# Anisotropy of schists: Contribution of crustal anisotropy to active source seismic experiments and shear wave splitting observations 

Nicola J. Godfrey ${ }^{1}$<br>Earth Science Department, University of Southern California, Los Angeles<br>Nikolas I. Christensen<br>Department of Geology and Geophysics, University of Wisconsin-Madison

David A. Okaya
Earth Science Department, University of Southern California, Los Angeles


#### Abstract

We have made sets of five independent compressional and shear wave velocity measurements, which with density, allow us to completely characterize the transverse isotropy of samples from five metamorphic belts: the Haast schist terrane (South Island, New Zealand), Poultney slate, Chugach phyllite, Coldfoot schist, and Pelona schist (United States). These velocity measurements include compressional wave velocities for propagation parallel, perpendicular, and at $45^{\circ}$ to the symmetry axis, shear wave velocity for propagation and particle motion perpendicular to the symmetry axis, and shear wave velocity for propagation parallel to the symmetry axis. Velocity measurements were made up to pressures of $1 \mathrm{GPa}(\sim 35-\mathrm{km}$ depth $)$ where microcracks are closed and anisotropy is due to preferred mineral orientation. Our samples exhibit compressional wave anisotropy of $9-20 \%$ as well as significant shear wave splitting. Metamorphic terranes that are anisotropic to ultrasonic waves may also be anisotropic at the scale of active and passive seismic experiments. Our data suggest that a significant thickness ( $10-20 \mathrm{~km}$ ) of appropriately oriented (steeply dipping foliation) schist in the crust could contribute as much as $45 \%$ of observed shear wave splitting. Our data set can also be used to model the effects of crustal anisotropy for active source seismic experiments in order to determine if the anisotropy of the terrane is significant and needs to be taken into account during processing and modeling of the data.


## 1. Introduction

Most crustal lithologies exhibit some degree of anisotropy [e.g., Christensen and Mooney, 1995] Crustal anisotropy may be produced by a variety of means. Magma flow may preferentially orient elongate crystals in the direction of flow, tabular sediments may settle preferentially, and anisotropy may be enhanced by sediment compaction, recrystallization during metamorphism may form a pervasive foliation or schistosity, and cracks may open as a result of differential principal stresses [Helbig, 1994] In this study we are interested in the preferred mineral alignment observed in metamorphic terranes, which are often characterized by planar structures such as slaty cleavage, schistosity, and foliation at the centimeter scale as well as the meter or kilometer scale. These structures are often pervasive for tens to hundreds of kilometers. Laboratory investigations have shown that common rock-forming silicate minerals are anisotropic to ultrasonic waves at the scale of a single crystal [Alexandrov and Rhyzhova, 1961a, 1961b, 1962]. Laboratory investigations [e.g., Birch, 1960; Christensen, 1965, 1966; Christensen and Mooney, 1995; Siegesmund, 1996; Rabbel et al., 1998; Weiss et al., 1999] have also shown

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that most crustal metamorphic rocks are highly anisotropic to ultrasonic wave propagation as a result of lattice-preferred orientations of these silicate minerals. The preferred orientations of minerals in extensive metamorphic terranes also make metamorphic rocks highly anisotropic to lower-frequency seismic waves [Brocher and Christensen, 1990]. If the crustal anisotropy is appropriately oriented, it will affect the interpretation of seismic data collected in the region.

Data from active source experiments include near-vertical and wide-angle arrivals. Near-vertical and wide-angle rays sample different velocities in anisotropic rocks depending on whether the rays travel in slow, fast, or intermediate directions or most likely a cumulative combination of all these. A set of rays with a range of ray paths traveling through an anisotropic region will arrive at the surface with different travel times compared to a similar ray set passing through an equivalent isotropic region. If these travel time differences are significant, modeling techniques that assume isotropy, which are generally used to model crustal-scale seismic data, will not achieve a good fit to the data. In addition, even if a good fit to the data can be achieved, the resulting model will not be correct since it is underparameterized.

Passive teleseismic experiments often observe significant shear wave splitting (e.g., 2.2 s beneath South Island, New Zealand [Klosko et al., 1999]). Shear wave splitting measurements are usually attributed to anisotropy in the mantle [Savage, 1999]. If the degree of crustal anisotropy is significant and the orientation of the anisotropic body is appropriate, crustal anisotropy may contribute, either positively or negatively, to observed shear wave splitting values. If the crustal anisotropy is oriented with the fast


Figure 1. Schematic diagrams showing the velocity measurements required to characterize transverse isotropy (hexagonal) and orthorhombic symmetry rocks. Arrows represent propagation directions, "zig-zag" lines represent vibration directions. Solid arrows and lines indicate independent measurements. Shaded arrows and lines indicate measurements that are equivalent to those shown in black. (a) Orthorhombic case for a rock with lineations (such as those formed by mineral alignment) within the foliation. (b) Orthorhombic case for a rock that has crenulations within the foliation. (c) Transverse isotropy case for a rock with uniform foliation and no lineations, mineral alignment, or crenulations.
direction parallel to that of the mantle, the combined crustal and mantle anisotropy will result in greater total anisotropy than from the mantle alone. In this case, the mantle component of anisotropy determined from shear wave splitting will be less than the total estimated from the observed shear wave splitting data. This in turn affects the amount of mantle strain estimated from the mantle anisotropy or the thickness of the anisotropic slab. If the crustal and mantle anisotropies are misaligned, the total anisotropy determined from shear wave splitting measurements may be less than the actual amount of mantle anisotropy or double splitting will be observed [Silver and Savage, 1994]. This in turn may lead to an underestimation of mantle strain in the region [Silver
and Chan, 1991]. A summary of worldwide shear wave splitting observations and their modeling, including why the contribution of crustal anisotropy is assumed to be negligible, is given by Savage [1999].

