Forward modeling of the development of seismic anisotropy in the upper mantle

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Abstract

Development of seismic anisotropy in response to upper mantle flow is approached through an integrated numerical model. This model allows to predict the splitting parameters for a shear wave propagating across an upper mantle which deformed in response to a given geodynamic process. It consists of (1) thermo-mechanical modeling of the finite strain field, (2) modeling olivine lattice-preferred orientation (LPO) generated by this strain field, (3) calculation of the 3-D elastic properties associated with this LPO, and (4) estimation of the shear-wave splitting parameters: the time lag between the fast and slow split shear wave arrivals (δt) and the polarization azimuth of the fast wave (φ). Modeled olivine LPO are constrained relative to LPO measured in naturally and experimentally deformed peridotites. Comparison of predicted shear-wave splitting parameters with seismological data allows us to quantify the possible contribution of the modeled upper mantle flow to the measured splitting values and, hence, to constrain the interpretation of shear-wave splitting data in terms of upper mantle flow. We use this forward model to investigate the seismic anisotropy generated in ocean basins by a velocity gradient between the plate and the deep mantle. Fast-shear wave polarizations calculated assuming a constant plate motion are in good agreement with both the SKS polarization and the fast propagation direction for P and Rayleigh waves observed in the Pacific and Indian oceans, suggesting that, away from mid-ocean ridges, seismic anisotropy in oceanic basins primarily results from asthenospheric deformation by resistive drag beneath the plate. Delay times are, however, overestimated. This may be ascribed to a stronger strain localization in nature or to partial erosion of the anisotropic layer by hotspots. Indeed, hotspot activity may explain the short length scale variations of δt in the southern Pacific. Finally, two-layer models that simulate a change in Pacific plate motion as suggested by the bend in the Hawaii–Emperor chain fail to reproduce the observed shear-wave splitting. This is consistent with previous suggestions that the Emperor chain track may not faithfully record the Pacific plate absolute motion before 43 Ma.

Keywords: upper mantle; deformation; petrofabrics; numerical models; plate tectonics; asthenosphere; Pacific Plate; hot spots

1. Introduction

Seismic anisotropy measurements, especially SKS splitting, have emerged as an unique tool to probe the upper mantle deformation fabric, since this anisotropy results primarily from deformation-induced lattice-preferred orientation of upper mantle minerals, in particular olivine [1–3]. Yet, the interpretation of the upper mantle deformation fabric from teleseismic shear-wave splitting data remains a
poorly constrained problem. The delay time between arrivals of the two quasi-orthogonally polarized shear waves (\( \delta t \)) and the direction of polarization of the fast shear wave (\( \phi \)) are usually interpreted assuming that the anisotropy is localized within a single transverse isotropic layer with a horizontal axis of symmetry [4]. Under this assumption, \( \delta t \) is proportional to both the thickness and the intrinsic anisotropy of the layer and \( \phi \) is parallel to the preferred orientation of the [100] axis of olivine within this layer [2]. For large strains, the [100] axis of olivine tends to parallel the flow direction. However, the deformation geometry and intensity cannot be uniquely determined, due to the trade-off between intrinsic anisotropy and thickness of the deformed layer. The location of the anisotropic layer at depth also cannot be resolved; the measured anisotropy may result from a frozen lithospheric fabric, from the current asthenospheric flow, or from a superposition of both. Finally, a more complex mantle structure (multiple anisotropic layers or a dipping structure) can only be determined if measurements with a large range of initial polarization directions and arrival angles are available. Unfortunately the irregular distribution of earthquakes and the limited time span of observation available for many stations (in particular for portable experiments) usually hinder the acquisition of data with a good azimuthal coverage.

