Flow and deformation in Earth’s deepest mantle from geodynamic modeling and implications for seismic anisotropy

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Abstract

The dynamics of Earth’s D” layer at the base of the mantle plays an essential role in Earth’s thermal and chemical evolution. Mantle convection in D” is thought to result in seismic anisotropy; therefore, observations of anisotropy may be used to infer lowermost mantle flow. However, the connections between mantle flow and seismic anisotropy in D” remain ambiguous. Here we calculate the present-day mantle flow field in D” using 3D global geodynamic models. We then compute strain, a measure of deformation, outside the two large-low velocity provinces (LLVPs) and compare the distribution of strain with previous observations of anisotropy. We find that, on a global scale, D” material is advected towards the LLVPs. Strain is highest at the core-mantle boundary (CMB) and decreases with height above the CMB. Material outside the LLVPs mostly undergoes lateral stretching, with the stretching direction often, but not always, aligning with mantle flow direction. Strain generally increases towards the LLVPs and reaches a maximum at their edges, although models that consider recrystallization suggest that anisotropy may actually be weaker near LLVP edges. The depth-averaged strain in D” is >1.5 in almost all regions, consistent with widespread observations of seismic anisotropy. The mantle flow field and strain in D” outside of LLVPs are not very sensitive to LLVP density but are strongly controlled by local density and viscosity variations outside the LLVPs. Flow directions inferred from anisotropy observations often (but not always) align with predictions from geodynamic modeling calculations.
(a) Lateral strain = 3.0 cm/yr

(b) Frequency (%)

Angle (degrees)

Strain

- CMB
- 45 km
- 90 km
- 135 km
- 180 km
- 225 km

Strain

1 2 3 4 5 6 7 8

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(a) Strain

(b) Stretch angle (degree)

(c) Strain

(d) Strain
(a) case 1

(b) case 2

(c) case 3

(d) case 4

(e) case 5

(f) case 6

Radial velocity (cm/yr)

min 0.0 max

min 3 cm/yr min/max = -0.5/0.5

min 3 cm/yr min/max = 0.5/0.5

min 3 cm/yr min/max = -0.5/0.5

min 3 cm/yr min/max = -0.5/0.5

min 3 cm/yr min/max = -0.5/0.5

Radial velocity (cm/yr)

min 0.0 max
(a) case 1
(b) case 2
(c) case 3
(d) case 4
(e) case 5
(f) case 6

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Strain

(a)
(b)
(c)
(d)
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Table 1.docx available at https://authorea.com/users/554336/articles/722908-flow-and-deformation-in-earth-s-deepest-mantle-from-geodynamic-modeling-and-implications-for-seismic-anisotropy
Flow and deformation in Earth’s deepest mantle from geodynamic modeling and implications for seismic anisotropy

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Key points:

- Geodynamic models show that D” materials are strongly deformed, consistent with observations of seismic anisotropy.
- Strain in D” generally increases with depth and increases towards the large low velocity provinces.
- Flow directions inferred from anisotropy observations often align with that from geodynamic calculations.
Abstract

The dynamics of Earth’s D” layer at the base of the mantle plays an essential role in Earth’s thermal and chemical evolution. Mantle convection in D” is thought to result in seismic anisotropy; therefore, observations of anisotropy may be used to infer lowermost mantle flow. However, the connections between mantle flow and seismic anisotropy in D” remain ambiguous. Here we calculate the present-day mantle flow field in D” using 3D global geodynamic models. We then compute strain, a measure of deformation, outside the two large-low velocity provinces (LLVPs) and compare the distribution of strain with previous observations of anisotropy. We find that, on a global scale, D” material is advected towards the LLVPs. Strain is highest at the core-mantle boundary (CMB) and decreases with height above the CMB. Material outside the LLVPs mostly undergoes lateral stretching, with the stretching direction often, but not always, aligning with mantle flow direction. Strain generally increases towards the LLVPs and reaches a maximum at their edges, although models that consider recrystallization suggest that anisotropy may actually be weaker near LLVP edges. The depth-averaged strain in D” is >1.5 in almost all regions, consistent with widespread observations of seismic anisotropy. The mantle flow field and strain in D” outside of LLVPs are not very sensitive to LLVP density but are strongly controlled by local density and viscosity variations outside the LLVPs. Flow directions inferred from anisotropy observations often (but not always) align with predictions from geodynamic modeling calculations.

 Plain Language Summary

The Earth’s deep mantle deforms and moves at a geological timescale. This movement is called mantle convection, which controls plate tectonics. Of particular importance is the flow in the lowermost few hundred kilometers of the mantle, which is called the D” layer. The dynamics of D” plays an essential role in Earth’s thermal and chemical evolution. Direct observation of D” flow is not possible, but D” flow causes deformation of minerals that can align in preferential directions, leading to variable seismic velocities along different directions. This feature is called seismic anisotropy. In this study, we use numerical simulations to investigate D” flow and its connection to rock deformation and seismic anisotropy. We find that D” materials are strongly deformed, consistent with observations of seismic anisotropy in this layer. The strength of rock deformation in D” generally increases with depth and increases towards regions beneath the Central Pacific and Africa, where two continental-sized seismic anomalies exist. It is encouraging to find that flow directions inferred from anisotropy observations often align with our numerical simulations. This study thus improves our understanding on the dynamics of the D” layer.

1. Introduction

1. Introduction

A grand challenge in solid Earth science is to understand the Earth’s mantle flow field, which controls deep mantle structures, the generation and mechanism of plate tectonics, and the Earth’s long-term thermal and chemical evolution. The lowermost mantle flow is of particular interest because it regulates the heat flux at the core-mantle boundary (CMB) [e.g., Nakagawa and Tackley, 2008; Li and McNamara, 2018; Li et al., 2018] which is critical for generating the magnetic field [e.g., Larson and Olson, 1991; Zhang and Zhong, 2011; Olson et al., 2014] and is essential for Earth’s thermal evolution [e.g., Christensen, 1985; Korenaga, 2008]. It dictates the formation of mantle plumes [Steinberger and O’Connell, 1998; Li and Zhong, 2017; Li and Zhong, 2019; Li, 2023a] that cause surface volcanism [Morgan, 1971]. Furthermore, the lowermost mantle greatly influences the morphology and internal structure of seismic anomalies such as the large
low velocity provinces (LLVPs) and ultra-low velocity zones (ULVZs) [e.g., McNamara et al., 2010; Li et al., 2017; Pachhai et al., 2021; Yuan and Li, 2022b]. It causes topography on the CMB [e.g., Yoshida, 2008; Lassak et al., 2010; Deschamps et al., 2018] and also affects the process of core-mantle reaction [e.g., Manga and Jeanloz, 1996; Kanda and Stevenson, 2006; Ko et al., 2022]. Moreover, it controls the advection, distribution, mixing, and accumulation of compositional reservoirs in the Earth's deep interior [e.g., McNamara and Zhong, 2005; Zhang et al., 2010; Tackley, 2011; Li, 2021; Li and McNamara, 2022; Hansen et al., 2023; Li, 2023b] and thus plays an essential role in Earth’s chemical evolution.

The present-day mantle flow field has been widely studied using geodynamic simulations. By solving the conservation equations of mass and momentum, the instantaneous mantle flow field at the present day can be computed, based on a specified mantle density and viscosity structure. This predicted flow field has been used in many applications, such as estimating the driving forces of plates [e.g., Lithgow-Bertelloni and Richards, 1995; Conrad and Lithgow-Bertelloni, 2004], calculating the dynamic topography at the surface and the CMB [e.g., Flament et al., 2013; Yang and Gurnis, 2016; Deschamps et al., 2018], and modeling the geoid [Hager and Richards, 1989; Zhong and Davies, 1999; Liu and Zhong, 2016]. However, less attention has been paid to the characteristic of the lowermost mantle flow field itself; images of lowermost mantle flow fields are only presented in a small number of publications [e.g., Steinberger and Holme, 2008; Yoshida, 2008; Walker et al., 2011].

The large-scale structure of Earth’s lowermost mantle is dominated by two large low velocity provinces (LLVPs) [e.g., Li and Romanowicz, 1996; Grand, 2002; Ritsema et al., 2004; Garnero et al., 2016]. Surrounding the LLVPs are regions with generally higher-than-average seismic velocities, which are often interpreted as former, relatively cold, subducted slabs [e.g., Lithgow-Bertelloni and Richards, 1998]. It has been suggested that flow in the lowermost mantle, on a global scale, moves away from subduction zones towards the LLVPs [Steinberger and Holme, 2008; Yoshida, 2008; Walker et al., 2011]. However, understanding lowermost mantle flow at a scale smaller than a few hundred kilometers is more challenging, due to the influence of regional density and viscosity variations that are not yet well understood [e.g., Li, 2023a].

When deformation is accommodated in the dislocation creep regime, mantle flow causes the alignment of individual crystals in an aggregate, leading to a variation of seismic velocity with propagation and polarization direction. Therefore, observations of seismic anisotropy at the base of the mantle can offer crucial insights into the deep mantle flow field [e.g., Romanowicz and Wenk, 2017]. Global seismic tomography has shown that the bulk of the lower mantle is (almost) isotropic, while strong seismic anisotropy can be found in D” [Auer et al., 2014; French and Romanowicz, 2014; Moulik and Ekstrom, 2014; Chang et al., 2015; Romanowicz and Wenk, 2017; Nowacki and Cottaar, 2021]. These global results are supported by regional investigations, which identify seismic anisotropy in many regions of D” [e.g., Garnero and Lay, 1997; Long, 2009; Nowacki et al., 2010; 2011; Asplet et al., 2020; Wolf and Long, 2022; Asplet et al., 2023; Wolf et al., 2023b]. In some regions, strong seismic anisotropy is observed outside and near the edges of the LLVPs, with weaker or absent anisotropy within the LLVPs [To et al., 2005; Cottaar and Romanowicz, 2013]. Observations of anisotropy in the D” layer are often interpreted as being caused by mantle flow and deformation. However, inferring mantle flow direction from measurements of seismic anisotropy remains challenging, because the connections among flow, deformation, and anisotropy are not well understood.
This study utilizes global geodynamic models to calculate the present-day mantle flow field and explore how variations in model parameters impact the prediction of flow velocities in the lowermost mantle. We introduce passive tracers into the model domain to track deformation history and compare the distribution of strain with seismic observations of anisotropy in D” outside the LLVPs. Through these numerical experiments, we aim to gain a better understanding of the mantle flow field in D” and its connection to observations of seismic anisotropy.

2. Methods

2.1. Setup of convection models

We solve the following non-dimensional conservation equations of mass and momentum and calculate the instantaneous mantle flow under the Boussinesq approximation:

\[ \nabla \cdot \mathbf{u} = 0, \]  
\[ -\nabla P + \nabla \cdot (\eta \dot{\varepsilon}) + [\text{Ra}(T - B_c C - B_{ppv} \Gamma)] \hat{r} = 0, \]  

where \( \mathbf{u}, P, \eta, \dot{\varepsilon}, \text{Ra}, \) and \( T \) are, respectively, the velocity, dynamic pressure, viscosity, strain rate, Rayleigh number, and temperature. \( B_c \) and \( C \) are respectively the buoyancy number and the fraction of an intrinsically dense compositional component. \( B_{ppv} \) is the buoyancy number that represents the density increase due to the Bridgmanite (Bdg) to post-Perovskite (pPv) phase transition and the \( \Gamma \) is the phase function. \( \hat{r} \) is a unit vector in the radial direction. Equations (1-2) are solved using the CitcomS code \cite{Zhong2008}. The model domain ranges from the CMB to the top surface and is divided into 12 equal-volume caps with each cap containing \( 64 \times 64 \times 64 \) elements. Both the top surface and the CMB are free-slip boundaries.

The Rayleigh numbers \( \text{Ra} \) is defined as:

\[ \text{Ra} = \frac{\rho g \alpha \Delta T R^3}{\kappa \eta_0}, \]  

where \( \rho \) is density, \( g \) is gravitational acceleration, \( \alpha \) is thermal expansivity, \( \Delta T \) is the reference temperature (e.g., the non-adiabatic temperature increase from surface to CMB), \( R \) is Earth’s radius, \( \kappa \) is thermal diffusivity, and \( \eta_0 \) is reference viscosity.

The \( B_c \) and \( B_{ppv} \) are defined as:

\[ B_c = \frac{\Delta \rho_c}{\rho \alpha \Delta T}, \]  
\[ B_{ppv} = \frac{\Delta \rho_{ppv}}{\rho \alpha \Delta T}, \]  

where \( \Delta \rho_c \) the intrinsic density anomaly with respect to the background mantle and \( \Delta \rho_{ppv} \) is the density increase due to the pPv phase transition.

The phase function for the pPv phase transition is defined as:

\[ \Gamma(\pi) = 0.5 + 0.5 \tanh(\frac{\pi}{\delta}), \]  

where \( \delta \) is the width of the phase transition, and the excess pressure \( \pi \) is defined as:

\[ \pi = d - d_{ppv} - \gamma(T - T_{ppv}), \]
where \( d \) is the depth, \( d_{ppv} \) and \( T_{ppv} \) are the reference depth and temperature for the pPv phase transition, and \( \gamma \) is the Clapeyron slope. The phase function \( \Gamma \) varies from 0.0 for purely Bdg phase to 1.0 for purely pPv phase.

The viscosity depends on temperature, depth, and the presence of pPv phase, expressed as:

\[
\eta = \eta_r \exp \left[ E (0.5 - T) + \Gamma \ln(\eta_{ppv}) \right],
\]

where \( E \) is the activation coefficient for the temperature-dependent viscosity, \( \eta_{ppv} \) is the viscosity change due to the pPv phase transition, and \( \eta_r \) is a prefactor that controls the depth-dependence of viscosity. The \( E, \eta_{ppv} \) and \( \eta_r \) are free parameters tested in this study.