In order to study the effects of crustal anisotropy on both passive and active source seismic data and to be able to incorporate anisotropy into the modeling and processing of seismic data, we need to characterize the anisotropy completely. This paper presents a more complete petrophysical analysis of schist samples than has previously been done. The most general anisotropic elastic solid (triclinic symmetry) requires 21 independent elastic stiffness constants to describe the equations of motion that govern the propagation of elastic waves, including compressional and shear waves, through the solid [Nye, 1985]. Many crustal metamorphic rocks behave as elastic solids with orthorhombic symmetry [e.g., Christensen, 1965] which require nine independent constants to describe their elastic tensors (Figures 1a and 1b). Lowto medium-grade (greenschist to amphibolite facies) rocks, however, commonly have well-developed foliation and behave as transversely isotropic (hexagonal symmetry) media if lineations are not present (Figure 1c). The stiffness tensor of a transversely isotropic elastic solid can be described by five independent constants, which can be obtained from five independent velocity measurements (two compressional wave velocities, one quasicompressional wave velocity, and two shear wave velocities) and density. The required compressional wave velocity measurements are velocities perpendicular and parallel to the symmetry axis (which in the case of our samples is perpendicular to the foliation). The quasi-compressional wave velocity is measured at $45^{\circ}$ to the symmetry axis. The two required shear wave velocities are for propagation parallel to the symmetry axis and for propagation and particle motion perpendicular to the symmetry axis. We could also use shear wave velocities measured perpendicular to the symmetry axis with particle motion parallel to the symmetry axis for one shear wave and particle motion perpendicular to the symmetry axis for the other shear wave (Figure 1c). We present this suite of measurements for 14 samples from a variety of schist belts. Geological field evidence from the New Zealand and Alaska terranes suggests that our samples are representative of these terranes over several kilometers. We then discuss a number of observations pertaining to the laboratory data in relation to seismic field studies. Our data set allows for the future modeling of such schist terranes with regard to active source seismic analysis and data processing as well as passive seismological analysis.

## 2. Geologic Setting of Terranes Studied

We have analyzed samples from the Haast schist terrane of South Island, New Zealand, Pelona schist from southern California, Coldfoot schist and Chugach phyllite from Alaska, and Poultney slate from Vermont.

### 2.1. Haast Schist Terrane, South Island, New Zealand

One of the most prominent schist belts in the world is the Haast schist terrane of South Island, New Zealand (Figure 2). It is exposed southeast of the Pacific-Australian plate boundary (the Alpine fault) where it is upturned with foliation approximately vertical [Wellman, 1979; Norris et al., 1990]. The Torlesse graywacke terrane (Figure 2), has been interpreted as a Mesozoic accretionary prism and is believed to be the protolith of the Haast schist. The Haast schist terrane is thought to extend beneath the Torlesse terrane across most of South Island with an extent of up to $80,000 \mathrm{~km}^{2}$. The Haast schist has been divided into several metamorphic zones and subzones. In order of increasing metamorphic grade the Haast schist terrane consists of chlorite subzones II (sample A-33), III (samples A-18 and A-35), and IV (samples A-16 and A-43); the biotite zone (sample A-13); and, adjacent to the Alpine Fault, the garnet-oligoclase zone (samples


Figure 2. (a) Terrane map of South Island, New Zealand, including the locations of samples collected for petrophysical analysis in this study (solid circles). (b) Cross section across the center of South Island [after Wellman, 1979]. Open circle shows the location of shear wave splitting observations referred to in the text.

A-1, A-5, and A-12). Initial ultrasonic velocity measurements of different Haast schist samples revealed strong compressional wave velocity anisotropy of up to 17\% [Okaya et al., 1995], where

$$
\begin{equation*}
\mathrm{A}=\left[\left(\mathrm{V}_{\max }-\mathrm{V}_{\min }\right) / \mathrm{V}_{\text {ave }}\right] \times 100 \tag{1}
\end{equation*}
$$

and A is percentage anisotropy, $\mathrm{V}_{\text {max }}$ is the maximum velocity, $V_{\text {min }}$ is the minimum velocity and $V_{\text {ave }}$ is the average of three orthogonal velocities.

Foliation within the exposed Haast schist is parallel to the A1pine fault and is consistent over many kilometers. The anisotropy of the Haast schist is significant with regard to the interpretations of both an active source seismic reflection/refraction experiment (SIGHT) [Davey et al., 1997] and a passive seismic experiment (SAPSE) [Anderson et al., 1997] conducted across South Island, New Zealand, in 1996.

### 2.2. Poultney Slate, Vermont

Poultney slate (sample VT-1, Figure 3c) is part of the Lower Ordovician Mount Hamilton Group [Zen, 1961; Drake et al., 1989]. It is a red-colored slate with well-developed cleavage [Christensen, 1965]. Sediments deposited during the Early and

Middle Ordovician were metamorphosed to produce the slaty cleavage during "event $2^{\prime \prime}$ of the Taconic orogeny as a result of low-grade regional metamorphism [Zen, 1972].

### 2.3. Chugach Phyllite, Alaska

Alaska is made up of a number of terranes that were accreted to the North American continent (Figure 3b). Southern Alaska consists of the Southern Margin Composite terrane [Pflaker et al., 1994], which includes the Chugach terrane. Samples TA-2 and TA-23 come from the Upper Cretaceous Valdez Group of the Chugach terrane [Pflaker et al., 1994]. After accretion the Chugach terrane was metamorphosed by the emplacement of early Eocene anatectic granite plutons [Pflaker et al., 1989] The Valdez Group consists of multiply deformed, interbedded metagraywacke and fine-grained schistose to mylonitic pelitic phyllites in the region of our samples [Pflaker et al., 1989]. Foliation observed in outcrop is steeply dipping [Nokleberg et al., 1989]. The Chugach terrane is arcuate in shape, and its foliation, which is consistent over many kilometers, curves with the terrane. The anisotropy of the schists in the Chugach terrane is significant with respect to a seismic experiment (Trans-Alaska Crustal Transect or TACT) conducted across these terranes in 1986 [Brocher et al., 1989; Brocher and Christensen, 1990, 1991].


