The origin of shear wave splitting beneath Iceland

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SUMMARY

The origin of shear wave splitting (SWS) in the mantle beneath Iceland is examined using numerical models that simulate 3-D mantle flow and the development of seismic anisotropy due to lattice-preferred orientation (LPO). Using the simulated anisotropy structure, we compute synthetic SKS waveforms, invert them for fast polarization directions and split times, and then compare the predictions with the results from three observational studies of Iceland. Models that simulate a mantle plume interacting with the Mid-Atlantic Ridge in which the shallow-most mantle has a high viscosity due to the extraction of water with partial melting, or in which C-type olivine LPO fabric is present due to high water content in the plume, produce the largest chi-squared misfits to the SWS observations and are thus rejected. Models of a low-viscosity mantle plume with A-type olivine fabric everywhere, or with the added effects of E-type fabric in the plume below the solidus produce lower misfits. The lowest misfits are produced by models that include a rapid (~50 km Myr^{-1}) northward regional flow (NRF) in the mid-upper mantle, either with or without a plume. NRF was previously indicated by a receiver function study and a regional tomography study, and is shown here to be a major cause of the azimuthal anisotropy beneath Iceland. The smallest misfits for the models with both a plume and NRF are produced when LPO forms above depths of 300–400 km, which, by implication, also mark the depths above which dislocation creep dominates over diffusion creep. This depth of transition between dislocation and diffusion creep is greater than expected beneath normal oceanic seafloor, and is attributed to the unusually rapid strain rates associated with an Iceland plume and the NRF.

Key words: Numerical solutions; Microstructure; Plasticity, diffusion, and creep; Seismic anisotropy; Dynamics: convection currents, and mantle plumes; Atlantic Ocean.

1 INTRODUCTION

Seismic anisotropy in the mantle can result from lattice-preferred orientation (LPO) of olivine, and is widely interpreted as indicating the sense of deformation and flow within the mantle. Three studies have focused on characterizing azimuthal seismic anisotropy beneath the Iceland hotspot (Bjarnason et al. 2002; Li & Detrick 2003; Xue & Allen 2005), an area of particular interest owing to its association with a mantle plume rising beneath the Mid-Atlantic Ridge (MAR; e.g. Hart et al. 1973; White & McKenzie 1995; Wolfe et al. 1997; Allen et al. 2002). Shear wave splitting (SWS) results of Bjarnason et al. (2002) and Li & Detrick (2003) showed an overall pattern of shear wave fast polarization directions oriented NNW in eastern and southern Iceland, and NNE in northern and western Iceland (Fig. 1). These authors interpreted the results as reflecting LPO due to the mantle shear below the two spreading plates caused by a northward, regional upper mantle flow. In contrast, the SWS results of Xue & Allen (2005) revealed that the NNW fast orientations previously seen in east Iceland on the Eurasian plate are also prominent just west of the MAR on the North American Plate. This finding contradicts Bjarnason et al.’s (2002) model of northward regional flow (NRF), which predicts fast orientations that are distinct between, and uniform on the two plates. Xue & Allen (2005) attributed their results as revealing channelled flow of plume material along the curved plate boundary that bisects Iceland.

The story is further complicated by Rayleigh wave azimuthal anisotropy detected by Li & Detrick (2003). Their tomography study revealed that the fast Rayleigh wave directions differ significantly between east, west, and central Iceland, and changes rapidly with wave period, suggesting the anisotropy changes rapidly with depth. The shorter period (20–40 s) waves, which are most sensitive to the shallowest 50 km, showed a NNE fast direction in central Iceland, roughly parallel to the ridge axis. Different explanations were considered, but the favoured explanation was that of LPO due to channelled flow of plume material along the ridge axis in the shallow-most mantle. This interpretation is broadly consistent with that of Xue & Allen (2005), but details such as the direction of the inferred plume flow in northern Iceland differ between the two studies. Li & Detrick (2003) also found little consistency between the Rayleigh wave anisotropy and their SWS results; they attributed
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Figure 1. Topography (Smith & Sandwell 1997) and results of the three main shear wave splitting studies carried out on Iceland (Bjarnason et al. 2002; Li & Detrick 2003; Xue & Allen 2005). Coloured bars indicate the fast polarization directions (\(\phi\)), which have lengths proportional to the corresponding split time (\(t\)), and are colour coded according to their respective seismic study. Dotted lines mark the offshore location of the Mid-Atlantic Ridge (MAR); onshore dotted lines outline the young zone of faulting on crust of ages \(<0.8\) Myr (http://en.ni.is/geology/geological-maps/). The circle marks the approximate centre of the Iceland mantle plume (e.g. Wolfe et al. 1997). The Western Volcanic Zone (WVZ), Eastern Volcanic Zone (EVZ) and Northern Volcanic Zone (NVZ) are labelled.

This result to the SWS being less influenced by the shallow LPO due to plume flow than the deeper LPO caused by the NRF. Overall, the above studies of azimuthal anisotropy below Iceland produced differing and contradicting conclusions that have yet to be resolved.

The interpretations of the seismic results in the above studies were conceptual in nature, and relied on the simple assumption that the seismically fast directions parallel the preferred orientations of the olivine a-axes as well as the direction of mantle shear. This assumption is valid in the presence of simple shear after large (\(\sim 150\) per cent) accumulated strain (Zhang & Karato 1995) and for the olivine A- and E-type fabrics, which occur for low and moderate water contents, respectively (Karato 2007). Whereas under more general conditions (Ribe 1989; Wenk et al. 1991; Zhang & Karato 1995), especially when deformation changes rapidly in space, such as in a mantle plume (Kaminski & Ribe 2002), the relationship between mantle flow and LPO is expected to be variable. In addition, the elevated water contents associated with mantle plumes can alter olivine fabric type, and hence the relationship between LPO fabric and mantle deformation (Jung & Karato 2001; Karato 2007).