The intrinsic upper mantle anisotropy and, thus, the interpretation of shear-wave splitting data in terms of upper mantle fabric and deformed layer thickness are usually constrained using average or ‘typical’ upper mantle seismic properties calculated using LPO measured in naturally deformed peridotites [2,5,6]. However, this method allows only a crude estimation of the upper mantle flow geometry and deformation intensity. In this paper, we develop a forward model to simulate the splitting of a shear wave propagating across an upper mantle deformed in response to different geodynamic processes. The innovation of the present approach and of the one developed by Blackman and coworkers [7] is that, through an association of thermo-mechanical finite-element models and simulations of the development of lattice-preferred orientations, we can predict the olivine LPO and elastic anisotropy that will develop in response to a given upper mantle flow (e.g., the asthenospheric deformation due to resistive drag beneath a plate). The relationships between upper mantle deformation, LPO development and seismic anisotropy are explicitly taken into account. This allows a quantitative estimation of the seismic anisotropy generated by a given upper mantle deformation that may be used to constrain the interpretation of shear-wave splitting measurements.

2. Integrated modeling technique

The model consists of four coupled numerical simulations: (1) thermo-mechanical modeling of the finite strain field, (2) modeling of the crystallographic fabrics generated by this strain field, (3) calculation of the resulting 3-D elastic properties, and (4) estimation of the shear-wave splitting parameters \( \phi \) and \( \delta t \).

2.1. Upper mantle flow and finite strain

The finite strain field generated by a given geodynamic process is calculated using CREEP, a 2D finite-element program in which a Lagrangian formulation of the equations of continuum mechanics is solved for either plane stress or plane strain formulations [8]. Within this program, deformation of the lithosphere (or asthenosphere) is modeled as a Stokes flow of a non-linear, incompressible, viscoplastic material. Thermal transfers follow a 2D, time-dependent heat conduction equation.

The geodynamical process is defined by a set of kinematic and thermal boundary conditions and by an initial temperature field. The upper mantle is simulated by a homogeneous material able to deform by dislocation creep following a dry dunite constitutive relation [9]. It displays a temperature- and pressure-dependent viscosity. The resulting upper mantle flow is described by the evolution of the velocity gradient and finite strain fields through time.

2.2. Development of olivine lattice-preferred orientations

Development of olivine lattice-preferred orientations (LPO) during deformation may be simulated using polycrystal plasticity models. For olivine, purely kinematic [10–12], relaxed Taylor [13,14],
Sachs [7,15], constrained-hybrid [16], and isotropic viscoplastic self-consistent approaches [14,17] have been used. With the exception of the kinematic ones, all these models are based on the assumption that the polycrystal behavior may be calculated by an appropriate average of the grains’ response, the lower and upper bounds being represented by the Sachs [18] and Taylor [19] models, that impose respectively homogeneous stress or strain within the aggregate. In the following simulations, we use the viscoplastic self-consistent (VPSC) model developed by Molinari and coworkers [20] and extended to anisotropic materials by Lebensohn and Tomé [21]. Within this model, both the microscopic stress and strain rate may differ from the corresponding macroscopic quantities. Strain compatibility and stress equilibrium are ensured at the aggregate scale. A complete description of the model may be found in Lebensohn and Tomé [21].

At the grain scale, deformation is accommodated by dislocation glide; other deformation mechanisms, like dynamic recrystallization, are not taken into account. The shear rate in a slip system $s$ is related to the local deviatoric stress tensor $\sigma$ by a viscoplastic law

$$\dot{\gamma}_s = \dot{\gamma}_0 \left( \frac{\tau_s}{\tau_0} \right)^n = \dot{\gamma}_0 \left( \frac{r_s S_{ij} \dot{\varepsilon}_{ij}}{r_0} \right)^n$$

(1)

where $\dot{\gamma}_0$ is a reference strain rate, taken as $1 \text{s}^{-1}$, and $n$, $r_s$, and $r_0$ are, respectively, the stress exponent, the resolved shear stress, and the critical resolved shear stress for the system $s$, whose orientation relative to the macroscopic axes is expressed by its Schmid tensor $r^s$.