We assume a linear relationship between the density anomaly caused by thermal expansion \((\Delta \rho_T)\) and the anomaly of seismic shear-wave velocity \((\delta V_s)\), given by:

\[
\frac{\Delta \rho_T}{\rho} = C_{v_s-\rho} \frac{\delta V_s}{V_s},
\]

where \( C_{v_s-\rho} \) is the conversion factor. The change of temperature is related to thermal density anomaly via thermal expansivity by:

\[
\delta T = - \frac{\Delta \rho_T}{\rho \alpha}.
\]

The non-dimensional form of the temperature anomaly is given by:

\[
\delta T' = \frac{\delta T}{\Delta T},
\]

Finally, the non-dimensional temperature field in our models is calculated by:

\[
T = T_{ave} + \delta T',\]

where the \( T_{ave} \) is the 1D laterally averaged profile of the non-dimensional temperature taken from present-day temperature field of a previous 3D global geodynamic model in [Li and Zhong, 2019] and is given in Supporting Information Figure S1a. In many cases, the LLVPs are treated as compositionally distinct materials. For this condition, the buoyancy field also includes the effect of intrinsic density anomalies in LLVPs in most models. Table 1 summarizes the physical parameters used in this study, which remain constant throughout the analysis.

Table 1. Physical parameters used in this study

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Reference value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Earth’s radius ( R )</td>
<td>6371 km</td>
</tr>
<tr>
<td>Mantle thickness</td>
<td>2890 km</td>
</tr>
<tr>
<td>Reference density ( \rho )</td>
<td>3300 kg/m³</td>
</tr>
<tr>
<td>Gravitational acceleration ( g )</td>
<td>9.8 m/s²</td>
</tr>
<tr>
<td>Thermal expansivity ( \alpha )</td>
<td>( 1.0 \times 10^{-5} ) K⁻¹</td>
</tr>
<tr>
<td>Thermal diffusivity ( \kappa )</td>
<td>( 1.0 \times 10^{-6} ) m²/s</td>
</tr>
<tr>
<td>Temperature change across the mantle ( \Delta T )</td>
<td>2500 K</td>
</tr>
<tr>
<td>Reference viscosity ( \eta_0 )</td>
<td>( 1.04 \times 10^{21} ) Pa s</td>
</tr>
</tbody>
</table>
2.2. Strain calculation

Mantle flow causes deformation and, under certain conditions, orientation of minerals which can develop seismic anisotropy. Specifically, lattice preferred orientation (LPO) develops when deformation is accommodated via dislocation creep, which is generally favored by lower temperatures, higher stresses, and larger grain sizes [e.g., Frost and Ashby, 1982]. In this work, we assume that the base of the mantle is deforming in the dislocation creep regime [e.g., Karato, 1998]. This assumption is consistent with previous modeling studies [McNamara et al., 2003] as well as the observation that seismic anisotropy in the deep mantle is constrained to the lowermost few hundred kilometers of the mantle [e.g., Auer et al., 2014; French and Romanowicz, 2014]. The nature of seismic anisotropy caused by LPO is controlled by the geometry and amount of deformation. Here, the amount of deformation is measured by strain, which is defined as the ratio between the final length and the original length.

Passive tracers are introduced into the model domain and are advected by the instantaneous mantle flow field. The tracers track the strain along their paths. We use the method in [McNamara et al., 2003] to calculate strain on tracers. More detailed descriptions of this method are also given in [Mckenzie, 1979] and [Spencer, 1980]. Strain is calculated as the positive square root of the maximum eigenvalue of the left Cauchy-Green deformation tensor \( G \) which is defined as:

\[
G = F \cdot F^T, \quad (13)
\]

where \( F \) is the deformation tensor and \( F^T \) is its transpose. The \( F \) is initially a unit tensor. It changes along the trajectories of tracers and is determined by integrating the velocity gradient tensor \( L \) by:

\[
\frac{\partial F_{ij}}{\partial t} = L_{ik} F_{kj}, \quad (14)
\]

The trajectory of tracers is a function of time and is controlled by the initial location of tracers. In this study, we assume that tracers start to accumulate strain at the time when they first sink across the top of the D’’ layer, which we defined as 300 km above the CMB. Therefore, we initially introduce tracers at the top of the D’’ layer. We aim to obtain such an initial distribution of tracers that it allows the tracers to be advected into and sample all regions in D’’. To achieve this goal, we take the following steps. First, we evenly distribute tracers at their final positions in the D’’ layer. Then, these tracers are advected backward-in-time until they reach their initial locations at the top of D’’ (again, 300 km above the CMB). We denote the time that a tracer takes to travel from its final to initial location as \( t^* \). After we obtain the initial location of all tracers, they are advected forward-in-time with a time of \( t^* \) to their final positions in D’’. Note that each tracer takes different path and thus has a different value of \( t^* \). Strain on tracers is calculated during the forward-in-time process starting from the initial locations of tracers, and all tracers begin with unity strain.

3. Results

3.1. Flow and deformation for the reference model

We first present the results for our reference model, Case 1. In this case, the density and temperature structures are converted from tomography model S40RTS [Ritsema et al., 2011] using a conversion factor \( C_{\nu_s-\rho} = 0.4 \) between thermal density anomaly and S-wave velocity anomaly (Eq. 9). Similar to previous studies [e.g., Hager and Richards, 1989; Liu and Zhong, 2016], the density anomaly in the topmost 300 km of the mantle is removed because its relationship with \( V_s \) anomaly
may not be well captured by a linear scaling law, due to the presence of strong composition variations. The LLVPs have been suggested to be made of compositionally distinct material that is intrinsically denser than the surrounding mantle [e.g., McNamara and Zhong, 2005; Mulyukova et al., 2015; Jones et al., 2020; Li and McNamara, 2022; Yuan and Li, 2022a; Yuan et al., 2023]. To account for this, we assign a buoyancy number of $B_c=0.3$ in regions where the $V_s$ anomaly is $<-0.5\%$ in the lowermost 600 km of the mantle. The Rayleigh number is $Ra=2 \times 10^8$. The viscosity of case 1 follows Eq. (7). The activation energy is $E=9.21$. The depth-dependent viscosity prefactor $\eta_r$ is 1.0 in the upper mantle and 70.0 in the lower mantle. Parameters related to the pPv phase transition are $B_{pPv}=0.4$, $\gamma=0.1456$, $d_{pPv}=0.42386$, $T_{pPv}=0.5$, and $\eta_{pPv}=0.01$. The laterally averaged viscosity profile is shown in Supporting Information Figure S1b.

Because temperature scales linearly with $V_s$, the LLVP regions with lower $V_s$ are generally hotter than surroundings (Figure 1a). However, most regions within the LLVPs remain negatively buoyant (Figure 1b). This is because the LLVPs in this model are assumed to be made of materials that are intrinsically denser than the background mantle. The temperature field controls the Bdg-pPv phase transition. At 2,800 km depth, regions outside the LLVPs are dominated by pPv phase, whereas no pPv phase exists within the hot LLVPs (Figure 1c). Note that the fraction of pPv phase, as represented by the phase function $\Gamma$, varies with depth. For example, compared to at 2,800 km depth, the fraction of pPv phase is smaller at 45 km above the CMB and at depths near the top of D” (Supporting Information Figure S2).
Figure 1. The distribution of (a) non-dimensional temperature anomaly after the horizontal average is removed, (b) residual buoyancy, and (c) the fraction of pPv phase as represented by the phase function $\Gamma$ defined in Eq. (6), at 2,800 km depth for case 1. The cyan contours correspond to a $V_s$ anomaly at -0.5% in the S40RTS tomography model [Ritsema et al., 2011], indicating the LLVP regions.

Flow in D" generally converge towards the LLVPs (Figure 2a). Downwelling flows with negative radial velocity mostly occur far from LLVPs (e.g., beneath the circum-Pacific and Indian Ocean), whereas upwelling flows mainly occur within and just outboard of the LLVPs (Figure 2b). As expected, horizontal flow dominates near the bottom boundary of the mantle outside LLVPs (Figure 2a). The radial velocity is about 10 times lower than the lateral velocity at 2,800 km depth outside the LLVPs (Figure 2b).
We calculate the 2nd invariant of the strain rate tensors. High strain rate (Figure 2c) mostly occurs in regions with high lateral flow velocity (Figure 2a, 2c). Regions with the largest strain rate include outside the southern boundary of the African LLVP, the area surrounding the Perm anomaly (northern Siberia), just west of Central and South America, and outboard the southern and northwest boundaries of the Pacific LLVP (Figure 2c).

**Figure 2.** The distribution of (a) lateral flow velocity, (b) radial flow velocity, and (c) 2nd invariant of the strain rate at 2,800 km depth for case 1. The cyan contours $V_s$ anomaly at -0.5% in the S40RTS tomography model, indicating the regions of LLVPs.

We first examine strain at 90 km above the CMB. We introduce 3,072 evenly distributed tracers at this depth, and then advect these tracers backward-in-time until they reach the top of D". We find that most tracers when they arrive at the top of D" are in regions with downwelling flows (Figure 2c) around the Pacific Ocean and the ancient Tethys Ocean (green colors in Figure 3a).

Then, for tracers at the top of D", we set their strain to be unity, advect them forward-in-time, and track their strains along their paths until they reach their final locations at 90 km above the CMB. Note that we define the time it takes for a tracer to be advected from the top of D" to their final
locations, e.g., 90 km above the CMB, as $t^*$. As shown in Figure 3b, the $t^*$ is shortest in regions with negative radial flow velocities (e.g., downwelling regions in Figure 2b), whereas it is longer in regions with positive radial flows (e.g., near LLVP edges) where the tracers are typically advected laterally and take a long path to reach their final locations. The $t^*$ for most tracers outside the LLVPs is <100 Myr (Figure 3b). Tracers with $t^*>100$ Myr mainly occur around the edges of the LLVPs (Figure 3b). The strain of tracers, after beginning with unity at downwelling locations, generally (but not always) increases along the tracer trajectories (Figure 3c).

Figure 3. (a) The trajectories of 3,072 tracers from the top of D” to 90 km above the CMB for case 1 with colors showing the heights of these tracers above the CMB. This panel shows the locations of the tracers over the time duration of these tracers. (b) The time duration $t^*$ for these tracers to be advected from the top of D” to 90 km above the CMB. The white color show regions where $t^*$ is larger than 200 Myr. (c) Strain for tracers along their D” trajectories. Only the trajectories of 20% tracers are shown in this panel (c) to avoid overlapping of tracer paths. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS, indicating the LLVP regions.
Figure 4a shows the maximum stretches of tracers at the final locations at 90 km above the CMB outside the LLVPs together with mantle flow velocity. We quantify the angle between the direction of maximum stretch and the local flow velocity at each tracer's final location. We observe that ~50% of tracers have strain smaller than 3.0 and the other tracers have strain larger than 3.0. For tracers with strain smaller than 3.0, approximately 63% of them exhibit an angle less than 30 degrees (Figure 4b), and this slightly increases to 66% for tracers with strain larger than 3.0. Therefore, generally speaking, the stretching direction of these tracers at their final locations is similar to the mantle flow direction in these regions, and this similarity slightly increases with the amount of stretching.

**Figure 4.** (a) The lateral flow velocities (black arrows) and the projections of the strains in the lateral directions (red bars) at 90 km above the CMB for case 1 outside the LLVPs (blue contours) where the $V_s$ anomaly is less than -0.5%. (b) The statistical distribution of the angles between the flow directions and the maximum stretch directions for strain larger than 3.0 (gray) and for strain smaller than 3.0 (red).
Next, we study the distribution of strain in D\'' in greater detail. The computational domain in our geodynamic models is divided into 64 spherical shells with equal thickness and each shell is divided into 49,152 elements with equal volume. For each of the lowermost 6 spherical shells from the CMB to 270 km above it, we place one tracer at the center of each element in this shell. This gives 49,152 tracers within each shell that are evenly distributed. We first find the locations of these tracers at the top of D\'' by advecting these tracers backward-in-time, and then advect these tracers forward-in-time from the top of D\'' to their final locations within the D\'' and track strain along tracer paths. After obtaining strain of tracers at their final locations, we interpolate it to the grids at 6 depths within the D\'' for better visualization, including the CMB and 45 km, 90 km, 135 km, 180 km, and 225 km above the CMB. As shown in Figure 5, strain is generally the highest at the CMB and decreases with the height above the CMB. Almost all CMB regions have strain >4.0, with most regions having strain >7.0 (Figure 5a). Most regions at 45 km above the CMB have strain >3.0 (Figure 5b). For comparison, strain at depths of 90 – 225 km above the CMB is much lower and regions with high strain typically have linear shapes (Figure 5c-5f). We also find that, as the distance above the CMB increases, we observe a shift in the distribution of high strains. At the CMB, high strains occur in all regions outside the LLVPs, but with the increase in height above the CMB, they become concentrated along the edges of the LLVPs (Figure 5d-f). At depths of 135-225 km above the CMB, strain >6.0 only occurs in a few localized regions that typically extend from outside the LLVPs to the edges of the LLVPs (Figure 5d-f).

Figure 5. The distribution of the strain in the maximum stretched direction for case 1 (a) at the CMB and at (b) 45 km, (c) 90 km, (d) 135 km, (e) 180 km, and (f) 225 km above the CMB. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS, indicating the LLVP regions.

Motivated by the possibility that shear wave splitting measurements typically represent a path-integrated measure of anisotropy throughout the D'' layer, we calculate the depth-averaged strain from the 6 depth shells within D'' (that is, at the CMB and at 45 km, 90 km, 135 km, 180 km, and 225 km above the CMB). Average strain >2.0 is observed everywhere outside the LLVPs; furthermore, the regions with an average strain >4.0 often have a linear shape, extending from outside the LLVPs to the margins of the LLVPs and often from the edges one LLVP to the edges of the other LLVP (Figure 6a). The large strains (in Figure 6a) preferentially occur in regions with
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tracers that have been advected for long duration of $t^* > 100$ Myr (see Figure 3b), which is not surprising because strain generally increases along the path of tracers (Figure 3c). We find that, in most regions outside the LLVPs, the directions of stretch are within 15 degrees from lateral directions (Figure 6b), indicating that materials are mainly stretched laterally in the D” layer outside the LLVPs.

