The calculated seismic properties of quartz mylonites with typical fabrics: relationship to kinematics and temperature

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SUMMARY
Quartz mylonites produced by intense ductile deformation in shear zones are often invoked as seismic reflectors in the lower continental crust. The seismic properties ($V_p$, $V_s$, $\Delta V_s$ and polarization planes of $V_s$ and $\Delta V_s$) have been calculated for five quartz mylonites that display typical lattice preferred orientations (LPOs). All specimens display considerable seismic anisotropy with $V_p$ anisotropy coefficients of between 8 and 12 per cent. There are good correlations between the LPO of the $a$-axes and the minimum $V_p$ direction, and the $c$-axes and the maximum $V_p$ direction. The other seismic properties are not reliable indicators of LPO. As the $a$-axis is the major flow direction in quartz, a correlation between the ductile flow direction and the $V_p$ minimum can be made.

With increasing temperature and pressure alpha-quartz transforms to beta-quartz. In the beta field quartzites are seismically transverse isotropic. Furthermore there is a considerable increase in $V_p$ from 6 km s$^{-1}$ in the alpha field to nearly 7 km s$^{-1}$ in the beta field. The transition is instantaneous and hence will result in a sharp seismic discontinuity with a reflection coefficient of $-0.08$ between pure aggregates of alpha and beta phases. However, under typical conditions of the continental crust the seismological importance of this phase transition will be limited by the presence of secondary minerals and partial melting.

Key words: quartz mylonites, seismic properties, temperature effects.

INTRODUCTION
Over the last few years the in situ seismic characteristics of the lower continental crust have been measured by seismic reflection profiling (BIRPS, CALCRUST, COCORP, CROP, DECORP, ECORS, NSF20: Mooney & Brocher 1987). The geological interpretation of the profiles is based on the geometry of the reflectors and seismic characteristics ($V_p$, $V_s$, $\Delta V_s$) of lower crustal rocks defined by laboratory measurements (e.g. Christensen 1965; Fountain 1976; Kern & Richter 1981), and constrained, in some cases, by additional geophysical measurements such as gravity (e.g. Fuchs et al. 1987; Bayer et al. 1989).

The characterization of seismic properties by laboratory measurement is usually limited to three orthogonal directions as a function of hydrostatic pressure. In some cases the measurements are made as a function of temperature (e.g., Kern 1978); however it is very rare that measurements are made in more than three directions (Christensen & Crosson 1968; Seront, Mainprice & Christensen 1989). Using lattice preferred orientation (LPO) data it is possible to calculate the elastic constants of the rock and hence the seismic properties in any direction (e.g. Crosson & Lin 1971; Peselnick, Nicolas & Stevenson 1974). The LPO-derived seismic properties may be extrapolated to any temperature and pressure if the appropriate derivatives of the single crystal elastic constants are known. The aggregate elastic constants provide an ideal basis for the description of the 3-D seismic properties in synthetic seismograms (Gajewski & Pšenčík 1988) for a more quantitative comparison with seismic profiles in several directions (Lueschen et al. 1987). A further advantage of the LPO technique is the possibility to investigate the correspondence between crystallographic pole figures and the directional variation of seismic properties. It is now well established that LPO can provide reliable kinematic indicators such as sense of shear, coaxial or non-coaxial deformation [see Mainprice & Nicolas (1989) for a review], hence the establishment of a relationship between seismic anisotropy and LPO will help to establish a link between seismic properties and flow kinematics. A link between seismic anisotropy and the direction of plastic flow has been
established for the upper mantle [see review by Nicolas & Christensen (1987)]. Such a correlation may be possible in the continental crust; however the greater mineralogical diversity and heterogeneity of deformation (e.g. folds) may render a correlation difficult on a regional scale. In this paper we present the seismic properties of quartz mylonites that show microstructural evidence of extensive plastic deformation and have typical c-axes pole figures. The deformation mechanisms (dislocation glide systems) of these specimens have been previously studied by Schmid & Casey (1986). We present the fabrics, their seismic properties, and the possible correlation between seismic properties and the kinematics of deformation.