Thus, if the microscopic constitutive relation, the crystallographic orientation, and the imposed macroscopic velocity gradient $L$ are known, the problem lies in the calculation of a microscopic state $(s, \dot{\varepsilon})$ for each grain, whose volume average (denoted by angle brackets) determines the response of the polycrystal $(\overline{\mathbf{S}}, \overline{\mathbf{E}})$:

$$\langle s \rangle = \overline{\mathbf{S}} \quad \text{and} \quad \langle \dot{\varepsilon} \rangle = \overline{\mathbf{E}}$$

(2)

The 1-site approximation of Molinari et al. [20] is used in the anisotropic VPSC formulation. Influence of neighboring grains is therefore not taken into account. Interactions between each in-situ grain and its surroundings are successively replaced by the interaction between an ellipsoidal inclusion and an infinite homogeneous equivalent medium (HEM), whose behavior is the weighted average of the behavior of all constituent grains. This treatment leads to an interaction equation that relates local stresses and strain rates to the macroscopic quantities:

$$\dot{\varepsilon}_{ij} - \dot{\varepsilon}_{ij} = -\dot{M}_{ijkl}(\delta_{ij} - \overline{\Sigma}_{ij})$$

(3)

where $\dot{M}$ is the interaction tensor that depends on the rheological properties of the aggregate and on grain shape. Once numerical convergence is achieved, grain shapes are updated and the reorientation of each grain is calculated. The lattice rotation for each grain (relative to the macroscopic axes) is given by:

$$\dot{\Omega}_{ij} = \dot{\Omega}_{ij} + \Pi_{ijkl}(S_{lm}^{-1} \dot{\varepsilon}_{mn} - \mathbf{Q}_{mn})$$

$$- \sum_I \frac{1}{2}(b_{ij} n_j - b_{jn} j_i) \dot{\varepsilon}^I$$

(4)

where $\dot{\Omega}$ is the antisymmetric component of the imposed macroscopic velocity gradient $L$, the second term is the reorientation of the associated ellipsoidal inclusion that depends on the difference in strain rate between the grain and the polycrystall and increases with ellipsoid distortion [22], and the last term is the antisymmetric component of the plastic distortion rate.

The VPSC model calculates the evolution of the aggregate yield strength, the activity of slip systems, and LPO development for a given macroscopic strain history. Potentially active slip systems for olivine, their critical resolved shear stresses (CRSS), and stress exponent $n$ (Table 1) are evaluated from TEM observations on naturally and experimentally deformed peridotites and single-crystal deformation experiments [23–25]. A detailed study of the validity of this approach to simulate olivine LPO evolution during upper mantle deformation is presented elsewhere [26]. We just summarize the main results of this work.

‘1-site’ anisotropic VPSC models yield, in a first approximation, good predictions of LPO development in naturally deformed peridotites and in olivine polycrystals submitted to axial compression and single shear under experimental conditions (Fig. 1). Both measured and modeled LPO are characterized by a [100] maximum close to the flow direction and a [010] maximum roughly normal to the flow plane,
Table 1
Slip systems data for VPSC simulations of LPO development in dunites

<table>
<thead>
<tr>
<th>Slip system</th>
<th>CRSS</th>
<th>( n^i )</th>
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<tbody>
<tr>
<td>(010)[100]</td>
<td>1</td>
<td>3.5</td>
</tr>
<tr>
<td>(001)[100]</td>
<td>1</td>
<td>3.5</td>
</tr>
<tr>
<td>(010)[001]</td>
<td>2</td>
<td>3.5</td>
</tr>
<tr>
<td>(100)[001]</td>
<td>3</td>
<td>3.5</td>
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<tr>
<td>(011)[100]</td>
<td>4</td>
<td>3.5</td>
</tr>
<tr>
<td>(031)[100]</td>
<td>4</td>
<td>3.5</td>
</tr>
<tr>
<td>(110)[001]</td>
<td>6</td>
<td>3.5</td>
</tr>
<tr>
<td>'complementary system' ^a (111)[110]</td>
<td>1000</td>
<td>3.5</td>
</tr>
</tbody>
</table>

^a Since only dislocations with \( b = [100] \) or [001] have been observed, olivine does not have five independent slip systems. Five independent slip systems are however needed for the VPSC simulation, since it converges locally over five variables. Thus a pyramidal system is added. This 'complementary system' accommodates less than 1% deformation in all simulations.