Figure 6. (a) The depth-averaged strain within D” for case 1. Strain is averaged from that at 6 different depths including depths at the CMB and at 45 km, 90 km, 135 km, 180 km, and 225 km above the CMB. (b) The angle between the horizontal plane and the direction of stretch for the depth-averaged strain. The gray regions have $V_S$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS, indicating the LLVP regions.

In the real Earth, D” may contain a mixture of the Bdg and pPv phases. Materials in D” can be advected from the Bdg stability region to the pPv stability region, and vice versa. Materials experience recrystallization during phase changes, although the degree to which aggregates may retain fabrics inherited across a phase transition, known as topotaxy, remains imperfectly known [but may be important; e.g., Dobson et al., 2013; Walker et al., 2018; Chandler et al., 2021]. Recrystallization may also occur when the strain of materials reaches a critical value [e.g., Wenk et al., 1997]. Importantly, recrystallization may cause the strength of preferred orientation, and thus the anisotropy of materials, to decrease. However, the amount of anisotropy reduction due to recrystallization in lowermost mantle aggregates remains unclear. In order to approximate the process recrystallization and its effects on anisotropy, we assume a simplified, endmember scenario in which we consider the amount of strain as a proxy for anisotropy strength and parameterize the effects of recrystallization (including due to phase transitions) by resetting the strain to be unity for each tracer that experiences recrystallization. For comparison, Figure 7a shows the reference depth-averaged strain in D” if we do not consider recrystallization. When recrystallization is considered (via our strain reduction parameterization), the depth-averaged strain is significantly reduced (Figure 7b-d). Specifically, if we set the critical value of strain for recrystallization to be 10, the magnitude of the depth-averaged strain is reduced mostly in regions where the reference average strains are large, but the distribution of strain is not significantly affected (Figure 7b). If we set this critical strain value to be 5, the depth-averaged strain is reduced by about half from the reference (Figure 7c). In this condition, the regions with high strain (e.g., $> 2.5$) no longer have linear shapes, and these regions mainly occur outboard of the edges of the LLVPs (Figure 7c). If we set the maximum strain to be 5 and also consider recrystallization due to phase transitions, the average strain in the D” layer is further reduced (Figure 7d), and the
preferential occurrence of high strains near LLVPs edges becomes less evident (Figure 7d) compared to what is observed for other cases (Figure 7a-c). Note that some regions at the edges of LLVPs (such as that outboard of the southeastern boundary of the African LLVP) actually exhibit relatively lower strain (in this exercise a proxy for anisotropy strength) compared to other regions, after considering strain reduction in case of recrystallization events.

**Figure 7.** The distribution of the depth-averaged strain in the D” layer for case 1 under conditions: (a) when recrystallization is not considered (same as Figure 6a); (b) when recrystallization occurs after reaching a strain of 10.0; (c) when recrystallization occurs after reaching a strain of 5.0; and (d) as in (c) with additional recrystallization when materials are advected from the Bdg stability region (defined with $\Gamma < 0.5$) to the pPv stability region (defined with $\Gamma > 0.5$). Strain is reset to be unity when recrystallization occurs. The gray regions have $V_S$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS, indicating the LLVP regions. Note the different maximum strain values in the color bars. Strain is averaged from that at 6 different depths including depths at the CMB and at 45 km, 90 km, 135 km, 180 km, and 225 km above the CMB.
3.2. Investigating the role of model parameter variations

Deep mantle flow is controlled by the density and viscosity structures, both of which involve significant uncertainties [Li, 2023a]. Therefore, we carry out models with different mantle density and viscosity structures to study how the lowermost mantle flow and strain change with these structures. The key parameters cases presented in this study are shown in Table 2. In case 2, the LLVPs are treated as purely thermal structures, with a buoyancy number \( B = 0.0 \). In case 3, we reduce the viscosity prefactor \( \eta_r \) in the lower mantle to 10; in other words, the case 3 has 7 times lower viscosity in the lower mantle than case 1. In case 4, we use \( E = 0 \) or remove the temperature dependence of viscosity. In case 5, we use \( \eta_{ppv} = 1.0 \), which effectively removes the change of viscosity due to the pPv phase transition. In case 6, the thermal density anomaly is derived from a different tomography model, SEMUCB-WM1 [French and Romanowicz, 2014]. Except for these changes, all other parameters for cases 2-6 are the same as that of case 1. In what follows, we compare D” mantle flow, strain rate, and strain cases 2-6 and the reference case 1.

Table 2. Parameters of cases 1-6. Parameters different from case 1 are bolded.

<table>
<thead>
<tr>
<th>Case</th>
<th>Nature of LLVPs</th>
<th>( \eta_r ) in lower mantle</th>
<th>( E )</th>
<th>( \eta_{ppv} )</th>
<th>Tomography model</th>
</tr>
</thead>
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<tr>
<td>1</td>
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<td>70.0</td>
<td>9.21</td>
<td>0.01</td>
<td>S40RTS</td>
</tr>
<tr>
<td>2</td>
<td><strong>Purely thermal</strong></td>
<td>70.0</td>
<td>9.21</td>
<td>0.01</td>
<td>S40RTS</td>
</tr>
<tr>
<td>3</td>
<td>Thermochemical</td>
<td><strong>10.0</strong></td>
<td>9.21</td>
<td>0.01</td>
<td>S40RTS</td>
</tr>
<tr>
<td>4</td>
<td>Thermochemical</td>
<td>70.0</td>
<td>9.21</td>
<td><strong>0.00</strong></td>
<td>S40RTS</td>
</tr>
<tr>
<td>5</td>
<td>Thermochemical</td>
<td>70.0</td>
<td>9.21</td>
<td><strong>1.00</strong></td>
<td>S40RTS</td>
</tr>
<tr>
<td>6</td>
<td>Thermochemical</td>
<td>70.0</td>
<td>9.21</td>
<td>0.01</td>
<td><strong>SEMUCB-WM1</strong></td>
</tr>
</tbody>
</table>

First, we compare the mantle flow velocities in the lowermost mantle between cases 2-6 and case 1. Figure 8a shows the mantle velocities at 2,800 km depth for case 1 as a reference, and Figures 8b-f show that for other cases. We observe that the removal of the intrinsic density anomaly of the LLVPs in case 2 does not induce significant changes to the lowermost mantle flow field (Figure 8b). By reducing the lower mantle viscosity by a factor of 7 in case 3, the magnitude of the lateral flow velocities is increased by ~4-5 times (Figure 8c; notice the different color-bar used in panel c). The removal of temperature dependence of viscosity in case 4 results in a minor reduction in velocity magnitudes and a smoother pattern of the mantle flow velocities compared to case 1 (Figure 8d). By removing the viscosity change in regions with pPv phase in case 5, the magnitudes of mantle flow velocities are also reduced (Figure 8e). By deriving the density field from a different tomography model of SEMUCB-WM1 in case 6, the mantle flow field changes in localized regions, but the overall magnitudes of the flow velocities in the D” layer remain similar (Figure 8f). Despite these distinctions, the overarching convection pattern remains similar to the first order across all 6 cases. In each case, we observe mantle flow velocities moving from downwelling regions toward the two LLVPs, and regions around the edges of the LLVPs frequently exhibit positive upwelling flows (Figure 8).
Figure 8. The mantle flow velocities for (a) case 1, (b) case 2, (c) case 3, (d) case 4, (e) case 5, and (f) case 6, at 2,800 km depth. The arrows show lateral flow velocity, and the background colors show radial flow velocity with reddish colors showing upward and blueish colors showing downward. The gray regions have $V_\text{s}$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS (a-e) and tomography model SEMUCB-WM1 (f), indicating the LLVPs. See Table 2 for model parameters for cases 1-6.

We next compare the strain rate of cases 2-6 with that of case 1 at 2,800 km depth (Figure 9). We find that the strain rate of case 2 is similar to case 1 (Figure 9b), which is not surprising, as the mantle flow fields between the two cases are similar (Figures 8a-b). In case 3, which has reduced lower mantle viscosity, the strain rate is $\sim5$ times larger than case 1; however, both cases have similar relative geographical distribution of high and low strain rates. For case 4, in which the viscosity does not depend on temperature, the strain rate is reduced ($\sim60\%$ that of case 1) but its spatial distribution remains similar (Figure 9d). As shown in Figure 9e, the strain rate of case 5, in which no viscosity change is applied to the pPv phase transition, is $\sim5$ times lower than case
1. The high strain rates found in case 1 outside the northwest boundary of the Pacific LLVP, outside the southeast boundary of the African LLVP, and off the west coast of northernmost South America, are rather small in case 5. In contrast to case 1, in which the LLVP regions have smaller strain rates than outside, large strain rates are found within and at the edges of LLVPs in case 5. In both case 1 and case 5, relatively high strain rates are observed in regions beneath the subduction zones in the western Pacific. The distribution of strain rate for case 6, which is based on a different tomography model, is shown in Figure 9f. Compared to case 1, the spatial distribution of strain rate in case 6 is less smooth and has much more short wavelength patterns (Figure 9f). Specifically, the strain rate in case 6 is smaller outside the northwest boundary of the Pacific LLVP, and larger beneath Greenland, compared to case 1 (Figure 9f). Despite these differences, the LLVPs are surrounded by regions with high strain rates in both cases and high strain rates frequently occur outside at or in close distance from the edges of the LLVPs.

Figure 9. The strain rate for (a) case 1, (b) case 2, (c) case 3, (d) case 4, (e) case 5, and (f) case 6, at 2,800 km depth. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS (a-e) and tomography model SEMUCB-WM1 (f), indicating the LLVPs. See Table 2 for model parameters for cases 1-6.
Finally, we compare the strain in the D" layer between cases 2-6 and case 1 in Figure 10. Note that the calculations of strain shown in Figure 10 do not include the effects of effective strain reduction due to recrystallization. The strain of case 2 is almost identical to case 1 (Figure 10b), which is not surprising as both the mantle flow field and strain rate are similar between these two cases (Figures 8a-b, 9a-b). Interestingly, the strains of case 3 (with reduced lower mantle viscosity) are also similar to that of case 1 (Figure 10c), although the former has 5 times larger strain rates than the latter (Figure 9c). This may be because although the strain rates in case 3 are increased, the mantle flow velocities are also increased, such that tracers travel through the D" layer more quickly. As a result, the accumulated amount of strain for each tracer, which is calculated by the product of the strain rate and the time, remains largely unchanged. The strain in case 4 is also similar to case 1 (Figure 10d). Although case 4 has lower strain rate than case 1 (Figure 9d), the former also has reduced mantle flow velocities (Figure 8d) and thus longer travel times for tracers in the D" layer. Therefore, the accumulated strain for case 4 remains similar in many regions, except that the large strains found west of the African LLVP and northwest of the Pacific LLVP in case 1 are much smaller in case 4. The strain distribution of case 5, in which no viscosity change is applied to pPv phase transition, is quite different from and is much lower than that of case 1 (Figure 10e). This is not surprising, because the distribution of strain rate of case 5 is rather different from case 1 (Figure 9e). The distribution of strain of case 6 differs from that of case 1 in some localized regions, but there are many similarities between them as well (Figure 10f). For example, in both cases 6 and case 1, high strains are observed outside the north and south boundaries of the Pacific LLVP, around the African LLVP, beneath regions near north pole, and beneath regions around Antarctica (Figure 10f). However, the high strains outside the northwest boundary of the Pacific LLVP and North America in case 1 are reduced in case 6.

It needs to mention that, for simplicity, we use a constant scaling factor \( C_{vs-\rho} \) of 0.4 between \( V_s \) anomaly and density anomaly. By increasing or decreasing the value of \( C_{vs-\rho} \) will respectively increase or decrease the magnitude of density anomaly and thus the magnitude of mantle flow velocity, but the flow direction would remain the same. The comparison of modeling results between case 3 and case 1 suggests that higher mantle flow velocity will result in larger strain rate (Figure 9a, 9c), but the strain may be less affected (Figure 10a, 10c). In reality, \( C_{vs-\rho} \) can vary with pressure, temperature, and composition [e.g., Steinberger and Calderwood, 2006]. Steinberger and Holme [2008] considered these complexities and derived density anomaly from seismic velocities using mineral physical models. They showed that D" flow generally moves towards the LLVP regions on a global scale, which is consistent with our models. This feature of D" flow has been reported by other studies as well [Yoshida, 2008; Walker et al., 2011].
Figure 10. The depth-averaged strain in the D'' layer for (a) case 1, (b) case 2, (c) case 3, (d) case 4, (e) case 5, and (f) case 6. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS (a-e) and tomography model SEMUCB-WM1 (f), indicating the LLVPs. See Table 2 for model parameters for cases 1-6.
4. Discussion

4.1. Mantle flow field, strain rate, and strain in D”

In this study, we calculate the present-day instantaneous mantle flow field in the mantle with a focus on regions outside the LLVPs in D”. We test different lower mantle viscosity and density structures. We find that the lowermost mantle flow velocities vary from model to model in localized regions; however, in all models examined in this study, mantle flow velocities in D” move towards the two LLVPs (Figure 8) which is consistent with D” flow field reported in previous studies [e.g., Steinberger and Holme, 2008; Yoshida, 2008; Walker et al., 2011]. We find that the spatial distribution of the strain rate in D” outside the LLVPs is similar among different models, with relatively high strain rates frequently occurring at or in close proximity to LLVP edges (Figure 9). An interesting finding is that D” mantle flow field and strain rate outside the LLVPs are nearly the same when the LLVPs are treated as intrinsically dense materials or purely thermal structure (Figures 8a-b, 9a-b), suggesting that it is the downwelling flows (such as that associated with plate subduction) that control mantle dynamics in these regions, not the LLVPs.

