Ubiquitous lower mantle anisotropy beneath subduction zones

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Seismic anisotropy provides key information to map the trajectories of mantle flow and understand the evolution of our planet. While the presence of anisotropy in the uppermost mantle is well-established, the existence and nature of anisotropy in the transition zone and uppermost lower mantle are still debated. Here we use 3-D global seismic tomography images based on a large data set sensitive to this region to show the ubiquitous presence of anisotropy in the lower mantle beneath subduction zones. Whereas above the 660-km seismic discontinuity slabs are associated with faster SV anomalies up to about 3%, in the lower mantle faster SH anomalies of about 2% persist near slabs down to about 1,000-1,200 km. These observations are consistent with 3-D

numerical models of deformation from subducting slabs and the associated lattice preferred orientation of bridgmanite produced in the dislocation creep regime in areas subjected to high stresses. This study provides evidence that dislocation creep may be active in the Earth's lower mantle, providing new constraints on the debated nature of deformation in this key but inaccessible component of the deep Earth.

The Earth's upper and lower mantle have quite distinct physical properties, with the characteristics of material exchange between them being a long-debated issue. Progress in global seismic tomography in the 1990s[1,2] showed that the upper and lower mantle interact mainly via subducting slabs and mantle plumes, albeit subject to the presence of strong resistance along the upper-lower mantle boundary at $\sim$660 km depth. More recently, enhanced tomography images showed that amongst the slabs that penetrate into the lower mantle, many of them stagnate down to about $\sim$1,000 km depth[3]. Conversely, mantle plumes rising from the deep lower mantle seem to deflect laterally when they reach this region[4]. However, the uppermost lower mantle, located at depths of $\sim$660–1,000 km, remains an enigmatic part of the Earth. It has been suggested that compositional layering[5,6] or a viscosity increase[7,8] may cause flow stagnation in this region, but its rheology and role in mantle convection are poorly understood.

The stagnation of subducting slabs at $\sim$660 km depth and their penetration into the lower mantle lead to intense strain and deformation around the slabs, which in turn can align mineral aggregates. Since the most abundant lower mantle mineral (bridgmanite) is anisotropic, observable seismic anisotropy should develop when considering a dislocation creep deformation mechanism[9,10,11]. However, apart from the D" region in the lowermost mantle[12], the presence of seismic anisotropy in the lower mantle is uncertain and debated[13,14,15], with most previous seismological models suggesting that the bulk of the uppermost lower mantle is radially isotropic in shear wavespeed[16]. In order to resolve this paradox, it has been proposed that the dominant deformation mechanisms in the lower mantle may not produce anisotropy, such as, e.g., through superplastic flow[17] or a pure climb creep mechanism[18].

Observations of anisotropy in the uppermost lower mantle

Some recent regional shear wave splitting studies suggest the presence of anisotropy in the transition zone and uppermost lower mantle near some subduction zones[19−21]. However, the limited depth resolution and azimuthal coverage in regional studies together with the difficulty in isolating lower mantle anisotropy from upper mantle effects can restrict the interpretation of these studies. While illuminating mostly large-scale features, global anisotropy tomography overcomes these issues by mapping the whole mantle, which is key to interpret large-scale processes and global mantle flow in a unified way[22,23]. Nevertheless, several issues such as the use of different data and modelling approaches, notably when handling crustal effects[24−26], led to poor agreement between past global mantle anisotropy models.