suggesting predominance of the (010)[100] system. Slight differences between modeled and measured LPO patterns are probably due to dynamic recrystallization, not taken into account in the simulations. During simple shear, in particular, dynamic recrystallization seems to favor the development of a single maximum LPO characterized by a parallelism between the dominant slip system and the macroscopic shear at low shear strains (compare Fig. 1b and c). In the simulations, this parallelism is only achieved for shear strains higher than 10. Angular misorientations between [100] maxima in model aggregates and natural samples remain, however, weak (less than 15° for a shear strain of 1). In both models (Fig. 1d) and experiments [27,28], LPO develops at low strains (30% shortening, or shear strains of 1). However, whereas modeled LPOs evolve continuously with increasing strain, in nature, recrystallization through subgrain rotation and/or nucleation induces local misorientations that weaken the LPO [27,29] and favor the development of a steady-state LPO. This suggests that, in highly strained natural rocks, LPO intensity records only part of strain accommodated by the aggregate. Thus VPSC simulations predict correctly LPO patterns, but overestimate LPO intensity for high strains.

Coupling between the thermo-mechanical and VPSC simulations is achieved through the association of an olivine polycrystalline aggregate composed of 200 grains to each nodal point of the finite-element model. The deformation of each aggregate is described by the velocity gradient and finite strain tensors that characterize the flow at this point in the thermo-mechanical model. Since VPSC simulations are highly time-consuming, in the following models we only simulate the LPO development for a lim-
ited number of points of the finite element model (a vertical profile through the upper mantle).

2.3. Upper mantle elastic properties and seismic anisotropy

If single crystal elastic stiffness coefficients, the volume fraction of the different mineral phases composing an aggregate, and their crystallographic preferred orientation are known, the elastic properties of the aggregate may be calculated using a Voigt–Reuss–Hill average of the stiffnesses \( C_{ijkl} \) over all crystal orientations [30]. Seismic properties calculated for modeled olivine polycrystals agree well with those calculated using olivine LPO measured in naturally deformed peridotites (Fig. 2a and b). Angular misorientations decrease with increasing strain and are less than 15° for shear strains higher than 1. Moreover, overestimation of LPO intensity in VPSC simulations for highly strained aggregates does not affect seismic anisotropy estimates, because seismic anisotropy for both P- and S-waves does not display a linear dependence on LPO intensity or strain, but tends to limiting values for high shear strains (Fig. 3).

However, comparison between the seismic properties of the modeled, 100% olivine aggregate and those of a natural peridotite with a typical upper mantle composition (70% olivine and 30% pyroxene) shows that the anisotropy for both P- and S-waves of the model aggregate is overestimated relative to the natural sample (Fig. 2a and c). Simulations of LPO development in peridotitic aggregates [17] as well as LPO data from naturally deformed peridotites suggest that the presence of enstatite does not modify olivine LPO patterns. The main effect of enstatite is to weaken the aggregate seismic anisotropy. LPO simulations for enstatite-bearing aggregates are limited to low strains because of the high plastic anisotropy of enstatite crystals [17]. Thus, in the following models, we calculate the seismic properties for model aggregates composed of 70% olivine (simulated LPO) and 30% enstatite that displays a typical high-temperature deformation LPO measured in a naturally deformed harzburgite from the Oman ophiolite (Fig. 1a).

![Fig. 2. Calculated seismic properties: P-wave velocity (0.1 km/s contours; left), fast S-wave polarisation (center) and S-wave anisotropy \( A\% = \left( \frac{v_{S1} - v_{S2}}{v_{Smean}} \right) \times 100/v_{Smean} \times 1\% \) contours; right), for (a) a 100% olivine aggregate displaying the modeled LPO from Fig. 1d; (b) a 100% olivine aggregate displaying the same olivine LPO as the naturally deformed harzburgite in Fig. 1a, and (c) the naturally deformed harzburgite (70% olivine, 30% enstatite). Equal-area projections, lower hemisphere. Full line indicates the shear plane, dashed line the foliation (XY finite strain plane).](image)

![Fig. 3. Calculated P-wave anisotropy (%) for a model aggregate (100% olivine) submitted to simple shear as a function of strain.](image)
3. Case study: seismic anisotropy development in ocean basins by resistive drag of the sublithospheric mantle

Asthenospheric deformation by resistive drag beneath an oceanic plate was chosen as a case study to develop this integrated modeling approach because (1) the flow geometry is simple and two-dimensional (a horizontal shear zone), and (2) since the oceanic lithosphere develops by progressive accretion of asthenosphere as the plate cools, away from active margins or hotspots, a velocity gradient between the plate and the deep mantle should be the main cause for the development of a structured fabric in the uppermost mantle.