We find that strain increases along the trajectories of the tracers in D” (Figure 3c). Strain also varies with depth. Specifically, strain at the CMB is highest and it decreases with the distance away from the CMB (Figure 5). In our models, materials in D” outside the LLVPs are mainly stretched in the lateral direction. The directions of the maximum stretches are often, but not always, parallel to mantle flow directions (Figure 4). Note that strains result from the time-integrated deformation along the trajectories of tracers, whereas the mantle flow velocities represent an instantaneous snapshot of the present-day mantle dynamics. The discrepancy between the two directions in some areas therefore may indicate significant variation of mantle flow velocity directions along the trajectories of tracers in these regions. The depth-averaged strain in the D” layer increases towards the edges of the LLVPs, consistent with the fact that strain increases along the flow trajectories (which are generally) towards the two LLVPs. These characteristics of strain distribution in the D” layer outside the LLVPs is found in all 6 models, except that in case 5 without considering the viscosity reduction of pPv phase (Table 2), the strain outside the LLVPs is significantly lower than other cases with weak pPv phase (Figure 10e). Therefore, the viscosity of pPv phase plays an important role in D” deformation.

4.2. Uncertainties in the strain calculation

The calculations of strain in D” from mantle flow field suffer from several uncertainties. First, the presence of pPv phase in the lowermost mantle depends on temperature, with pPv phase preferentially occurring in relatively cold regions [e.g., Houser, 2007], but the lowermost mantle temperature is not well constrained. We also find that viscosity of pPv phase strongly affects D” flow and strain (Figure 10e). However, the change of viscosity for the Bdg- pPv phase transition remains under debated, with some studies suggested decreasing of viscosity [e.g., Hunt et al., 2009; Ammann et al., 2010] whereas others suggested increasing of viscosity [Karato, 2010].

Second, there are a number of uncertainties regarding our simplified treatment of recrystallization, which occurs when materials experience a phase transition or reach a critical amount of strain. We have implemented a simplified parameterization of recrystallization processes and found that anisotropy strength (as approximated via strain in our models) can be reduced at or near the edges of the LLVPs due to recrystallization. However, the details depend on the values of critical strain at which recrystallization starts, which are poorly known. In this study, we test different values of the maximum strain. We find that as the critical strain is reduced from...
10 to 5, only regions with strain larger than 5 are affected. If the critical strain is even smaller, materials can achieve this critical strain more easily, so recrystallization could occur more easily; as a result, there would be more frequent spatial variation of strain, which would eventually become spatially more random.

Third, we assume strain (as a proxy for anisotropy strength) is effectively reset (with value unity) at the top of D". It is reasonable to argue that materials may have already been deformed before descending into D". However, as suggested by McNamara et al. [2003], anisotropy is formed in regions where deformation is controlled by dislocation creep, which likely includes the D" layer, but is not formed (and may be erased) in regions with diffusion creep such as in the mid-lower mantle. Therefore, assuming that the strain relevant for developing texture and thus anisotropy begins with unity at the top of D" is likely a good first-order assumption. Furthermore, mantle material passing through the Bdg-pPv phase transition at the top of D" may experience a resetting of its texture, although some degree of crystallographic alignment may be inherited via topotaxy [e.g., Walker et al., 2018]. We also note that the strain information carried by tracers that are advected to relatively deeper depths in D" is tightly controlled by the trajectories of the tracers in D", making the initial strain of these tracers less important.

A fourth source of uncertainty in our modeling is our calculation of the deformation history using the instantaneous mantle flow field at the present day. This is due to the fact that mantle flow in Earth’s past history is not well constrained [e.g., Li, 2023a]. In our calculations, each tracer takes a different amount of time to migrate from its initial location at the top of the D" layer to its final location within D". We find that for case 1, this duration is typically less than 100 Myr in regions outside the LLVPs and is less than 50 Myr in downwelling regions (e.g., Figure 3b). It is longer at the edges of the LLVPs but still less than 200 Myr (e.g., Figure 3b). The LLVPs have been suggested to be relatively stable in their locations at least during the past 200 Ma [Torsvik et al., 2010]. Thus, the first-order aspects of the lowermost mantle flow field over the past 200 Ma may be similar to the present day, at least at large scales over the globe. The less we go back in time from present-day, the more closely the lowermost mantle flow field resembles that at the present-day [e.g., Li, 2023a]; therefore, the calculated strain in our models may be more robust in regions beneath downwellings, where tracers are advected for shorter durations.

4.3. Comparison between geodynamic modeling results and seismic anisotropy observations

Because deformation of anisotropic minerals in the D" layer can cause seismic anisotropy, we carry out a comparison between our predicted strain distributions and global observations of lowermost mantle anisotropy. In Figure 11, the red polygons show regions where D" anisotropy has been reported in previous studies, as compiled by [Wolf et al., 2023b]. Note that regions outside the red polygons could be places that have not yet been surveyed by seismic studies, and thus we do not intend to imply that they correspond to either isotropic or anisotropic D" regions, as discussed in detail by [Wolf et al., 2023b]. Seismic anisotropy has been found in regions outside, at the edges, and within the LLVPs (Figure 11). We compare these anisotropic locations to the distributions of the depth-averaged D" strain for case 1. We find that whether or not recrystallization-induced strain reduction is included when calculating strains, almost all regions outside the LLVPs show strains > 1.5, which is consistent with observations of seismic anisotropy in regions either near the edges of the LLVPs [e.g., Wang and Wen, 2007; Lynner and Long, 2014; Deng et al., 2017; Reiss et al., 2019; Wolf and Long, 2023] or far away from the LLVPs (Figure 11) [e.g., Garnier and Lay, 1997; Wookey et al., 2005; Long, 2009; Grund and Ritter, 2018; Asplet et al., 2020]. For each individual panel of Figure 11, both higher-than-average and lower-than-
average strains are found in regions with observed seismic anisotropy. There are some regions where relatively high strains are found in all four panels of Figure 11 and these regions also show seismic anisotropy, including the regions beneath the north of the Pacific LLVP \cite{Long2009, Asplet2020, Wolf2023a}, west of the African LLVP \cite{Pisconti2019}, and near the eastern boundary of the African LLVP \cite{Wang2007, Ford2015, Reiss2019}. We infer that in these regions, anisotropic minerals are significantly deformed to high strains and develop LPO and thus seismic anisotropy.

Figure 11. Previous observations of seismic anisotropy (red polygons, Wolf et al., 2023) and the distribution of the depth-averaged strain in the D” layer for case 1 under conditions: (a) when recrystallization is not considered (same as Figure 6a); (b) when recrystallization occurs after reaching a strain of 10.0; (c) when recrystallization occurs after reaching a strain of 5.0; and (d) as in (c) with additional recrystallization when materials are advected from the Bdg stability region (defined with $\Gamma < 0.5$) to the pPv stability region (defined with $\Gamma > 0.5$). Strain is reset to be unity when recrystallization occurs. The gray regions show the LLVPs as defined in Figure 1. Note the different maximum strain values in the color bars. This figure is the same as Figure 7 except that the previous observations of seismic anisotropy are added on each panel.

Seismic observations of anisotropic directions are often interpreted as flow directions at the present-day \cite[e.g.,][]{Nowacki2010, Ford2015, Wolf2023}. In Figure 12, we compare the direction of mantle flow in the D” layer inferred from seismic anisotropy observations and that calculated from the geodynamic calculation in case 1. We find that the directions predicted from the two different approaches agree on a global scale, both pointing...
toward the two LLVPs (Figure 12). Specifically, the horizontal flow directions suggested by seismic studies in regions of #3, #7, #8, #9, #11 and #17 agree with the geodynamic modeling results. The upwelling flows suggested by #4, #7, #15, and #16 are consistent with the positive radial velocity (darker shading in the figure) in the geodynamic model. Wolf et al. [2019] predicted converging and upwelling flows beneath Iceland in region #7, which also agrees with that from geodynamic calculation. It is encouraging that two completely independent approaches predict similar mantle flow directions in these regions. However, there are some significant disagreements between seismologically and geodynamically inferred flow directions, such as in regions of #2, #10, and #18. Further research will be required to assess why results differ in these regions.

Figure 12. Comparison between mantle flow directions in the D" layer based on previous seismic anisotropy studies (with orange arrows showing lateral flow and orange circles showing upwelling flows) and the flow velocities for case 1 at 2,800 km depth (with black arrows showing lateral velocities and the background showing regions with positive radial velocities). Numbers in yellow boxes refer to citations of 1 = [Suzuki et al., 2021], 2 = [Asplet et al., 2023], 3 = [Wolf and Long, 2022], 4 = [Kawai and Geller, 2021; Wolf and Long, 2023], 5 = [Creasy et al., 2017], 6 = [Vanacore and Liu, 2011], 7 = [Wolf et al., 2019], 8 = [Pisconti et al., 2019], 9 = [Pisconti et al., 2023], 10 = [Creasy et al., 2021], 11 = [Pisconti et al., 2023], 12, 13, 14 = [Reiss et al., 2019], 15 = [Cottaar and Romanowicz, 2013], 16, 17 = [Ford et al., 2015; Reiss et al., 2019], 18 = [Wolf et al., 2023].

There are some significant caveats when comparing geodynamic modeling results on the mantle flow field and strain distribution with seismic observations of anisotropy and their interpretation of flow directions. They include: (1) Whereas seismic anisotropy has been observed within LLVPs interiors in some regions [e.g., Wang and Wen, 2007; Reiss et al., 2019], we do not calculate strains in these regions. This is because internal density and viscosity structures, and thus the nature of small-scale internal convections within the LLVPs, are not well constrained, so predicting mantle flow within the LLVPs would be far less certain with our approach and thus
requires future exploration. (2) For many regions with high strains in our models, no D” anisotropy has been reported in previous studies. For many of these regions, this can be explained by a lack of seismic ray coverage, especially in the southern hemisphere [Wolf et al., 2023b]. In other cases where sampling is good, regions in which strain is high may be dominated by less anisotropic minerals, potentially Bdg [Romanowicz and Wenk, 2017]. (3) In this study, we assume that seismic in D” is caused by LPO, but seismic anisotropy could be caused by shape-preferred orientation (SPO) of elastically distinct materials as well. (4) Seismic anisotropy contains information about the time-integrated deformation history of minerals and is affected by mantle flow history; therefore, the mantle flow directions interpreted from seismic anisotropy observations may not always agree with the present-day mantle flow direction (Figure 4). This may explain the disagreement of mantle flow directions in some regions of Figure 12. (5) Our models demonstrate that the strains in D” often vary with depth, which agrees with the finding of [Ritsema et al., 1998], who showed that anisotropy may vary with depth. However, most D” anisotropy observations reflect shear wave splitting due to anisotropy integrated along the D” portion of the raypath. Depending on the specific seismic phases used in anisotropy measurements, seismic waves in D” may sample anisotropy along a horizontal or oblique path, whereas the strains shown in Figure 11 are averaged along the radial (vertical) direction. (6) It remains unclear how large the strain of a lowermost mantle rock must be for it to develop anisotropy that is strong enough to be seismically detectable. A number of different minerals may contribute to lowermost mantle anisotropy [e.g., Creasy et al., 2020], including post-perovskite [e.g., Yamazaki et al., 2006; Miyagi et al., 2008; Hirose et al., 2010], bridgmanite [e.g., Miyagi and Wenk, 2016], and ferropericlase [e.g., Karki et al., 1999; Merkel et al., 2002], but the degree to which they contribute to seismic observations is not well constrained. Ameliorating this uncertainty requires a more comprehensive understanding of the physical properties of lowermost mantle minerals, which can then be incorporated into predictions of anisotropy from calculated mantle flow fields. (7) The geodynamically calculated D” flow field has uncertainties as well, particularly at relatively small scales [Li, 2023a], because the mantle density and viscosity structures remain uncertain, even for the present-day Earth.

Because of the caveats and uncertainties discussed above, it remains difficult to make more specific and quantitative comparisons between geodynamic modeling of the D” mantle flow field and flow inferred from seismic anisotropy studies. Reducing these uncertainties requires better constraints on the deep mantle viscosity and density structure, improvements in the seismic coverage of D” anisotropy studies, and a more comprehensive understanding of the distribution of minerals in D” as well as their elastic properties and deformation behavior. From an observational point of view, seismic characterization of regions with weak or absent seismic anisotropy, in addition to those that exhibit strong anisotropy, is essential. Future improvements in all these areas will lead to a greatly increased understanding of the patterns and drivers of flow at the base of Earth’s mantle.
5. Conclusion

In this study, we explore the character of instantaneous mantle flow in the present day, with a focus on flow at the base of the mantle. We show that for a range of geodynamic models with different density and viscosity structures, the flow typically moves away from regions beneath downwelling centers and towards the two LLVPs on a global scale, though the details of the D” flow field vary across models in local regions. Relatively high strain rates are frequently found in regions surrounding the two LLVPs. Neither the mantle flow velocities nor the strain rates in the lowermost mantle outside the LLVPs are very sensitive to the density of the LLVPs, suggesting that the LLVPs do not have a large influence on the mantle flow fields in D” outside the LLVPs themselves. However, the magnitude and distribution of the strain rate strongly depend on the lowermost mantle viscosity: as the viscosity in the lowermost mantle decreases, flow velocities and strain rates increase. For models where the viscosity of pPv in D” is reduced by a factor of 100, the strain rates outside the LLVPs are approximately five times higher than models that do not consider the viscosity reduction due to the pPv phase.

We find that D” materials outside LLVPs are mainly laterally stretched, as might be expected in a horizontal boundary layer. The maximum stretch directions are often, but not always, similar to the mantle flow directions. These results suggest that some caution may be warranted when interpreting seismically inferred anisotropy geometry at the base of the mantle as indicating present-day flow directions. We also find that the strain is highest at the CMB and decreases with the distance above the CMB. High-strain regions outside the LLVPs often show linear shapes, extending from regions far away from the LLVPs to their edges. We approximate the effects of recrystallization by resetting the strain to be unity in our models when materials undergo phase transitions or reach a critical strain. When this recrystallization-induced strain reduction is considered, anisotropy strength (as represented by our strain proxy) can be larger in regions away from the LLVPs than at the LLVPs’ margins. All regions outside the LLVPs show strain >1.5, consistent with globally widespread observations of D” anisotropy. Many regions with observations of seismic anisotropy also show higher-than-average strains, suggesting that anisotropic minerals develop lattice-preferred orientation and thus seismic anisotropy in these regions, but seismic anisotropy has been observed in regions where smaller-than-average strain is predicted as well.