Figure 3. (a) Map of the United States. Solid lines are state/province boundaries, bold lines are national boundaries. Dark shaded areas are the regions enlarged in Figures 3b-3d. Open circles show the locations of shear wave splitting observations referred to in the text. Open stars show sample locations. (b) Maps of (top) northern and (bottom) southern Alaska showing terrane locations. (c) Map of Vermont showing the location of the Poultney slate sample used in this study. (d) Map of southern California showing the locations of the Pelona, Rand, and Orocopia schists (solid regions).

### 2.4. Coldfoot Schist, Alaska

The Brooks Range region of northern Alaska is a collisional orogenic belt made up of several terranes including the Coldfoot subterrane [Moore et al., 1997a]. The Coldfoot subterrane (sample TA-80, Figure 3b) is an allochthonous terrane made up of passive margin strata [Moore et al., 1997a], which have been metamorphosed. Retrograde greenschist deformation in the quartz-mica schist making up most of the Coldfoot subterrane has totally obliterated any original sedimentary features, leaving only a pervasive nearly vertical dipping foliation [Moore et al., 1997a]. The Coldfoot schist belt is $15-50 \mathrm{~km}$ wide and extends for up to 800 km along the southern Brooks Range [Moore et al., 1997a; 1997b]. The schist has a structural thickness of $3-12 \mathrm{~km}$ [Moore et al., 1997a]. Foliation of the Coldfoot schist is parallel to the Coldfoot subterrane (east-west) and is consistent over many kilometers.

### 2.5. Pelona Schist, California

Pelona (sample LA-1), Rand, and Orocopia schists, which are believed to be equivalent [Ehlig, 1968; Haxel et al., 1987], out-
crop in southern California (Figure 3d) and southwestern Arizona. Taken together, these schists may make up a sizeable portion of the crust in this region. The protoliths of the Pelona and Orocopia schists are thought to be Jurassic or Cretaceous deep marine sediments deposited on oceanic crust, and they are thought to have been metamorphosed by thrusting no later than the Late Cretaceous [Haxel et al., 1985]. In the Chocolate Mountains the foliation of the Orocopia schist is antiformal in structure. The anisotropy of the Pelona and related schists is significant with respect to active source and passive data collected during the Los Angeles Region Seismic Experiment (LARSE) in 1994 and 1999 [Fuis et al., 1996; Kohler and Davis, 1997; Fuis et al., 1999].

## 3. Petrophysical Data

Nine Haast schist samples collected as part of the SIGHT experiment from South Island, New Zealand (Figure 2), and five samples from the U.S. schist terranes (Figure 3) were analyzed in this study. In order to characterize transverse isotropy, a minimum of five velocity measurements are required (Figure 1c). For these measurements we removed four $5-\mathrm{cm}$-long cores from each


Figure 4. Schematic diagrams of the relationship between sample fabric, the cores taken for velocity analysis, and the terminology used in the text. Terminology is defined for horizontally propagating waves. (a) Schematic diagram showing the four cores taken in this study (top). The measurements taken for this study are shown. Note these measurements oversample the transverse isotropy case and undersample the orthorhombic case. (b) Schematic diagram showing the three cores and measurements required to characterize a truly transversely isotropic sample.
sample using a $2.54-\mathrm{cm}$ diamond-coring bit. Cores were taken from single blocks of each rock. The well-defined and uniform foliations of the rocks allowed core axes to be oriented relative to the foliation and to one another to within $\pm 1^{\circ}$. Bulk densities of the four cores from each rock agree to within $3 \%$, indicating sample uniformity. One core was taken normal to the foliation, two mutually perpendicular cores were oriented with their axes within the foliation, and the fourth core was oriented at $45^{\circ}$ to the foliation (Figure 4a). The two cores taken with their axes within the foliation were oriented parallel and perpendicular to lineation if present.

The core ends were trimmed and ground flat and parallel to within 0.07 cm on a diamond grinding disk. The volume of each core was measured from the length and diameter. The cores were weighed and densities were calculated from their weights and dimensions. Each core was then fitted with a soldered copper jacket to prevent penetration of high-pressure oil into the rock samples. To make velocity measurements, $1-\mathrm{MHz}$ transducers were fixed to both core ends. Gum rubber tubing was placed over the sample assembly as a further prevention of oil leakage.

While many techniques can be used to obtain the elastic coefficients of single crystals, e.g., transit time measurements, pulse echo methods, acoustic interferometry [Schreiber et al., 1973], whole rock analyses are generally done by transit time methods such as the pulse-transmission technique. Velocities were measured at room temperature and hydrostatic pressures up to 1000 MPa ( 10 kbar , equivalent to $\sim 35 \mathrm{~km}$ depth) using the pulse transmission technique described by Christensen [1985]. A pulse generator produced a $50-\mathrm{V}$ square wave, which was simultaneously sent to a transducer on the sample and to a calibrated vari-able-length mercury delay line. Output signals were viewed on a dual-trace oscilloscope screen. The length of the mercury delay line was varied until the first breaks of the arrivals of the signals coincided, corresponding to equal transit times. The mercury delay line readings, the known velocity in mercury, and the measured length of the sample permit the calculation of the acoustic velocity in the sample. Readings were taken at $20,40,60,80$, $100,200,400,600,800$, and 1000 MPa and were repeated for downgoing pressures. Compressional wave velocities were measured in this way for the differently oriented cores for each rock
type. Shear wave measurements were made using the same technique but with AC quartz as a sending transducer and lead zirconate titanate (shear mode) as a receiver. The sending transducer for shear wave measurements generates shear waves with known orientation with respect to the sample. It is important to ensure that the sending and receiving transducers are aligned parallel to one another since the vibration directions of shear waves propagating through anisotropic media are as important as the propagation directions [Christensen, 1985].