3.1. Thermo-mechanical modeling

The finite strain field generated by this process was calculated using the vertical plane-strain finite-element model presented by Tommasi et al. [5]. The oceanic plate and underlying upper mantle are simulated by a homogeneous material able to deform by dislocation creep following a dry dunite constitutive relation (Table 2). Kinematic boundary conditions simulate coherent motions of the plate and of the deep mantle (same direction), but the plate moves with a higher velocity. The upper boundary of the model is submitted to a constant velocity (1.5, 3, or 6 cm/yr), whereas the lower boundary (at 500 km depth) is kept fixed. The upper and lower boundaries have fixed temperatures of 273 K and 1773 K, respectively, and a null heat flow is prescribed at the lateral boundaries. The initial temperature field consists of a 2 Myr old oceanic plate geotherm calculated using a half-space cooling model and an adiabatic gradient in the sublithospheric mantle. Convective heat transfer in the sublithospheric mantle is simulated through an enhanced thermal conductivity (Table 2).

Strain is localized in a horizontal shear zone several tens of kilometers wide between an almost rigid mechanical lithosphere and a mildly deformed upper mantle. Shear strain accumulates progressively with plate displacement (Fig. 4). For young oceanic lithosphere, cooling induces a migration of the maximum shear strain rate towards deeper levels. This migration is accompanied by a progressive widening of the shear zone (Fig. 4a). Another effect of cooling is the freezing of the upper part of the deformed layer. The fabric responsible for the measured anisotropy is therefore contained in two distinct layers: a frozen lithospheric layer and an asthenospheric layer that records the current plate motion. The thickness of the deformed layers depends on the time elapsed (or displacement, Fig. 4b), but it also depends on the velocity of the plate, because plates with different velocities have different shear strain rate profiles: for a similar displacement, faster plates accumulate larger shear strains over thinner layers [5].

3.2. Development of olivine lattice-preferred orientations and seismic anisotropy

Olivine LPO were calculated at each nodal point of a vertical profile in the central domain of the finite element model for different plate ages and velocities. Analysis of the shear strain profile for a 100 Myr old slow plate \( v = 1.5 \) cm/yr and corresponding [100] axis preferred orientations (Fig. 5) shows that a clear LPO pattern is already present at shear strains lower than 1. Strong LPO develop at shear strains higher than 5, and [100] orients parallel to the flow direction. This results in fast shear-wave polarizations as well as fast propagation directions for both P (Fig. 2) and Rayleigh waves [31] parallel to the plate motion direction (APM).

Using the modeled olivine LPO and the measured enstatite LPO from Fig. 1a, we calculated the depth distribution of anisotropy for a vertically propagating shear-wave (e.g., SKS) beneath a 100 Myr old plate moving with different velocities. Shear wave
Fig. 4. Shear strain rate (a) and finite shear strains profiles (b) at time intervals of 10 Myr for a model with a low velocity gradient between the plate and the deep mantle (\(dv = 1.5 \text{ cm/yr}\)).

Fig. 5. Shear strain profile for a 100 Myr old slow plate (\(dv = 1.5 \text{ cm/yr}\)) and corresponding [100] axis preferred orientations (100% olivine aggregates, 200 grains).
anisotropy does not depend linearly on finite strain. It increases quickly for shear strains lower than 5 and then tends to a limiting value around 4.8% (Fig. 6).

Estimated delay times range from 2.1 s for small velocity gradients between the plate and the underlying mantle (1.5 cm/yr) to 2.7 s for a velocity gradient of 6 cm/yr).

