Petrofabric derived seismic properties of a mylonitic quartz simple shear zone: implications for seismic reflection profiling

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Abstract: The link between petrofabric (LPO) and seismic properties of an amphibolite facies quartzofeldspathic shear zone is explored using SEM/EBSD. The shear zone LPO develops by a combination of slip systems and dauphine twinning with a-maximum parallel to lineation (X) and c-maximum normal to mylonitic foliation (XY). The LPO are used to predict elastic parameters from which the three dimensional seismic properties of different shear zone regions are derived. Results suggest that LPO evolution is reflected in the seismic properties but the precise impact is not simple. In general, the P-wave velocity (Vp) minimum is parallel to the a-axis maximum, the direction of maximum shear wave splitting (AVs) is parallel to mylonitic foliation and the Vp maximum and AVs minimum are parallel to the c-axis maximum. The seismic anisotropy predicted is significant and increases from shear zone wall rock to mature mylonite. P-wave anisotropy ranges from 11-13%, fast and slow shear waves anisotropy range from 6-15% and the magnitude of shear wave splitting ranges from 9-16%. Nevertheless, such anisotropy requires a considerable thickness of rock with this LPO before it becomes seismically visible (i.e. 100's m for local earthquakes, 10's m for controlled source experiments). However, reflections and mode conversions provide much better resolution, particularly across tectonic boundaries. The low symmetry and strong anisotropy due to the LPO suggest that multi-azimuth wide-angle reflection data may be useful in the determination of the deformation characteristics of deep shear zones.

A number of factors may produce elastic anisotropy in rocks, including variations in the spatial distribution of mineral phases, layering, grain size and shape fabrics, grain boundary properties and the presence of oriented pores or fractures (e.g., Kern & Wenk 1985; Mainprice et al. 2003). In addition, because elastic properties vary with respect to direction in single crystals, the majority of rock-forming minerals are elastically anisotropic (e.g. Christensen 1966). Thus, deformation processes, such as dislocation creep, which produce strong crystal lattice preferred orientations (LPO) in anisotropic minerals, may also induce bulk elastic anisotropy. It is no surprise therefore that petrophysical properties, such as seismic velocities, are often highly anisotropic in bulk rock aggregates (e.g. Babuska & Cara 1991).

Understanding of the impact of microstructural variables on elastic anisotropy has been obtained typically from experimental measurements of elastic properties of minerals or theoretical models that incorporate microstructural variables (e.g. Burlini & Kern 1994; Wendt et al. 2003). Experimental studies have focused primarily on the influence of fractures or LPO on the elastic properties of minerals and rocks. In particular, the seismic properties can be measured via direct laboratory methods at ambient or elevated temperatures and pressures (e.g. Kern 1982), although it is often difficult or impossible to obtain the complete three dimensional property orientation distribution from a single experiment (however, see Pros et al. 2003). Furthermore, experimental investigation of seismic properties usually involves rocks in which all microstructural fabric elements contribute, effectively masking any LPO component.

An alternative approach is to calculate the seismic properties of a rock aggregate from individual crystal orientation measurements, incorporating the single crystal elastic constants and the LPO for each mineral weighted according to its modal content (e.g. Mainprice & Humbert 1993). The seismic phase velocities and anisotropies then can be calculated in every direction (e.g. Mainprice & Silver 1993). Recently, the advent of electron backscattered diffraction (EBSD) in the scanning electron microscope (SEM) has made it possible to measure LPO in statistically viable numbers from different minerals within a rock aggregate whilst maintaining a one-to-one relationship with other microstructural elements (e.g. Mainprice et al. 1993, Lloyd 2000; Mauler et al. 2000; Bascou et al. 2001). In this contribution, the seismic properties of an amphibolite facies, mylonitic quartz shear zone are predicted by averaging the single crystal elastic properties according to the SEM/EBSD derived LPO in order to study the relationship between microstructure, petrofabric and seismic properties. This approach illustrates the usefulness of integrating geological studies of rock microstructures, and particularly petrofabrics, with geophysical studies of seismic properties to obtain a fuller understanding of geodynamic processes.

Specimen details

The sample used in this study was collected from a 30cm wide deformed planar quartz vein (Fig. 1a)
Fig. 1. Summary of Torridon simple shear zone microstructure and LPO. All pole figures equal area, upper hemispheres viewed towards ENE; contour intervals multiples of mean uniform distribution. (a) SEM electron channelling (see Lloyd 1987 for details) orientation contrast image of microstructure, indicating positions of SEM/EBSD analyses: SZWR, shear zone wall rock (analysis G 030600); SZM, shear zone margin (G040800); MM, mature mylonite (G270700); and MD, mylonite/shear zone detail (G050800). Amended from Lloyd (in press). (b) Specimen co-ordinates for all LPO diagrams (after Law et al. 1991). (c) Universal stage optical c-axis pole figure; 1815 measurements (after Law et al. 1991). (d) X-ray pole figures for c, m, a, r, z, π and π' orientations (after Law et al. 1991). (e) Manual SEM electron channelling pole figures for c, m, a, r, z, π and π' orientations after Lloyd et al. 1992).
The vein was deformed by crystal plastic processes to form a dextral shear zone with an intense mylonitic foliation and lineation (Fig. 1a, region MM). The foliation locally displays the orientation variations expected for heterogeneous simple shear (e.g. Ramsay & Graham 1970; Ramsay 1980): X parallel to lineation, XY parallel to foliation and Z normal to foliation. The shear zone wall rock (Fig. 1a) comprises coarsely crystalline quartz and plagioclase feldspar. Foliation at the shear zone margins (Fig. 1a) is oriented at ~45° to the vein walls but curves rapidly with a dextral shear sense to become sub-parallel to the walls. In the XZ section analysed in this contribution, the vein microstructure is that of a classic Type II S-C mylonite (Lister & Snoke 1984) and consists of two planar domains, A (grain size <5µm) and B (grain size <100µm), aligned parallel to the macroscopic mylonitic foliation (S). The obliquity between grain axes (S, S) in the two domains is consistent and indicates a dextral shear sense. The microstructures observed are indicative of constant volume, strongly non-coaxial, essentially plane strain deformation that closely approximates to simple shear.