The accuracy of velocity measurements using the technique described above is within $1 \%$ and depends principally on establishing the onset of the first motion from the sample, the accurate length measurement of the sample, and the calibration of the mercury delay line [Christensen, 1985] The velocities reported here are determined at regular pressure intervals from an empirical curve fit that best matches the measured data. Using the empirical formula an entire velocity-pressure data set can be reduced to four constants [Wepfer and Christensen, 1991].

## 4. Results

All the measured data are presented in Tables 1 and 2. All the samples that we analyzed show significant compressional wave anisotropy (Figure 5). There is no systematic increase in the degree of anisotropy with metamorphic grade (compare Haast schist samples, Figure 5). The different Haast schist samples are assumed to have the same Torlesse graywacke protolith, which is isotropic (Figure 5a). The Poultney slate, which is a fairly lowgrade metamorphic rock, shows the highest percentage anisotropy (Figure 5b). Its sedimentary protolith may have been either an isotropic mudstone or an anisotropic shale.

In addition to the velocity and density measurements for each core, thin sections were cut from each sample, and the mineralogy of most of the samples was determined (Table 3). Samples VT-1, TA-2, and TA-23 were too fine grained for mineralogical identification by petrographic techniques. We calculate the isotropic Voigt and Reuss averages [Christensen, 1966] for all samples for which we have modal mineralogy. These values, based on the $P$ wave velocities of the constituent minerals in each sample (Table 3), give an intrinsic isotropic velocity for each rock that may be compared with our measurements (Figure 5c).

All the tested samples show some degree of orthorhombic symmetry (Figure 5). Eight samples show a significant degree of orthorhombic symmetry (Figure 5b and Table 2), and they are not considered further in this study. They indicate that two additional quasi-compressional wave velocity measurements at $45^{\circ}$ to the symmetry axes are required for full orthorhombic characterization of these schists (Figures la and 1b). Six samples show only a small degree of orthorhombic symmetry (Figure 5a and Table 1). They show two fast compressional waves, those propagating perpendicular to the symmetry axis, and one slow compressional wave propagating parallel to the symmetry axis (perpendicular to foliation) (Figures 5 a and 6a). The quasi-compressional wave velocity measured at $45^{\circ}$ to the symmetry axis lies between the fast and slow values (Figure 6a).

The shear waves show two slow directions and one fast direction (Figure 6a). The fastest shear wave is that which propagates perpendicular to the symmetry axis with the particle motion also perpendicular to the symmetry axis and within the foliation. The slowest shear wave is that which propagates parallel to the symmetry axis. The faster of the two slow shear waves propagates perpendicular to the symmetry axis with particle motion parallel to the symmetry axis (Figure 6a). Although samples A-1, LA-1, TA-2, TA-23, TA-80, and VT-1 are truly orthorhombic, they approximate well to transverse isotropy (Figure 5a). Orthorhombic symmetry is described by nine elastic constants, which would require nine velocity measurements (Figures 1a and 1b). By approximating the symmetry to transverse isotropy, we only need the five velocity measurements that we have made to describe the
anisotropy (Figures 1c and 4b). In the true transverse isotropy case the two fast compressional velocities (perpendicular to the symmetry axis) would be equal as would the two slowest shear wave velocities (parallel to the symmetry axis and perpendicular to the symmetry axis with particle motion parallel to the symmetry axis). We average the two fast compressional velocities ( $V_{\text {Pave }}$, Table 1) and the two slowest shear wave velocities ( $V_{\text {svave }}$, Table 1) in order to approximate transverse isotropy for these samples (Figure 6b). This gives us the five velocities, which, along with density, can be used to characterize transverse isotropy. The measurements we have made undersample orthorhombic symmetry and oversample transverse isotropy (Figure 4).

### 4.1. Velocity Variation With Propagation Direction

The two compressional wave velocities, one quasicompressional wave velocity, two shear wave velocities, and density allow us to calculate the five independent elastic stiffness constants (abbreviated tensor notation $C_{i j}$ [Auld, 1990; Winterstein, 1990]) that describe a transversely isotropic medium [Neighbours and Schacher, 1967; Musgrave, 1970; Mavko et al., 1998]:

$$
\begin{gather*}
C_{11}=\rho V_{P}\left(90^{\circ}\right)^{2},  \tag{2}\\
C_{33}=\rho V_{P}\left(0^{\circ}\right)^{2},  \tag{3}\\
C_{44}=\rho V_{s v}\left(0^{\circ}\right)^{2}=\rho V_{s t h}\left(0^{\circ}\right)^{2}=\rho V_{s v}\left(90^{\circ}\right)^{2},  \tag{4}\\
C_{12}=C_{11}-\left[2 \rho V_{s H}\left(90^{\circ}\right)^{2}\right],  \tag{5}\\
C_{13}=-C_{44}+\left[\left(4 \rho^{2} V_{P}\left(45^{\circ}\right)^{4}\right)-\left(2 \rho V_{P}\left(45^{\circ}\right)^{2}\right)\left(C_{11}+C_{33}+2 C_{44}\right)\right. \\
\left.+\left(C_{11}+C_{44}\right)\left(C_{33}+C_{44}\right)\right]^{12},  \tag{6}\\
C_{66}=\left[C_{11}-C_{12}\right] / 2 . \tag{7}
\end{gather*}
$$