3.3. Comparison to seismic anisotropy measurements in ocean basins

Our models suggest that old plates (~100 Myr) should be anisotropic down to ca. 250–300 km. This result is consistent with the depth distribution of anisotropy in the Pacific and Indian oceans inferred from surface-wave analysis [32–34]. Moreover, Montagner [35] observes that anisotropy in the Pacific seems to involve deeper layers when plate age increases, in agreement with the progressive thickening of the sheared layer observed in our models (Fig. 4). Nishimura and Forsyth [33], on the other hand, only detected anisotropy in the upper 50 km of the mantle in regions of the Pacific older than 80 Myr. However, these results were calculated using an a-priori model based on the fossil seafloor spreading direction, which, in the northwestern Pacific, differs from the present-day APM by ca. 30° (Fig. 7).

Few shear-wave splitting measurements are available in ocean basins. Moreover, most of them come from volcanic islands in the western Pacific (Fig. 7), a plate that presumably underwent a marked change in its motion (inferred from the bend in the Hawaii–Emperor seamounts chain) and whose upper mantle has been at least partially modified by hotspots. Fast polarization directions in old domains of the Pacific from SKS splitting [36,37] are exactly (TPT) or approximately (RAR, AFI, TBI, KIP) parallel to the APM direction [38]. These results are consistent with splitting of PS [39], ScS [40], and SS phases [36] in the Pacific plate. Measured delay times range from 0.9 to 1.45 s, with exception of Papeete, Tahiti, which appears to be isotropic (Fig. 7). They are much lower than our estimated delay times. This suggests a thinner anisotropic layer beneath the stations. Several processes may be invoked: (1) a stronger strain localization in nature due to either partial melting or development of a strain-induced mechanical anisotropy, (2) a partial or total destruction of the LPO, (3) a slower velocity gradient between the plate and the mesosphere, or (4) a transition from dislocation creep to grain-sensitive deformation mechanisms in the deep upper mantle.

Hotspot activity, small-scale convection, or a change in APM with time may all induce destruction or modification of the resistive-drag upper mantle fabric. Perturbation of the anisotropy pattern is evidenced near La Réunion hotspot in the Indian Ocean using surface waves [34]. Hotspot activity may also explain small wavelength variations in splitting parameters in the southern Pacific. Buoyant vertical flow should occur beneath an active hotspot; very little or no splitting would be detected for an SKS in such a case (Fig. 2). Isotropy in Papeete is consistent with significant lithospheric thinning leading to total destruction of the resistive-drag LPO in the region immediately beneath the plume conduit [41], while the fast shear-wave polarization roughly parallel to the APM observed in older islands could be related to accumulation and drag by the plate motion of hot plume material in the asthenosphere [42].

Seismic anisotropy in old domains of ocean basins may be primarily related to differential motion between the plate and the deep mantle. Near oceanic ridges, however, a different pattern can be observed: SKS splitting measurements [36,43,44] yield δτ ranging between 0.7 and 2 s. Numerical simulations indicate that even high velocity gradients between the plate and the deep mantle plate cannot generate delay times higher than 0.5 s for plate displacements lower than 500 km. Moreover, fast shear-wave polarizations at Iceland and Easter Island do not parallel either APM or the spreading di-
Fig. 7. SKS, PS, and SS splitting data in the Pacific Ocean [36,37,39]. SKS measurements are best receiver splitting parameters at each station for a suite of events calculated using the stacking method of Wolfe and Silver [36]. Arrows indicate the APM relative to hotspots [38]. \( TPT \) = Tiputa, Tuamotu Islands; \( RAR \) = Rarotonga, Cook Islands; \( AFI \) = Afiamalu, Samoa; \( TBI \) = Tubuai, Austral Island; \( KIP \) = Oahu, Hawaii; \( PPT \) = Papeete, Tahiti.

rection. Measured splitting parameters near oceanic ridges should therefore record the active, probably convection-related, flow of the upper mantle.