SGLOBE-rani is a recent whole-mantle shear-wave radially anisotropic model based on a large seismic data set of over 43M seismic measurements with complementary sensitivity to the entire Earth's mantle. It models simultaneously crustal thickness and mantle structure to reduce artefacts in the retrieved anisotropic structure[27,28]. The use of a huge set of over 10M surface wave overtone measurements, which have sensitivity down to $\sim$1,000 km depth (Supplementary Fig. S1), enables good data coverage in the transition zone (Supplementary Fig. S2-S4). Below that, a large set of body-wave travel time measurements assures good data coverage in the remainder of the lower mantle (Supplementary Fig. S3). However, the poor balance between SV- and SH-sensitive travel-time data in existing body-wave data sets leads to poorly resolved lowermost mantle anisotropy and leakage effects[28], in agreement with the findings from other previous whole mantle anisotropy studies[29,30]. Thus, we take the conservative approach of not interpreting any anisotropic structures below $\sim$1,400 km depth. Chang et al.[28] compared SGLOBE-rani with other recent global anisotropy models and, as expected, found better correlations between the isotropic part of the models than between the anisotropic structure. Yet, a correlation of about 0.5 was found between the anisotropic structure in SGLOBE-rani and in the recent model *Savani*[31], which was built with a different modelling scheme. This is a substantial improvement in global anisotropy tomography, since correlations with and between other models tend to not exceed about 0.3[29−31]. The correlation between *Savani* and SGLOBE-rani is improved possibly because both models use, among other data sets, a large number of surface wave overtone dispersion measurements. In this study, we focus primarily in SGLOBE-rani, for which we know the full details of the modelling procedure. Given the improved correlation with *Savani* compared to other models, we also show comparisons of key features in these two models.

Anisotropy beneath subduction zones

We systematically study the patterns of anisotropy in SGLOBE-rani around major global subduction zones. Average 1-D profiles of radial anisotropy beneath subduction zones show deviations from the global 1-D average of SGLOBE-rani (Fig. 1). Above 660 km depth slabs are mostly associated with average faster SV anomalies ($V_{SH} < V_{SV}$) up to 3%, whereas in the lower mantle average faster SH anomalies ($V_{SH} > V_{SV}$) up to 2% seem to be mostly confined in the 660-1,200 km depth range. Depth cross-sections of SGLOBE-rani in various subduction zones (Fig. 2, Supplementary Fig. S5) show the clear fast V_S signature of slabs and a good match to subduction zone seismicity. Similar to previous tomography models, SGLOBE-rani shows slabs in various stages: (i) ponding at the bottom of the upper mantle, such as beneath Honshu (Fig. 2), Bonin and Northern Chile (Supplementary Fig. S5); (ii) trapped in the region between 660 km and $\sim$1,000 km depth (e.g., beneath the Northern Kuriles, Kermadec, Eastern Java; see Fig. 2 and Supplementary Fig. S5); and, (iii) penetrating below 1,000 km, such as beneath Central America, Western Java and possibly Northern Peru (Fig. 2 and Supplementary Fig. S5). As seen in the 1-D profiles shown in Fig. 1, in general the anisotropy cross-sections show faster SV anomalies near the slabs in the upper mantle (Fig. 2), whose strength reduces as depth increases towards the 660-km discontinuity. Below 660 km depth, fast SH anomalies are imaged near the slabs. These anomalies tend to be observed beneath the steeply dipping section of the slab and below the slab's tip (Fig. 2). They appear both in regions with good data coverage, such as the Western Pacific (Fig. 2 and Supplementary Fig. S5) as well as in regions with more limited coverage and resolution, such as South America (Supplementary Fig. S5). Since the resolution of radial anisotropy is lower than for isotropic structure, and the anisotropic images are susceptible to smearing effects (e.g., Fig. S6), we carried out resolution tests to verify the robustness of the features imaged (Supplementary Figs. S3, S4, S6), particularly beneath the Western Pacific, and we verified that the images of uppermost lower mantle anisotropy are not affected by leakage effects from isotropic structure (Supplementary Fig. S7–S9). Moreover, a F-test suggests that the radial anisotropy anomalies below 660 km depth are significant (see the supplementary information). Compared with the $Savani$ model and a high-resolution P-wave model[3], SGLOBE-rani agrees very well with their isotropic structure and shows similar long-wavelength anisotropic features near subduction zones (Supplementary Fig. S10). In addition, in order to assess in an objective, quantitative way the anisotropic anomalies mapped in SGLOBE-rani, we performed a statistical K-means cluster analysis[32−34] of the radial anisotropy profiles over the entire globe in the mantle down to 1,200 km depth (Fig. 3). Subduction zones are clearly captured in the analysis as regions of anisotropy anomalies in the transition zone and in the uppermost lower mantle, which shows that the observed anomalies near slabs make a distinct class of features in SGLOBE-rani. When examining differences in anisotropy as a function of slab stagnation depth we do not find any clear trends. Future higher resolution, regional studies of specific subduction slabs with different stagnation depths may enable such detailed analyses.