FABRICS

Because a detailed presentation of these samples has been given in Schmid & Casey (1986), we will restrict ourselves here to a brief description of the pole figures. For further details the reader is referred to their paper. The complete LPO of the samples was determined at ETH Zurich by X-ray texture gonimetry using eight pole figures (10.2), (10.1), (20.1), (10.0), (21.1), (11.0), (11.1) and (11.2). The appropriate corrections were made for defocusing in the reflection mode. The data were processed using a set of FORTRAN computer programs developed and described by Casey (1981). The Orientation Distribution Functions (ODFs) were calculated by pole figure inversion using the spherical harmonic method described by Bunge & Wenk (1977) and Bunge (1982). All pole figures (Fig. 1) are presented with the foliation aligned E–W (vertical) and the lineation horizontal.

![Figure 1](image_url)

Figure 1. Pole figures of the c-axes (c), second-order prisms (a), first-order prisms (m), combined positive and negative rhombs (r+z), positive rhombs (r), and negative rhombs (z). All pole figures calculated from the ODF. The contours are given in multiples of a uniform distribution, the contour interval is 0.5 or 1.0 and the order of expansion used is 8–12 (see figures below specimen label). Stippled areas have a density of less than 0.5 or 1.0 respectively. Foliation is N–S (horizontal line) and the lineation E–W (black squares).

![Figure 2](image_url)

Figure 2. A schematic fabric transition diagram showing the variation of the critical resolved shear stress ($\tau_{crt}$) to activate (a) direction, slip on rhombohedral (r), basal (c) and prism (m) planes in quartz for a constant strain rate as a function of temperature. The dashed line represents the strength and fabric transitions of a hypothetical polycrystal.

For this study, we have selected five fabrics from the specimens studied by Schmid & Casey (1986). They can be classified into two groups; a low-temperature (LT) class of c-axes girdles fabrics and a high-temperature (HT) class of point maxima c-axes pole figures (Figs 1 and 2). The LT c-axes pole figures (Fig. 1) of specimens CC1, RL8215 and RL8330 illustrate the transition in the girdle fabric between a single inclined girdle (CC1) to an asymmetric cross-girdle.
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(RL8215) to a symmetric cross-girdle (RL8233). Such fabric transitions are typical of those associated with the change from non-coaxial (CC1) to coaxial (RL8330) deformation regime that occurs with increasing sample distance from a thrust contact (Law 1987). Complementary behaviour is also shown in the a-axes pole figures with a strong asymmetric concentration near the lineation in CC1 becoming more symmetrically distributed about the foliation in RL8330. The HT fabrics tend to be dominated by point maxima c-axes poles figures because recovery processes, in particular dislocation climb, allow single slip to dominate. This process is further aided by the elimination of unfavourable orientations for glide by recrystallization processes (e.g. Bouchez & Duval 1982; Schmid & Casey 1986). These HT fabrics probably form under lower amphibolite or granulite facies conditions as evidence of hexagonal symmetry has been found in the fabric analysis [see sample Gran 133 in Schmid, Casey & Starkey (1981)] indicating the beta-quartz stability field for such fabrics. Two c-axes pole figures show a maximum normal to the lineation in the foliation (Fig. 2, P248 and Gran 133); P248 is slightly asymmetric whereas no asymmetry is observed in Gran 133, a tendency that is confirmed in the a-axes pole figures. Note the hexagonal character of the a-axes pole figure of Gran 133, also seen in the z rhombohedral pole figure (Fig. 1) (Schmid et al. 1981). P248 represents a non-coaxial deformation whereas Gran 133 appears to be perfectly coaxial.

METHOD OF CALCULATION OF SEISMIC PROPERTIES

To calculate the seismic properties from a polycrystal one needs to evaluate the elastic properties of the aggregate. In the case of polycrystal with a random fabric this is easily achieved using the classical Voigt or Reuss averages for an axial elastic property, such as the shear modulus (G). For example the Voigt average \( V_z \) is given by

\[
V_z = \sqrt{G/v/p}
\]

where \( p \) is the density,

\[
G/v = (C_{11} + C_{22} + C_{33} - (C_{12} + C_{13} + C_{23}) + 3(C_{44} + C_{55} + C_{66})/15,
\]

and \( C_{ij} \) are the elements of the single crystal elastic stiffness tensor.