Recent numerical modelling studies of mantle plumes (Marquart et al. 2007; Fu et al. 2012; Ito et al. 2014) have indeed shown that the relationships between SWS fast polarization directions, LPO, and mantle flow are complex. Addressing the SWS at Iceland, Fu et al. (2012) and Ito et al. (2014) found that models of a low-viscosity mantle plume produced fast polarization directions that resembled the observed directions in eastern and far western Iceland; however, the models failed to explain the observed directions in southern and central Iceland. While these studies simulated the full 3-D flow field of a mantle plume, the development of LPO, as well as SWS, they fell short in various aspects specific to Iceland. For example, the models simulated an idealized straight ridge axis unlike the geometry of the MAR, they considered only the dry (A-type) olivine fabric, and did not address the possibility of regional mantle flow.

The main objective of this study is to determine the origin of the SWS observed on Iceland and reconcile the differences in interpretations that exist to date. First, we combine the results of the three SWS splitting studies of Iceland to estimate true fast polarization directions with standard errors. Then, following Ito et al. (2014), we use geodynamic models to simulate mantle flow, the development of LPO, and SWS parameters for comparison to the observational splitting parameters. The models include a realistic representation of relative plate motion and plate boundary geometry in various situations: a low-viscosity plume being channelled along the ridge, a plume with high-viscosities due to the extraction of water with partial melting, the absence and presence of NRF, varying depths above which LPO forms, and different olivine fabric types. Chi-squared statistics and F tests are used to identify the best fitting models and to guide the conclusions.

2 ANALYSIS OF SWS OBSERVATIONS

2.1 Estimates of the mean SWS parameters

The three SWS studies (Bjarnason et al. 2002; Li & Detrick 2003; Xue & Allen 2005) used SKS and other core reflected and refracted
2.2 Standard errors of the true expected SWS parameters

It is equally important to estimate the standard errors of the expected parameters $\phi$ and $\bar{t}$. The standard error for $\phi$ was computed according to

$$\sigma_\phi = \frac{180}{\pi} \sqrt{\frac{1}{\kappa r^2} + \frac{c_\phi^2}{m}}^{1/2}. \tag{1}$$

The first term under the radical represents the variance between the $m$ measurements near each station. It assumes noise about $\phi$ follows a Von Mises distribution (e.g. Davis 2002), and is therefore a function of the length $r$ of the vector sum of the $m$ unit vectors for measurements at the $m$ seismic stations, as well as the estimated concentration parameter $\kappa$ of the Von Mises distribution (which depends on $r$ and $m$, see e.g. Circstat: A MATLAB Toolbox for Circular Statistics). The second term in the radical is the mean square measurement error as reported by the three SWS studies. The standard error for $\bar{t}$ was computed according to

$$\sigma_\bar{t} = \left(\frac{(m-1)}{m^2} \sigma_t^2 + \frac{c_t^2}{m}\right)^{1/2}. \tag{2}$$

The first term under the radical is a function of the variance of $\bar{t}$ among the $m$ measurements near each station, and the second term is the mean square measurement error as reported by the three seismic studies (Bjarnason et al. 2002; Li & Detrick 2003; Xue & Allen 2005). The mean splitting parameters and associated standard errors (Fig. 2) at the station locations are used for evaluating the fits of the geodynamic model predictions.
has been propagating southward (Einarsson 1991; LaFemina et al. 2005) since 2–3 Ma (Ivarsson 1992). For simplicity, we approximate seafloor spreading as occurring entirely along the longer-lived and stably configured WVZ (Fig. 3) in southern Iceland. For northern Iceland, the models simulate rifting along the Northern Volcanic Zone (NVZ). The relative direction and half rate (9.6 km Myr$^{-1}$, in the model x-direction) of seafloor spreading is based on the Nuvel 1A-HS3 absolute plate motion model (Gripp & Gordon 1990, www.unavco.org). The other five boundaries of the model are open to flow through them. In some cases, a northward regional mantle flow is imposed as a uniform horizontal velocity on the base of the model.

Potential temperature was set at 0 °C at the surface and 1350 °C at the base, except in a circular patch at the base where a plume was imposed in all but one model. The excess plume temperature peaks at the centre of the patch at $\Delta T = 200$ °C, and decays as a Gaussian function of radial distance to a value of $\Delta T / e$ at a distance of 80 km, which defines the plume radius. The buoyancy flux of the plume as measured in the middle of the box is $\sim 3$ Mg s$^{-1}$. The dissipation number of 0.2638 controls the adiabatic gradient with the extended Boussinesq approximation, which includes the loss of latent heat with melting (in Models 2, 4c.i and 4c.ii) but not viscous heating. We considered four different mantle flow models (Models 1–4) described below and summarized in Table 1.

The imposed Rayleigh number determines the effective mantle viscosity at the base of the model, $\eta_0$. Viscosity varies according to depth, $z$, absolute temperature $T$ and mantle water content $C_w$ relative to the starting content $C_{w0}$ according to

$$\eta = \eta_0 \left(\frac{C_w}{C_{w0}}\right)^{-1} \exp \left(\frac{E + \rho_0 g z V}{RT} - \frac{E + \rho_0 g D V}{RT_0}\right). \quad (3)$$

Here $T_0$, $\rho_0$, $g$, $D$ and $R$ are the reference ambient mantle temperature at the base of the model, reference mantle density (3300 kg m$^{-3}$), acceleration of gravity (9.8 m s$^{-2}$), total model depth (660 km) and ideal gas constant, respectively. Activation energy and volume are $E = 350$ kJ mol$^{-1}$ and $V = 6 \times 10^{-6}$ m$^3$ mol$^{-1}$ (Hirth & Kohlstaedt 2003). The dependence on water content $(C_w/C_{w0})^{-1}$ is present only in Model 2, in which case dehydration with partial melting (Hirth & Kohlstaedt 1996, see Bianco et al. 2011 for details of implementation) causes $(C_w/C_{w0})^{-1}$ to increase from unity at the very onset of (hydrous) melting (which is $\sim 150$ km below the surface, above the hottest centre of the plume stem) to its maximum allowable value of 10$^2$ over a height of about 25 km (water is treated as an incompatible element extracted by fractional melting with bulk partition coefficient of 0.01). In the other three models (Models 1, 3 and 4) $(C_w/C_{w0})^{-1}$ is held fixed at unity, making the viscosity only depend on $T$ and $P$.