3.4. Effect of a change in APM

The bend in the Hawaii–Emperor hotspot trail in the northwestern Pacific (Fig. 7) is usually interpreted as the record of a marked change in the Pacific APM at ca. 43 Ma. A similar shift in APM is suggested by the Louisville chain in the south Pacific [45]. A change in plate motion would result in a different LPO in the asthenospheric material added subsequently to the base of the lithosphere. Thus, old oceanic domains, as those sampled by most shear-wave splitting measurements, might display a suite of anisotropic layers. In particular, petrofabrics in the frozen lithosphere and those in the currently deforming asthenosphere should display different orientations. Gradients between these layers will depend on how sudden or gradual plate motions changed and on the strain intensity needed to reorient the old LPO.

In the Pacific, surface wave analysis highlights variations in azimuthal anisotropy with depth; maximum Rayleigh wave velocity directions are more closely related to the present absolute plate motion at 200 km depth than at shallower levels [46]. Similar observations were performed in the Indian Ocean; a good correlation between surface-waves azimuthal anisotropy and the APM is observed between 100 and 200 km, whereas in the uppermost 100 km fast velocity directions display a complex pattern and do not correlate simply with the present-day [34]. Since maximum Rayleigh wave velocity directions are parallel to the preferred orientations of the [100] axis of olivine [31], these observations denote changes in flow direction with time.

The effect of a change in the APM direction on the upper mantle fabrics was investigated using the forward model scheme. Fig. 8 displays olivine LPO
Fig. 8. Right: lattice-preferred orientations developed in 200 grains, 100% olivine aggregates that sample the upper mantle beneath a 100 Myr plate that instantaneously changed its motion from N10°W to N70°W at 40 Ma. Left: solid lines show the C11 (X1 is the flow direction for the present-day motion, APM2), C22 (X2 is the vertical direction), and C33 components of the elastic stiffness tensor calculated for the same model. For comparison, dashed lines show elastic stiffnesses for a model with a constant APM (parallel to APM2).

developed at different depths for a model in which the plate moved towards N10°W with a velocity of 6 cm/yr for 60 Myr. and then changed instantaneously to a N70°W direction during the last 40 Myr, i.e., for an APM evolution similar to the one suggested for the northern Pacific by the Emperor–Hawaii hotspot track. The uppermost mantle displays frozen lattice-preferred orientations related to the old APM. The deeper levels display LPO related to the current plate motion. Between these two layers there is a thick transitional layer in which a slow rotation of the LPO between the two orientations occurs. Reorientation of LPO in this model is slower than LPO development from a initially random aggregate, because such a large APM change places the [001] axes of olivine near to the new shear direction, inducing simultaneous activation of (010)[001] and (010)[100] slip systems. Thus, even a instantaneous APM change may result in a gradational reorientation of olivine LPO over a thick layer.

We approach this continuous variation by a model with two discrete layers. In the lower layer, fast shear-wave polarization is parallel to the Hawaii trend (φ = −70°) whereas, in the upper layer, it is parallel to the Emperor trend (φ = −10°). Using this model, we estimate the apparent splitting parameters that would be measured assuming a single anisotropic layer [47]. Apparent fast shear-wave polarization and delay time were calculated for two different depths of the interface (Fig. 9): 105 km (δt1 = 1.72 s; δs1 = 0.86 s) and 153 km (δt1 = 1.25 s; δs1 = 1.33 s). Both models fail to reproduce the E–W to NW–SE fast shear-wave polarizations measured in old domains of the Pacific (Fig. 7).