The quartz petrofabric (Figs. 1c-e and 2) is characterized by an <a> axis maximum sub-parallel to X, which occupies a pole position to the corresponding single girdle (c) axis fabric, with a (c) axis maximum oriented sub-normal to the XY foliation plane and normal to the average of the quartz basal plane defined by the LPO. These associations suggest a simple relationship between shear zone geometry, simple shear kinematic framework and orientation of crystal slip systems responsible for shear zone formation. Most quartz grains are preferentially oriented to exploit slip on systems that utilise <a> as the slip direction (such as {-r}<a>, [z]<a>, (c)<a> and (m)<a>). Such observations can be interpreted in terms of a lower resistance to slip on negative forms compared to positive forms and/or the occurrence of significant dauphine twinning. The latter is consistent also with the {r} point maximum position within the XZ plane at ~35° to the foliation, subparallel to the NW-SE trending maximum principle stress direction inferred from the simple shear kinematic framework (e.g. Tullis & Tullis 1972). The petrofabric therefore approximates that of a ‘dauphine twinned single crystal’.

SEM studies of the Torridon shear zone have focused attention on the misorientations between adjacent grains and hence upon the grain boundary...
Table 2. Seismic properties analysis for single crystal quartz (left) and plagioclase (right)

<table>
<thead>
<tr>
<th></th>
<th>Quartz</th>
<th>Plagioclase</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Symmetry:</strong></td>
<td>trigonal, space group P3221</td>
<td>triclinic, space group P-1</td>
</tr>
<tr>
<td><strong>Unit cell dimensions (Å):</strong></td>
<td>a = 4.9130, b = 4.9130, c = 5.5040</td>
<td>a = 8.1553, b = 12.8206, c = 7.1397</td>
</tr>
<tr>
<td><strong>Unit cell angles (degrees):</strong></td>
<td>α = 90.0, β = 90.0, γ = 120.0</td>
<td>α = 93.95, β = 116.47, γ = 89.62</td>
</tr>
<tr>
<td><strong>Euler angle triplet:</strong></td>
<td>Φ&lt;sub&gt;max&lt;/sub&gt; = 180°, φ&lt;sub&gt;max&lt;/sub&gt; = 180°, φ&lt;sub&gt;2-max&lt;/sub&gt; = 120°</td>
<td>Φ&lt;sub&gt;max&lt;/sub&gt; = 360°, φ&lt;sub&gt;max&lt;/sub&gt; = 180°, φ&lt;sub&gt;2-max&lt;/sub&gt; = 360°</td>
</tr>
<tr>
<td><strong>Density:</strong></td>
<td>density = 2.6473 g.cm&lt;sup&gt;-3&lt;/sup&gt;</td>
<td>density = 2.6100 g.cm&lt;sup&gt;-3&lt;/sup&gt;</td>
</tr>
</tbody>
</table>

Methodology

**LPO determination**

SEM/EBSD (e.g. Venables & Harland 1973; Dingley 1984) provides the precise crystallographic orientation at the point of incidence of the electron beam on the sample surface, with a spatial resolution of ~1 μm and angular resolution of ~1° (e.g. Prior et al. 1999). A one-to-one correlation is achieved therefore between crystal orientation and microstructural position. The crystal orientation is defined by three spherical Euler angles (Φ, ϕ, φ) with respect to the sample reference frame (Bunge 1982). EBSD analysis can be performed either manually or automatically, but in both cases the diffraction patterns are indexed via computer programs; the system used in this study was the HKL Technology Channel 5 software (e.g. Schmidt & Olesen 1989). Automated EBSD
analysis enables very large (i.e. \(>10^6\)) orientation data sets to be acquired from samples \(>10\times10\text{mm}\) area, which can be used to test and/or interpret seismic anisotropy (e.g. Burlini & Kunze 2000; Lloyd 2000; Mauler et al. 2000).

Three overlapping large-scale auto-EBSD analyses (G030100, G040800, G270700) were performed extending from the shear zone wall rock to the mature mylonite and a detailed auto-EBSD analysis (G050800) of the mature mylonite (Lloyd in press; see Fig. 1a for locations and Table 1 for experimental details). The LPO derived from the SEM/EBSD experiments (Fig. 2) provide a part of the input data for the seismic properties determinations described below. Important aspects of the LPO are as follows. (1) The shear zone wall rock (Fig. 2a) and shear zone margin (Fig. 2b) comprise relatively few, large quartz (and plagioclase in the former) grains and hence their LPO consist of isolated concentrations associated with individual grain orientations. (2) The mylonite LPO (Fig. 2c, d) approximates that of a ‘dauphine twinned single crystal’, consistent with results derived from other methods (e.g. Fig. 1c-e), which maintains (strengthens?) the \(\{m\}\) positions but interchanges the \(\{r\}\) and \(\{z\}\) positions. (3) The differences between LPO from the shear zone margin and mature mylonite indicate a rapid migration of wall rock crystal pole directions towards the mature mylonite LPO, in agreement with microstructural observations (see Fig. 1a). (4) The shear zone shear strain gradient and dynamic recrystallisation rates therefore are expected to have been steep and/or rapid.

**Seismic properties determination**

The propagation of seismic waves generates a short-term deformation in a medium, such that the velocity and polarisation direction of the waves depend on the elastic parameters (i.e. the elastic stiffness matrix) of the medium and the nature of the deformation. Thus, knowledge of the elastic parameters can be used to predict seismic velocities and propagation directions. The general relationship between elastic parameters and seismic waves is given by (e.g. Nye 1957, Kendall 2000),

\[
\begin{align*}
(C_{ijkl}X_lX_j - \delta_{kl}\rho V^2)U_\ell &= 0, \\
\end{align*}
\]

where \(C_{ijkl}\) is the fourth-order proportionality tensor that relates stress to strain, \(X_l\) expresses the propagation direction of the wave (\(C_{ijkl}XX_l\) is the 3x3 Christoffel matrix), \(\delta_{kl}\) is the Kronecker delta, \(\rho\) is the density of the medium, \(V\) is the phase velocity of the wave in a given direction and \(U_\ell\) is the displacement.

A non-trivial solution for \(U_\ell\) requires that the determinant of the system in (1) vanishes and results in the Christoffel equation (e.g. Nye 1957),

\[
\text{det}\left[C_{ijkl}X_lX_j - \delta_{kl}\rho V^2\right] = 0. \tag{2}
\]

The Christoffel equation has three possible solutions representing one compressional (P) and two shear (\(S_1, S_2\)) waves. Due to the symmetry of the elastic tensor, the Christoffel matrix is symmetrical also, which means that the three solutions have mutually perpendicular displacement vectors. In isotropic media, seismic wave velocities are independent of their propagation direction and their polarisation depends only on the type of wave and the nature of the source. In anisotropic media, seismic velocities depend locally on the propagation direction and their polarisation depends not only on the type of wave but also on the local symmetry of the elastic properties (i.e. \(C_{ijkl}\)).