3.2 Calculations of LPO and mantle anisotropy

Following Ito et al. (2014), we assume that seismic anisotropy occurs only due to LPO of olivine and enstatite in the mantle. To simulate the development of LPO, we used the numerical code D-Rex (Kaminski & Ribe 2001; Kaminski et al. 2004), which uses the macroscopic mantle flow field to kinematically impose the microscopic deformation of individual grains within an aggregate of olivine (70 per cent) and enstatite (30 per cent). Deformation of the aggregate is accommodated by, and LPO develops in response to, intracrystalline slip, dynamic recrystallization and grain boundary sliding. The different, water-content-dependent, olivine fabric types ($A$-, $E$-, $C$-type) describe the LPO (and anisotropy) that develops for a given sense accumulated strain. The different fabric types are

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Figure 3. 3-D perspective views of plume models (a) Model 1, low-viscosity plume beneath sloping lithosphere, (b) Model 2, high viscosity shallow-most mantle due to dehydration, which acts like a relatively flat, compositional lithosphere and (c) Model 4, low-viscosity plume with northward regional mantle flow (long arrows). Potential temperature at the base and two vertical edges of the model box are coloured (redish orange = ambient potential temperature of 1350 °C, blue = 1300 °C, orange = 1400 °C). The 1400 °C isosurface shows the deep plume stem and a layer of ponded plume material spreading beneath the lithosphere and the model plate boundary (white lines). Double arrows show the direction of seafloor spreading imposed on surface boundary. Iceland is outlined in black.
Table 1. Summary of key aspects of models and \( \chi^2 \) statistics for fit to observations.

<table>
<thead>
<tr>
<th>Model</th>
<th>Key characteristics</th>
<th>Model parameters, ( p )</th>
<th>Degrees of freedom, ( v )</th>
<th>( \chi^2_{\phi}, \chi^2_{t}, \bar{\chi}^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model 1 (Fig. 4)</td>
<td>Low-viscosity plume ((C_{\alpha}/C_{\omega0} = 1)), no regional flow, A-type olivine LPO fabric, ( R_A = 2 \times 10^8, D_{LP0} = 200 \text{ km} )</td>
<td>19</td>
<td>32</td>
<td>17.4, 34.4, 25.9</td>
</tr>
<tr>
<td>Model 2a</td>
<td>Same as Model 1, but with high-viscosity shallow most mantle due to dehydration</td>
<td>20</td>
<td>31</td>
<td>96.3, 12.3, 54.3</td>
</tr>
<tr>
<td>Model 2b (Fig. 5)</td>
<td>Same as Model 2a, but with northward regional mantle flow (NRF), and ( D_{LP0} = 400 \text{ km} )</td>
<td>22</td>
<td>29</td>
<td>47.1, 33.3, 40.2</td>
</tr>
<tr>
<td>Model 3 (Fig. 6)</td>
<td>No plume, no water dependence of viscosity, flow is driven entirely passively by the spreading plates ((R_A = 0)), NRF, A-type olivine LPO, ( D_{LP0} = 200 \text{ km} )</td>
<td>16</td>
<td>35</td>
<td>6.2, 25.3, 15.7</td>
</tr>
<tr>
<td>Models 4a, 4b, 4c, and 4e (Figs 7 and 8)</td>
<td>Same as Model 1 but with NRF and varying ( D_{LP0} ) ((150,200,300,400,600 \text{ km}) )</td>
<td>21</td>
<td>30</td>
<td>a: 28.2, 42.8, 35.5 b: 21.9, 31.7, 26.8 c: 12.7, 24.3, 18.5 d: 11.4, 32.0, 21.7 e: 13.1, 74.9, 44.0</td>
</tr>
<tr>
<td>Models 4c.i and 4c.ii (Fig. 9)</td>
<td>Same as Model 4c ((D_{LP0} = 300 \text{ km})), but olivine fabric is E ((\text{Model 4c.i})) or C ((\text{Model 4c.ii})) in the plume ((T_P &gt; 1360 ^\circ \text{C})) and where extent of partial melting is ( \leq 3 ) per cent. Elsewhere olivine LPO is A-type</td>
<td>22</td>
<td>29</td>
<td>c.i: 11.6, 29.5, 20.6 c.ii: 114.9, 23.3, 69.1</td>
</tr>
</tbody>
</table>

Simulated by changing the relative strengths of the different olivine slip systems \( \text{(Kaminski 2002)} \) using the values of Becker et al. \( \text{(2002)} \). The values of other D-Rex parameters are from Kaminski et al. \( \text{(2004)} \). LPO, the bulk anisotropic elastic tensors \( (\text{Voigt average of the hexagonally symmetric elastic tensors of the crystals in the aggregate}) \), and shear wave anisotropy were computed at every other finite element node of the numerical model.

An important consideration is the depth range over which LPO and anisotropy develop. Mantle deformation is described generally as the sum of strain rates due to both diffusion and dislocation creep, and LPO develops. Mantle deformation is described generally as the sum of strain rates due to both diffusion and dislocation creep, and LPO develops when the latter dominates. The depths in this study is too simplistic to self-consistently predict the maximum depth where dislocation creep dominates; therefore we assumed it is constant as defined by a model parameter, \( D_{LP0} \). A reference value of \( D_{LP0} = 200 \text{ km} \) was used in Models 1, 2a, 3, 4b, and we examined different values in Models 2b, 4a, c, d, e.

3.3 SWS predictions and comparisons with observations

After computing the solutions for mantle flow and anisotropy, we used the anisotropy structure to predict SWS directions and times. Synthetic waveforms were created by first setting the initial polarization directions of the incoming model SKS waves based on the distributions of the back azimuths of the real events used by Xue & Allen \( \text{(2005)} \) (taking into account the \( 180^\circ \) redundancy of these directions, six waves were simulated with incoming polarization azimuths of \( 45^\circ, 37^\circ, 29^\circ, 20^\circ, 13^\circ \) and \(-38^\circ\)). Then, after Rümpker & Silver \( \text{(1998)} \) (see also Fischer et al. \( \text{(2000)} \)), we assumed vertically travelling \( S \) waves and integrated the effects of consecutive splitting operators for each anisotropic layer along the ray path to compute the final waveform at the surface. With the synthetic split \( S \) waveforms, we then solved for the model station-averaged splitting direction \( \phi_M \) and time \( t_M \), using Silver and Chan’s \( \text{(1991)} \) and Wolfe & Silver’s \( \text{(1998)} \) methods, just as was done in the three SWS studies.

