscosity in the lower mantle (Fig. 4, e.g., (Lau et al. 2016)). In contrast, if oxygen diffusion controls the rate of diffusion creep, viscosity will change with depth more than ~ 10 orders of magnitude or more (Figs. 8, 9).

But how does this model make sense? Diffusion creep in a compound involves diffusive mass transport of all species (Mg (Fe) and O for (Mg,Fe)O), and mass transport should maintain the composition of the compound. Consequently, it is the diffusion of the species that has smaller diffusion coefficient that controls the rate of diffusion creep (e.g., Chapter 8 of (Karato 2008) (see Fig. 8.5)), and oxygen diffusion is slower than Mg diffusion in a crystal of MgO (e.g., (Ando 1989)). So the above model (control of diffusion creep of (Mg,Fe)O by (volume) diffusion of Mg(Fe)) would seem odd. However, in ceramics (MgO, Al_2O_3), diffusion creep is in most cases controlled by the volume diffusion of Mg or Al and this is attributed to the fact that oxygen diffusion is enhanced much by grain-boundaries but Mg (or Al) diffusion is not (e.g., (Gordon 1973a)) (a similar discussion was presented for olivine by (Karato 1989a)). Therefore it is plausible deformation of (Mg,Fe)O in the lower mantle is controlled by extrinsic diffusion of Mg(Fe). It is important to confirm this by high-pressure experiments.

How about the power-law dislocation creep? Experimental studies summarized before show relatively small pressure dependence ($V^* = 2\text{–}3 \times 10^{-6}$ m^3/mol for MgO), see Fig. 6). Power-law creep is controlled by the recovery involving dislocation climb, and therefore pressure dependence comes mainly from the pressure dependence of relevant diffusion coefficient (and of jog density). For MgO, since dislocation energy is small (because of a short Burgers vector), a dislocation is likely saturated with jogs and jog density is independent of pressure. Therefore, the activation volume for creep is activation volume for diffusion. Observed small activation volume for power-law creep suggests that rate-controlling diffusion is diffusion of Mg (Fe). It is possible that oxygen diffusion is enhanced along dislocation lines and hence, like grain-boundaries in diffusion creep, mass transport among dislocations is controlled by Mg (Fe) diffusion.

7 Grain-size

Grain-size plays an important role in controlling the rheological properties when deformation occurs by diffusion creep. Since viscosity is sensitive to grain-size in diffusion creep, variation of grain-size will have an important effect on the variation in viscosity. Given a notion that diffusion creep plays an important role in the lower mantle (Karato et al. 1995), (Yamazaki and Karato 2001) estimated a grain-size of ~ 3 mm in the lower mantle. Also a few papers were published where they discussed the distribution of grain-size in the lower mantle due to grain-growth and grain-size reduction by phase transformations (e.g., Okamoto and Hiraga 2024; Solomatov et al. 2002; Solomatov and Reese 2008)). Both papers show a modest effect of grain-size on the radial and lateral variations in viscosity. However, there remains a large uncertainty on the kinetics of grain-growth in the lower mantle.

In contrast, there are other studies on grain-size distribution where grain-size is assumed to be controlled by dynamic recrystallization in the lower mantle in addition to the phase transformation across the 660-km discontinuity (e.g., Dannberg et al. 2017; Liu et al. 2024; Paul et al. 2024; Rozel 2012; Schierjott et al. 2020)). However, the validity of assuming dynamic recrystallization is questionable because of the absence of seismic anisotropy that implies the operation of diffusion creep (Karato et al. 1995). Dynamic recrystallization occurs by dislocation creep where strain is produced by dislocation glide. When deformation is by dislocation glide, lattice-preferred orientation (LPO) is formed causing seismic anisotropy. Therefore, the assumption of dynamic recrystallization is inconsistent with the absence of seismic anisotropy. Furthermore, when deformation is by dislocation creep, the influence of grain-size on the strength is small (~ 20–30% change in strength) (e.g., Chen and Argon 1979; Grossman and Ashby 1975) (note that this is opposite to what (Hirth and Kohlstedt 2003) proposed)).

8 Hydrogen effects

For olivine (and wadsleyite), the influence of hydrogen on plastic deformation is well documented (e.g., Karato et al. 1986; Mei and Kohlstedt 2000)). This is due to the fact that hydrogen solubility in these minerals is high compared to the point defect concentration in hydrogen-free minerals. Much less is known about the role of hydrogen in lower mantle minerals (Fp, Br), but in both Fp (Otsuka and Karato 2015; Otsuka et al. 2013) and Br (Fu et al. 2019), the solubility of hydrogen is substantially less than the concentrations of point defects formed through the reaction with oxygen (Fe^3+) and/or Al^{3+} (in case of Br). Consequently, it is unlikely that hydrogen has an important effect on plastic deformation in the lower mantle.