Rayleigh waves and Pn azimuthal anisotropy in the southern Pacific [46,48,49] indicate a coherent upper mantle fabric at all depth ranges from the sub-Moho mantle to the asthenosphere. Similarly, in the vicinity of Hawaii, fast propagation directions for both P-waves [50] and Rayleigh waves at 91 s [51] point to an E–W alignment of the olivine [100] axis. Even though hotspots were active in both areas, a complete destruction or reorientation of the lithospheric LPO beneath all stations seems improbable. Moreover, the fast polarization at the SS bounce point north of Hawaii is also E–W (Ref). Lack of evidence of plate motion parallel to the Emperor chain track may be explained if this seamount chain was not produced by a stationary hotspot and therefore does not record the absolute motion of the Pacific plate before 43 Ma. This hypothesis is supported by the success of our one-layer models in simulating shear-wave splitting data in the Pacific, that, together
Fig. 9. Apparent fast shear-wave polarization (a) and delay time (b) dependencies on the initial polarization of the shear wave calculated for a two-layer model in which the upper layer displays $\phi = N10^\circ W$ (Emperor seamounts chain trend) and the lower layer has $\phi = N70^\circ W$ (Hawaiian Islands trend). The full line shows apparent splitting parameters for a model with the interface between the two anisotropic layers at 105 km depth ($\delta t_1 = 1.72$ s; $\delta t_0 = 0.86$ s) and the dashed line for a model with a 153 km deep interface ($\delta t_1 = 1.25$ s; $\delta t_0 = 1.33$ s). Frequency 0.125 Hz.

with the above-presented seismic data, suggest that old Pacific APM directions were similar to the present-day APM.

Indeed, although the great bend in the Hawaiian–Emperor chain at around 43 Ma is commonly assumed to represent a sudden change in the Pacific plate motion, the bend itself is apparently the only clear evidence for such a change [52]. Lack of significant plate reorganization and tectonic events at 43 Ma in the Pacific and surrounding plates has already led Norton [53] to propose a non–stationary hotspot origin for the Emperor chain. Moreover, calculated positions of the Iceland, Tristan da Cunha, Réunion, and Kerguelen hotspots using the history of the Pacific plate motion over the Hawaiian hotspot differ from the actual hotspots trails by several hundreds of kilometers for times older than 40 Ma [54]. Finally, geodynamic simulations of plate motions in terms of mantle buoyancy forces arising from subducted lithosphere and lithosphere thickening or from velocity anomalies mapped by seismic tomography also cannot account for such dramatic changes in plate motions [52]; most candidate mechanisms, as subduction initiation, are unlikely to generate new buoyancy forces in less than 5 Myr.

4. Conclusions

The development of upper mantle seismic anisotropy is approached through an integrated numerical model that allows the prediction of the splitting of shear waves propagating through an upper mantle region that deformed in response to a given geodynamic process. The model steps are: (1) thermo-mechanical modeling of the finite strain field, (2) modeling of the crystallographic fabrics generated by this strain field, (3) calculation of the resulting 3D elastic properties, and (4) estimation of the shear-wave splitting parameters $\delta t$ and $\phi$.

Comparison of modeled olivine lattice-preferred orientations (LPO) to olivine LPO measured in naturally and experimentally deformed peridotites shows that, to a first approximation, there is a good agreement between measured and modeled LPOs. Agreement between seismic properties calculated for model olivine polycrystals and natural upper mantle rocks is still better, because seismic anisotropy does not depend linearly upon LPO intensity. Seismic anisotropy predictions are therefore not affected by overestimation of LPO intensities in the models.

This forward model puts strong constraints on the interpretation of seismic anisotropy data; the possible contribution of the modeled upper mantle flow process to the measured splitting values may be isolated and quantified. Application of this forward model to the investigation of the asthenospheric deformation by resistive drag beneath an oceanic plate suggests that a velocity gradient between the plate and the deep mantle may generate the seismic anisotropy measured in oceanic basins. However, other processes, such as hotspot activity, must be invoked to explain the lateral variation in splitting parameters over relatively short distances in the southern Pacific. On the other hand, analysis of SKS splitting measurements near oceanic ridges in the light
of model predictions suggests that the measured splitting should record active, probably convection-related, flow in the sublithospheric mantle. Finally, two-layer models that simulate an abrupt change in Pacific plate motion as suggested by the bend in the Hawaii–Emperor seamounts chain fail to reproduce shear-wave splitting observed in Pacific stations. Fast shear-wave polarizations in Hawaii and the southern Pacific are better reproduced by a one-layer model. A coherent orientation of the [100] axis of olivine for the whole upper mantle in both areas is also suggested by surface- and P-waves anisotropy determinations. Both model predictions and observations are consistent with previous suggestions [53] that the Emperor chain track may not faithfully record the Pacific plate absolute motion before 43 Ma.

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