To quantify the fit of the model predictions to the observations, we computed the misfits between the model and observed mean split parameters at each station,

\[ \Delta \phi = |\bar{\phi} - \phi_M| \quad \text{and} \quad \Delta t = |\bar{t} - t_M| \]

(4)

(180° ambiguity in direction was accommodated by using the values that minimized the difference). The misfits normalized by the standard errors at each station are

\[ z_\phi = \frac{\Delta \phi}{\sigma_\phi} \quad \text{and} \quad z_t = \frac{\Delta t}{\sigma_t}. \]

(5)

Finally, the overall fit of a given model was quantified as chi-squared, scaled by the associated number of degrees of freedom. The scaled chi-squared quantities are

\[ \tilde{\chi}^2_\phi = \sum_{i=1}^n \frac{z^2_{\phi_i}}{v_i} \quad \text{and} \quad \tilde{\chi}^2_t = \sum_{i=1}^n \frac{z^2_{t_i}}{v_i}. \]

(6a)

for the fast polarization directions and delay times, respectively, where the number of degrees of freedom, \( v \), is the difference between \( n = 51 \) unique station locations and the number of model parameters, \( p \). The values of \( v \) and \( p \) for each model are given in Table 1, and the justifications for our estimates of \( p \) are given in Table 2. Finally, the combined chi-squared value is

\[ \tilde{\chi}^2 = \frac{(\tilde{\chi}^2_\phi + \tilde{\chi}^2_t)}{2}. \]

(6b)


4 MODEL RESULTS

4.1 Model 1: Plume with temperature-dependent viscosity

Model 1 (Table 1) simulates a low-viscosity plume owing to the temperature-dependence of viscosity without the effects of water content (eq. 3). Also assumed are no NRF and an olivine A-type LPO, which is the low-water content fabric commonly assumed for the ambient upper mantle. The plume rises from the model base and ponds in a layer of ~100 km in thickness, beneath the lithosphere where it spreads laterally away from the plume stem (Fig. 3). The viscosity of the hottest plume material just below the ridge axis is ~6 × 10^19 Pa s, less than a tenth of the viscosity of the ambient mantle, and 10^9 times lower than that of the lithosphere. As a result, the lateral flow of the ponding plume material tends to be more rapid and roughly parallel to the plate boundary in the shallow-most mantle (e.g. 50-km depth) where the plume is most influenced by the topography of the base of the lithosphere (see elongated isosurface in Fig. 3a and length of flow arrows in Fig. 4a). In this respect, this model represents a case of channelled plume flow along the plate boundary as proposed by Xue & Allen (2005). At 50 km depth, the horizontal components of the olivine a-axes (and directions of fast shear, Vp) tend to be (but are not everywhere) roughly parallel to the flow (Fig. 4a) because the LPO is controlled by shear between the plume and the overlying lithosphere.

In contrast, in the deeper portion of the plume layer, the flow is less influenced by the base of the lithosphere and therefore is directed radially away from the plume stem (Fig. 4b, 150 km depth). The olivine a-axes, however, tend to orient perpendicular to the flow in response to a combination of circumferential stretching (radially diverging flow lines) and radial shortening (flow rate decreases with radial distance). This behaviour was also seen in the numerical models of Ito et al. (2014) and Marquart et al. (2007), as well as in the laboratory experiments of Druken et al. (2013) (although the experiments measured directions of the integrated extensional strain, not of the crystallographic axes).

The depth-integrated effects of the model anisotropy structure are expressed as SWS predictions in Figs 4(c)–(f). The model split directions best match the observed directions in eastern and western Iceland (Figs 4c and e) but show large misfits (Δφ) in northern and southern Iceland. The misfits normalized by the standard errors (zφ) also show low values in most of eastern and western Iceland, and relatively high values in southern Iceland (Fig. 4e). The predicted delay times are generally shorter than observed: the mean model split time of t = 0.55 is a little more than half the observed mean (t = 0.99 s). The normalized misfits to delay time (z) are highest in the far east and lowest in the southwest (Fig. 4f). The overall misfits are zφ = 17.4, z = 34.4, χ2 = 25.9 (Table 1).

4.2 Model 2a: Plume with a stiff dehydrated layer

Model 2a is just like Model 1 but incorporates the added dependence of viscosity on mantle water content (Figs 3b and 5). Model 2a is motivated by a prior modelling study (Ito et al. 1999) finding that low-viscosities in the melting zone lead to an overprediction of crustal thickness, whereas high-viscosities due to dehydration lead to crustal thickness predictions more consistent with those measured beneath Iceland. The dehydrated, high-viscosity (~10^21 Pa s) layer in Model 2a extends from below the thermal lithosphere to the dry solidus (~120 km depth at the plume centre) and acts like a thick, relatively flat, compositional lithosphere moving with the plates. Hence, the plume spreads out in all directions and is not preferentially channelled along the ridge axis (Fig. 3b). Throughout most of the (DlPO =) 200-km-thick anisotropic layer, the flow and LPO are dominated by shear between the plume and the thick compositional lithosphere: the olivine a-axes tend towards the shear direction, being directed mostly radially outward from the plume centre in the lower part of the hot plume layer (e.g. 150 km depth, Fig. 5b) and directed mostly with the plate motion in the shallower part of the layer (e.g. 50 km depth, Fig. 5a).

The resulting SWS directions tend to radiate outward away from the plume centre, and thus are at a high angle to the observed directions. The absolute (Δφ) and normalized (zφ) misfits are much larger than those of Model 1 over most of Iceland (Figs 5c and e), and χ2 = 96.3. The split times, however, show smaller misfit χ2 = 12.3 because the times are greater on average (tM = 0.87) than those of Model 1. The larger split times are a consequence of the dehydrated mantle creating a thicker (100–200 km) layer where the directions of the olivine a-axes are more horizontal and less variable in the anisotropic layer. Nonetheless, the larger combined χ2 value of 54.3 indicates Model 2a is inferior to Model 1.