9 Deformation of a ferropericlase-bridgmanite mixture by diffusion creep: possibility of shear localization

Our analyses show that the lower mantle is made of a volumetrically minor and rheologically weaker Fp and a volumetrically major and rheologically stronger phase (Br). This contrasts with the upper mantle where a volumetrically major mineral (olivine) is weaker than volumetrically minor mineral such as orthopyroxene. In the latter case (the upper mantle), deformation is mostly by olivine and hence the strength is nearly independent of strain. In contrast, in the former case (the lower mantle), deformation of a weaker phase is more than deformation of a stronger phase, and when strain of a weaker phase is large, the weaker phase grains may be linked together to form an interconnected weak layer (IWL; (Handy 1994)). This would lead to shear localization (e.g., Chapter 16 of (Karato 2008)).

A few papers were published to address this issue. (Thielmann et al. 2020) reported a similar result on numerical modeling of a two-phase mixture, but they did not consider strain dependence of rheological contrast. (de Montserrat et al. 2021) conducted three-dimensional numerical modeling of a two-phase mixture and reported a smaller degree of strain weakening and interpreted this

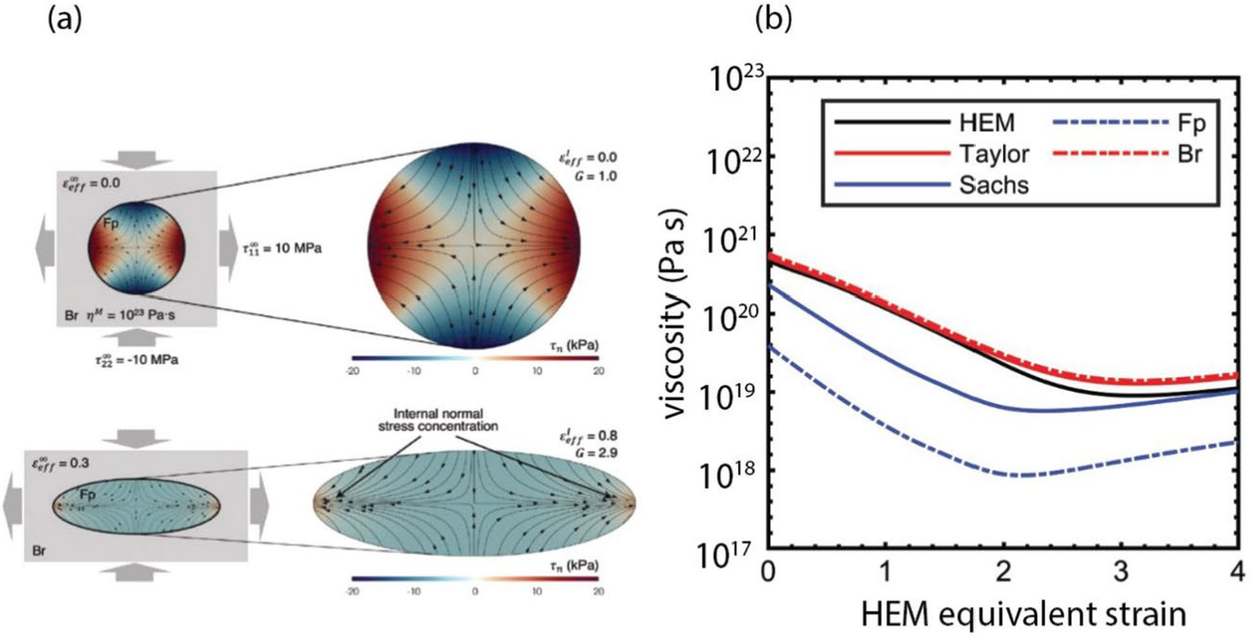

Fig. 10 Effect of strain on diffusional mass flux and resultant strain-dependent viscosity of a mixture of ferropericlase and bridgmanite. **a** Stress concentration and its effect on diffusion flux showing finite strain leads to higher stress concentration that promotes diffusional mass transport (from (Cho and Karato 2022)). **b** Viscosity-strain relationship of strain weakening of deformation of a Fp and Br mixture (Fp: 10%, Br: 90%) showing strain weakening (from (Cho and Karato 2024)) HEM: self-consistent model, Taylor: homogeneous strain model, Sachs: homogeneous stress model

result by a smaller degree of interconnection in three dimension than in two dimension.