In (2)-(7) the notation is based on a horizontally traveling wave and the angle between the wave propagation direction and the symmetry axis (Figure 4). $V_{P}\left(90^{\circ}\right)$ is the compressional wave propagating perpendicular to the symmetry axis (within the foliation). $V_{P}\left(0^{\circ}\right)$ is the compressional wave propagating parallel to the symmetry axis (perpendicular to foliation). $V_{P}\left(45^{\circ}\right)$ is the quasicompressional wave propagating at $45^{\circ}$ to the symmetry axis. $V_{s H}\left(0^{\circ}\right)$ is the shear wave propagating parallel to the symmetry axis (particle motion within the foliation). $V_{S H}\left(90^{\circ}\right)$ is the shear wave propagating perpendicular to the symmetry axis with particle motion also perpendicular to the symmetry axis (within the foliation) (Figure 4b). For a transversely isotropic material $V_{s H}\left(0^{\circ}\right)$ $=V_{s v}\left(0^{\circ}\right)=V_{s v}\left(90^{\circ}\right)$ (Figure 4b). Density is denoted by $\rho$. Note that only five of the above expressions are independent. Stiffness coefficient $C_{66}$ is dependent on $C_{11}$ and $C_{12}$.

We have calculated the stiffness coefficients for each sample and the results are presented in Table 1. The elastic stiffness constants can in turn be used to calculate compressional $\left(q V_{p}\right)$ and shear waves ( $q V_{S V}$ and $V_{S H}$ ) for all angles of incidence $(\theta)$ with respect to the symmetry axis [Auld, 1990; Mavko et al., 1998]. The $q V_{s V}$ is the horizontally propagating shear wave whose particle motion is vertical, and $V_{S H}$ is the horizontally propagating shear wave whose particle motion is horizontal (Figure 7):

$$
\begin{gather*}
q V_{P}^{2}=\left[C_{11} \sin ^{2} \theta+C_{11} \cos ^{2} \theta+C_{44}+\mathrm{M}^{1 / 2}\right] / 2 \rho,  \tag{8}\\
q V_{S V}{ }^{2}=\left[C_{11} \sin ^{2} \theta+C_{33} \cos ^{2} \theta+C_{44}-\mathrm{M}^{1 / 2}\right] / 2 \rho,  \tag{9}\\
V_{S H}^{2}=\left[C_{66} \sin ^{2} \theta+C_{44} \cos ^{2} \theta\right] / \rho,  \tag{10}\\
\mathrm{M}=\left[\left(C_{11}-C_{44}\right) \sin ^{2} \theta-\left(C_{33}-C_{44}\right) \cos ^{2} \theta\right]^{2}+\left(C_{13}+C_{44}\right) \sin ^{2} 2 \theta . \tag{11}
\end{gather*}
$$

Equations (8)-(11) are linearized formulations of the otherwise more complex descriptions of the relationship between elastic stiffness and phase velocity. In the laboratory we measure phase velocities since the sources are plane waves [Dellinger and Vernik, 1994; Johnston and Christensen, 1995]. In active source field experiments, however, we measure group velocities since the waves are generated from point sources [Johnston and Chris-