4.3 Model 3: Northwest regional mantle flow (NRF) without a plume

We now consider cases with a NRF. The first case (Model 3) has no plume and simulates mantle flow driven only by the spreading

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Table 2. Model parameters that significantly influence shear wave splitting predictions.

<table>
<thead>
<tr>
<th>Rheology</th>
<th>Parameter</th>
<th>Model 1</th>
<th>Model 2a (2b)</th>
<th>Model 3</th>
<th>Model 4 (4c.i, 4c.ii)</th>
</tr>
</thead>
<tbody>
<tr>
<td>η0, E/R, P/R</td>
<td>3</td>
<td>3</td>
<td>2</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>D-Resp</td>
<td>fabric type, M_{0}, \eta_{i}, D_{LPO}</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td>4</td>
</tr>
<tr>
<td>Ridge geometry</td>
<td>3 main ridge segments × 2 (spreading direction and speed)</td>
<td>8</td>
<td>8</td>
<td>8</td>
<td>8</td>
</tr>
<tr>
<td>Plume</td>
<td>Radius, D, x, y location</td>
<td>4</td>
<td>4</td>
<td>0</td>
<td>4</td>
</tr>
<tr>
<td>Regional mantle flow</td>
<td>Direction and speed</td>
<td>0</td>
<td>0 (2)</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>Water-dependent viscosity</td>
<td>Presence/absence</td>
<td>0</td>
<td>1</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Distinct fabric in plume</td>
<td>Extent of melting the fabric transitions to A (dry) fabric</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0 (1)</td>
</tr>
</tbody>
</table>

Total model parameters: 19 (20 (22)) 16 21 (22)
Shear wave splitting beneath Iceland

Figure 4. Results of Model 1: low-viscosity plume. Top row shows horizontal components of mantle flow (arrows, length proportional to horizontal flow rate) and preferred orientations of olivine a-axes (red bars) at depths of (a) 50 km and (b) 150 km shown at every 6th finite element node. Middle row shows absolute misfits between the predicted and observationally expected (c) fast polarization directions, and (d) split times. Bottom row shows normalized misfits for (e) fast polarization directions and (f) split times. Bars (length proportional to split time) show observationally expected (thick beige) and model (red) SWS results. In (d) red bars show model predictions at every 6th finite element node. Colours in (e) and (f) are saturated at $z = 6$, but $z$ values in the map can be larger.

plates ($Ra = 0$; no buoyancy driven flow) and NRF imposed at the model base. The rate and direction of the NRF was determined following Bjarnason et al. (2002). First, the (grand) mean measured split directions were computed for northwestern (North American Plate) and eastern (Eurasian Plate) Iceland separately ($\bar{\phi}_{NA} = 13.00^\circ$ and $\bar{\phi}_{EA} = -22.00^\circ$, respectively, see inset of Fig. 6). The grand mean $\bar{\phi}_{NA}$ was assumed to parallel the vector difference (and shear direction) between the motion of North American Plate and the NRF, and $\bar{\phi}_{EA}$ was assumed to parallel the vector difference between the motion of the Eurasian Plate and the NRF. From the two vector difference equations and the known relative plate motion vectors, a unique solution for the relative direction and magnitude of the NRF was obtained. In the model reference frame in which the MAR is fixed, the solution is 29.4 km Myr$^{-1}$ at an azimuth of $-6.05^\circ$ (long blue arrows in inset of Fig. 6). In the fixed hotspot reference frame, $0^\circ$ is defined as a southwestward flow. In the model reference frame, the flow is $\sim 48$ km Myr$^{-1}$ at an azimuth of $\sim -3^\circ$. The solution pertains to the flow at the base of the anisotropic layer; but because the regional flow is imposed at a depth of 660 km, it is given a faster rate of 69 km Myr$^{-1}$ in the model reference frame to achieve the approximate solution at the base of the anisotropic layer ($D_LPO = 200$ km).

The predicted flow and the olivine a-axes are largely horizontal and parallel to each other below the lithosphere at 150 km depth where simple shear dominates (Fig. 6b). Near the base of the lithosphere (50 km depth), the horizontal projections of olivine a-axes show variable orientations relative to flow (Fig. 6a). The SWS solutions recover what was originally sought in defining the regional flow vector: the split directions are oriented approximately with $\bar{\phi}_{EA}$ and $\bar{\phi}_{NA}$ in far eastern and northwestern Iceland, respectively (Fig. 6). However, the model fast directions are not uniform on each plate; they vary with the strain history beneath the plate, and primarily with proximity to the plate boundary. This
predicted variation leads to improved fits to the observed mean fast directions in northcentral Iceland, not the worse fits that Xue & Allen (2005) thought would occur given the assumption of uniform shear beneath each plate (after Bjarnason et al. 2002), which led them to reject the regional flow explanation. The misfits to $\bar{\phi}$ are lower than those of Model 1 (and Model 2a) and show low $z_\phi$ values over a greater area of Iceland (Fig. 6e), resulting in $\sum_\phi^2 = 6.2$. The model delay times decrease across Iceland from SE to NW...
due to the asymmetry in the shear rate between the two plates and the NRF (Fig. 6f). The mean split time of 1.13 s better matches the observed mean (again 0.99 s) and $\chi^2 = 25.3$ is relatively low. The combined $\chi^2$ of 15.7 turns out to be the smallest for all the models examined.

4.4 Models 4 and 2b: Plume and northwest regional mantle flow

We now examine several cases that include a plume plus the NRF (Fig. 3c). We first discuss a model that combines a low-viscosity plume, like in Model 1 (without NRF), with the NRF, as in Model 3 (no plume with NRF). In this Model 4b, the patterns of the horizontal components of mantle flow and LPO, at the two depths in Fig. 7, are similar to those of Model 1 (plume and without NRF); the main difference is at 150 km depth in the far southeast where the additional NRF in Model 4b inhibits flow away from the plume centre. Correspondingly, the resulting SWS predictions for Model 4b are similar to those of Model 1. The $\chi^2$ value of 26.8 is comparable to that of Model 1 (25.9) and higher than that of Model 3 (15.7, NRF without a plume). The similarity of the results to Model 1 and the contrast with Model 3, reflects the fact that in Model 4b, the ponding plume material spans most of the anisotropic layer (i.e. $D_{LPO} = 200$ km), and therefore the LPO is controlled primarily by the spreading plume material and is not significantly changed by the NRF. The anisotropic layer is simply too thin for the NRF to appreciably impact the SWS predictions.