The flow direction has been inferred from seismic anisotropy observation in a few different locations in D”. It is encouraging to find that the flow direction in the majority of these locations agrees with that predicted from geodynamic modeling calculation. However, significant uncertainties remain in the predictions of strain and its connection to mantle flow field and the interpretations of seismic observations of anisotropy in terms of lowermost mantle flow; therefore, disagreement regarding the D” flow direction from the two completely different approaches is to be expected. These uncertainties will be lessened as we improve our understanding of deep mantle density and viscosity structures, the distribution of mineral phases in the D” layer, and their elastic properties, and as we obtain more observations of lowermost mantle anisotropy.
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Open research

The code and input files for geodynamic modeling, and the data used in each figure of this paper are available at https://doi.org/10.6084/m9.figshare.25330288.v1.

References


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Figure 5.
Figure 7.
(a) case 1
(b) case 2
(c) case 3
(d) case 4
(e) case 5
(f) case 6

Radial velocity (cm/yr)
Figure 9.
(a) case1
max = 5.0 \times 10^{-15}

(b) case2
max = 5.0 \times 10^{-15}

(c) case3
max = 25.0 \times 10^{-15}

(d) case4
max = 3.0 \times 10^{-15}

(e) case5
max = 1.0 \times 10^{-15}

(f) case6
max = 5.0 \times 10^{-15}

Strain rate (s^{-1})
Figure 10.
Figure 11.
Radial velocity (cm/yr)

Horizontal flow
Upwelling flow

3 cm/yr
**Table 1. Physical parameters used in this study**

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Reference value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Earth’s radius $R$</td>
<td>6371 km</td>
</tr>
<tr>
<td>Mantle thickness</td>
<td>2890 km</td>
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<tr>
<td>Reference density $\rho$</td>
<td>3300 kg/m³</td>
</tr>
<tr>
<td>Gravitational acceleration $g$</td>
<td>9.8 m/s²</td>
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<tr>
<td>Thermal expansivity $\alpha$</td>
<td>$1.0 \times 10^{-5}$ K⁻¹</td>
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<tr>
<td>Thermal diffusivity $\kappa$</td>
<td>$1.0 \times 10^{-6}$ m²/s</td>
</tr>
<tr>
<td>Temperature change across the mantle $\Delta T$</td>
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<td>Reference viscosity $\eta_0$</td>
<td>$1.04 \times 10^{21}$ Pa s</td>
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Table 2. Parameters of cases 1-6. Parameters different from case 1 are bolded.

<table>
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<tr>
<th>Case</th>
<th>Nature of LLVPs</th>
<th>$\eta_r$ in lower mantle</th>
<th>$E$</th>
<th>$\eta_{ppv}$</th>
<th>Tomography model</th>
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<td>SEMUCB-WM1</td>
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</table>
Flow and deformation in Earth’s deepest mantle from geodynamic modeling and implications for seismic anisotropy

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Key points:

- Geodynamic models show that D” materials are strongly deformed, consistent with observations of seismic anisotropy.
- Strain in D” generally increases with depth and increases towards the large low velocity provinces.
- Flow directions inferred from anisotropy observations often align with that from geodynamic calculations.
Abstract

The dynamics of Earth’s D” layer at the base of the mantle plays an essential role in Earth’s thermal and chemical evolution. Mantle convection in D” is thought to result in seismic anisotropy; therefore, observations of anisotropy may be used to infer lowermost mantle flow. However, the connections between mantle flow and seismic anisotropy in D” remain ambiguous. Here we calculate the present-day mantle flow field in D” using 3D global geodynamic models. We then compute strain, a measure of deformation, outside the two large-low velocity provinces (LLVPs) and compare the distribution of strain with previous observations of anisotropy. We find that, on a global scale, D” material is advected towards the LLVPs. Strain is highest at the core-mantle boundary (CMB) and decreases with height above the CMB. Material outside the LLVPs mostly undergoes lateral stretching, with the stretching direction often, but not always, aligning with mantle flow direction. Strain generally increases towards the LLVPs and reaches a maximum at their edges, although models that consider recrystallization suggest that anisotropy may actually be weaker near LLVP edges. The depth-averaged strain in D” is >1.5 in almost all regions, consistent with widespread observations of seismic anisotropy. The mantle flow field and strain in D” outside of LLVPs are not very sensitive to LLVP density but are strongly controlled by local density and viscosity variations outside the LLVPs. Flow directions inferred from anisotropy observations often (but not always) align with predictions from geodynamic modeling calculations.

Plain Language Summary

The Earth’s deep mantle deforms and moves at a geological timescale. This movement is called mantle convection, which controls plate tectonics. Of particular importance is the flow in the lowermost few hundred kilometers of the mantle, which is called the D” layer. The dynamics of D” plays an essential role in Earth’s thermal and chemical evolution. Direct observation of D” flow is not possible, but D” flow causes deformation of minerals that can align in preferential directions, leading to variable seismic velocities along different directions. This feature is called seismic anisotropy. In this study, we use numerical simulations to investigate D” flow and its connection to rock deformation and seismic anisotropy. We find that D” materials are strongly deformed, consistent with observations of seismic anisotropy in this layer. The strength of rock deformation in D” generally increases with depth and increases towards regions beneath the Central Pacific and Africa, where two continental-sized seismic anomalies exist. It is encouraging to find that flow directions inferred from anisotropy observations often align with our numerical simulations. This study thus improves our understanding on the dynamics of the D” layer.

1. Introduction

A grand challenge in solid Earth science is to understand the Earth’s mantle flow field, which controls deep mantle structures, the generation and mechanism of plate tectonics, and the Earth’s long-term thermal and chemical evolution. The lowermost mantle flow is of particular interest because it regulates the heat flux at the core-mantle boundary (CMB) [e.g., Nakagawa and Tackley, 2008; Li and McNamara, 2018; Li et al., 2018] which is critical for generating the magnetic field [e.g., Larson and Olson, 1991; Zhang and Zhong, 2011; Olson et al., 2014] and is essential for Earth’s thermal evolution [e.g., Christensen, 1985; Korenaga, 2008]. It dictates the formation of mantle plumes [Steinberger and O’Connell, 1998; Li and Zhong, 2017; Li and Zhong, 2019; Li, 2023a] that cause surface volcanism [Morgan, 1971]. Furthermore, the lowermost mantle greatly influences the morphology and internal structure of seismic anomalies such as the large
low velocity provinces (LLVPs) and ultra-low velocity zones (ULVZs) [e.g., McNamara et al., 2010; Li et al., 2017; Pachhai et al., 2021; Yuan and Li, 2022b]. It causes topography on the CMB [e.g., Yoshida, 2008; Lassak et al., 2010; Deschamps et al., 2018] and also affects the process of core-mantle reaction [e.g., Manga and Jeanloz, 1996; Kanda and Stevenson, 2006; Ko et al., 2022]. Moreover, it controls the advection, distribution, mixing, and accumulation of compositional reservoirs in the Earth's deep interior [e.g., McNamara and Zhong, 2005; Zhang et al., 2010; Tackley, 2011; Li, 2021; Li and McNamara, 2022; Hansen et al., 2023; Li, 2023b] and thus plays an essential role in Earth’s chemical evolution.

The present-day mantle flow field has been widely studied using geodynamic simulations. By solving the conservation equations of mass and momentum, the instantaneous mantle flow field at the present day can be computed, based on a specified mantle density and viscosity structure. This predicted flow field has been used in many applications, such as estimating the driving forces of plates [e.g., Lithgow-Bertelloni and Richards, 1995; Conrad and Lithgow-Bertelloni, 2004], calculating the dynamic topography at the surface and the CMB [e.g., Flament et al., 2013; Yang and Gurnis, 2016; Deschamps et al., 2018], and modeling the geoid [Hager and Richards, 1989; Zhong and Davies, 1999; Liu and Zhong, 2016]. However, less attention has been paid to the characteristic of the lowermost mantle flow field itself; images of lowermost mantle flow fields are only presented in a small number of publications [e.g., Steinberger and Holme, 2008; Yoshida, 2008; Walker et al., 2011].

The large-scale structure of Earth’s lowermost mantle is dominated by two large low velocity provinces (LLVPs) [e.g., Li and Romanowicz, 1996; Grand, 2002; Ritsema et al., 2004; Garnero et al., 2016]. Surrounding the LLVPs are regions with generally higher-than-average seismic velocities, which are often interpreted as former, relatively cold, subducted slabs [e.g., Lithgow-Bertelloni and Richards, 1998]. It has been suggested that flow in the lowermost mantle, on a global scale, moves away from subduction zones towards the LLVPs [Steinberger and Holme, 2008; Yoshida, 2008; Walker et al., 2011]. However, understanding lowermost mantle flow at a scale smaller than a few hundred kilometers is more challenging, due to the influence of regional density and viscosity variations that are not yet well understood [e.g., Li, 2023a].

When deformation is accommodated in the dislocation creep regime, mantle flow causes the alignment of individual crystals in an aggregate, leading to a variation of seismic velocity with propagation and polarization direction. Therefore, observations of seismic anisotropy at the base of the mantle can offer crucial insights into the deep mantle flow field [e.g., Romanowicz and Wenk, 2017]. Global seismic tomography has shown that the bulk of the lower mantle is (almost) isotropic, while strong seismic anisotropy can be found in D” [Auer et al., 2014; French and Romanowicz, 2014; Moulik and Ekstrom, 2014; Chang et al., 2015; Romanowicz and Wenk, 2017; Nowacki and Cottaar, 2021]. These global results are supported by regional investigations, which identify seismic anisotropy in many regions of D” [e.g., Garnero and Lay, 1997; Long, 2009; Nowacki et al., 2010; 2011; Asplet et al., 2020; Wolf and Long, 2022; Asplet et al., 2023; Wolf et al., 2023b]. In some regions, strong seismic anisotropy is observed outside and near the edges of the LLVPs, with weaker or absent anisotropy within the LLVPs [To et al., 2005; Cottaar and Romanowicz, 2013]. Observations of anisotropy in the D” layer are often interpreted as being caused by mantle flow and deformation. However, inferring mantle flow direction from measurements of seismic anisotropy remains challenging, because the connections among flow, deformation, and anisotropy are not well understood.
This study utilizes global geodynamic models to calculate the present-day mantle flow field and explore how variations in model parameters impact the prediction of flow velocities in the lowermost mantle. We introduce passive tracers into the model domain to track deformation history and compare the distribution of strain with seismic observations of anisotropy in D” outside the LLVPs. Through these numerical experiments, we aim to gain a better understanding of the mantle flow field in D” and its connection to observations of seismic anisotropy.

2. Methods

2.1. Setup of convection models

We solve the following non-dimensional conservation equations of mass and momentum and calculate the instantaneous mantle flow under the Boussinesq approximation:

\[ \nabla \cdot \mathbf{u} = 0, \]  
\[ -\nabla P + \nabla \cdot (\eta \dot{\epsilon}) + [Ra(T - B_c C - B_{ppv} \Gamma)] \hat{r} = 0, \]  

where \( \mathbf{u}, P, \eta, \dot{\epsilon}, Ra, \) and \( T \) are, respectively, the velocity, dynamic pressure, viscosity, strain rate, Rayleigh number, and temperature. \( B_c \) and \( C \) are respectively the buoyancy number and the fraction of an intrinsically dense compositional component. \( B_{ppv} \) is the buoyancy number that represents the density increase due to the Bridgmanite (Bdg) to post-Perovskite (pPv) phase transition and the \( \Gamma \) is the phase function. \( \hat{r} \) is a unit vector in the radial direction. Equations (1-2) are solved using the CitcomS code [Zhong et al., 2008]. The model domain ranges from the CMB to the top surface and is divided into 12 equal-volume caps with each cap containing \( 64 \times 64 \times 64 \) elements. Both the top surface and the CMB are free-slip boundaries.

The Rayleigh numbers \( Ra \) is defined as:

\[ Ra = \frac{\rho g \alpha \Delta T R^3}{\kappa \eta_0}, \]  

where \( \rho \) is density, \( g \) is gravitational acceleration, \( \alpha \) is thermal expansivity, \( \Delta T \) is the reference temperature (e.g., the non-adiabatic temperature increase from surface to CMB), \( R \) is Earth’s radius, \( \kappa \) is thermal diffusivity, and \( \eta_0 \) is reference viscosity.

The \( B_c \) and \( B_{ppv} \) are defined as:

\[ B_c = \frac{\Delta \rho_c}{\rho \alpha \Delta T}, \]  
\[ B_{ppv} = \frac{\Delta \rho_{ppv}}{\rho \alpha \Delta T}, \]  

where \( \Delta \rho_c \) the intrinsic density anomaly with respect to the background mantle and \( \Delta \rho_{ppv} \) is the density increase due to the pPv phase transition.

The phase function for the pPv phase transition is defined as:

\[ \Gamma(\pi) = 0.5 + 0.5 \tanh \left( \frac{\pi}{\delta} \right), \]  

where \( \delta \) is the width of the phase transition, and the excess pressure \( \pi \) is defined as:

\[ \pi = d - d_{ppv} - \gamma(T - T_{ppv}), \]
where \( d \) is the depth, \( d_{ppv} \) and \( T_{ppv} \) are the reference depth and temperature for the pPv phase transition, and \( \gamma \) is the Clapeyron slope. The phase function \( \Gamma \) varies from 0.0 for purely Bdg phase to 1.0 for purely pPv phase.

The viscosity depends on temperature, depth, and the presence of pPv phase, expressed as:

\[
\eta = \eta_r \exp\left[ E (0.5 - T) + \Gamma \ln(\eta_{ppv}) \right], \quad (8)
\]

where \( E \) is the activation coefficient for the temperature-dependent viscosity, \( \eta_{ppv} \) is the viscosity change due to the pPv phase transition, and \( \eta_r \) is a prefactor that controls the depth-dependence of viscosity. The \( E, \eta_{ppv} \) and \( \eta_r \) are free parameters tested in this study.