Although \(C_{ijkl}\) has 81 components, these can be reduced to a symmetrical \(6 \times 6\) matrix, \(C_{ij}\), because of symmetry in the stress and strain matrices (Babuska & Cara 1991). Consideration of the energy function of a strained crystal, which depends only on the strain components, reduces the stiffness matrix to a maximum of 21 independent coefficients. Furthermore, elastic parameters of crystalline media depend ultimately on chemical composition and atomic arrangement of the crystal structure and hence are characteristic for each mineral. Thus, the elastic parameters are closely related to the strength of interatomic bonds in corresponding directions of the crystal structure (e.g. elastic moduli are larger in the direction in which the structure has the strongest bonds). Consequently, because elastic parameters are the same in crystal symmetry related directions, the number of independent elastic coefficients for single crystals can be reduced further still depending on crystal symmetry (see Babuska & Cara 1991, table 2.1).

**Single crystal seismic properties**

Understanding the seismic behaviour of the individual constituent crystals is critical to any interpretation of whole rock seismic behaviour, as exhibited for example in mylonitic shear zones. As an example of the impact of crystal structure on seismic properties and the methodology used to predict whole rock seismic properties, the variations in compressional and shear waves velocities with direction have been calculated for quartz and plagioclase single crystals, the minerals of interest in the Torridon shear zone.

Figure 4a shows common morphological single crystal forms of quartz and plagioclase and their relationship to directions of maximum and minimum seismic velocities (V) and anisotropy (A), based on experimentally determined values.
Figure 4. Seismic property distributions (Vp, Vs1, Vs2, AVs, dVs, Vs1 polarisation planes – see text for details) of quartz and plagioclase single crystals based on LPO-averaging (see Table 2 for input data). All pole figures equal area, upper hemispheres projections; contours multiples of mean uniform distribution, as indicated. (a) Quartz and plagioclase single crystal forms (after Dana & Ford, 1951) and seismic property variations (after Babuska & Cara 1991). (b) Quartz single crystal LPO generated from an individual Euler angle triplet. (c) Variation in quartz single crystal seismic properties with crystal direction. (d) Plagioclase single crystal LPO generated from an individual Euler angle triplet. (e) Variation in plagioclase single crystal seismic properties with crystal direction.

summarized by Babuska & Cara (1991). Although the principal crystal directions for quartz (i.e. $a(1120)$, $m(100)$, $c(0001)$, $r(1011)$, $z(0111)$) and plagioclase (i.e. $a(100)$, $b(010)$, $c(001)$) can be represented on a single stereographic projection, petrofabric analysis typically plots individual ‘pole figures’ for each form. Appropriate Euler angle triplets (e.g. Casey 1981) for quartz and plagioclase (Table 2a) were used therefore to calculate individual pole figure diagrams (Fig. 4b, d) representative of the single crystal forms via the program Pfch5 (Mainprice, 2003).

The single crystal elastic tensor, $C_{ijkl}$, is generally defined in the crystal reference frame, whereas crystal orientations are generally defined via the Euler angles in the specimen reference frame. The former can be rotated into the latter via (Babuska & Cara 1991),

$$ C_{ijkl}(g) = g_{im} g_{jn} g_{ko} g_{lp} C_{mnop}^0, $$

(3)

where $g_{ij} = g(\phi, \varphi, \phi_2)$ is the crystal to sample reference frame rotation matrix and $C_{mnop}^0$ is the single crystal elastic stiffness tensor in the crystallographic reference frame.

The single crystal orientations, as defined by the Euler angle triplets in (3), were combined via (2) with the experimentally determined single crystal elastic stiffness matrix and density for each mineral (i.e. quartz, McSkimin et al. 1965; plagioclase, Aleksandrov et al. 1974; see Table 2b) to calculate the single crystal seismic property distributions using the program ANISch5 (Mainprice 2003; see also Mainprice 1990; Mainprice & Humbert 1994). Stereographic projections (Fig. 4c, e) of the seismic properties have been plotted in sample coordinates via the program VpG (Mainprice 2003). The specific property distributions calculated are the compressional (Vp) and shear (Vs1, Vs2) wave phase velocities and the degree of shear wave splitting for a given direction, represented as either the absolute difference in shear wave velocities ($dVs = Vs1 – Vs2$) or the shear wave anisotropy (AVs), which is conventionally calculated via (e.g. Mainprice & Silver 1993),

$$ AVs\% = 100(Vs1 – Vs2)/[(Vs1 + Vs2)0.5]. $$

(4)

In addition, the absolute anisotropy of Vp, Vs1 and Vs2 have been calculated by substituting their appropriate maximum and minimum values for Vs1 and Vs2 respectively in (4). Finally, as shear wave splitting analysis of real data estimates the degree of splitting and the orientation of the fast shear-wave for a given ray direction (e.g. Kendall 2000), the polarisations of the fast shear waves (Vs1) are also shown in stereographic projection.

The trigonal crystal symmetry of quartz is reflected in its seismic property distributions, although AVs (and dVs) also exhibit 6-fold...
Table 1. Summary of Torridon shear zone auto-EBSD experiments (see Lloyd, in press, for further details)