In the case of an aggregate with a fabric, the anisotropy of the elastic properties of the single crystal must be taken into account for the calculation of the tensorial properties. For each volume fraction measured at an orientation \( g \) defined by the three Euler angles \( (\psi_1, \Phi, \psi_2) \), the single crystal elastic properties must be rotated into the specimen coordinate frame using a rotation matrix \( (g_{ij}) \) using

\[
C(g) = \sum_{i=1}^{n} \sum_{j=0}^{m} \sum_{k=0}^{5} \sum_{l=0}^{0} C(g^*)
\]

where

\[
C(g) = \text{elastic property in specimen coordinates},
\]

\[
g_{ij} = g(\psi_1, \Phi, \psi_2) = \text{crystal to specimen coordinates rotation matrix},
\]

\[
g = (\psi_1, \Phi, \psi_2) = \text{measured crystal orientation in sample coordinates},
\]

\[
g^* = (\psi_1 = 0, \Phi = 0, \psi_2 = 0) = \text{crystal reference orientation in sample coordinates}.
\]

Using the continuous ODF, the elastic properties are averaged over all possible orientations within the asymmetric unit defined by the crystal and specimen symmetry. The elastic constants may be calculated by integration over all orientations of the ODF, or calculated from the ODF coefficients of the harmonic method (Wenk, Johnson & Matties 1988). In the general case of triclinic crystal and specimen symmetry the integration is given by Bunge (1985) as

\[
\bar{C} = \frac{1}{8\pi^2} \int C(g)\hat{f}(g) \sin \Phi dg
\]

where \( \bar{C} \) is the elastic property of the aggregate and \( \hat{f}(g) \) is the even part of the ODF.

As the elastic properties are fourth-order centrosymmetric tensorial properties we have verified by numerical calculation that one need only expand the even part of the ODF to the fourth order of the \( L \) index. The previous statement has four important implications; first numerical calculations can be limited to the \( L \geq 4 \) of the series expansion with considerable saving of calculation time; second the number of pole figures needed to determine the ODF (with \( L \geq 4 \)) is greatly reduced (only two pole figures for trigonal crystal symmetry and triclinic specimen symmetry); third the limitation to \( L \geq 4 \) means that only the smooth or ‘long range’ part of the ODF contributes to the elastic properties; and fourth only the even part of the ODF is required. The even part of the ODF is directly obtained by pole figure inversion in the harmonic method whereas the odd part of the ODF can only be obtained by this method using various assumptions (Bunge 1982). The final step in the calculation is to evaluate the seismic velocities by solution of the Christoffel equation (e.g., Crosson & Lin 1971; Peselnick et al. 1974).

The procedures used here are capable of calculating the Voigt, Reuss or Voigt–Reuss–Hill (V–R–H) elastic averages. The Voigt average assumes a homogeneous elastic deformation throughout the polycrystal, whereas the Reuss average assumes a homogeneous stress throughout. Neither is physically realistic and Hill (1952) suggested taking the mean of the two values (V–R–H) which has no physical justification, but often produces values close to those measured in laboratory experiments. Measurements on the Twin Sisters Dunite by Crosson & Lin (1971) have shown that seismic velocities calculated using the Voigt average give the closest agreement between petrofabric derived and laboratory measured seismic velocities. A similar study on an anorthosite by Seront et al. (1989) also showed that the Voigt average was in closest agreement, hence we have elected to use Voigt averages in all calculations.

P-WAVE CHARACTERISTICS OF QUARTZ MYLONITES

The low-temperature girdle fabrics (Fig. 1) exhibit a similar \( V_p \) distribution with a broad ridge of high \( V_p \) normal to the lineation (Fig. 3). The minimum \( V_p \) corresponds to the maximum density in the a-axis pole figure. The simplest
c-axes pattern, a single girdle (CC1), gives the greatest anisotropy coefficient \[ A = \frac{(V_{\text{max}} - V_{\text{min}})}{V_{\text{max}}}/100 \] of 11.9 per cent. Of the two cross-girdle fabrics, (RL8215, RL8330) of which RL8215 has a strong asymmetric character, it is RL8330 that has an asymmetric \( V_p \) distribution. The high-temperature point maxima fabrics (Fig. 3) show a band of low \( V_p \) that correlates with the \( a \)-axes pole figure. The \( V_p \) minimum band is a plane containing the lineation and foliation normal (P248, Gran 133). The simple topological distribution of the c-axes results in relatively high anisotropy coefficients (11 to 16.4 per cent).

In broad terms we can classify these \( V_p \) distributions into two distinct patterns. The low-temperature \( V_p \) pattern with a broad band of high \( V_p \) normal to the lineation (CC1, RL8215, RL8330), and a moderate- to high-temperature \( V_p \) pattern with a point maximum approximately normal to the lineation in the foliation plane (P248, Gran 133). A good correlation can be made between the low \( V_p \) velocity direction and lineation in all fabrics.