We assume a linear relationship between the density anomaly caused by thermal expansion (\( \Delta \rho_T \)) and the anomaly of seismic shear-wave velocity (\( \delta V_s \)), given by:

\[
\frac{\Delta \rho_T}{\rho} = C_{v_s-\rho} \frac{\delta V_s}{V_s}, \quad (9)
\]

where \( C_{v_s-\rho} \) is the conversion factor. The change of temperature is related to thermal density anomaly via thermal expansivity by:

\[
\delta T = - \frac{\Delta \rho_T}{\rho \alpha}. \quad (10)
\]

The non-dimensional form of the temperature anomaly is given by:

\[
\delta T' = \frac{\delta T}{\Delta T}, \quad (11)
\]

Finally, the non-dimensional temperature field in our models is calculated by:

\[
T = T_{ave} + \delta T', \quad (12)
\]

where the \( T_{ave} \) is the 1D laterally averaged profile of the non-dimensional temperature taken from present-day temperature field of a previous 3D global geodynamic model in [Li and Zhong, 2019] and is given in Supporting Information Figure S1a. In many cases, the LLVPs are treated as compositionally distinct materials. For this condition, the buoyancy field also includes the effect of intrinsic density anomalies in LLVPs in most models. Table 1 summarizes the physical parameters used in this study, which remain constant throughout the analysis.

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<td>Reference viscosity ( \eta_0 )</td>
<td>(1.04 \times 10^{21} ) Pa s</td>
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</table>
2.2. Strain calculation

Mantle flow causes deformation and, under certain conditions, orientation of minerals which can develop seismic anisotropy. Specifically, lattice preferred orientation (LPO) develops when deformation is accommodated via dislocation creep, which is generally favored by lower temperatures, higher stresses, and larger grain sizes [e.g., Frost and Ashby, 1982]. In this work, we assume that the base of the mantle is deforming in the dislocation creep regime [e.g., Karato, 1998]. This assumption is consistent with previous modeling studies [McNamara et al., 2003] as well as the observation that seismic anisotropy in the deep mantle is constrained to the lowermost few hundred kilometers of the mantle [e.g., Auer et al., 2014; French and Romanowicz, 2014]. The nature of seismic anisotropy caused by LPO is controlled by the geometry and amount of deformation. Here, the amount of deformation is measured by strain, which is defined as the ratio between the final length and the original length.

Passive tracers are introduced into the model domain and are advected by the instantaneous mantle flow field. The tracers track the strain along their paths. We use the method in [McNamara et al., 2003] to calculate strain on tracers. More detailed descriptions of this method are also given in [Mckenzie, 1979] and [Spencer, 1980]. Strain is calculated as the positive square root of the maximum eigenvalue of the left Cauchy-Green deformation tensor $\mathbf{G}$ which is defined as:

$$\mathbf{G} = \mathbf{F} \cdot \mathbf{F}^T,$$

(13)

where $\mathbf{F}$ is the deformation tensor and $\mathbf{F}^T$ is its transpose. The $\mathbf{F}$ is initially a unit tensor. It changes along the trajectories of tracers and is determined by integrating the velocity gradient tensor $\mathbf{L}$ by:

$$\frac{\partial F_{ij}}{\partial t} = L_{ik}F_{kj},$$

(14)

The trajectory of tracers is a function of time and is controlled by the initial location of tracers. In this study, we assume that tracers start to accumulate strain at the time when they first sink across the top of the D” layer, which we defined as 300 km above the CMB. Therefore, we initially introduce tracers at the top of the D” layer. We aim to obtain such an initial distribution of tracers that it allows the tracers to be advected into and sample all regions in D”. To achieve this goal, we take the following steps. First, we evenly distribute tracers at their final positions in the D” layer. Then, these tracers are advected backward-in-time until they reach their initial locations at the top of D” (again, 300 km above the CMB). We denote the time that a tracer takes to travel from its final to initial location as $t^*$. After we obtain the initial location of all tracers, they are advected forward-in-time with a time of $t^*$ to their final positions in D”. Note that each tracer takes different path and thus has a different value of $t^*$. Strain on tracers is calculated during the forward-in-time process starting from the initial locations of tracers, and all tracers begin with unity strain.

3. Results

3.1. Flow and deformation for the reference model

We first present the results for our reference model, Case 1. In this case, the density and temperature structures are converted from tomography model S40RTS [Ritsema et al., 2011] using a conversion factor $C_{\upsilon_s-\rho}=0.4$ between thermal density anomaly and S-wave velocity anomaly (Eq. 9). Similar to previous studies [e.g., Hager and Richards, 1989; Liu and Zhong, 2016], the density anomaly in the topmost 300 km of the mantle is removed because its relationship with $\upsilon_s$ anomaly
may not be well captured by a linear scaling law, due to the presence of strong composition
variations. The LLVPs have been suggested to be made of compositionally distinct material that
is intrinsically denser than the surrounding mantle [e.g., McNamara and Zhong, 2005; Mulyukova
et al., 2015; Jones et al., 2020; Li and McNamara, 2022; Yuan and Li, 2022a; Yuan et al., 2023].
To account for this, we assign a buoyancy number of $B_c = 0.3$ in regions where the $V_s$ anomaly is
$< -0.5\%$ in the lowermost 600 km of the mantle. The Rayleigh number is $Ra = 2 \times 10^8$. The viscosity
of case 1 follows Eq. (7). The activation energy is $E = 9.21$. The depth-dependent viscosity
prefactor $\eta_r$ is 1.0 in the upper mantle and 70.0 in the lower mantle. Parameters related to the pPv
phase transition are $B_{ppv} = 0.4$, $\gamma = 0.1456$, $d_{ppv} = 0.42386$, $T_{ppv} = 0.5$, and $\eta_{ppv} = 0.01$. The laterally averaged viscosity profile is shown in Supporting Information Figure S1b.

Because temperature scales linearly with $V_s$, the LLVP regions with lower $V_s$ are generally
colder than surroundings (Figure 1a). However, most regions within the LLVPs remain negatively
buoyant (Figure 1b). This is because the LLVPs in this model are assumed to be made of materials
that are intrinsically denser than the background mantle. The temperature field controls the Bdg-
pPv phase transition. At 2,800 km depth, regions outside the LLVPs are dominated by pPv phase,
whereas no pPv phase exists within the hot LLVPs (Figure 1c). Note that the fraction of pPv phase,
as represented by the phase function $\Gamma$, varies with depth. For example, compared to at 2,800 km
depth, the fraction of pPv phase is smaller at 45 km above the CMB and at depths near the top of
D" (Supporting Information Figure S2).
Figure 1. The distribution of (a) non-dimensional temperature anomaly after the horizontal average is removed, (b) residual buoyancy, and (c) the fraction of pPv phase as represented by the phase function $\Gamma$ defined in Eq. (6), at 2,800 km depth for case 1. The cyan contours correspond to a $V_s$ anomaly at -0.5% in the S40RTS tomography model [Ritsema et al., 2011], indicating the LLVP regions.

Flow in D" generally converge towards the LLVPs (Figure 2a). Downwelling flows with negative radial velocity mostly occur far from LLVPs (e.g., beneath the circum-Pacific and Indian Ocean), whereas upwelling flows mainly occur within and just outboard of the LLVPs (Figure 2b). As expected, horizontal flow dominates near the bottom boundary of the mantle outside LLVPs (Figure 2a). The radial velocity is about 10 times lower than the lateral velocity at 2,800 km depth outside the LLVPs (Figure 2b).
We calculate the 2nd invariant of the strain rate tensors. High strain rate (Figure 2c) mostly occurs in regions with high lateral flow velocity (Figure 2a, 2c). Regions with the largest strain rate include outside the southern boundary of the African LLVP, the area surrounding the Perm anomaly (northern Siberia), just west of Central and South America, and outboard the southern and northwest boundaries of the Pacific LLVP (Figure 2c).

**Figure 2.** The distribution of (a) lateral flow velocity, (b) radial flow velocity, and (c) 2nd invariant of the strain rate at 2,800 km depth for case 1. The cyan contours $V_s$ anomaly at -0.5% in the S40RTS tomography model, indicating the regions of LLVPs.

We first examine strain at 90 km above the CMB. We introduce 3,072 evenly distributed tracers at this depth, and then advect these tracers backward-in-time until they reach the top of D”. We find that most tracers when they arrive at the top of D” are in regions with downwelling flows (Figure 2c) around the Pacific Ocean and the ancient Tethys Ocean (green colors in Figure 3a). Then, for tracers at the top of D”, we set their strain to be unity, advect them forward-in-time, and track their strains along their paths until they reach their final locations at 90 km above the CMB. Note that we define the time it takes for a tracer to be advected from the top of D” to their final
locations, e.g., 90 km above the CMB, as $t^*$. As shown in Figure 3b, the $t^*$ is shortest in regions with negative radial flow velocities (e.g., downwelling regions in Figure 2b), whereas it is longer in regions with positive radial flows (e.g., near LLVP edges) where the tracers are typically advected laterally and take a long path to reach their final locations. The $t^*$ for most tracers outside the LLVPs is $<100$ Myr (Figure 3b). Tracers with $t^* > 100$ Myr mainly occur around the edges of the LLVPs (Figure 3b). The strain of tracers, after beginning with unity at downwelling locations, generally (but not always) increases along the tracer trajectories (Figure 3c).

**Figure 3.** (a) The trajectories of 3,072 tracers from the top of D'' to 90 km above the CMB for case 1 with colors showing the heights of these tracers above the CMB. This panel shows the locations of the tracers over the time duration of these tracers. (b) The time duration $t^*$ for these tracers to be advected from the top of D'' to 90 km above the CMB. The white color show regions where $t^*$ is larger than 200 Myr. (c) Strain for tracers along their D'' trajectories. Only the trajectories of 20% tracers are shown in this panel (c) to avoid overlapping of tracer paths. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS, indicating the LLVP regions.
Figure 4a shows the maximum stretches of tracers at the final locations at 90 km above the CMB outside the LLVPs together with mantle flow velocity. We quantify the angle between the direction of maximum stretch and the local flow velocity at each tracer's final location. We observe that ~50% of tracers have strain smaller than 3.0 and the other tracers have strain larger than 3.0. For tracers with strain smaller than 3.0, approximately 63% of them exhibit an angle less than 30 degrees (Figure 4b), and this slightly increases to 66% for tracers with strain larger than 3.0. Therefore, generally speaking, the stretching direction of these tracers at their final locations is similar to the mantle flow direction in these regions, and this similarity slightly increases with the amount of stretching.

![Figure 4a](image1.png)

**Figure 4.** (a) The lateral flow velocities (black arrows) and the projections of the strains in the lateral directions (red bars) at 90 km above the CMB for case 1 outside the LLVPs (blue contours) where the $V_s$ anomaly is less than -0.5%. (b) The statistical distribution of the angles between the flow directions and the maximum stretch directions for strain larger than 3.0 (gray) and for strain smaller than 3.0 (red).
Next, we study the distribution of strain in D'' in greater detail. The computational domain in our geodynamic models is divided into 64 spherical shells with equal thickness and each shell is divided into 49,152 elements with equal volume. For each of the lowermost 6 spherical shells from the CMB to 270 km above it, we place one tracer at the center of each element in this shell. This gives 49,152 tracers within each shell that are evenly distributed. We first find the locations of these tracers at the top of D'' by advecting these tracers backward-in-time, and then advect these tracers forward-in-time from the top of D'' to their final locations within the D'' and track strain along tracer paths. After obtaining strain of tracers at their final locations, we interpolate it to the grids at 6 depths within the D'' for better visualization, including the CMB and 45 km, 90 km, 135 km, 180 km, and 225 km above the CMB. As shown in Figure 5, strain is generally the highest at the CMB and decreases with the height above the CMB. Almost all CMB regions have strain >4.0, with most regions having strain >7.0 (Figure 5a). Most regions at 45 km above the CMB have strain >3.0 (Figure 5b). For comparison, strain at depths of 90 – 225 km above the CMB is much lower and regions with high strain typically have linear shapes (Figure 5c-5f). We also find that, as the distance above the CMB increases, we observe a shift in the distribution of high strains. At the CMB, high strains occur in all regions outside the LLVPs, but with the increase in height above the CMB, they become concentrated along the edges of the LLVPs (Figure 5d-f). At depths of 135-225 km above the CMB, strain >6.0 only occurs in a few localized regions that typically extend from outside the LLVPs to the edges of the LLVPs (Figure 5d-f).

![Figure 5](image)

Figure 5. The distribution of the strain in the maximum stretched direction for case 1 (a) at the CMB and at (b) 45 km, (c) 90 km, (d) 135 km, (e) 180 km, and (f) 225 km above the CMB. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS, indicating the LLVP regions.

Motivated by the possibility that shear wave splitting measurements typically represent a path-integrated measure of anisotropy throughout the D'' layer, we calculate the depth-averaged strain from the 6 depth shells within D'' (that is, at the CMB and at 45 km, 90 km, 135 km, 180 km, and 225 km above the CMB). Average strain >2.0 is observed everywhere outside the LLVPs; furthermore, the regions with an average strain >4.0 often have a linear shape, extending from outside the LLVPs to the margins of the LLVPs and often from the edges one LLVP to the edges of the other LLVP (Figure 6a). The large strains (in Figure 6a) preferentially occur in regions with
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Tracers that have been advected for long duration of \( t^* > 100 \) Myr (see Figure 3b), which is not surprising because strain generally increases along the path of tracers (Figure 3c). We find that, in most regions outside the LLVPs, the directions of stretch are within 15 degrees from lateral directions (Figure 6b), indicating that materials are mainly stretched laterally in the D” layer outside the LLVPs.

![Figure 6](image)

**Figure 6.** (a) The depth-averaged strain within D” for case 1. Strain is averaged from that at 6 different depths including depths at the CMB and at 45 km, 90 km, 135 km, 180 km, and 225 km above the CMB. (b) The angle between the horizontal plane and the direction of stretch for the depth-averaged strain. The gray regions have \( V_s \) anomalies lower than \(-0.5\%\) at 2,800 km depth in tomography model S40RTS, indicating the LLVP regions.