Geodynamical mantle modelling and fabric calculations

In order to interpret our seismic images, we performed a series of 3-D petrological-thermo-mechanical simulations and mantle fabric calculations using the modelling strategy of ref. 35. At high pressures and temperatures, viscous deformation is accommodated by a combined diffusion-dislocation creep mechanism. For microflow simulations of strain-induced lattice preferred orientation (LPO), we consider only the fraction of deformation accommodated in the dislocation creep regime together with a harzburgitic composition in the upper mantle and a pyrolitic transition zone and lower mantle (see the supplementary information). Slab stagnation at the 660-km discontinuity is promoted by fast trench retreat and the negative Clapeyron slope of the 660-km discontinuity[36,37] (Fig. 4e, g, Supplementary Movie S1). On the other hand, slab penetration into the lower mantle (Fig. 4f, h, Supplementary Movie S2) is achieved by reducing the viscosity contrast between the upper and lower mantle (see the supplementary information and Table S1). The rheological model considered includes a viscosity hill in the mid-mantle around 1,500 km depth. This is consistent with the recently proposed viscosity jump in the mid-mantle[7] and can also be compatible with compositional layering in this region[5], such as the proposed bridgmanite-enriched ancient mantle structures (BEAMS) in the lower mantle[38]. This feature helps to localise the deformation and thus the fabrics in our modelling of the 660 –1,000 km depth range for stagnating slabs. The radially anisotropic shear wave structures obtained from the fabric calculations show a good agreement with the seismic tomography images (Fig. 4a, b). The rigid slab is isotropic because of the low deformation. Previous numerical studies have calculated the development of lower mantle fabric within the slab, finding that significant anisotropy can be generated in the lowermost mantle[39,40]. Here, we show that in the uppermost lower mantle anisotropy develops mostly around the rigid slab, not within it, which is due to high (subduction-induced) strain in the surrounding hot mantle. In the upper mantle, fast SV anomalies are observed near the slab, whereas the bottom of the transition zone ($\sim$520–660 km) is isotropic since the dominant phases in these P-T conditions (ringwoodite and garnet) are nearly isotropic[41] (Fig. 4e–h). In the lower mantle, faster SH anomalies are observed beneath the slab. As explained previously, these features are also observed in SGLOBE-rani (Fig. 4a, b). Remarkably, our geodynamical modelling seems to reproduce well the two lower mantle faster SH anomalies often observed in SGLOBE-rani beneath the steeply dipping section of the slab and below its tip (Figs. 2 and 3a, b; Supplementary Movie S1). In these two areas, dynamic models predict that a dislocation creep mechanism is activated by the high deviatoric stresses resulting from the early interaction of the slab's tip with the lower mantle, and successively from the load transmitted by the dipping portion of the slab, which is migrating away from the slab's tip (see the supplementary information). We tested these results with different sets of elastic constants and found the same conclusions (Fig. S13, see the supplementary information). In models where the slab penetrates directly into the lower mantle, a single positive radial anisotropy anomaly was found around the slab tip, which is consistent with the observations in Kermadec I (Fig. 2), Northern Central America and Northern Peru (Fig. S5). Recently, Girard et al.[42] reported grain-scale fabrics at relatively low strains in a bridgmanite-MgO mixture deformed at uppermost lower mantle conditions, with the weak MgO crystal elongating quickly along the shear direction. Further deformation would increase the grain shape-preferred orientation (SPO), possibly generating a penetrative schistosity and associated apparent seismic anisotropy. We have thus calculated the radial anisotropy resulting from grain-scale SPO at relatively high deformation (see the supplementary information). Due to the large contrast in mineral isotropic elastic properties in the region between the post-spinel and post-garnet reactions, apparent radial anisotropy goes up to $\sim$2.2% just below the 660-km discontinuity. However, at greater depths the calculated radial anisotropy decreases to about half of the observed positive anomalies due to the majorite-bridgmanite phase transformation (Fig. S14). Hence, grain-scale SPO could potentially explain the radial anisotropy found below stagnating slabs, while the anisotropy observed below penetrating slabs requires an additional source. Given that the modelled apparent anisotropy provides an upper bound estimate (see the supplementary information), these results suggest that bridgmanite LPO is present beneath the slab.