**S-WAVE CHARACTERISTICS OF QUARTZ MYLONITES**

For \( S \)-waves, we define \( V_{S1} \) as the faster wave (Fig. 4) and \( V_{S2} \) as the slower wave (Fig. 5). The \( S \)-wave velocities are about 30 per cent slower than the \( P \)-wave velocities. The \( V_{S1} \) distributions (Fig. 4) have no obvious correlations with the pole figure topologies (Fig. 1). The \( V_{S2} \) distributions (Fig. 5) show a broad correlation with the c-axes pole figures (Fig. 1). The shear wave splitting (\( \Delta V_s \)), which is defined as the difference between the shear wave velocities (\( \Delta V_s = V_{S1} - V_{S2} \)), has complex distributions (Fig. 6) with no correlation with the pole figures. The polarization planes of the \( S \)-waves (Figs 7 and 8) also have complex distributions. In Fig. 7 the stereonets are presented in the same orientation as the previous figures, but in Fig. 8 we rotated the reference frame such that the origin is now the pole to the foliation. In this orientation one can see more clearly the polarization of \( S \)-waves with a vertical propagation direction when the foliation is horizontal in the crust. The vertical \( V_{S1} \) polarization plane (Fig. 8) is a plane that contains the c-axes maximum in fabrics with a simple c-axes topology (RL8215, P248, Gran 133), but some local complexity occurs at the origin in the other fabrics (CC1 and RL8330).

**THE EFFECT OF TEMPERATURE**

The major discontinuity in the elastic and seismic properties of quartz in the pressure range of the continental crust (0–1 GPa) is associated with the polymorphic phase transition of alpha-quartz to beta-quartz with increasing temperature. Alpha-quartz has trigonal symmetry and its...
elastic stiffness is described by six constants whereas beta-quartz is hexagonal and its elastic properties are described by five constants. Hexagonal elastic symmetry is often referred to as transverse isotropic symmetry in the seismological literature (e.g. Christensen & Crosson 1968). Kern (1978, 1979, 1982) has previously suggested the alpha–beta quartz transition has important seismological implications.

To quantitatively investigate the effect of temperature on the 3-D seismic properties of quartz we calculated the $V_p$ distributions at high temperature for a quartz single crystal (Fig. 9) and polycrystal Gran 133 (Fig. 11), which is known to have experienced beta-quartz stability conditions (Schmid et al. 1981). The calculations used the single crystal elastic constants at 25°C (Zubov & Firsova 1962; McSkimin, Anreatch & Thurston 1965) at temperatures of 510°C and 571°C, (Kammer, Pardue & Frissel 1948) at 580°C, 600°C and 800°C. The densities for quartz as a function of temperature were calculated on the basis of the room temperature value of McSkimin et al. (1965) and the thermal expansion data given by Skinner (1966). The effect of pressure was investigated using the compressibility data of Olinger & Halleck (1976) and found to introduce a correction of 2 per cent in density at 800 MPa. Hence for the present investigation the pressure effect could be ignored without loss of generality.

The $V_p$ anisotropy and general seismic characteristics of an alpha-quartz single crystal (Fig. 9) do not change significantly between 25°C and 510°C at room pressure. However, at 571°C, (2°C below the transition), the $V_p$ velocity anisotropy has significantly decreased. At 580°C in the beta-quartz stability field, the $V_p$ values have increased by 0.6 km s$^{-1}$ or more, and the anisotropy has decreased to 2.4 per cent. The velocity distribution no longer has the three-fold symmetry around the c-axis (the origin in Fig. 9) as in the alpha-quartz field. The $V_p$ distribution is circular about the c-axis, i.e. the symmetry is axial or transverse isotropic. The symmetry does not change with increasing temperature above 580°C, but the $V_p$ increases with temperature (Fig. 10).