In the real Earth, D” may contain a mixture of the Bdg and pPv phases. Materials in D” can be advected from the Bdg stability region to the pPv stability region, and vice versa. Materials experience recrystallization during phase changes, although the degree to which aggregates may retain fabrics inherited across a phase transition, known as topotaxy, remains imperfectly known [but may be important; e.g., Dobson et al., 2013; Walker et al., 2018; Chandler et al., 2021]. Recrystallization may also occur when the strain of materials reaches a critical value [e.g., Wenk et al., 1997]. Importantly, recrystallization may cause the strength of preferred orientation, and thus the anisotropy of materials, to decrease. However, the amount of anisotropy reduction due to recrystallization in lowermost mantle aggregates remains unclear. In order to approximate the process recrystallization and its effects on anisotropy, we assume a simplified, endmember scenario in which we consider the amount of strain as a proxy for anisotropy strength and parameterize the effects of recrystallization (including due to phase transitions) by resetting the strain to be unity for each tracer that experiences recrystallization. For comparison, Figure 7a shows the reference depth-averaged strain in D” if we do not consider recrystallization. When recrystallization is considered (via our strain reduction parameterization), the depth-averaged strain is significantly reduced (Figure 7b-d). Specifically, if we set the critical value of strain for recrystallization to be 10, the magnitude of the depth-averaged strain is reduced mostly in regions where the reference average strains are large, but the distribution of strain is not significantly affected (Figure 7b). If we set this critical strain value to be 5, the depth-averaged strain is reduced by about half from the reference (Figure 7c). In this condition, the regions with high strain (e.g., \( >2.5\)) no long have linear shapes, and these regions mainly occur outboard of the edges of the LLVPs (Figure 7c). If we set the maximum strain to be 5 and also consider recrystallization due to phase transitions, the average strain in the D” layer is further reduced (Figure 7d), and the...
preferential occurrence of high strains near LLVPs edges becomes less evident (Figure 7d) compared to what is observed for other cases (Figure 7a-c). Note that some regions at the edges of LLVPs (such as that outboard of the southeastern boundary of the African LLVP) actually exhibit relatively lower strain (in this exercise a proxy for anisotropy strength) compared to other regions, after considering strain reduction in case of recrystallization events.

**Figure 7.** The distribution of the depth-averaged strain in the D” layer for case 1 under conditions: (a) when recrystallization is not considered (same as Figure 6a); (b) when recrystallization occurs after reaching a strain of 10.0; (c) when recrystallization occurs after reaching a strain of 5.0; and (d) as in (c) with additional recrystallization when materials are advected from the Bdg stability region (defined with $\Gamma < 0.5$) to the pPv stability region (defined with $\Gamma > 0.5$). Strain is reset to be unity when recrystallization occurs. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS, indicating the LLVP regions. Note the different maximum strain values in the color bars. Strain is averaged from that at 6 different depths including depths at the CMB and at 45 km, 90 km, 135 km, 180 km, and 225 km above the CMB.
3.2. Investigating the role of model parameter variations

Deep mantle flow is controlled by the density and viscosity structures, both of which involve significant uncertainties [Li, 2023a]. Therefore, we carry out models with different mantle density and viscosity structures to study how the lowermost mantle flow and strain change with these structures. The key parameters cases presented in this study are shown in Table 2. In case 2, the LLVPs are treated as purely thermal structures, with a buoyancy number of $B=0.0$. In case 3, we reduce the viscosity prefactor $\eta_r$ in the lower mantle to 10; in other words, the case 3 has 7 times lower viscosity in the lower mantle than case 1. In case 4, we use $E=0$ or remove the temperature dependence of viscosity. In case 5, we use $\eta_{ppv}=1.0$, which effectively removes the change of viscosity due to the pPv phase transition. In case 6, the thermal density anomaly is derived from a different tomography model, SEMUCB-WM1 [French and Romanowicz, 2014]. Except for these changes, all other parameters for cases 2-6 are the same as that of case 1. In what follows, we compare D” mantle flow, strain rate, and strain cases 2-6 and the reference case 1.

Table 2. Parameters of cases 1-6. Parameters different from case 1 are bolded.

<table>
<thead>
<tr>
<th>Case</th>
<th>Nature of LLVPs</th>
<th>$\eta_r$ in lower mantle</th>
<th>$E$</th>
<th>$\eta_{ppv}$</th>
<th>Tomography model</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
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<td>70.0</td>
<td>9.21</td>
<td>0.01</td>
<td>S40RTS</td>
</tr>
<tr>
<td>2</td>
<td>Purely thermal</td>
<td>70.0</td>
<td>9.21</td>
<td>0.01</td>
<td>S40RTS</td>
</tr>
<tr>
<td>3</td>
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<td>10.0</td>
<td>9.21</td>
<td>0.01</td>
<td>S40RTS</td>
</tr>
<tr>
<td>4</td>
<td>Thermochemical</td>
<td>70.0</td>
<td>0.00</td>
<td>0.01</td>
<td>S40RTS</td>
</tr>
<tr>
<td>5</td>
<td>Thermochemical</td>
<td>70.0</td>
<td>9.21</td>
<td>1.00</td>
<td>S40RTS</td>
</tr>
<tr>
<td>6</td>
<td>Thermochemical</td>
<td>70.0</td>
<td>9.21</td>
<td>0.01</td>
<td>SEMUCB-WM1</td>
</tr>
</tbody>
</table>

First, we compare the mantle flow velocities in the lowermost mantle between cases 2-6 and case 1. Figure 8a shows the mantle velocities at 2,800 km depth for case 1 as a reference, and Figures 8b-f show that for other cases. We observe that the removal of the intrinsic density anomaly of the LLVPs in case 2 does not induce significant changes to the lowermost mantle flow field (Figure 8b). By reducing the lower mantle viscosity by a factor of 7 in case 3, the magnitude of the lateral flow velocities is increased by ~4-5 times (Figure 8c; notice the different color-bar used in panel c). The removal of temperature dependence of viscosity in case 4 results in a minor reduction in velocity magnitudes and a smoother pattern of the mantle flow velocities compared to case 1 (Figure 8d). By removing the viscosity change in regions with pPv phase in case 5, the magnitudes of mantle flow velocities are also reduced (Figure 8e). By deriving the density field from a different tomography model of SEMUCB-WM1 in case 6, the mantle flow field changes in localized regions, but the overall magnitudes of the flow velocities in the D” layer remain similar (Figure 8f). Despite these distinctions, the overarching convection pattern remains similar to the first order across all 6 cases. In each case, we observe mantle flow velocities moving from downwelling regions toward the two LLVPs, and regions around the edges of the LLVPs frequently exhibit positive upwelling flows (Figure 8).
Figure 8. The mantle flow velocities for (a) case 1, (b) case 2, (c) case 3, (d) case 4, (e) case 5, and (f) case 6, at 2,800 km depth. The arrows show lateral flow velocity, and the background colors show radial flow velocity with reddish colors showing upward and blueish colors showing downward. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS (a-e) and tomography model SEMUCB-WM1 (f), indicating the LLVPs. See Table 2 for model parameters for cases 1-6.

We next compare the strain rate of cases 2-6 with that of case 1 at 2,800 km depth (Figure 9). We find that the strain rate of case 2 is similar to case 1 (Figure 9b), which is not surprising, as the mantle flow fields between the two cases are similar (Figures 8a-b). In case 3, which has reduced lower mantle viscosity, the strain rate is ~5 times larger than case 1; however, both cases have similar relative geographical distribution of high and low strain rates. For case 4, in which the viscosity does not depend on temperature, the strain rate is reduced (~60\% that of case 1) but its spatial distribution remains similar (Figure 9d). As shown in Figure 9e, the strain rate of case 5, in which no viscosity change is applied to the pPv phase transition, is ~5 times lower than case
1. The high strain rates found in case 1 outside the northwest boundary of the Pacific LLVP, outside the southeast boundary of the African LLVP, and off the west coast of northernmost South America, are rather small in case 5. In contrast to case 1, in which the LLVP regions have smaller strain rates than outside, large strain rates are found within and at the edges of LLVPs in case 5. In both case 1 and case 5, relatively high strain rates are observed in regions beneath the subduction zones in the western Pacific. The distribution of strain rate for case 6, which is based on a different tomography model, is shown in Figure 9f. Compared to case 1, the spatial distribution of strain rate in case 6 is less smooth and has much more short wavelength patterns (Figure 9f). Specifically, the strain rate in case 6 is smaller outside the northwest boundary of the Pacific LLVP, and larger beneath Greenland, compared to case 1 (Figure 9f). Despite these differences, the LLVPs are surrounded by regions with high strain rates in both cases and high strain rates frequently occur outside at or in close distance from the edges of the LLVPs.

Figure 9. The strain rate for (a) case 1, (b) case 2, (c) case 3, (d) case 4, (e) case 5, and (f) case 6, at 2,800 km depth. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS (a-e) and tomography model SEMUCB-WM1 (f), indicating the LLVPs. See Table 2 for model parameters for cases 1-6.
Finally, we compare the strain in the D” layer between cases 2-6 and case 1 in Figure 10. Note that the calculations of strain shown in Figure 10 do not include the effects of effective strain reduction due to recrystallization. The strain of case 2 is almost identical to case 1 (Figure 10b), which is not surprising as both the mantle flow field and strain rate are similar between these two cases (Figures 8a-b, 9a-b). Interestingly, the strains of case 3 (with reduced lower mantle viscosity) are also similar to that of case 1 (Figure 10c), although the former has 5 times larger strain rates than the latter (Figure 9c). This may be because although the strain rates in case 3 are increased, the mantle flow velocities are also increased, such that tracers travel through the D” layer more quickly. As a result, the accumulated amount of strain for each tracer, which is calculated by the product of the strain rate and the time, remains largely unchanged. The strain in case 4 is also similar to case 1 (Figure 10d). Although case 4 has lower strain rate than case 1 (Figure 9d), the former also has reduced mantle flow velocities (Figure 8d) and thus longer travel times for tracers in the D” layer. Therefore, the accumulated strain for case 4 remains similar in many regions, except that the large strains found west of the African LLVP and northwest of the Pacific LLVP in case 1 are much smaller in case 4. The strain distribution of case 5, in which no viscosity change is applied to pPv phase transition, is quite different from and is much lower than that of case 1 (Figure 10e). This is not surprising, because the distribution of strain rate of case 5 is rather different from case 1 (Figure 9e). The distribution of strain of case 6 differs from that of case 1 in some localized regions, but there are many similarities between them as well (Figure 10f). For example, in both cases 6 and case 1, high strains are observed outside the north and south boundaries of the Pacific LLVP, around the African LLVP, beneath regions near north pole, and beneath regions around Antarctica (Figure 10f). However, the high strains outside the northwest boundary of the Pacific LLVP and North America in case 1 are reduced in case 6.

It needs to mention that, for simplicity, we use a constant scaling factor $C_{V_s-\rho}$ of 0.4 between $V_s$ anomaly and density anomaly. By increasing or decreasing the value of $C_{V_s-\rho}$ will respectively increase or decrease the magnitude of density anomaly and thus the magnitude of mantle flow velocity, but the flow direction would remain the same. The comparison of modeling results between case 3 and case 1 suggests that higher mantle flow velocity will result in larger strain rate (Figure 9a, 9c), but the strain may be less affected (Figure 10a, 10c). In reality, $C_{V_s-\rho}$ can vary with pressure, temperature, and composition [e.g., Steinberger and Calderwood, 2006]. Steinberger and Holme [2008] considered these complexities and derived density anomaly from seismic velocities using mineral physical models. They showed that D” flow generally moves towards the LLVP regions on a global scale, which is consistent with our models. This feature of D” flow has been reported by other studies as well [Yoshida, 2008; Walker et al., 2011].
Figure 10. The depth-averaged strain in the D" layer for (a) case 1, (b) case 2, (c) case 3, (d) case 4, (e) case 5, and (f) case 6. The gray regions have $V_s$ anomalies lower than $-0.5\%$ at 2,800 km depth in tomography model S40RTS (a-e) and tomography model SEMUCB-WM1 (f), indicating the LLVPs. See Table 2 for model parameters for cases 1-6.
4. Discussion

4.1. Mantle flow field, strain rate, and strain in D”

In this study, we calculate the present-day instantaneous mantle flow field in the mantle with a focus on regions outside the LLVPs in D”. We test different lower mantle viscosity and density structures. We find that the lowermost mantle flow velocities vary from model to model in localized regions; however, in all models examined in this study, mantle flow velocities in D” move towards the two LLVPs (Figure 8) which is consistent with D” flow field reported in previous studies [e.g., Steinberger and Holme, 2008; Yoshida, 2008; Walker et al., 2011]. We find that the spatial distribution of the strain rate in D” outside the LLVPs is similar among different models, with relatively high strain rates frequently occurring at or in close proximity to LLVP edges (Figure 9). An interesting finding is that D” mantle flow field and strain rate outside the LLVPs are nearly the same when the LLVPs are treated as intrinsically dense materials or purely thermal structure (Figures 8a-b, 9a-b), suggesting that it is the downwelling flows (such as that associated with plate subduction) that control mantle dynamics in these regions, not the LLVPs.

We find that strain increases along the trajectories of the tracers in D” (Figure 3c). Strain also varies with depth. Specifically, strain at the CMB is highest and it decreases with the distance away from the CMB (Figure 5). In our models, materials in D” outside the LLVPs are mainly stretched in the lateral direction. The directions of the maximum stretches are often, but not always, parallel to mantle flow directions (Figure 4). Note that strains result from the time-integrated deformation along the trajectories of tracers, whereas the mantle flow velocities represent an instantaneous snapshot of the present-day mantle dynamics. The discrepancy between the two directions in some areas therefore may indicate significant variation of mantle flow velocity directions along the trajectories of tracers in these regions. The depth-averaged strain in the D” layer increases towards the edges of the LLVPs, consistent with the fact that strain increases along the flow trajectories (which are generally) towards the two LLVPs. These characteristics of strain distribution in the D” layer outside the LLVPs is found in all 6 models, except that in case 5 without considering the viscosity reduction of pPv phase (Table 2), the strain outside the LLVPs is significantly lower than other cases with weak pPv phase (Figure 10e). Therefore, the viscosity of pPv phase plays an important role in D” deformation.