<table>
<thead>
<tr>
<th>DETAILS</th>
<th>JOB (format: username-day-month-year)</th>
<th>Shear zone wall rock</th>
<th>Mature mylonite</th>
<th>Mature mylonite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Specific comments</td>
<td>*blown filament; ended by user break</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Area analysed</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dimensions (mm)</td>
<td>10.0 x 13.0</td>
<td>17.0 x 6.0</td>
<td>13.5 x 7.25</td>
<td>1.5 x 1.0</td>
</tr>
<tr>
<td>No. data pts.</td>
<td>136640</td>
<td>120537</td>
<td>228825</td>
<td>150001</td>
</tr>
<tr>
<td>Grid step (µm)</td>
<td>50</td>
<td>25</td>
<td>20</td>
<td>1</td>
</tr>
<tr>
<td>Index rate (sec/ebsp)</td>
<td>1.1</td>
<td>0.88</td>
<td>0.65</td>
<td>0.53</td>
</tr>
<tr>
<td>Minerals indexed</td>
<td>quartz and plagioclase (An16)</td>
<td>quartz</td>
<td>quartz</td>
<td>quartz</td>
</tr>
<tr>
<td>MAD</td>
<td>&lt;1.5</td>
<td>&lt;1.5</td>
<td>&lt;1.5</td>
<td>&lt;1.5</td>
</tr>
<tr>
<td>% Low BC</td>
<td>*46.6</td>
<td>2.2</td>
<td>5.1</td>
<td>1.3</td>
</tr>
<tr>
<td>% Low BN</td>
<td>0.1</td>
<td>13.0</td>
<td>0.2</td>
<td>2.1</td>
</tr>
<tr>
<td>% Not indexed</td>
<td>27.0</td>
<td>68.5</td>
<td>87.2</td>
<td>74.8</td>
</tr>
<tr>
<td>% Indexed</td>
<td>26.19</td>
<td>16.21</td>
<td>7.53</td>
<td>21.77</td>
</tr>
<tr>
<td>No. indexed</td>
<td>35789</td>
<td>19541</td>
<td>17233</td>
<td>32651</td>
</tr>
<tr>
<td>No. phases</td>
<td>2</td>
<td>1</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Phase 1</td>
<td>quartz</td>
<td>quartz</td>
<td>quartz</td>
<td>quartz</td>
</tr>
<tr>
<td>Total good pts</td>
<td>24881</td>
<td>19541</td>
<td>17233</td>
<td>32651</td>
</tr>
<tr>
<td>Good pts %</td>
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<td>16.21</td>
<td>5.53</td>
<td>21.77</td>
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<tr>
<td>Vol. fraction %</td>
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<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
</tr>
<tr>
<td>Phase 2</td>
<td>plagioclase</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Total good pts</td>
<td>10908</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Good pts %</td>
<td>7.98</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Vol. fraction</td>
<td>30.48</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Key: MAD, mean angular deviation; BC, band contrast; BN, band number; in all but G030100 plagioclase was ignored

Symmetry close to the basal plane (Fig. 4c). Single crystal quartz is highly anisotropic in terms of all of its seismic properties. The seismic property distributions for plagioclase are more complex, reflecting the triclinic crystal symmetry, although a seismic symmetry plane normal to b (010) is apparent (Fig. 4e). Single crystal plagioclase is almost as anisotropic as quartz in terms of all of its seismic properties.

Elastic parameters are known to vary with composition. It is likely therefore that seismic properties for plagioclase vary with anorthite (An) content. The experimentally determined density and elastic stiffness matrix data used to derive Fig. 4e was for plagioclase An16 specifically because electron microprobe analysis of the plagioclase in the Torridon shear zone showed a composition range of An15-23. Babuska and Cara (1991, table 3-1) have summarized experimentally measured seismic properties for four other plagioclase compositions (An0, An29, An53 and An100). In an attempt to constrain quantitatively the variations in seismic properties with plagioclase composition, the four sets of experimental values together with the single crystal values for An16 derived above have been plotted in Fig. 5. The velocities vary linearly across the range of compositions, showing slight to moderate increases with An content. In contrast, both the anisotropy of Vp and shear wave splitting are almost invariant with An content.

Fig. 5. Variation in seismic properties (Vp, AVp, Vs1, Vs2, AVs – see text for details) and density (ρ) with An content in plagioclase (n.b. anisotropy represented from 0 - 1 rather than 0 - 100%). Data for An10, An20, An11, An10 from Babuska & Cara (1991); Data for An16, derived from LPO-averaged values using experimentally measured elastic parameters (see Table 2). Broken lines are best fit trends through and R² is the correlation coefficient. The shaded box indicates the range of An values measured in Torridon shear zone.
Polycrystal seismic properties

Mineral grains in a rock contribute to the overall seismic properties according to their single crystal elastic parameters, crystallographic orientation distribution (LPO) and volume fraction. However, although the effect of LPO on seismic wave velocities has long been recognized (e.g. Kaarsberg, 1959; Hess, 1964), the significance of LPO on the seismic properties of rocks depends on how the seismic velocities are related to the crystal directions and the origin of the LPO. There must be a significant degree of crystal alignment to produce a seismic anisotropy, whilst randomly oriented crystals generate an isotropic bulk rock. Most LPO causing mechanisms (e.g. dislocation creep) considered in seismic studies are induced by tectonic stresses in highly deformed rocks (e.g., Babuska and Cara 1991; Blackman et al. 1996).

The methodology used to calculate polycrystal seismic property distributions is the same as that employed in the single crystal calculations described above. For each mineral grain orientation, measured via SEM/EBSD, the single crystal parameters (Table 2b) need to be rotated into the sample reference frame using (3), such that the elastic parameters of the polycrystal are derived by integration over all possible orientations in the three-dimensional orientation distribution function. The three seismic phase velocities (Vp, Vs1 and Vs2), and related anisotropy (AVp, AVs), can be obtained then in every direction via solution of (2). However, due to stress/strain compatibility assumptions, three different averaging schemes are possible. The Voigt (V) average (Voigt 1928) assumes a constant strain approximation, whilst the Reuss (R) average (Reuss 1929) assumes a constant stress approximation in calculating the compliance tensor, the inverse of the stiffness tensor (e.g. Crosson & Lin 1971). The Voigt and Reuss averages represent idealized situations and provide upper and lower bounds respectively for the real elastic parameters (Bunge et al. 2000). Thus, the mean or Hill (H) average (Hill 1952) of the two values is often taken as the best estimate (VRH) of the average elastic parameters.

Finally in this section, it must be emphasised that the prediction of seismic properties using the LPO averaging approach requires accurate data on single crystal elastic parameters of different minerals. Such data are usually determined by measuring the ultrasonic velocities of single crystals in different directions and deriving the elasticity tensor via the Cristoffel equation (e.g. McSkimin et al. 1965), although a more recent approach uses Brillouin scattering and is more accurate (e.g. Sinogeikin & Bass 2000). Nevertheless, relevant data for many minerals are still lacking, although the situation is rapidly improving (e.g. Ji et al. 2002). Furthermore, single crystal elastic properties vary with pressure (P) and temperature (T) but the PT-derivative values are often unavailable. Following the current convention established for mantle minerals (e.g. Mainprice et al. 2000), we assume the elastic parameters of quartz and feldspar determined under ambient conditions (McSkimin et al. 1965; Aleksandrov et al. 1974).
The procedures described in the previous section have been applied to derive the LPO-dependent seismic properties of the Torridon mylonitic shear zone. In general, quartz dominates the shear zone region in terms of modal content, although plagioclase is common in the wall rock (see Fig. 1a). Thus, whilst the seismic properties of the mylonitic shear zone are likely to be derived from the quartz LPO alone, those for the wall rock represent a combination of quartz and plagioclase LPO. The relevant LPO have been combined with the single crystal elastic parameters of quartz and/or feldspar in their appropriate modal proportions, as measured by SEM/EBSD, to determine the seismic property distributions for different regions of the Torridon simple shear zone.