We therefore examine four additional cases with different anisotropic layer thicknesses: $D_{LPO} = 150$, 300, 400 and 600 km (Fig. 8). The misfits are larger for the thinner anisotropic layer
Figure 7. Same as Fig. 6 but for Model 4b: NRF with low-viscosity mantle plume and LPO developing within a depth of $D_{\text{max}} = 200$ km as in Models 1–3.

(Model 4a, $D_{\text{LPO}} = 150$ km), improve for thicker layers (Models 4c,d, $D_{\text{LPO}} = 300$ and 400 km, respectively), and increase again for the thickest layer ($D_{\text{LPO}} = 600$ km, Model 4e). The minimal $\chi^2$ of 18.5 for Model 4c ($D_{\text{LPO}} = 300$ km), is well below that for Model 1 (25.9, plume without NRF) and is closer to that of Model 3 (15.7, NRF without plume).

Finally we examine a case (Model 2b, Figs 5g and h) with a high-viscosity, dehydrated layer as in the plume Model 2a, but with NRF and a relatively thick anisotropic layer ($D_{\text{LPO}} = 400$). This value for $D_{\text{LPO}}$ produces the second lowest $\chi^2$ misfit among the models with a low-viscosity plume and allows us to examine the degree to which the fits to the observations are sensitive to the shallow viscosity structure of the plume in the presence of a relatively strong influence by the NRF. The total misfit of $\chi^2 = 40.2$ for this Model 2b is lower than that (54.4) for Model 2a (high-viscosity dehydrated layer without NRF), but is still much higher than those (18.5 and 21.7) of Models 4c and 4d (low-viscosity plume with NRF, $D_{\text{LPO}} = 300$ and 400 km, respectively). These results point towards low, not high viscosities in the shallowest $\sim 100$ km for cases with a plume and NRF.

4.5 Different olivine fabrics in the plume

The last two cases explore the possibility that the elevated water content of the Iceland plume (Nichols et al. 2002) causes different LPO fabrics (Karato 2007). The models are the same as Model 4c (low-viscosity plume + NRF, $D_{\text{LPO}} = 300$ km), which has the lowest $\chi^2$ of the plume models. E-type fabric is expected to occur at low stresses (i.e. low viscosities) and moderate water contents (Karato 2007). Model 4c.i therefore simulates E-type fabric in the plume ($T_p \geq 1360$ °C) prior to it losing its water due to melt extraction (extent of partial melting $\leq 3$ per cent), and A-type fabric elsewhere ($T_p < 1360$ °C and extent of melting $> 3$ per cent). The depth at which the fabric switches from E-type (below) to A-type (above) is $\sim 120$ km in the hottest (deepest solidus) centre of the plume. The SWS predictions (Figs 9a and b) are qualitatively indistinguishable.
Figure 8. Results of Models 4a, c, d and e: NRF with low-viscosity plume and LPO developing within different maximum depths ($D_{LPO}$). Colours show $z$ values for (left-hand column) split directions $\phi$ and (right-hand column) times $t$ as in Figs 4(e) and (f). Thick beige bars show mean observed SWS results; thin red bars show model SWS predictions at each seismic station (left-hand column) and every 6th finite element node (right-hand column).

from those of Model 4c (A-type fabric only). This result might have been anticipated because the $a$-axes of olivine are oriented similarly for E-type and A-type fabric during simple shear, with the LPO differing only in the orientation of the $b$- and $c$-axes (Fig. 9). The $\bar{\chi}^2$ value of 20.6 for Model 4c.i is comparable to that (18.5) of Model 4c.

Model 4c.ii simulates C-type fabric in the plume for $F \leq 3$ per cent; and A-type fabric elsewhere. C-type fabric is expected for low stresses and high water contents and was hypothesized by Karato (2007) to occur in mantle plumes. C-type fabric differs from A-type fabric in a more profound way: during simple shear, the $a$-axes of olivine tend to lie in the shear plane (as for A-type fabric), but are nearly orthogonal to the shear direction. As a result, Model 4c.ii predicts that in the lower part of the plume layer where circumferential stretching controls LPO, the $a$-axes tend to parallel the flow and form a radial pattern, rather than the circumferential
Figure 9. Models 4c.i with E-type olivine fabric in the plume ($T_p > 1360$ °C) where extent of melting $F < 3$ per cent and A-type fabric elsewhere; and (bottom row) Model 4c.ii with C-type fabric in plume and $F < 3$ per cent and A-type fabric elsewhere. Format of maps are the same as in Fig. 8. Pole figures show predicted orientation distribution functions of olivine $a$-, $b$-, and $c$-axes for the different fabric types after 100 per cent simple shear shown by the double arrows.

pattern produced with E- or A-type fabrics. The predicted SWS pattern thus shows dominantly radially oriented fast directions. In fact, the pattern resembles that produced by the poorest fitting Model 2a (high-viscosity dehydrated layer, without NRF, Fig. 5), in which the anisotropy pattern due to A-type fabric was dominated by radial shear effects. The misfits of Model 4c.ii are correspondingly large ($\bar{\chi}_\phi^2 = 114.9, \bar{\chi}_t^2 = 23.3, \bar{\chi}_J^2 = 69.1$).