If we apply an identical temperature sequence to the polycrystalline quartzite Gran 133 (Figs 11 and 12) we see the same trends. The c-axis maximum of this fabric (Fig. 2) is in the centre of the diagram which facilitates comparison with the single crystal case (Figs 9 and 10). The $V_p$ values and anisotropy coefficients decrease gradually with increasing temperature in the alpha field. In the beta field the aggregate is virtually isotropic ($A = 1.1–2.6$ per cent) and $V_p$ increases with temperature (Fig. 12). We can compare our calculated values with the experimental measurements at 200 MPa by Kern (1979) on a quartzite with no marked LPO (Fig. 12). Kern's data fall between our maximum and

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**Figure 4.** $V_p$ velocity diagrams with the foliation normal N–S and the lineation E–W. Contour interval 0.1 km s$^{-1}$ (CC1, P248 Gran 133) or 0.25 km s$^{-1}$ (RL 8125, RL 8330). Minimum contour is dashed.
Figure 5. $V_{s2}$ velocity diagrams with the foliation normal N–S and the lineation E–W. Contour interval 0.1 km s$^{-1}$ (CC1, P248, Gran 133) or 0.05 km s$^{-1}$ (RL 8125, RL 8300). Minimum contour is dashed. Horizontal line is the trace of the foliation.

Figure 6. Shear wave splitting diagrams ($\Delta V_s$) with foliation normal N–S and the lineation E–W. Contour interval 0.1 km s$^{-1}$ in all diagrams.
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**DISCUSSION**

Extensive calculations have been made on the seismic properties of quartz with commonly reported LPOs. These LPOs are characteristic of quartz mylonites which form ductile shear zones or ductile faults in the middle to lower continental crust. The majority of these zones occur in greenschist to amphibolite facies rocks where quartz is in the alpha stability field. The deep continental seismic reflection programs (e.g. COCORP, ECORS, BIRPS, etc.) have imaged reflectors that are interpreted to correspond to such features (Fountain, Hurich & Smithson 1984; Jones & Nur 1984; Klemperer 1987; McDonough & Fountain 1988).

The $V_p$ anisotropy coefficients of the typical quartz fabric pattern (Fig. 3) are between 8 and 14 per cent, which minimum values with a general trend identical to our calculations. The minimum $V_p$ value of Kern's data is displaced by 80°C with respect to our calculations. We return to this observation in the discussion.

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**Alpha-beta transition in quartz single crystal**

**Figure 9.** $V_p$ diagrams for a single crystal of quartz as a function of temperature. The $c$-axis is at the origin (centre), $a$-axis is N–S and the $m$-axis is E–W. Contour intervals 0.2 (25°C, 510°C) 0.1 (57°C) 0.01 (580°C), and 0.05 (600°C, 700°C, 800°C) in km s⁻¹.

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**Figure 7.** Shear wave polarization diagrams with foliation normal N–S and lineation E–W. Longer trace is the polarization plane of $V_{52}$ and the shorter trace is $V_{52}$.

**Figure 8.** Shear wave polarization diagrams with foliation normal at the origin (centre) of the diagram, lineation E–W. Longer trace is the polarization plane of $V_{52}$ and the shorter trace is $V_{52}$.
Alpha-Beta Quartz Transition: single crystal

![Graph showing the relationship between temperature and Vp](image)

Figure 10. $V_p$ as a function of temperature for a single crystal of quartz.

indicates that anisotropy can increase the seismic reflector potential of such rocks. The minimum $V_p$ velocity is associated with the orientation of the $a$-axes (Figs 2 and 3) and the maximum $V_p$ velocity with the $c$-axes concentrations. These correlations are rather broad but consistent, suggesting that the variation in $V_p$ with orientation could be used to infer pole figures in quartzites. Such an inversion between seismic properties and LPO will allow some interpretation of the kinematics of movement as it is generally accepted that the $a$-axis is the main direction of flow in quartzites (e.g., Bouchez 1978). However the 3-D seismic properties would need to be determined for such kinematic studies, which is rarely achieved in practice.

The S-wave behaviour of the quartzites is rather complex with an anisotropy coefficient between 14 and 5 per cent for $V_{S1}$ and 16 and 6 per cent for $V_{S2}$. There are no simple correlations between the pole figures and $V_{S1}$. The maxima of $V_{S2}$ are broadly correlated with $c$-axes pole figures (Figs 2 and 5). The topology of the $\Delta V_s$ diagrams (Fig. 6) is extremely complex, although the actual values can be high (0.8 km s$^{-1}$). The orientation of the polarization planes (Figs 7 and 8) may provide some indication of orientation of the $c$-axes with the plane of $V_{S1}$ containing the $c$-axes maximum. In general the orientations of the S-waves do not provide good indicators for the orientations of the LPO.