The first step is to derive the elastic parameters (stiffness matrix) for each region from the relevant LPO using the VRH averaging approach. Results are shown in Table 3. These data are used then to derive the seismic velocities and polarisations in three dimensions for each region by solving (2). Results are shown as stereographic projections oriented according to the sample reference frame (Fig. 6). The maximum and minimum phase velocities and anisotropy for each region, and also for single crystal quartz and feldspar (based on Fig. 4), are summarized in Fig. 7, whilst Table 4 summarizes the principle orientation relationships.

**Table 4. Summary of single crystal and Torridon shear zone seismic properties (see Figs. 6 and 8)**

<table>
<thead>
<tr>
<th>Property</th>
<th>Single crystal</th>
<th>SZ wall rock</th>
<th>SZ margin</th>
<th>Mature mylonite</th>
<th>Mylonitic detail</th>
</tr>
</thead>
<tbody>
<tr>
<td>max.</td>
<td>Q: // z</td>
<td>Q: broad // c</td>
<td>Q: broad // c</td>
<td>Q: sub-// c, r, z</td>
<td>Q: sub-// c, r, z</td>
</tr>
<tr>
<td>P: sub-// (101)</td>
<td>P: broad // (001)</td>
<td>T: broad sub-// σc</td>
<td>T: broad sub-// σc</td>
<td>T: sub-// Z</td>
<td>T: sub-// Z</td>
</tr>
<tr>
<td>Vp</td>
<td>Q: sub-// r</td>
<td>Q: // a, z ?</td>
<td>Q: // a &amp; basal plane</td>
<td>Q: // a &amp; basal plane</td>
<td></td>
</tr>
<tr>
<td>min.</td>
<td>P: // (001)</td>
<td>P: sub-// (100)</td>
<td>T: sub-// Y</td>
<td>T: // X, sub-// XY</td>
<td>T: // X, sub-// XY</td>
</tr>
<tr>
<td></td>
<td>max. P: sub-// (110)</td>
<td>T: // σ1, sub-// Y</td>
<td>T: sub-// σc, Y</td>
<td>T: // X, Y, Z</td>
<td>T: // X, Y, Z</td>
</tr>
<tr>
<td>Vs1</td>
<td>Q: // z</td>
<td>Q: broad sub-// a, m ?</td>
<td>Q: ?</td>
<td>Q: // z small circle</td>
<td>Q: // r, z small circle</td>
</tr>
<tr>
<td>min.</td>
<td>P: sub-// (101)</td>
<td>P: broad sub-// (010)</td>
<td>T: ?</td>
<td>T: // σ1 ?</td>
<td>T: // σ1 &amp; σ3 ?</td>
</tr>
<tr>
<td>Vs2</td>
<td>Q: // a</td>
<td>Q: // a ?</td>
<td>Q: // a</td>
<td>Q: // basal plane</td>
<td>Q: // basal plane</td>
</tr>
<tr>
<td>min.</td>
<td>P: // (101)</td>
<td>P: // (010)</td>
<td>T: sub-// Y</td>
<td>T: between X &amp; Y</td>
<td>T: sub-// XY</td>
</tr>
<tr>
<td></td>
<td>max. P: sub-// (110)</td>
<td>P: sub-// (010)</td>
<td>T: sub-// Y</td>
<td>T: between X &amp; Y</td>
<td>T: sub-// XY</td>
</tr>
<tr>
<td>AVs</td>
<td>Q: // c</td>
<td>Q: sub-// a, c ?</td>
<td>Q: sub-// a, c ?</td>
<td>Q: // basal plane</td>
<td>Q: // basal plane</td>
</tr>
<tr>
<td>and dVs</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Q: // e</td>
<td>Q: sub-// m ?</td>
<td>Q: sub-// m ?</td>
<td>Q: sub-// c &amp; r/t small circle</td>
<td>Q: sub-// c &amp; r/t small circle</td>
</tr>
<tr>
<td>min.</td>
<td>P: // (100)</td>
<td>P: sub-// (100), (001)</td>
<td>T: sub-// X</td>
<td>T: sub-// Z</td>
<td>T: sub-// Z</td>
</tr>
</tbody>
</table>

Q, quartz crystal directions (c, m, a, r, z); P, plagioclase crystal directions ((100), (010), (001)); T, tectonic axes (XZYZZ); σc and σt, inferred maximum and minimum principle stress directions, parallel to Qr (NW-SE) and Qr (NE-SW) respectively; //, parallel to; note, only shear zone wall rock contains significant plagioclase

Torridon shear zone seismic properties

The maximum and minimum velocities in Vp are fairly constant across the shear zone and range from 6.40-6.47 km/s and 5.66-5.79 km/s respectively (Figs. 6 and 7a), representing variations of only 1-2%. There is a suggestion of a slight drop in both velocities from the relatively undeformed wall rock and shear zone margin regions to the mature mylonite. However, the difference between Vp maximum and minimum values (i.e. the compressional wave anisotropy, AVp) increases from 11% in the shear zone wall rock to 12.5% in the mylonite (Figs. 6 and 7b), although the maximum value (12.8%) occurs in the
Figure 6. Seismic property distributions (Vp, Vs1, Vs2, AVs, dVs, Vs1 polarisation – see text for details) for different parts of Torridon shear zone based on LPO-averaging of elastic parameters constructed using Tables 2 and 3 and programs Anis2Ck and VpG2k (Mainprice 2003). Only shear zone wall rock includes quartz and plagioclase, the rest involve quartz alone. All pole figures equal area, upper hemispheres viewed towards ENE; contours multiples of mean uniform distribution, with tectonic X and Z directions oriented east-west and north-south respectively. See text for discussion. (a) Shear zone wall-rock (G030600). (b) Shear zone margin (G040800). (c) Mature mylonite (G050800). (d) Detail mature mylonite (G050800).

Shear zone margin.