4.6 Identifying the best-fitting models

Although the chi-squared residuals indicate the level of fit of the models to the observations, tests are needed determine which models are statistically distinguishable from the others, and thus which mantle processes are most likely to control the SWS observations. Model 3 (NRF, without a plume) produces the lowest $\bar{\chi}^2$ value; therefore the first test is to determine which residuals of the plume models are distinguishable from that of Model 3. The ratios of the residuals of a given model $M$ relative to that of Model 3 are

$$F_{\phi,M} = \frac{\bar{\chi}_\phi^2}{\bar{\chi}_\phi^3}, \quad F_{t,M} = \frac{\bar{\chi}_t^2}{\bar{\chi}_t^3}, \quad F_{J,M} = \frac{\bar{\chi}_J^2}{\bar{\chi}_J^3}. \quad (7)$$

for split direction $\phi$, delay time $t$, and the combination, respectively. The $F$ distribution for degrees of freedom $\nu_M$ and $\nu_3$ allows us to estimate the probability that the ratios of residuals arose due to Gaussian noise. The null hypothesis, $H_0$, states that the residuals of Models 1, 2 and 4 differ only due to random chance and are therefore indistinguishable from the residuals of Model 3. The results show that the $F$ values for Models 2a, 2b, 4a, 4e and 4c.ii are highly unlikely to have occurred by pure chance (Fig. 10), and $H_0$ is rejected at the $\alpha = 5$ per cent significance level. These models have detectably worse misfits than Model 3 and are least likely to explain the SWS observations. For Models 1, 4b, 4c, 4d and 4c.i, however, $H_0$ cannot be rejected at $\alpha = 5$ per cent, so there is no reason to distinguish their results from that of Model 3. A subsequent test does not allow us to distinguish their results from each other either.

A second test addresses whether the reduction in $\bar{\chi}^2$ due to the addition of NRF in the low-viscosity plume models 4b, 4c and 4d significantly improves the misfit over the case of a low-viscosity plume without NRF (Model 1). Here, $F$ is

$$F = \left( \frac{\sum_{i=1}^{n} z_i^2 - \sum_{i=1}^{n} z_{Mi}^2}{v_1 - v_M} \right) / \bar{\chi}_M^2, \quad (8)$$
where the numerator is the reduction of the sum of square, normalized misfits of model $M$ (i.e. $4b$, $4c$ and $4d$) compared to Model 1, scaled by the loss of the 2 degrees of freedom associated with the NRF. This chi-squared statistic is compared with the $\chi^2$ value for model $M$ in the denominator. The resulting $F$ values of 0.46, 7.35 and 4.08 for Models $4b$, $4c$ and $4d$, respectively, have probabilities of 63, 0.25 and 2.7 per cent of occurring randomly without any true differences in the sum of square misfits. Hence the addition of the NRF significantly improves the misfit for Models $4c$ and $4d$—but not for $4b$—at the $\alpha = 5$ per cent level. These results indicate that NRF is important and further support the need for $D_{LPO} = 300$–400 km.

5 MODEL LIMITATIONS

Whereas the above tests allow us to identify the best-fitting models, in fact, none of the model predictions fit all of the observations especially well ($\varepsilon$ values typically exceeding 1–2). Thus, there are other factors controlling the SWS observations that are not well addressed by the models.

One set of model limitations relates to the predictions of seismic anisotropy. We have used only one method (D-Rex) for computing LPO; whereas there are more mechanically complete methods (e.g. see Blackman 2007 and references therein) that show predictions that differ from those of D-Rex to varying degrees (Castelnau et al. 2009). Another simplification is associated with the approximate, see Blackman 2007 and references therein) that show predictions LPO; whereas there are more mechanically complete methods (e.g. they develop. More realistically, the depths over which LPO forms by dislocation creep-dominated deformation should vary appreciably due to large variations in strain rates (Podolefsky et al. 2004; Conder 2007) such as those associated with plume-ridge interaction. We have also assumed that the only source of seismic anisotropy is LPO due to melt-free deformation; however, it has been argued that the presence of melt can alter the sense of mantle deformation (Holtzman et al. 2003), or create its own seismic anisotropy (e.g. Blackman & Kendall 1997; Mainprice 1997; Holtzman & Kendall 2010).

The models also did not address some local geologic and tectonic factors. For example, the intrusion of dikes in the crust can create anisotropy that may influence SWS (Li & Detrick 2003; Fu et al. 2012), albeit to a minor degree given the short distance $S$ waves travel in the crust compared to in the mantle. Perhaps more significant may be remnant LPO in the mantle associated with the eastward shifts of the ridge segment now at the Northern Volcanic Zone at $\sim$7 and $\sim$15 Ma (Garcia et al. 2008), as well as newly developing LPO associated with the recent (2–3 Ma) southward propagation of the Eastern Volcanic Zone (Einarsdottir 1991; Ivarsson 1992). The two best-fitting plume models (Models $4c$, $4ci$) show large values of $\varepsilon_{\phi}$ in southern Iceland and large values of $\varepsilon_{z}$ in both southern and north central Iceland (Figs 8c, 8d, 9a, and 9b).

Finally, the processes that the models did simulate are associated with some $\sim$20 model parameters, whereas only a handful of parameters were varied. It is likely that more optimal models exist within the full parameter space, but finding them is impractical given the computational time required. The current models should therefore be viewed as representations of the few predominant influences (plume/no plume, NRF or not, fabric type, water-dependent rheology or not), and thus this study addresses which among these aspects are more and less likely to control the SWS beneath Iceland.

6 DISCUSSION

6.1 Water effects

The effects of water on mantle rheology and the development of E- and C-type LPO fabric were tested with Models 2a, 2b, 4c.i and 4c.ii, respectively. The large misfits and the $F$ test results of Models 2a and 2b indicate that a 10$^2$ time increase in viscosity in shallow-most mantle due to dehydration cannot explain the SWS observations beneath Iceland. This is true even in the presence of a relatively thick ($\sim$200 km) layer of anisotropic mantle below the ponding plume material within which the LPO is controlled by the NRF (Model 2b, $D_{LPO} = 400$ km). Thus, if dehydration really does strengthen the mantle beneath Iceland, then the effects must be much less than those modelled. For example the contrast between wet and dry rheology could be less (Fei et al. 2013), the weakening effects of partial melt beneath Iceland could partially counteract the effects of dehydration (Mei et al. 2002), power law, dislocation creep and enhanced strain-rates beneath Iceland could reduce the effective viscosity (Ito et al. 2010), or the maximum temperature...
of the plume could be lower (Kreutzmann et al. 2004) making the dehydrated layer thinner. Regarding olivine fabric type, Model 4c:ii was a test of Karato’s (2007) hypothesis that C-type fabric typically occurs in plumes. The large misfits of this model and the results of the first $F$ test suggest this hypothesis is incorrect for the Iceland plume. A plume with E-type fabric (Model 4c:i), which is expected at moderate water contents, produces low $\chi^2$ misfits and $F$-test results that do not allow us to reject the null hypothesis that this model is indistinguishable from the similar model with A-type fabric everywhere (Model 4c). Thus, E-type fabric is allowed within the Iceland plume, but is not necessarily favoured.