Table 1. Transverse Isotropy Approximation ${ }^{\text {a }}$

| $P$ | $V_{P}\left(0^{\circ}\right)$ | $V_{P}\left(90^{\circ}\right) \mathrm{a}$ | $V_{P}\left(90^{\circ}\right) \mathrm{b}$ | P 90 are | $V_{p}\left(45^{\circ}\right)$ | $V_{S H}\left(90^{\circ}\right)$ | $V_{S V}\left(0^{\circ}\right)$ | $V_{S V}\left(90^{\circ}\right)$ | $V_{S V}$ | $C_{11}$ | $\mathrm{C}_{12}$ | $C_{33}$ | $C_{44}$ | $C_{13}$ |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| Sample A-1 (Haast schist (garnet-oligoclase zone)), Density $=2.718 \mathrm{~g} / \mathrm{cm}^{3}$ |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 1 | 5.163 | 6.143 | 6.037 | 6.090 | 5.242 | 3.750 | 3.177 | 3.009 | 3.093 | 100.805 | 24.362 | 72.453 | 26.002 | 7.895 |
| 2 | 5.472 | 6.339 | 6.268 | 6.304 | 5.598 | 3.862 | 3.338 | 3.101 | 3.219 | 107.997 | 26.902 | 81.384 | 28.171 | 17.416 |
| 3 | 5.581 | 6.403 | 6.356 | 6.379 | 5.720 | 3.895 | 3.390 | 3.140 | 3.265 | 110.609 | 28.147 | 84.659 | 28.974 | 20.605 |
| 4 | 5.651 | 6.466 | 6.405 | 6.436 | 5.789 | 3.911 | 3.418 | 3.168 | 3.293 | 112.568 | 29.432 | 86.796 | 29.475 | 21.952 |
| 5 | 5.706 | 6.500 | 6.440 | 6.470 | 5.839 | 3.922 | 3.440 | 3.189 | 3.314 | 113.778 | 30.174 | 88.494 | 29.857 | 22.994 |
| 6 | 5.750 | 6.534 | 6.468 | 6.501 | 5.881 | 3.931 | 3.457 | 3.207 | 3.332 | 114.871 | 30.891 | 89.864 | 30.171 | 23.873 |
| 7 | 5.789 | 6.559 | 6.491 | 6.525 | 5.916 | 3.938 | 3.471 | 3.222 | 3.347 | 115.712 | 31.415 | 91.087 | 30.441 | 24.614 |
| 8 | 5.823 | 6.583 | 6.511 | 6.547 | 5.947 | 3.944 | 3.484 | 3.235 | 3.359 | 116.502 | 31.932 | 92.160 | 30.675 | 25.264 |
| 9 | 5.852 | 6.602 | 6.529 | 6.566 | 5.974 | 3.950 | 3.495 | 3.246 | 3.371 | 117.162 | 32.346 | 93.080 | 30.885 | 25.850 |
| 10 | 5.879 | 6.621 | 6.545 | 6.583 | 5.998 | 3.955 | 3.506 | 3.257 | 3.381 | 117.787 | 32.757 | 93.941 | 31.073 | 26.331 |
| Sample LA-l (Pelona schist), Density $=2.720 \mathrm{~g} / \mathrm{cm}^{3}$ |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 1 | 5.069 | 5.996 | N/A | 5.996 | 5.556 | 3.304 | 3.076 | 3.314 | 3.195 | 97.789 | 38.404 | 69.890 | 27.766 | 26.802 |
| 2 | 5.462 | 6.317 | N/A | 6.317 | 5.859 | 3.482 | 3.256 | 3.350 | 3.303 | 108.540 | 42.584 | 81.147 | 29.675 | 31.025 |
| 3 | 5.597 | 6.437 | N/A | 6.437 | 5.964 | 3.549 | 3.321 | 3.432 | 3.377 | 112.703 | 44.184 | 85.208 | 31.010 | 31.016 |
| 4 | 5.672 | 6.496 | N/A | 6.496 | 6.024 | 3.582 | 3.355 | 3.472 | 3.414 | 114.779 | 44.979 | 87.507 | 31.693 | 31.424 |
| 5 | 5.727 | 6.532 | N/A | 6.532 | 6.069 | 3.604 | 3.379 | 3.498 | 3.439 | 116.054 | 45.395 | 89.212 | 32.159 | 32.030 |
| 6 | 5.771 | 6.560 | N/A | 6.560 | 6.106 | 3.620 | 3.399 | 3.517 | 3.458 | 117.051 | 45.763 | 90.588 | 32.525 | 32.621 |
| 7 | 5.809 | 6.582 | N/A | 6.582 | 6.137 | 3.634 | 3.415 | 3.532 | 3.474 | 117.838 | 45.997 | 91.785 | 32.817 | 33.166 |
| 8 | 5.842 | 6.601 | N/A | 6.601 | 6.164 | 3.646 | 3.429 | 3.546 | 3.488 | 118.519 | 46.203 | 92.831 | 33.082 | 33.627 |
| 9 | 5.872 | 6.619 | N/A | 6.619 | 6.188 | 3.656 | 3.442 | 3.558 | 3.500 | 119.166 | 46.453 | 93.787 | 33.320 | 34.003 |
| 10 | 5.898 | 6.634 | N/A | 6.634 | 6.209 | 3.666 | 3.453 | 3.568 | 3.511 | 119.707 | 46.596 | 94.619 | 33.520 | 34.368 |
| Sample TA-2 (Chugach phyllite), Density $=2.723 \mathrm{~g} / \mathrm{cm}^{3}$ |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 1 | 5.272 | 6.057 | 6.142 | 6.100 | 5.401 | 3.725 | 3.216 | 3.131 | 3.174 | 101.306 | 25.740 | 75.683 | 27.424 | 13.567 |
| 2 | 5.637 | 6.271 | 6.341 | 6.306 | 5.734 | 3.828 | 3.337 | 3.291 | 3.314 | 108.282 | 28.478 | 86.525 | 29.906 | 20.687 |
| 3 | 5.798 | 6.358 | 6.446 | 6.402 | 5.894 | 3.872 | 3.395 | 3.363 | 3.379 | 111.604 | 29.955 | 91.539 | 31.090 | 24.541 |
| 4 | 5.887 | 6.445 | 6.508 | 6.477 | 5.983 | 3.899 | 3.428 | 3.402 | 3.415 | 114.216 | 31.425 | 94.370 | 31.756 | 26.299 |
| 5 | 5.947 | 6.482 | 6.548 | 6.515 | 6.041 | 3.917 | 3.450 | 3.426 | 3.438 | 115.578 | 32.021 | 96.304 | 32.185 | 27.662 |
| 6 | 5.994 | 6.519 | 6.575 | 6.547 | 6.084 | 3.931 | 3.466 | 3.445 | 3.456 | 116.717 | 32.561 | 97.832 | 32.514 | 28.556 |
| 7 | 6.032 | 6.542 | 6.596 | 6.569 | 6.118 | 3.941 | 3.479 | 3.459 | 3.469 | 117.502 | 32.918 | 99.076 | 32.768 | 29.337 |
| 8 | 6.065 | 6.565 | 6.613 | 6.589 | 6.147 | 3.950 | 3.490 | 3.472 | 3.481 | 118.219 | 33.248 | 100.163 | 32.996 | 29.954 |
| 9 | 6.095 | 6.583 | 6.628 | 6.606 | 6.173 | 3.958 | 3.500 | 3.483 | 3.492 | 118.812 | 33.496 | 101.157 | 33.195 | 30.542 |
| 10 | 6.121 | 6.600 | 6.640 | 6.620 | 6.195 | 3.965 | 3.508 | 3.493 | 3.501 | 119.334 | 33.716 | 102.022 | 33.366 | 31.017 |
| Sample TA-23 (Chugach phyllite), Density $=2.739 \mathrm{~g} / \mathrm{cm}^{3}$ |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 1 | 5.708 | 6.449 | 6.484 | 6.467 | 5.758 | 3.990 | 3.348 | 3.313 | 3.331 | 114.533 | 27.323 | 89.240 | 30.382 | 17.323 |
| 2 | 5.852 | 6.530 | 6.560 | 6.545 | 5.883 | 4.025 | 3.398 | 3.361 | 3.380 | 117.331 | 28.584 | 93.800 | 31.282 | 20.133 |
| 3 | 5.910 | 6.563 | 6.594 | 6.579 | 5.944 | 4.038 | 3.421 | 3.385 | 3.403 | 118.535 | 29.214 | 95.668 | 31.719 | 21.797 |
| 4 | 5.949 | 6.596 | 6.616 | 6.606 | 5.988 | 4.046 | 3.436 | 3.403 | 3.420 | 119.528 | 29.852 | 96.935 | 32.027 | 22.987 |
| 5 | 5.979 | 6.615 | 6.633 | 6.624 | 6.021 | 4.052 | 3.448 | 3.416 | 3.432 | 120.180 | 30.238 | 97.915 | 32.262 | 23.928 |
| 6 | 6.004 | 6.634 | 6.647 | 6.641 | 6.049 | 4.057 | 3.458 | 3.427 | 3.443 | 120.780 | 30.616 | 98.736 | 32.459 | 24.713 |
| 7 | 6.025 | 6.648 | 6.659 | 6.654 | 6.072 | 4.061 | 3.466 | 3.436 | 3.451 | 121.253 | 30.911 | 99.427 | 32.620 | 25.372 |
| 8 | 6.044 | 6.662 | 6.669 | 6.666 | 6.093 | 4.064 | 3.473 | 3.444 | 3.459 | 121.691 | 31.216 | 100.055 | 32.762 | 25.984 |
| 9 | 6.060 | 6.673 | 6.678 | 6.676 | 6.111 | 4.067 | 3.479 | 3.451 | 3.465 | 122.056 | 31.447 | 100.586 | 32.885 | 26.519 |
| 10 | 6.074 | 6.683 | 6.687 | 6.685 | 6.127 | 4.070 | 3.485 | 3.457 | 3.471 | 122.404 | 31.661 | 101.051 | 32.999 | 26.976 |
| Sample TA-80 (Coldfoot schist), Density $=2.657 \mathrm{~g} / \mathrm{cm}^{3}$ |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
|  | 5.077 | 5.913 | 6.007 | 5.960 | 5.457 | 3.786 | 3.161 | 2.905 | 3.033 | 94.381 | 18.211 | 68.487 | 24.442 | 26.301 |
| 2 | 5.418 | 6.127 | 6.187 | 6.157 | 5.763 | 3.892 | 3.322 | 3.041 | 3.182 | 100.723 | 20.229 | 77.996 | 26.894 | 32.260 |
| 3 | 5.577 | 6.203 | 6.262 | 6.233 | 5.858 | 3.923 | 3.378 | 3.093 | 3.236 | 103.209 | 21.427 | 82.640 | 27.815 | 32.938 |
| 4 | 5.673 | 6.278 | 6.306 | 6.292 | 5.952 | 3.936 | 3.404 | 3.120 | 3.262 | 105.189 | 22.864 | 85.510 | 28.272 | 35.608 |
| 5 | 5.744 | 6.315 | 6.337 | 6.326 | 6.006 | 3.943 | 3.420 | 3.140 | 3.280 | 106.329 | 23.710 | 87.664 | 28.585 | 36.858 |
| 6 | 5.800 | 6.351 | 6.362 | 6.357 | 6.059 | 3.948 | 3.432 | 3.155 | 3.294 | 107.356 | 24.529 | 89.381 | 28.821 | 38.477 |
| 7 | 5.847 | 6.376 | 6.383 | 6.380 | 6.096 | 3.952 | 3.442 | 3.167 | 3.305 | 108.135 | 25.139 | 90.836 | 29.014 | 39.418 |
| 8 | 5.889 | 6.401 | 6.401 | 6.401 | 6.133 | 3.956 | 3.450 | 3.178 | 3.314 | 108.865 | 25.701 | 92.146 | 29.181 | 40.513 |
| 9 | 5.925 | 6.421 | 6.417 | 6.419 | 6.163 | 3.959 | 3.457 | 3.188 | 3.323 | 109.478 | 26.188 | 93.276 | 29.331 | 41.339 |
| 10 | 5.958 | 6.441 | 6.431 | 6.436 | 6.192 | 3.962 | 3.464 | 3.197 | 3.331 | 110.059 | 26.642 | 94.318 | 29.472 | 42.180 |
| Sample VT-1 (Vermont slate), Density $=2.810 \mathrm{~g} / \mathrm{cm}^{3}$ |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 1 | 5.040 | 6.563 | 6.418 | 6.491 | 5.472 | 3.956 | 2.787 | 2.318 | 2.553 | 118.376 | 30.423 | 71.378 | 18.308 | 31.523 |
| 2 | 5.143 | 6.602 | 6.493 | 6.548 | 5.606 | 3.981 | 2.867 | 2.377 | 2.622 | 120.464 | 31.396 | 74.326 | 19.318 | 35.969 |
| 3 | 5.225 | 6.620 | 6.539 | 6.580 | 5.686 | 4.006 | 2.923 | 2.421 | 2.672 | 121.644 | 31.454 | 76.715 | 20.062 | 38.214 |
| 4 | 5.293 | 6.638 | 6.572 | 6.605 | 5.744 | 4.027 | 2.969 | 2.458 | 2.714 | 122.589 | 31.451 | 78.725 | 20.690 | 39.516 |
| 5 | 5.352 | 6.649 | 6.598 | 6.624 | 5.789 | 4.045 | 3.008 | 2.491 | 2.750 | 123.277 | 31.322 | 80.489 | 21.243 | 40.362 |
| 6 | 5.403 | 6.659 | 6.619 | 6.639 | 5.826 | 4.059 | 3.042 | 2.520 | 2.781 | 123.854 | 31.262 | 82.031 | 21.732 | 40.952 |
| 7 | 5.448 | 6.667 | 6.637 | 6.652 | 5.857 | 4.072 | 3.073 | 2.546 | 2.810 | 124.340 | 31.154 | 83.403 | 22.180 | 41.343 |
| 8 | 5.488 | 6.674 | 6.653 | 6.664 | 5.884 | 4.083 | 3.100 | 2.570 | 2.835 | 124.770 | 31.080 | 84.632 | 22.585 | 41.640 |
| 9 | 5.523 | 6.680 | 6.666 | 6.673 | 5.909 | 4.093 | 3.125 | 2.593 | 2.859 | 125.126 | 30.976 | 85.715 | 22.969 | 41.947 |
| 10 | 5.554 | 6.686 | 6.679 | 6.683 | 5.930 | 4.101 | 3.147 | 2.613 | 2.880 | 125.483 | 30.965 | 86.680 | 23.307 | 42.115 |