Maximum and minimum velocities in Vs1 range from 4.11-4.50 km/s and 3.83-4.02 km/s respectively, representing variations of 8.6% and 4.7% in each. The difference between Vs1 maximum and minimum values ranges from 7-
15%. Maximum and minimum velocities in Vs2 range from 3.93-4.38 km/s and 3.70-3.77 km/s respectively, representing variations of 10% and 2% in each. The difference between Vs2 maximum and minimum values ranges from 4-15%. Both shear wave velocities increase from the shear zone wall rock into the margin and mature mylonite (Fig. 6 and 7a). The difference, or anisotropy, between the maximum and minimum values of Vs1 increases from 7% in the wall rock to 14.7% in the margin, before decreasing to 9-10% in the mylonite (Fig. 6 and 7b). In contrast, the anisotropy of Vs2 increases progressively from 6.0% in the wall rock to 15.0% in the mylonite.

Obviously, the behaviour of shear waves is reflected in the shear wave splitting, measured as AVs or dVs. This is typically large across the shear zone and increases from 9.49% or 0.4 km/s in the wall rock to 16.15% or 0.7 km/s in the mylonite (Figs. 6 and 7b).

**Orientations**

The seismic property orientation distributions, and in particular the orientations of their maximum and minimum values, relative to the LPO and tectonic reference frame are crucial in terms of explaining and using seismic properties in geodynamic interpretations (see Figs. 6 and 8, and Table 4).

The seismic property orientation distributions for the shear zone wall rock and margins regions (Fig. 6a, b) result from relatively weak LPO development (Fig. 2a, b) and, in the case of the wall rock, the contrasting impact of quartz and plagioclase LPO. Nevertheless, the seismic properties do show some consistent orientation patterns, particularly relative to the quartz LPO and tectonic reference frame. The maximum in Vp is aligned sub-parallel to the quartz c-axis maximum and inferred position of σ1, whilst the minimum in Vp is sub-parallel to the tectonic Y direction (and perhaps also the inferred position of σ3 in the wall rock). The maximum in Vs1 has a broad distribution that encompasses the quartz c and r maximum directions, the tectonic Y direction and the inferred position of σ1, but the orientation of the minimum in Vs1 is poorly defined relative to crystal or tectonic directions. The maximum in Vs2 aligns also sub-parallel to the quartz c-axis maximum (and perhaps the inferred position of σ1), whilst the minimum in Vs2 appears to align parallel to the quartz a-axis maximum and sub-parallel to the tectonic Y direction. The orientation distributions of AVs (and dVs) have maxima oriented sub-parallel to quartz a-axis concentrations and the tectonic Y direction and...
minima oriented sub-parallel to the tectonic X direction and perhaps quartz m direction concentrations.

In contrast to the shear zone wall rock and margin, seismic property orientation distributions in the mature mylonite are exceptionally well ordered (Fig. 6c, d) and clearly reflect the quartz LPO (Fig. 2c, d). Several of the relationships recognized in the shear zone wall rock and margin regions persist into the mylonite. A broad maximum in Vp encompasses the quartz c-axis maximum and r/z direction small circle distributions and is sub-parallel to the tectonic Z direction. Vp has an absolute minimum parallel to the a-axis maximum and tectonic X direction and a great circle distribution of low values parallel to the quartz basal plane and sub-parallel to the XY tectonic plane. Although the absolute maximum in Vs1 is sub-parallel to the tectonic Y direction, similar values occur parallel to both the quartz basal plane and the c-axis maximum and hence encompass all three tectonic directions. The minima in Vs1 form small circle distributions that appear to reflect the r/z LPO (the absolute minimum in Vs1 may indicate the inferred directions of both σ1 and σ3). The maximum in Vs2 is parallel to the quartz c-axis maximum and sub-parallel to the tectonic Z direction, whilst minimum values in Vs2 occupy a great circle distribution parallel to the quartz basal plane and sub-parallel to the XY tectonic plane. Maximum values in AVs and dVs occupy great circle distributions parallel to the quartz basal plane and sub-parallel to the XY tectonic plane, whilst minimum values occur over broad areas that encompass the quartz c-axis maximum and r/z small circle distributions.

The polarisation behaviours of the fast shear waves (Vs1) for the three shear zone regions are somewhat similar, particularly for vertical wave propagation. More specifically, those for the shear zone margin rock (Fig. 6b) and the mylonite (Fig. 6c-d) are very similar.

Discussion

The results of the LPO-averaged seismic property determinations for different regions in the Torridon shear zone have significant implications for the interpretation and use of seismic velocities and anisotropy in geodynamic analysis. Firstly, it is worth comparing the single crystal seismic properties of quartz and feldspar with the LPO-averaged results obtained from the shear zone rocks to see how the former become modified in the latter. Secondly, the relationships between the LPO-averaged seismic properties and shear zone structure and tectonics are briefly considered, including the potential for using LPO-averaged seismic property results in seismic waveform modelling. Finally, the potential impact of the grain boundary microstructure elements on shear zone seismic properties is briefly considered.

Comparison between single crystal and polycrystal seismic properties

There are significant differences between the seismic properties of quartz and plagioclase single crystals and those of the three shear zone regions. The trigonal symmetry anisotropy parallel to the c-axis clearly seen in single crystal quartz seismic properties (see Fig. 4a, c) is dispersed by the LPO-averaging effect in polycrystal aggregates (Fig. 6). The compressional and shear wave velocities show wider variations (higher anisotropy) for the single crystals compared to the shear zone regions (Fig. 7). This is the most important difference between single crystal and polycrystal behaviours.

In terms of orientation, the overall effect of shear zone LPO (Fig. 2) is to either displace and/or disperse (see Fig. 6 and Table 4) the unique maximum and minimum seismic orientations recognized for the single crystals (Fig. 4). For example, the sharp orientation maximum in Vp parallel to the quartz single crystal (z) pole becomes broadened and parallel to the c-axis and/or {z}. Similarly, the quartz single crystal maxima in Vs1, AVs and dVs parallel to the a-axes are broadened into a great circle distribution parallel to the quartz basal plane.

The polarisations of the fast shear waves (Vs1) for all three shear zone regions (Fig. 6) are very different to those of the single crystals and particularly quartz (Fig. 4c, e). This is due to the fact that although the c-axes of quartz crystals are highly aligned (Fig. 2c-d), the other crystallographic directions are more dispersed in orientation. The polycrystal elasticity therefore is not trigonal in symmetry.