(Cho and Karato 2022, 2024) suggested that when deformation is by diffusion creep, the rheological contrast will increase with strain and the tendency for shear localization will be greater (Fig. 10). A key point is that in diffusion creep regime, the strength contrast becomes larger with strain and not constant. This is because in diffusion creep, the diffusional mass flux is created by the gradient in the normal stress in a grain and the stress gradient will evolve with strain due to the evolution of stress concentration. Cho and Karato (2022, 2024) showed that the degree of strain weakening depends strongly on the geometry of deformation: strain weakening is more pronounced for simple shear than pure shear. Their results suggest that deformation in the lower mantle likely leads to shear localization, and consequently, mixing (or stirring) of materials with different compositions will be limited. This provides an explanation for the long-term survival of geochemical reservoirs in the lower mantle that is difficult to explain if strong shear localization does not occur (van Keken et al. 2002). Also, a high degree of localization will result in high stress level in the boundary layers (the D" layer) that would cause deformation by dislocation creep leading to seismic anisotropy.

One major limitation of a study by Cho and Karato (2024) is the use of a self-consistent scheme. This approach tends to under-estimate the degree of strain softening. An alternative approach such as Mori–Tanaka homogenization method developed by (Mori and Tanaka 1973), which is more suitable for mixtures with interconnected weak grains, would have to be explored to obtain more realistic strain softening. In addition, effects of slippery grain boundary and surface tension need to be incorporated to explore more realistic deformation of the lower mantle mixture. Grain boundary is weak and easy to slide, thereby stress state of the grain would be different from the case with a tied boundary that is assumed in Cho and Karato (2024). Surface tension would play a role in reducing grain elongation and/or in leading to breakage of the deforming grain. It is also important to apply such a flow law to conduct a large-scale simulation to determine the degree of stress distribution and compare these results with a deformation mechanism maps to interpret the distribution of seismic anisotropy.

However, in our model where we show larger strain weakening, a weak phase is not interconnected. The larger degree of weakening in our model is caused by the increase in viscosity contrast cause by finite strain. This latter factor is not included in these two papers.

And finally, with a large-scale convection modeling, we should explore the extent of mixing (or stirring) to obtain a hint to explain the long-term preservation of geochemical reservoirs.