${ }^{\mathrm{a}} V_{P}\left(90^{\circ}\right) \mathrm{a}$ and $V_{P}\left(90^{\circ}\right) \mathrm{b}$ are P wave velocities for propagation perpendicular to the symmetry axis where $V_{P}\left(90^{\circ}\right) \mathrm{a}$ is orthoganal to $V_{P}\left(90^{\circ}\right) \mathrm{b} . V_{P}\left(90^{\circ}\right)_{\text {ave }}$ is the average compressional wave velocity for propagation perpendicular to the symmetry axis (derived from $V_{P}\left(90^{\circ}\right)$ a and $\left.V_{P}\left(90^{\circ}\right) \mathbf{b}\right) . V_{\text {SVave }}$ is the average shear wave velocity derived from $V_{S V}\left(0^{\circ}\right)$ and $V_{s v}\left(90^{\circ}\right)$. Velocities are in $\mathrm{km} / \mathrm{s} . P^{;}$, pressure in kbar. $C_{11}, C_{33}, C_{44}, C_{12}$ and $C_{13}$ are derived from $V_{P}\left(0^{\circ}\right), V_{P}\left(90^{\circ}\right)_{\text {ave }}$, $V_{P}\left(45^{\circ}\right), V_{S H}\left(90^{\circ}\right)$ and $V_{S V a v e}$ and are in MPa. Note that densities are the average of four measured cores for each sample.