Relationship between shear zone structure, tectonics and seismic properties

In general, the results of the LPO-averaging of shear zone seismic properties suggests that deformation (increasing strain) during shear zone formation and mylonitisation modifies the absolute seismic velocity values and increases the seismic anisotropy of the original wall rock (Fig. 7). To consider further the relationship between shear zone structure, tectonics and seismic properties, a summary stereographic projection of the orientations (see Table 4) of structural (i.e. LPO), tectonic (i.e. X≥Y≥Z) and seismic (i.e. Vp, Vs1, Vs2, AVs, dVs) properties in the field (i.e. geographic) reference frame is shown in Fig. 8. Recall that the mylonitic shear zone has a strong LPO (Fig. 2c, d) that approximates a dauphine twinned single crystal configuration, with c-axis maximum sub-parallel to Z and sub-normal to the...
Figure 9. Cross sections through slowness and wave surfaces for single crystals and shear zone regions. In each, left-hand-side shows slowness surfaces for each wave, including directions of particle motion (arrows); note, P-wave is innermost surface. The right-hand-side shows wave surfaces and can be viewed as a snapshot of the wavefront after 1 second; note ‘folding’ (F) on some S-wavesheets, which is never seen for P-waves. (a) Quartz single crystal parallel to plane normal to \( m \{010 \} \); note symmetry about quartz basal plane parallel to \( <a> \) direction. (b) Quartz single crystal parallel to plane normal to \( a \{110 \} \); note asymmetry about quartz basal plane that reflects maximum in \( V_p \) parallel to \( z \) and minimum in \( V_p \) between \( r \) and \( m \). (c) Plagioclase single crystal parallel to plane normal to \( b \{010 \} \); note overall (triclinic) asymmetry. (d) Shear zone wall rock, XZ section. (e) Shear zone margin, XZ section (note folding). (f) Mature mylonite, XZ section.

The results of the LPO-averaging of the seismic properties (Figs. 6 and 7) suggest that within the mylonitic shear zone the maximum in foliation (XY) plane and \( a \)-axis maximum parallel to X.
Vp and the minima in Vs2, AVs and dVs tend to align parallel to the c-axis maximum (i.e. normal to the mylonitic foliation), whilst the minimum in Vp aligns parallel to the a-axis maximum and hence defines the extension (X) direction. High values of Vs1, AVs and dVs and low values of Vp define the foliation (XY) plane, whilst the minimum in Vs1 either aligns sub-parallel to \( \{z\} \) or defines a weak small circle distribution about the c-axis maximum that includes \( \{z\} \).

The relationships between structure, tectonics and seismic properties described above may prove useful in the tectonic interpretation of seismic measurements. However, it is clear from the geographic representation of the various parameters in Fig. 8 that any analysis of seismic waves (e.g. see below) must be interpreted in terms of the appropriate structural and tectonic orientations in the field.

**Seismic waveform modelling**

Wave propagation is considerably more complicated in anisotropic media than it is in isotropic media. For example, wavefronts in homogeneous anisotropic media are no longer spherical and the direction of particle motion, ray direction and wavefront normal are in general not aligned. Furthermore, two shear-waves propagate in anisotropic media and they may exhibit folding, which leads to traveltime triplications (e.g. the fast shear wave arrives at three different times at a receiver oriented parallel to the a-axis of quartz, see Fig. 9a). Figure 9 illustrates these effects, showing cross-sections of the slowness and wave surfaces for the minerals and rocks of the Torridon shear-zone parallel to specific crystal (i.e. \(<\overline{1100}\>\) and \(<\overline{1120}\>\)) and tectonic (i.e. XZ) planes.

It is interesting to consider how seismic waves propagate through the shear zone and how they are affected by the enhanced deformation and hence seismic anisotropy. Teleseismic and regional earthquakes provide a passive means of imaging the deep crust with relatively low frequencies and hence low resolution. In contrast, wide-angle controlled-source surveys give better resolution. Shear-wave splitting is the most unambiguous indicator of anisotropy. However, whether or not this is observable depends on the magnitude of the anisotropy and its spatial extent. The rocks of the Torridon shear-zone show varying degrees of splitting (e.g. see Figs. 6 and 7), but in general the anisotropy must be uniform and persist over a large region with respect to seismic wavelength in order to accrue a measurable level of splitting. In earthquakes studies such a region must be kilometres in extent and therefore is only likely to be observed for regional scale shear zones or rock masses with constant petrofabric characteristics (e.g. gneisses). Fortunately, reflections and mode conversions at interfaces provide better vertical resolution (perhaps 10’s m for controlled source experiments) and can be very sensitive to changes in anisotropy. High degrees of anisotropy can enhance the seismic contrast across a boundary, both in terms of absolute values of anisotropy and also if the directional characteristics of the anisotropy change across the boundary.

If the crystals in the shear-zone rocks are randomly oriented, their elastic parameters are isotropic. Figure 10a, b shows the P-wave reflections and P-to-S-wave conversions at boundaries between such isotropic assemblages. These averages are based on the LPO-averaged elastic parameters for the Torridon rocks (see Tables 2, 3 and 5). The reflections are very weak, especially at near offsets (small angles of incidence).

Crystal alignment and hence anisotropy affects the reflection properties in a number of interesting ways. Figure 10c-h show the reflection coefficients for anisotropic interfaces as predicted from the petrophysical analysis described above. In general, more energy is reflected in the anisotropic case. This is due to higher vertical velocities in the underlying more-deformed layers. There is a significant level of mode-converted reflections at these interfaces. Both the fast and slow shear-waves are generated in the conversion, due to the complex anisotropic symmetries. There is a surprising amount of reflected converted-wave energy at normal incidence, which never exists in the isotropic case. This is due to the strong difference in the directions of group velocity (ray direction) and phase velocity (normal to the wavefront), again due to the low order of symmetry and strong anisotropy. There are also clear azimuthal variations in the reflections. This suggests that multi-azimuth wide-angle reflection

### Table 5. Isotropic aggregate properties of Torridon shear-zone rocks calculated from the measured elastic properties in Tables 2 and 3.