Reconciling seismic and mineral physics information

Our seismic tomography images provide comprehensive, compelling evidence of the ubiquitous presence of anisotropy near subducting slabs in the uppermost lower mantle, which has been thought to be mostly isotropic until recently. This reconciles previous contradictions between mineral physics and seismological observations of this key region of the Earth. Since it is quite challenging to fully quantify model uncertainty, currently it is not possible to quantify whether the anomalies imaged are beyond model errors. Nevertheless, a series of resolution and robustness tests show that the fast V_{SH} anomalies imaged near slabs in the uppermost lower mantle in SGLOBE-rani are robust. Moreover, a statistical clustering analysis demonstrates that these features make a distinct class of anomalies specific to global subduction zones. Anisotropy anomalies near subducted slabs in the deep mantle have been previously suggested in a few, scarce studies[19–21], but here we have been able to constrain their geometry globally in a quantitative way.

The imaged anisotropic anomalies in the lower mantle are mostly confined to the ∼660–1,200 km depth range, which may indicate the presence of dominant diffusion creep or a decrease in viscous anisotropy of mantle minerals below this depth. Comparisons with results from 3-D petrological-thermo-mechanical modelling show that the imaged anisotropic anomalies are consistent with strain-induced deformation of the lower mantle surrounding the slab where dislocation creep is predominant, yielding a LPO of anisotropic bridgmanite. Since significant uncertainties of the rheology and the thermodynamic and thermoelastic properties of lower mantle materials need to be considered, we tested different parameters suggested in the literature, and still found these same conclusions. In addition, we computed grain-scale SPO and found that it may partly explain the uppermost lower mantle radial anisotropy observed near slabs stagnant at the 660-km discontinuity. However, the seismic observations of slabs penetrating into the lower mantle require an additional source of anisotropy. Thus, our study brings new physically consistent seismological evidence that dislocation creep may be active in the Earth's lower mantle.

Methods

Here we describe the methodology of tomography models, 3-D petrological-thermo-mechanical simulations and mantle fabric calculations. For figures and tables, we refer the reader to the main text as well as the Supplementary Information.

Seismic tomography: resolution tests and model comparisons We carried out detailed data misfit analyses and resolution tests of SGLOBE-rani in previous studies[27–28], which showed good overall resolution of anisotropy in regions with good data coverage down to at least 1,400 km depth. Specifically, we found that the lateral resolution of our model is $\sim$1,000 km[28]. In this study we carry out further resolution tests specifically focused on the retrieved images near subduction zones in the mid-mantle.

Checkerboard tests for radial anisotropy including noise Figures S2-S4 show checkerboard tests conducted to assess the resolving power of the various datasets used in the construction of SGLOBE-rani (body wave travel-times, and surface wave fundamental mode and overtone dispersion data). Synthetic data are computed for a given input model with alternating anomalies using the same source-receiver configuration as in the real data inversions. These synthetic data are then inverted for isotropic and radially anisotropic structure to check how well the input model is recovered. Fig. S3 shows that while the overtone data are key to constrain radial anisotropy in the transition zone, the body-wave travel-time data help resolve the radial anisotropy in the uppermost lower mantle. In order to render the tests more realistic,

Fig. S4 shows the results of a checkerboard test with all the synthetic data considered, with 5% uncorrelated Gaussian random noise added. This test shows that the radial anisotropy in the transition zone and uppermost lower mantle are well recovered in the presence of noise.

Backus-Gilbert averaging kernels We compute Backus-Gilbert averaging kernels for twelve different locations in global subduction regions at a depth of 800 km (Fig. S6). These kernels[45,46] describe how the velocity or anisotropy perturbation at a given point is a spatial average of the real structure. Ideally Backus-Gilbert averaging kernels are delta functions, but in practice they have a finite spatial extent due to incomplete data coverage, model regularisation and the finiteness of model parametrisation. Fig. S6 shows horizontal and vertical cross-sections through these averaging kernels. As expected, isotropic structure (solid lines) is better resolved than anisotropy (dashed lines), with the anisotropy kernels being broader than the velocity kernels. Moreover, the regions with good data coverage are better resolved than those with poorer coverage (e.g., Honshu compared to Northern Chile; Figs. S5, S6), but overall the anisotropic structure is reasonably well resolved.