4.2. Uncertainties in the strain calculation

The calculations of strain in D” from mantle flow field suffer from several uncertainties. First, the presence of pPv phase in the lowermost mantle depends on temperature, with pPv phase preferentially occurring in relatively cold regions [e.g., Houser, 2007], but the lowermost mantle temperature is not well constrained. We also find that viscosity of pPv phase strongly affects D” flow and strain (Figure 10e). However, the change of viscosity for the Bdg- pPv phase transition remains under debated, with some studies suggested decreasing of viscosity [e.g., Hunt et al., 2009; Ammann et al., 2010] whereas others suggested increasing of viscosity [Karato, 2010].

Second, there are a number of uncertainties regarding our simplified treatment of recrystallization, which occurs when materials experience a phase transition or reach a critical amount of strain. We have implemented a simplified parameterization of recrystallization processes and found that anisotropy strength (as approximated via strain in our models) can be reduced at or near the edges of the LLVPs due to recrystallization. However, the details depend on the values of critical strain at which recrystallization starts, which are poorly known. In this study, we test different values of the maximum strain. We find that as the critical strain is reduced from
10 to 5, only regions with strain larger than 5 are affected. If the critical strain is even smaller, materials can achieve this critical strain more easily, so recrystallization could occur more easily; as a result, there would be more frequent spatial variation of strain, which would eventually become spatially more random.

Third, we assume strain (as a proxy for anisotropy strength) is effectively reset (with value unity) at the top of D”. It is reasonable to argue that materials may have already been deformed before descending into D”. However, as suggested by McNamara et al. [2003], anisotropy is formed in regions where deformation is controlled by dislocation creep, which likely includes the D” layer, but is not formed (and may be erased) in regions with diffusion creep such as in the mid-lower mantle. Therefore, assuming that the strain relevant for developing texture and thus anisotropy begins with unity at the top of D” is likely a good first-order assumption. Furthermore, mantle material passing through the Bdg-pPv phase transition at the top of D” may experience a resetting of its texture, although some degree of crystallographic alignment may be inherited via topotaxy [e.g., Walker et al., 2018]. We also note that the strain information carried by tracers that are advected to relatively deeper depths in D” is tightly controlled by the trajectories of the tracers in D”, making the initial strain of these tracers less important.

A fourth source of uncertainty in our modeling is our calculation of the deformation history using the instantaneous mantle flow field at the present day. This is due to the fact that mantle flow in Earth’s past history is not well constrained [e.g., Li, 2023a]. In our calculations, each tracer takes a different amount of time to migrate from its initial location at the top of the D” layer to its final location within D”. We find that for case 1, this duration is typically less than 100 Myr in regions outside the LLVPs and is less than 50 Myr is downwelling regions (e.g., Figure 3b). It is longer at the edges of the LLVPs but still less than 200 Myr (e.g., Figure 3b). The LLVPs have been suggested to be relatively stable in their locations at least during the past 200 Ma [Torsvik et al., 2010]. Thus, the first-order aspects of the lowermost mantle flow field over the past 200 Ma may be similar to the present day, at least at large scales over the globe. The less we go back in time from present-day, the more closely the lowermost mantle flow field resembles that at the present-day [e.g., Li, 2023a]; therefore, the calculated strain in our models may be more robust in regions beneath downwellings, where tracers are advected for shorter durations.

4.3. Comparison between geodynamic modeling results and seismic anisotropy observations

Because deformation of anisotropic minerals in the D” layer can cause seismic anisotropy, we carry out a comparison between our predicted strain distributions and global observations of lowermost mantle anisotropy. In Figure 11, the red polygons show regions where D” anisotropy has been reported in previous studies, as compiled by [Wolf et al., 2023b]. Note that regions outside the red polygons could be places that have not yet been surveyed by seismic studies, and thus we do not intend to imply that they correspond to either isotropic or anisotropic D” regions, as discussed in detail by [Wolf et al., 2023b]. Seismic anisotropy has been found in regions outside, at the edges, and within the LLVPs (Figure 11). We compare these anisotropic locations to the distributions of the depth-averaged D” strain for case 1. We find that whether or not recrystallization-induced strain reduction is included when calculating strains, almost all regions outside the LLVPs show strains > 1.5, which is consistent with observations of seismic anisotropy in regions either near the edges of the LLVPs [e.g., Wang and Wen, 2007; Lynner and Long, 2014; Deng et al., 2017; Reiss et al., 2019; Wolf and Long, 2023] or far away from the LLVPs (Figure 11) [e.g., Garnero and Lay, 1997; Wookey et al., 2005; Long, 2009; Grund and Ritter, 2018; Asplet et al., 2020]. For each individual panel of Figure 11, both higher-than-average and lower-than-
average strains are found in regions with observed seismic anisotropy. There are some regions where relatively high strains are found in all four panels of Figure 11 and these regions also show seismic anisotropy, including the regions beneath the north of the Pacific LLVP [Long, 2009; Asplet et al., 2020; Wolf et al., 2023a], west of the African LLVP [Pisconti et al., 2019], and near the eastern boundary of the African LLVP [Wang and Wen, 2007; Ford et al., 2015; Reiss et al., 2019]. We infer that in these regions, anisotropic minerals are significantly deformed to high strains and develop LPO and thus seismic anisotropy.

Seismic observations of anisotropic directions are often interpreted as flow directions at the present-day [e.g., Nowacki et al., 2010; Ford et al., 2015; Wolf and Long, 2023]. In Figure 12, we compare the direction of mantle flow in the D” layer inferred from seismic anisotropy observations and that calculated from the geodynamic calculation in case 1. We find that the directions predicted from the two different approaches agree on a global scale, both pointing
toward the two LLVPs (Figure 12). Specifically, the horizontal flow directions suggested by seismic studies in regions of #3, #7, #8, #9, #11, and #17 agree with the geodynamic modeling results. The upwelling flows suggested by #4, #7, #15, and #16 are consistent with the positive radial velocity (darker shading in the figure) in the geodynamic model. Wolf et al. [2019] predicted converging and upwelling flows beneath Iceland in region #7, which also agrees with that from geodynamic calculation. It is encouraging that two completely independent approaches predict similar mantle flow directions in these regions. However, there are some significant disagreements between seismologically and geodynamically inferred flow directions, such as in regions of #2, #10, and #18. Further research will be required to assess why results differ in these regions.

**Figure 12.** Comparison between mantle flow directions in the D” layer based on previous seismic anisotropy studies (with orange arrows showing lateral flow and orange circles showing upwelling flows) and the flow velocities for case 1 at 2,800 km depth (with black arrows showing lateral velocities and the background showing regions with positive radial velocities). Numbers in yellow boxes refer to citations of 1 = [Suzuki et al., 2021], 2 = [Asplet et al., 2023], 3 = [Wolf and Long, 2022], 4 = [Kawai and Geller, 2021; Wolf and Long, 2023], 5 = [Creasy et al., 2017], 6 = [Vanacore and Liu, 2011], 7 = [Wolf et al., 2019], 8 = [Pisconti et al., 2019], 9 = [Pisconti et al., 2023], 10 = [Creasy et al., 2021], 11 = [Pisconti et al., 2023], 12, 13, 14 = [Reiss et al., 2019], 15 = [Cottaar and Romanowicz, 2013], 16, 17 = [Ford et al., 2015; Reiss et al., 2019], 18 = [Wolf et al., 2023].

There are some significant caveats when comparing geodynamic modeling results on the mantle flow field and strain distribution with seismic observations of anisotropy and their interpretation of flow directions. They include: (1) Whereas seismic anisotropy has been observed within LLVPs interiors in some regions [e.g., Wang and Wen, 2007; Reiss et al., 2019], we do not calculate strains in these regions. This is because internal density and viscosity structures, and thus the nature of small-scale internal convections within the LLVPs, are not well constrained, so predicting mantle flow within the LLVPs would be far less certain with our approach and thus...
requires future exploration. (2) For many regions with high strains in our models, no D” anisotropy
has been reported in previous studies. For many of these regions, this can be explained by a lack
of seismic ray coverage, especially in the southern hemisphere [Wolf et al., 2023b]. In other cases
where sampling is good, regions in which strain is high may be dominated by less anisotropic
minerals, potentially Bdg [Romanowicz and Wenk, 2017]. (3) In this study, we assume that seismic
in D” is caused by LPO, but seismic anisotropy could be caused by shape-preferred orientation
(SPO) of elastically distinct materials as well. (4) Seismic anisotropy contains information about
the time-integrated deformation history of minerals and is affected by mantle flow history;
therefore, the mantle flow directions interpreted from seismic anisotropy observations may not
always agree with the present-day mantle flow direction (Figure 4). This may explain the
disagreement of mantle flow directions in some regions of Figure 12. (5) Our models demonstrate
that the strains in D” often vary with depth, which agrees with the finding of [Ritsema et al., 1998],
who showed that anisotropy may vary with depth. However, most D” anisotropy observations
reflect shear wave splitting due to anisotropy integrated along the D” portion of the raypath.
Depending on the specific seismic phases used in anisotropy measurements, seismic waves in D”
may sample anisotropy along a horizontal or oblique path, whereas the strains shown in Figure 11
are averaged along the radial (vertical) direction. (6) It remains unclear how large the strain of a
lowermost mantle rock must be for it to develop anisotropy that is strong enough to be seismically
detectable. A number of different minerals may contribute to lowermost mantle anisotropy [e.g.,
Creasy et al., 2020], including post-perovskite [e.g., Yamazaki et al., 2006; Miyagi et al., 2008;
Hirose et al., 2010], bridgmanite [e.g., Miyagi and Wenk, 2016], and ferropericlase [e.g., Karki et
al., 1999; Merkel et al., 2002], but the degree to which they contribute to seismic observations is
not well constrained. Ameliorating this uncertainty requires a more comprehensive understanding
of the physical properties of lowermost mantle minerals, which can then be incorporated into
predictions of anisotropy from calculated mantle flow fields. (7) The geodynamically calculated
D” flow field has uncertainties as well, particularly at relatively small scales [Li, 2023a], because
the mantle density and viscosity structures remain uncertain, even for the present-day Earth.

Because of the caveats and uncertainties discussed above, it remains difficult to make more
specific and quantitative comparisons between geodynamic modeling of the D” mantle flow field
and flow inferred from seismic anisotropy studies. Reducing these uncertainties requires better
constraints on the deep mantle viscosity and density structure, improvements in the seismic
coverage of D” anisotropy studies, and a more comprehensive understanding of the distribution of
minerals in D” as well as their elastic properties and deformation behavior. From an observational
point of view, seismic characterization of regions with weak or absent seismic anisotropy, in
addition to those that exhibit strong anisotropy, is essential. Future improvements in all these areas
will lead to a greatly increased understanding of the patterns and drivers of flow at the base of
Earth’s mantle.
5. Conclusion

In this study, we explore the character of instantaneous mantle flow in the present day, with a focus on flow at the base of the mantle. We show that for a range of geodynamic models with different density and viscosity structures, the flow typically moves away from regions beneath downwelling centers and towards the two LLVPs on a global scale, though the details of the D” flow field vary across models in local regions. Relatively high strain rates are frequently found in regions surrounding the two LLVPs. Neither the mantle flow velocities nor the strain rates in the lowermost mantle outside the LLVPs are very sensitive to the density of the LLVPs, suggesting that the LLVPs do not have a large influence on the mantle flow fields in D” outside the LLVPs themselves. However, the magnitude and distribution of the strain rate strongly depend on the lowermost mantle viscosity: as the viscosity in the lowermost mantle decreases, flow velocities and strain rates increase. For models where the viscosity of pPv in D” is reduced by a factor of 100, the strain rates outside the LLVPs are approximately five times higher than models that do not consider the viscosity reduction due to the pPv phase.

We find that D” materials outside LLVPs are mainly laterally stretched, as might be expected in a horizontal boundary layer. The maximum stretch directions are often, but not always, similar to the mantle flow directions. These results suggest that some caution may be warranted when interpreting seismically inferred anisotropy geometry at the base of the mantle as indicating present-day flow directions. We also find that the strain is highest at the CMB and decreases with the distance above the CMB. High-strain regions outside the LLVPs often show linear shapes, extending from regions far away from the LLVPs to their edges. We approximate the effects of recrystallization by resetting the strain to be unity in our models when materials undergo phase transitions or reach a critical strain. When this recrystallization-induced strain reduction is considered, anisotropy strength (as represented by our strain proxy) can be larger in regions away from the LLVPs than at the LLVPs’ margins. All regions outside the LLVPs show strain >1.5, consistent with globally widespread observations of D” anisotropy. Many regions with observations of seismic anisotropy also show higher-than-average strains, suggesting that anisotropic minerals develop lattice-preferred orientation and thus seismic anisotropy in these regions, but seismic anisotropy has been observed in regions where smaller-than-average strain is predicted as well.

The flow direction has been inferred from seismic anisotropy observation in a few different locations in D”. It is encouraging to find that the flow direction in the majority of these locations agrees with that predicted from geodynamic modeling calculation. However, significant uncertainties remain in the predictions of strain and its connection to mantle flow field and the interpretations of seismic observations of anisotropy in terms of lowermost mantle flow; therefore, disagreement regarding the D” flow direction from the two completely different approaches is to be expected. These uncertainties will be lessened as we improve our understanding of deep mantle density and viscosity structures, the distribution of mineral phases in the D” layer, and their elastic properties, and as we obtain more observations of lowermost mantle anisotropy.
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Open research

The code and input files for geodynamic modeling, and the data used in each figure of this paper are available at https://doi.org/10.6084/m9.figshare.25330288.v1.

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Supporting Information for

Flow and deformation in Earth’s deepest mantle from geodynamic modeling and implications for seismic anisotropy

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Figure S1. (a) Depth profile of the laterally averaged temperature. (b) Depth profile of the laterally averaged viscosity.
Figure S2. Fraction of pPv phase as represented by the phase function $\Gamma$ at depths of (a) 45 km, (b) 90 km, (c) 180 km, and (d) 270 km above the core-mantle boundary. Cyan contours show the $V_s$ anomaly at -0.5% in the S40RTS tomography model at 2,800 km depth, showing the two large low velocity provinces.