<table>
<thead>
<tr>
<th>Property</th>
<th>SZWR</th>
<th>SZM</th>
<th>MM</th>
</tr>
</thead>
<tbody>
<tr>
<td>P-wave velocity (km/s)</td>
<td>4.298</td>
<td>4.291</td>
<td>4.290</td>
</tr>
<tr>
<td>S-wave velocity (km/s)</td>
<td>2.748</td>
<td>2.911</td>
<td>2.906</td>
</tr>
<tr>
<td>Density (gm/cm³)</td>
<td>2.636</td>
<td>2.647</td>
<td>2.647</td>
</tr>
</tbody>
</table>

SZWR, shear zone wall rock; SZM, shear zone margin; MM, mature mylonite.
Figure 10. Seismic reflection coefficients. (a) and (b) Consider interfaces between isotropic aggregates (see Table 5) for P-wave reflections and P-S mode-converted reflections respectively. Solid line shows coefficients for isotropic shear-zone wall rock and isotropic shear-zone margin interface, whilst (barely visible) dashed line shows coefficients for interface between shear-zone wall rock and isotropic mylonite. (c) – (e) Consider an interface between anisotropic shear-zone wall rock and shear-zone margin (n.b. no azimuthal variation in reflection coefficient in isotropic cases; see (c) for key to azimuthal directions). (c) P-wave reflections as function of incidence angle and azimuthal angle. (d) Mode-converted reflections for P-S1-waves. (e) Mode-converted reflections for P-S2-waves. (f) – (h) Consider an interface between anisotropic shear-zone wall rock and mature mylonite (n.b. no azimuthal variation in reflection coefficient in isotropic cases; see (c) for key to azimuthal directions). (f) P-wave reflections as function of incidence angle and azimuthal angle. (g) Mode-converted reflections for P-S1-waves. (h) Mode-converted reflections for P-S2-waves.

data can be potentially used to study the sense of deformation in deep-rooted shear zones (see also Burlini et al. 1998; Khazanehdari et al. 1998). Such reflection effects have been used also to map fracture patterns in oil reservoirs (Hall and Kendall 2003).

Grain boundary microstructure

In addition to LPO, a number of other microstructural elements are known to impact on elastic anisotropy, such as cracks, fractures, pores, grain morphology and shape preferred orientation, and layering. The impact of many of these elements on elastic anisotropy has recently been considered by Wendt et al. (2003) and further discussion here is largely inappropriate. However, it is worth considering one particular microstructural element, the role of grain boundaries, that impacts directly on the present study.

Grain boundaries are important microstructural elements in rocks and may contribute a significant volume fraction, particularly in fine grained rocks such as mylonites. They represent regions of relatively high lattice distortion and defect concentration that form extended three-dimensional
networks and are likely to have different elastic parameters from those of the grains. In addition, as rock elasticity is dependent also on cohesion forces between grains (Kaarsberg, 1959), the contact area between grains should be considered. Furthermore, grain boundaries may form crack-like discontinuities in some rocks and consequently influence seismic anisotropy if the boundaries are preferentially aligned (e.g. Babuska and Cara, 1991). Thus, many rock properties, including the elastic parameters, may be influenced by the grain boundary network.

Unfortunately, there is little known about the exact impact of grain boundaries on seismic anisotropy (e.g. Wendt et al. 2003). However, a recent interpretation of the microstructural and petrofabric evolution of the Torridon shear zone (Lloyd in press) considered the formation and orientation of quartz grain boundaries. It was suggested that the grain boundary network consists of tilt boundaries parallel to quartz prism (and hence often YZ tectonic) planes and twist boundaries parallel to the XY foliation/basal-crystal planes and XZ tectonic plane (see Fig. 3).

The physical presence of grain boundaries was not considered in the LPO-averaging analysis described here and hence any direct impact due to grain boundary configuration on seismic properties cannot be assessed. However, comparison between grain boundary and seismic property orientations (Fig. 8; see also Fig. 6) may provide significant indications, as follows.

Twist boundaries that developed parallel to the XY foliation/basal-crystal planes due to a combination of \{\pi\}\!<a> + \{(c)\!<a> slip could contribute to the great circle distribution of high AVs, dVs and Vs1 and low Vp values. In addition, the broad concentration of high/maximum Vp, Vs2 and Vs1 and low/minimum AVs and dVs values parallel to the quartz e-axis maximum and sub-parallel to the tectonic Z direction may be influenced by two characteristics of the grain boundary network: 1. the intersection of tilt boundaries that develops due to a combination of dauphine twinning and/or slip on systems that exploit the quartz \!<a> axis as the slip direction; and 2 twist boundaries that develop parallel to the XZ tectonic plane due to combined slip on \{\pi\}\!<a> + \{m\}<a> systems. Similarly, the tendency for the minimum in Vp to occur parallel to the quartz a-axis maximum and sub-parallel to the tectonic X direction may be augmented by the intersection of twist boundaries that develop parallel to the XY foliation/basal crystal planes and XZ tectonic plane due to combined slip on \{\pi\}\!<a> + \{(c)\!<a> and \{\pi\}\!<a> + \{m\}\!<a> systems respectively. Thus, it appears that grain boundary configuration has the potential to impact on seismic property characteristics but much further work (i.e. direct measurements) is needed.

Conclusions

This contribution has explored the link between petrofabric (LPO) and petrophysical (seismic) properties of a quartzo-feldspathic shear zone developed under conditions typical of the mid-lower crust. Although it is clear that shear zone formation and petrofabric evolution from an initially quartzo-feldspathic gneissic wall rock to a mature quartz mylonite is reflected in the seismic properties, the precise impact of LPO on seismic properties is not simple. Several conclusions can be drawn, based on these observations.

1. The most obvious control of LPO on shear zone seismic properties results in the following relationships: Vp-min parallel to the a-maximum (i.e. tectonic X); AVs-max parallel to mylonitic foliation (i.e. XY mylonitic foliation plane); Vp-max and AVs-min parallel the c-maximum (i.e. foliation normal, Z); and Vs1-min parallel to the z-maxima.

2. The development of a strong LPO during shear zone evolution, and particularly during mylonitisation, is responsible for considerable seismic anisotropy. P-wave anisotropy varies between 11-13%, the fast and slow shear-waves show 6-15% anisotropy, and the magnitude of shear-wave splitting is 9.5-16%.

3. Although the degree of shear wave splitting exhibited by the shear zone due to LPO requires considerable thicknesses of rock (depending on the dominant wavelength of the source) with constant LPO characteristics before it becomes seismically visible, reflections and mode conversions provide much better resolution, particularly across boundaries (e.g. between the shear zone margin and the mature mylonite).

4. The low symmetry and strong anisotropy due to the petrofabric suggest that multi-azimuth wide-angle reflection data may be useful in the determination of the deformation characteristics of deep shear zones.

5. How well seismic data can be ultimately used to study anisotropy and hence lower crustal deformation remains to be seen. Data quality and spatial coverage will be crucial.

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