Leakage effects from isotropic into anisotropic structure in the mid-mantle We also tested potential leakage effects from isotropic to anisotropic structure in the uppermost lower mantle. Specifically, we carried out a synthetic test whereby the input isotropic structure is the same as in SGLOBE-rani and the input anisotropic structure corresponds to SGLOBE-rani in the upper mantle, with no anisotropy below 660 km depth. We then computed synthetic data for this model for all the source-receiver data pairs used to build SGLOBE-rani. These synthetic data were then inverted for isotropic and anisotropic structure in the whole mantle, using the exact same inversion procedure as that used to build SGLOBE-rani. Fig. S7 presents the results obtained, showing that, as expected, below 660 km depth there is no coherent pattern of anisotropic structure in the retrieved model. Thus, the fast SH anomalies observed in SGLOBE-rani beneath slabs in the uppermost lower mantle do not seem to be affected by leakage effects of isotropic structure or of upper mantle anisotropy. In addition, we have also performed additional leakage tests whereby we force the anisotropy to remain above 300 km depth and above 500 km depth (Figs. S8-S9). Similar to the results in Fig. S7, the test in Fig. S8 shows that when forcing the anisotropy to remain in the uppermost mantle there is no artificial anisotropy introduced in the mantle transition zone and uppermost lower mantle. Moreover, the test in Fig. S9 shows only weak smearing at 600 km depth from anisotropy anomalies at shallower depths.

Comparisons with other tomographic analyses Fig. S10 compares the isotropic structure in SGLOBE-rani beneath several subduction zones with that in the GAP_P4[3] and *Savani*[31] models, where GAP_P4 is a high-resolution isotropic P-wave model with good resolution beneath subduction zones[3]. Fig. S10 shows that SGLOBE-rani resolves subducting slabs well, showing very similar features to those in GAP_P4 and in *Savani*. The same figure compares the anisotropic structure in SGLOBE-rani and *Savani*, clearly showing that both models show faster SV anomalies near subducting slabs in the upper mantle and faster SH anomalies beneath the slabs in the uppermost lower mantle. These anomalies are constrained by the surface wave overtone data as well as by the body-wave travel-time measurements used in the tomographic inversions (Fig. S3).

Clustering analysis In order to assess the fast SH anomalies in the uppermost lower mantle in an objective, quantitative way, we performed a K-means cluster analysis[32–34] of the radial anisotropy profiles over the entire globe in the mantle down to 1,200 km depth in SGLOBE-rani (Fig. 3). K-means cluster analysis is an automatic statistical tool that partitions a set of objects into k groups of similar objects, where k is a pre-assigned integer, which in our case goes from 2 to 4. Since the analysis is easily reproducible and automated, clustering techniques are becoming increasingly popular in seismological studies[33,34]. Here, we apply the MATLAB implementation of the K-means cluster analysis to vertical profiles of radial anisotropy (ξ) in SGLOBE-rani. We use an iterative approach whereby the squared Euclidean distance measure is used to group ξ profiles according to their similarity; after five iterations convergence is achieved. We find that the analysis clearly captures clusters of anisotropic anomalies in the transition zone and in the uppermost lower mantle beneath subduction zones in SGLOBE-rani (blue regions), notably when $k = 3$. Different colours represent different regions identified (blue, red, brown and orange). Thus, the observed fast SH anomalies in the uppermost lower mantle make a distinct family of anomalies within SGLOBE-rani.