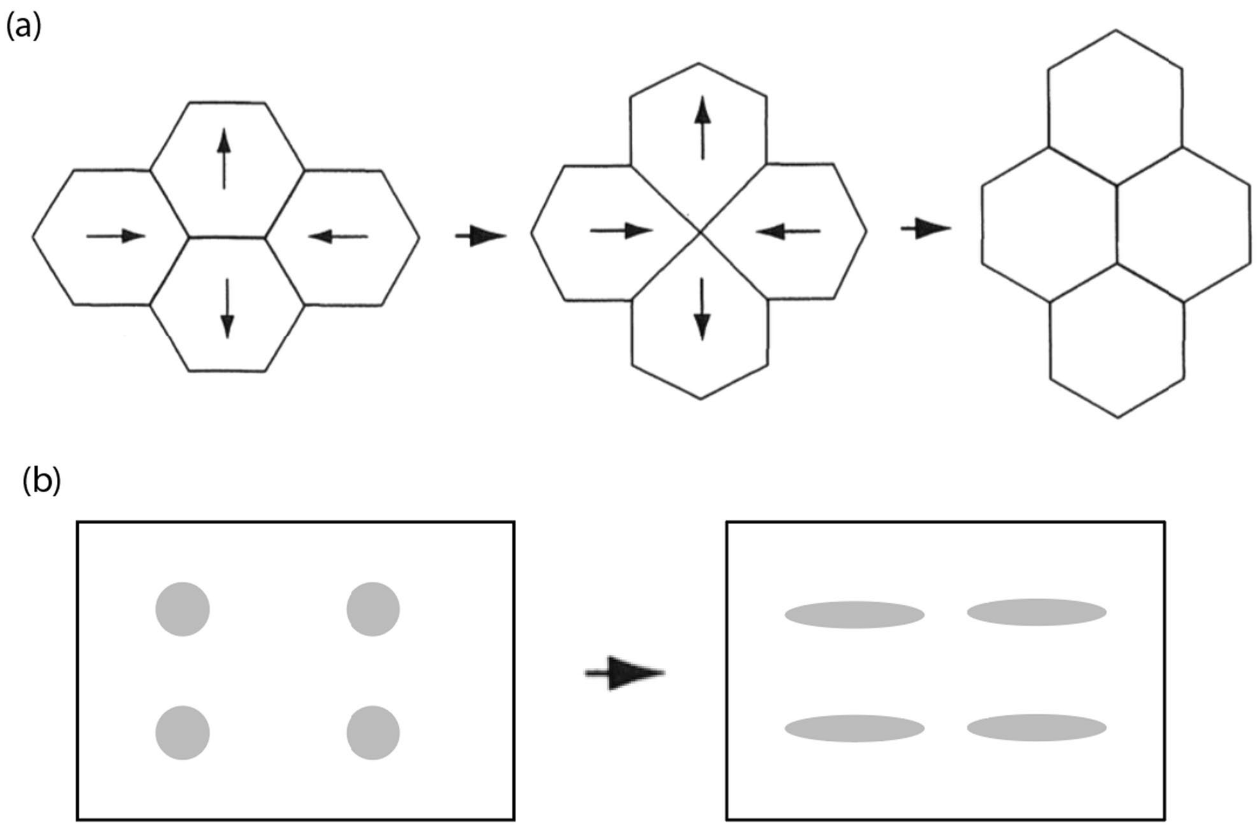

Fig. 11 Shape change during finite strain deformation **a** a case of a polycrystalline, single-phase material (after (Ashby and Verrall 1973)) Shape change is limited by grain switching, and grains remain nearly equant shape **b** a case of deformation of inclusions. A larger degree of shape change of inclusions will occur

It is often considered that grain elongation is limited by grain switching. In case of deformation of single-phase polycrystals, after ~ 55% strain, grain switching events will occur and grain shape is nearly equant even at a large strain (Ashby and Verrall 1973) (Fig. 11a). However, in case of the lower mantle, the weak phase (Fp) is present as a minor isolated phase. Therefore, grain-grain switching events will not occur and therefore high degree of elongation is possible (Fig. 11b). In other words, strain weakening in the lower mantle assembly occurs because the weaker phase is a minor phase.

10 Summary and concluding remarks

We reviewed models of deformation mechanisms in the lower mantle with special attention to recently proposed athermal creep and pure climb creep models. These models predict that a volumetrically major phase, i.e., bridgmanite (Br), is weaker than a volumetrically minor phase, opposite to some of previous models or experimental observations (e.g., Girard et al. 2016; Yamazaki and Karato 2001)). These new models would predict that deformation in the lower mantle is not as much localized as an alternative model where Fp is weaker than Br. These models also predict that the entire lower mantle should be seismically isotropic. Based on a comparison to experimental results and theoretical model of plastic deformation as well as to the known distribution of seismic anisotropy, we conclude that these models are inconsistent with well-established experimental observations. Therefore, we believe that the validity of these models is highly questionable.

By reviewing existing experimental results on deformation of Fp and Br in the power-law dislocation creep regime with a scaling analysis for the extrapolation to the geological strain-rate, we conclude that Fp is weaker than Br in the lower mantle in power-law dislocation creep regime. For diffusion creep regime, we review theoretical and experimental results on diffusion coefficients and conclude that Fp is weaker than Br because of the enhancement of Mg(Fe) diffusion by ferric Fe.

A new model is proposed on the role of ferric Fe on oxygen diffusion. The model predicts that oxygen diffusion is reduced by the presence of ferric Fe, and the pressure dependence of oxygen diffusion is large (diffusion coefficient of oxygen will increase more than ~ 10 orders of magnitude in the lower mantle). We suggest that to explain geophysically inferred only modest increase in viscosity in the lower mantle (by 1–2 orders of magnitude), an important role of Mg(Fe) diffusion in the extrinsic regime, not diffusion of oxygen, needs to be invoked as a rate-controlling process of deformation. Experimental observations to support this model are reviewed and the future direction of experimental (and theoretical) studies is suggested.

Supplementary Information

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Supplementary material 1

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Author contributions

SK has conceived the major ideas on this paper and wrote most of it. JG prepared Figs. 6 and 7 and discussions on them. HEC prepared Fig. 10 and discussions on it.

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Data availability

All the data related to this paper can be obtained by contacting the authors.

Declarations

Competing interests

The authors declare no competing interest.

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