The effect of temperature, in particular the alpha–beta quartz transition, shows that in the beta field there is no significant polycrystalline anisotropy (Figs 11 and 12). As the transition can be correlated with the amphibole-granulite facies boundary (Frost & Chacko 1989; Fig. 13) we can state that the quartzites in the granulite facies will be seismically isotropic. Further the $V_p$ values will be nearly 7 km s$^{-1}$ in the granulite facies, roughly 0.5 km s$^{-1}$ higher than in the amphibolite facies. We noted earlier that the minimum $V_p$ value associated with the transition is 80°C higher in the experimental determination of Kern (1978) (Fig. 12) than in our calculations. The shift of the transition

**Alpha - Beta Phase Transition in Quartzite GRAN 133**

![Graph showing Vp as a function of temperature for Quartzite GRAN 133](image)

Figure 11. $V_p$ diagrams for quartzite Gran 133 as a function of temperature. The foliation normal is N–S and the lineation E–W. Contour intervals 0.1 (25°C, 510°C) 0.05 (571°C), 0.01 (580°C, 600°C), and 0.02 (700°C, 800°C) in km s$^{-1}$. 
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GRAN133 Quartzite

Figure 12. $V_p$ as a function of temperature for quartzite Gran 133. The data of Kern (1979) for a quartzite at 200 MPa are plotted for comparison; note displacement of minimum $V_p$ value at 650°C.

temperature in real polycrystalline quartz is caused by internal stresses that arise from the anisotropy of thermal expansion (Skinner 1986) in individual grains. The internal stresses cause the shift of the transition temperature (Coe & Paterson 1969) of individual grains and hence the polycrystal (Kern 1979; Van der Molen 1981). Our calculations did not take into account the internal stresses of the aggregate; such a calculation would require knowledge of individual grain orientations and their neighbours in a rock with strong LPO. Such information can be obtained by techniques such as electron channelling.

The alpha–beta quartz transformation is a displacive inversion involving a low degree of transport (e.g. Spry 1969, p. 88), hence it is instantaneous. Depending on the thermal and pressure gradient, the transition will occur at a precise depth in the continental crust. Taking the sample Gran 133 as an example (Fig. 11), the vertical $V_p$ values for a horizontally layered quartzite would be 5.1 km s$^{-1}$ in the alpha field at 571°C and 6.1 km s$^{-1}$ in the beta field at 580°C. Taking into account the densities in the alpha and beta fields (2.5514 and 2.5321 g cm$^{-3}$ respectively) the vertical reflection coefficient ($R_v$) would be $-0.0855$. Such a value of $R_v$ will result in a strong seismic reflection in the case of pure quartzite. Several factors will limit the seismic importance of this transition. Pure quartzite sequences although common in the continental crust, are rarely over a hundred metres thick. The transition will only occur at crustal depths if the geothermal gradient is relatively high (e.g. Basin & Range geotherm in Fig. 13). The effect of the transition on seismic properties will be reduced by the presence of other minerals, e.g. albite and orthoclase. Furthermore the presence of such minerals and water will cause melting (e.g. Winkler 1974) in the alpha-quartz stability field (Fig. 13).

CONCLUSIONS

Pure quartz mylonites with strong LPOs have a $V_p$ anisotropy coefficient ($A$) of between 8 and 11 per cent. There is a good correlation between the $V_p$ minimum and $V_p$ maximum with the maximum intensities in the a-axes and c-axes pole figures respectively. Variation in $V_p$ with orientation in mylonites allows the correlation with the plastic flow direction. In contrast shear waves appear to be unreliable indicators of crystallographic orientation. Shear wave anisotropy is important ($A = 5–14$ per cent) and shear wave splitting can attain high values ($0.8$ km s$^{-1}$).

At deep crustal depths (~25 km) in the granulite facies the quartz will be in the beta stability field. The beta-quartz rocks are seismically transverse isotropic with a $V_p$ approximately 0.6 km s$^{-1}$ or more above the value in the alpha field. The localized polymorphic phase boundary will result in a strong $V_p$ reflection in pure quartz rocks with reflection coefficient of $-0.08$. The presence of secondary minerals and partial melting will limit the seismological importance of this phase transition.

ACKNOWLEDGMENTS

D. M. would like to thank the program 'ATP Accompagnement ECORS 89' for financial support. The authors thank the reviewers for their constructive comments which helped improve the manuscript.

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