Data misfit considerations Chang et al.[28] reported that when including lateral variations in radial anisotropy and crustal thickness perturbations in the construction of SGLOBE-rani led to a misfit reduction of 8% compared to an inversion with no anisotropy beyond that in PREM, and with exactly the same number of free parameters. Since globally on average the uppermost lower mantle is isotropic and the transition zone is only slightly anisotropic (Fig. 1), there is only a slight reduction in the overall data misfit compared to isotropic inversions when including these regions in global tomographic inversions[29,30]. Nevertheless, we performed a statistical F-test on SGLOBE-rani by comparing it to a model obtained by inverting for isotropic and crustal thickness structure without allowing lateral variations in anisotropy below 660 km depth, which we refer to as SGLOBE-rani-B. We performed the F-test by computing: $F = \frac{[\chi^2(r)-\chi^2(p)]/(p-r)}{\chi^2(p)/(n-p)}$, where χ^2 is the misfit function, n is the number of data points used in the inversions, and p and r are the numbers of model parameters for the two models which are under comparison ($p > r$). By calculating the integral probability $P_F(F; p-r, n-p)$ we can get the probability of how much the two models are different statistically[47]. Since errors in the data and in the modeling employed in our inversions are essentially unknown, here we use a misfit function χ^2 based on the sum of the squares of the residuals (variance) following, e.g., Forsyth, 1975[48]. The variance reductions of SGLOBE-rani and SGLOBE-rani-B are 0.763 and 0.754, respectively. The probability that the improved fit of SGLOBE-rani compared to SGLOBE-rani-B occurs by chance is less than 1% (i.e., the two models differ at the p=0.01 level). Thus, the F-tests support the significance of the radial anisotropy anomalies below 660 km depth.

In order to ensure good global data coverage, source-receiver paths used in the analyses are typically long, which makes it difficult to isolate the effect of a specific region, such as the uppermost lower mantle, on the data fit. Future regional studies performing forward modelling of the imaged anisotropy in subduction regions should enable us to quantitatively assess how well our images of uppermost lower mantle anisotropy fit regional data. However, such analyses involve intense numerical calculations and are thus well beyond the scope of this study.

3-D petrological-thermo-mechanical simulations We use the same modelling scheme as in ref. 35 and in ref. 23, with some modifications. For conciseness and to avoid unnecessary repetition, here we briefly explain the technique used; for further details, we refer the reader to the studies of ref. 35 and ref. 23.

The simulations were produced using I3MG, which is a 3D geodynamic framework based on a mixed Eulerian-Lagrangian finite difference scheme[49]. The model domain was defined as 6,000 $\times$ 3,000 $\times$ 3,000 km with 293 $\times$ 293 $\times$ 69 Eulerian nodes (x, y, z co-ordinates, respectively). Subduction is driven self-consistently by a 80-Myr-old, 90 km-thick oceanic plate slab, 3,260 km long and 1,000 km wide, juxtaposed on a 1 Myr-old mantle. An adiabatic geotherm of 0.5 K/km is applied below 90 km, while above this depth we define the thermal structure with a half-space cooling model. The stagnancy and penetration behaviour of slabs is strongly affected by mantle phase transitions, which are included in the modelling via density and enthalpy maps obtained from PERPLE_X[50] as a function of pressure and temperature for pyrolite.

A composite visco-plastic rheology is used in which the effective viscosity combines low-T Peierls creep and high-T dislocation and diffusion creep mechanisms. The mantle effective viscosity μ_{eff} is computed as:

$$\mu_{eff} = \left(\frac{1}{\mu_{diff}} + \frac{1}{\mu_{disl}} + \frac{1}{\mu_{Peierls}}\right)^{-1} \quad , \tag{1}$$

where:

$$\mu_{diff} = \frac{1}{2} A_{diff} \exp\left(\frac{H_{diff}}{RT}\right) \tag{2}$$

$$\mu_{disl} = \frac{1}{2} A_{disl} \sigma_{II}'^{\,1-n} \exp\left(\frac{H_{disl}}{RT}\right) \tag{3}$$

$$\mu_{Peierls} = \frac{1}{2} A_{Peierls} \sigma_{II}'^{\,-1} \exp\left(\frac{H_{Peierls}}{RT}\left(1 - \left(\frac{\sigma_{II}'}{\sigma_{Peierls}}\right)^{p}\right)^{q}\right) \quad (\text{for}\, T < 1,373\, K) \tag{4}$$