6.2 Regional flow

The second $F$ test showed that the low-viscosity plume Models 4c and 4d with NRF produce distinguishably better misfits than the low-viscosity plume model without NRF (Model 1). This argues for NRF as having a strong influence on the SWS at Iceland. Furthermore, the results for Models 3 (NRF, without a plume), 4c, and 4d suggest that the layer of mantle within which simple shear associated with the NRF controls the LPO is relatively thick (200–300 km). Such a layer thickness is present in Model 3 without a plume because $D_{LPO} = 200$ km, and is present in Models 4c and 4d below the ponding plume material because $D_{LPO}$ is much thicker (300–400 km). Model 3, however, must ultimately be rejected given the wealth of evidence for a plume-like upwelling beneath Iceland (Wolfe et al. 1997; Shen et al. 1998; Allen et al. 2002). We therefore conclude that a low-viscosity plume interacting with the MAR controls the LPO in the shallowest 100 km of the mantle, and that the NRF controls the LPO below the plume material in the mid-upper mantle between the depths of ~100 km and 300–400 km. This conclusion argues against Bjarnason et al.’s (2002) interpretation of NRF without a plume, as well as Xue and Allen’s (2005) interpretation of channelled plume flow without NRF. Rather, the conclusion supports Li & Detrick’s (2003) general interpretations of both a plume and NRF with strong layering in anisotropy.

The results also shed light on the factors controlling the depths at which LPO develops via dislocation creep. The thickness over which dislocation creep dominates over diffusion creep is expected to increase with strain rate. Recent studies predict that the strain rates associated with typical rates of plate motion lead to $D_{LPO}$ of ~200 km (Podolefsky et al. 2004), which is broadly consistent with seismic constraints (Ekström & Dziewonski 1998). Thus, the results of $D_{LPO} = 300–400$ km of our favoured models (4c and 4d) represent an anomaly, especially when considering the locally slow plate motion. The implication is that the combination of laterally flowing plume material and NRF leads to sufficiently large strain rates to as much as double the thickness over which dislocation creep dominates beneath Iceland.

Other studies have provided independent evidence for northward regional mantle flow around Iceland. Recent global models of azimuthal anisotropy show some supporting evidence to varying degrees (Debayle & Ricard 2013; Schaeffer & Lebedev 2013; Yuan & Beghein 2013). Stronger supporting evidence comes from a surface wave study using data from a regional seismic array around the North Atlantic (Pildiu et al. 2005). This study found fast Rayleigh directions near Iceland that are directed NW–SE across the range of depths resolved by the study (75–350 km). A receiver function study of data from the ICEMELT and HOTSPOT stations on Iceland found that an anomalously thin mantle transition zone, attributed to the Iceland plume, was centred well south of the tomographically imaged seismic anomaly in the upper mantle (Shen et al. 2002). This finding was interpreted as indicating a southward tilt (looking downward) of the Iceland plume in the upper mantle, similar to the prediction of Model 4 (Fig. 3c). Finally, a regional tomography study of the North Atlantic imaged distinct plume-like, low-velocity bodies beneath both the Iceland and the Jan Mayen hotspots that protrude downward into the lower mantle with striking southward tilts (Rickers et al. 2013). Such a tilt is consistent with an appreciable northward component of mantle flow in the upper mantle relative to the deep source of the Iceland plume.

The cause of the purported NRF has not been explored, but differences between models of global mantle flow reveal some hints. Little or no northward flow near the base of the asthenosphere around Iceland is predicted by models in which global mantle flow is driven by plate motion plus thermal buoyancy (inferred from seismic tomography) with viscosity being depth- or temperature-dependent (Becker 2006; Mihalffy et al. 2007; Conrad & Behn 2010). More northerly (but still too slow) flow is predicted in models with mantle buoyancy if viscosity also varies moderately with strain rate in the asthenosphere (Becker 2006), or in a model without thermal buoyancy, with a low-viscosity asthenosphere (Conrad & Behn 2010). Thus the viscosity structure in the upper mantle appears to be an important factor. But a key limitation is resolution. Models that include tomographic information (for thermal buoyancy or viscosity) are influenced by seismic heterogeneity related to the Iceland plume. This factor plus the relatively coarse resolution of the models themselves make it unlikely that the global mantle flow models are adequately distinguishing the flow driven by the plates and larger scale heterogeneity, from the local short-wavelength flow driven by the Iceland plume. Addressing the cause of the NRF around Iceland may require some combination of global and adequately resolved regional models of mantle flow as well as seismic tomography.

7 Conclusions

This study addresses the origin of SWS on Iceland using numerical simulations of 3-D mantle flow, the development of anisotropy due to LPO, and synthetic SWS. Models that simulate the effects of water (either via a 10$^2$-fold increase in mantle viscosity with its extraction during melting, or via the presence of olivine C-type fabric within the plume) produce the largest chi-squared misfits to the observations and are thus rejected. Models with a rapid (~50 km Myr$^{-1}$) northwest regional mantle flow (NRF) in the mid-upper mantle produce distinguishably better misfits than the best-fitting case without NRF (Model 1). The favoured models predict that a low-viscosity Icelandic mantle plume interacting with the Mid-Atlantic ridge controls the LPO in the shallowest ~100 km and NRF heavily influences the LPO between depth of ~100 to 300–400 km. The unusually large total depth (300–400 km) over which LPO is predicted to develop by dislocation creep is inferred to be caused by the anomalously rapid strain rates associated with the Iceland plume and the NRF. NRF near Iceland is evident in other regional seismic studies of the North Atlantic and understanding its origin will require models that adequately distinguish the local flow associated with the Iceland plume from the global scale mantle flow.

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