The pseudo-plastic viscosity is computed as:

$$\mu_{yield} = \frac{\tau_{yield}}{2\dot{\varepsilon}_{II}} \quad , \tag{5}$$

where the yield stress is given by the Drucker-Prager criterion:

$$\tau_{yield} = C_{DP} + \mu P \quad . \tag{6}$$

Table S1 defines all the physical parameters involved in equations (1)–(6) and shows all the values used in the simulations for both scenarios of slab stagnation at the 660-km discontinuity and of penetration into the lower mantle. The effective viscosity and the fraction of deformation accommodated by dislocation creep along the initial geotherm are plotted in Fig. S11. The high-temperature viscous rheological model based on a depth-dependent activation enthalpy and pre-exponential factor[51,52] yields a low viscosity asthenosphere, a viscosity jump at 660 km depth at medium to high stresses and a viscosity hill in the mid mantle around 1,500 km depth[7]. Dislocation creep accommodates most of the deformation in the upper mantle and transition zone[53], while in the lower mantle it is active only at deviatoric stresses above 10-20 MPa (Fig. S12). Given the relatively low horizontal and vertical resolution ($\sim$20 and 10 km along the x- and y-co-ordinates, respectively), the plate and background mantle have a 30 km thick crust with a strain-dependent low coefficient of friction, which is required to ensure the plate interface's lubrication at shallow depths. The low coefficient of friction of the plate's crust is then increased at 100 km, simulating slab dehydration. Density in the crust is computed in the same way as for the mantle.

Mantle fabric modelling After the geodynamical simulations, the mantle polycrystalline aggregates are advected using the velocity field obtained in the macro-flow calculations. At each time step we compute fabric development of each aggregate using a modified version of D-REX[23,35], which takes into account non-steady-state deformation and deformation history[53,54], combined diffusion-dislocation creep mechanisms and strain-induced LPO of the mid-mantle aggregate. The fabric is computed only for the fraction of viscous deformation accommodated by dislocation creep and only for mineral phases that have substantial single-crystal visco-elastic anisotropy, such as olivine, enstatite, wadsleiyte and bridgmanite. Thus, for aggregates of cubic phases such as ringwoodite, garnet and MgO-periclase, which are mostly isotropic above 1,500 km depth[55], we keep their crystal orientation random. We use the normalised Critical Resolved Shear Stress (CRSS) for slip systems of olivine, enstatite and wadsleiyte reported in laboratory experimental studies[23]. Regarding bridgmanite, we have tested the case of [001](100) easy slip system, a condition which was recently found in simple shear deformation experiments in Kawai type deformation-DIA apparatus reported by Tsujino et al.[10]. The lower mantle fabrics were calibrated according to those of ref. 10 by imposing a CRSS five times lower than for all the others slip systems[55]. The full elastic tensor of each aggregate is then computed as a function of the single crystal elastic constants (scaled at local P-T conditions), crystal orientation and volume fraction. For bridgmanite, we have tested elastic constants from two independent ab-initio simulations: those reported by Wentzcovitch et al.[56], and those obtained from interpolation of data reported by Zhang et al.[57] (see Fig. S13). The two sets of elastic constants yield similar patterns of shear wave anisotropy in the uppermost lower mantle. However, the magnitude of shear wave anisotropy is about 40-50% higher when using the elastic constants reported in ref. 53 (Fig. S13).

Extrinsic (apparent) radial anisotropy resulting from grain-scale SPO has been calculated with the effective medium theory developed by Backus[58], thus assuming planar schistosity for those mantle aggregates with a bulk Finite Strain Ellipsoid (FSE) characterised by the arbitrary $\ln(\frac{a_1}{a_3}) > 4$, where a_1 and a_3 are the maximum and minimum FSE semi-axes. This logarithmic ratio corresponds to a bulk simple shear strain $\varepsilon_b = 3.6$ accumulated by mantle aggregates throughout the simulation. Since the uppermost lower mantle fabrics are mostly formed within the entrained subslab asthenosphere, this implies assuming that pre-existing micro-structures are not significantly destroyed by phase transformations occurring across the transition zone. The smooth transversely isotropic long-wavelength equivalent (STILWE) medium has been computed using the P-T-dependent isotropic elastic moduli and volume fractions of the different crystal components in a pyrolitic mantle calculated using the mantle thermodynamics model HeFESTo[59]. The resulting elastic tensor is then rotated so that the layering is normal to the shortest axis of the FSE (Fig. S13). It is important to note that since a perfectly layered medium is assumed, the STILWE provides an upper bound estimate of the anisotropy. Thus, considering also that the micro-structures could be substantially weakened by the different phase transformation within the entrained subslab asthenosphere, the strength of the micro-fabrics and of the associated apparent anisotropy on Earth are probably lower than this study's calculations.

Data availability. The data that support the findings of this study are available from the corresponding author upon request. The tomography model SGLOBE-rani used in this study is available in the IRIS Data Services Products (http://ds.iris.edu/ds/products/emc-earthmodels/).

Code availability. The large scale subduction models were built using the code I3MG, which is not freely available and was kindly provided by Taras Gerya.

The mantle fabrics calculations used a modified version of the code D-REX available at http://www.ipgp.fr/~kaminski/web_doudoud/DRex.tar.gz .

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Figure captions

Figure 1. Comparison of 1-D averages of radial anisotropy beneath various subduction zones. Left: Location of the profiles used to compute the depth-dependent 1-D average of radial anisotropy in nine different subducting regions (N. Kurile, S. Kurile, Honshu, S. Bonin, E. Java, W. Java, Kermadec, C. America and N. Peru). The coloured lines correspond to locations close to the cross-sections shown in Fig. 2 and Supplementary Fig. S5. Right: Depth-dependent 1-D average of radial anisotropy ($\xi = \frac{V_{SH}^2}{V_{SV}^2}$) in the model SGLOBE-rani[28] for the nine subduction regions considered. The black line shows the global average in SGLOBE-rani. The averages of radial anisotropy are computed over 1-D depth profiles every 2^o along the coloured lines shown for each region on the left. The locations of the cross-sections are based on those used in ref. 3.

Figure 2. Cross-sections of perturbations in Voigt average and anisotropic structure. Cross-sections of the global tomography model SGLOBE-rani beneath Honshu, N. Kurile, W. Java and Kermadec. V_S denotes perturbations in the Voigt average model ($V_{Voigt}^2 = \frac{2V_{SV}^2 + V_{SH}^2}{3}$) with respect to PREM[43] down to the core-mantle boundary and ξ denotes perturbations in radial anisotropy ($\xi = \frac{V_{SH}^2}{V_{SV}^2}$) down to 1,400 km depth (below this depth the resolution is more limited, ref. 28). Focal depths from EHB data[44] with an upper bound of 60 km are superimposed in the cross-sections as grey circles. The depths of 410 km, 660 km and 1,000 km are represented by solid black lines. For reference, we use the same geographical locations and codes (Honshu D, W. Java G, N. Kurile I and Kermadec I) as in ref. 3.

Figure 3. K-means clustering analysis of the radially anisotropic structure in SGLOBE-rani. K-means clustering analysis of the radially anisotropic structure in SGLOBE-rani whereby the radial anisotropy ξ is grouped into k families[32−34], where k is a pre-assigned integer, which in our case goes from 2 to 4. We find that clusters comprising subduction zones are captured in

the transition zone and in the uppermost lower mantle, notably when using $k = 3$ (blue regions). Different colours represent different regions identified (blue, red, brown and orange).

Figure 4. Cross-sections of perturbations in anisotropic structure and corresponding geodynamic models. Comparison between cross-sections of radially anisotropic structure from SGLOBE-rani[28] beneath Honshu (**a, c**) and W. Java (**b, d**), and results from geodynamical simulations for a slab stagnant at the 660-km discontinuity (left column; **e, g**) and a slab penetrating into the lower mantle (right column; **f, h**). We calculated radial anisotropy resulting from: (i) LPO due to dislocation creep (**e, f**); and (ii) grain-scale SPO (**g, h**) assuming a perfectly layered medium for a pyrolitic medium (see Supplementary Information). The slip system considered for bridgmanite in this example of LPO calculations is [001](100), which is reported in ref. 10. The green contours in (**a, b**) correspond to an outline of Voigt average fast anomalies in the range of 1.25%-1.5%. For each fabric in (e) and (f) we show in the bottom right corner dVs (km/s) calculated at lower mantle P-T conditions and for a 80:20=Brd:MgO mixture deformed in horizontal simple shear (ε = 1.0); red is minimum; blue is maximum. Cubic MgO crystals are random. The bars in the dVs maps indicate the polarisation of the fast shear wave component for different propagation directions. The cartoons in the bottom right corner of (g) and (h) highlight that SPO is computed using 1-D Backus averaging for perfectly horizontally layered media.