Table 2. Orthorhombic Samples ${ }^{a}$

| $P$ | $V_{P}\left(0^{\circ}\right)$ | $V_{P}\left(90^{\circ}\right) \mathrm{a}$ | $V_{P}\left(90^{\circ}\right) \mathrm{b}$ | $V_{P}\left(45^{\circ}\right)$ | $V_{S H}\left(90^{\circ}\right)$ | $V_{S V}\left(0^{\circ}\right)$ | $V_{S V}\left(90^{\circ}\right)$ |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| Sample |  |  |  |  |  |  | A-5 (Haast schist (garnet zone $)$ ), Density |
| $=2.656$ | g/cm |  |  |  |  |  |  |
| 1 | 4.905 | 6.300 | 5.901 | 5.294 | 3.457 | 3.127 | 2.385 |
| 2 | 5.207 | 6.041 | 6.045 | 5.552 | 3.620 | 3.303 | 2.594 |
| 3 | 5.354 | 6.259 | 6.094 | 5.658 | 3.671 | 3.373 | 2.695 |
| 4 | 5.454 | 6.476 | 6.123 | 5.723 | 3.691 | 3.409 | 2.752 |
| 5 | 5.532 | 6.497 | 6.145 | 5.772 | 3.702 | 3.433 | 2.789 |
| 6 | 5.596 | 6.517 | 6.163 | 5.811 | 3.710 | 3.452 | 2.817 |
| 7 | 5.651 | 6.532 | 6.178 | 5.844 | 3.716 | 3.467 | 2.840 |
| 8 | 5.699 | 6.546 | 6.191 | 5.874 | 3.722 | 3.481 | 2.859 |
| 9 | 5.742 | 6.558 | 6.203 | 5.899 | 3.727 | 3.492 | 2.875 |
| 10 | 5.780 | 6.569 | 6.213 | 5.923 | 3.731 | 3.503 | 2.890 |

Sample A-12 (Haast schist (garnet zone)), Density $=2.719 \mathrm{~g} / \mathrm{cm}^{3}$

| 1 | 5.066 | 5.926 | 5.537 | 5.107 | 3.469 | 3.062 | 2.673 |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| 2 | 5.536 | 6.303 | 5.986 | 5.576 | 3.715 | 3.289 | 2.820 |
| 3 | 5.678 | 6.386 | 6.122 | 5.712 | 3.795 | 3.350 | 2.864 |
| 4 | 5.757 | 6.469 | 6.192 | 5.784 | 3.834 | 3.379 | 2.890 |
| 5 | 5.816 | 6.503 | 6.242 | 5.836 | 3.860 | 3.399 | 2.910 |
| 6 | 5.865 | 6.537 | 6.283 | 5.880 | 3.881 | 3.415 | 2.927 |
| 7 | 5.907 | 6.561 | 6.318 | 5.916 | 3.898 | 3.428 | 2.941 |
| 8 | 5.943 | 6.584 | 6.348 | 5.949 | 3.913 | 3.440 | 2.953 |
| 9 | 5.975 | 6.603 | 6.375 | 5.977 | 3.926 | 3.450 | 2.964 |
| 10 | 6.005 | 6.621 | 6.399 | 6.003 | 3.938 | 3.460 | 2.974 |


| Sample |  |  |  |  |  |  |  |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| A-13 (Haast schist (biotite zone), Density | $=2.703$ | $\mathrm{~g} / \mathrm{cm}^{3}$ |  |  |  |  |  |
| 1 | 5.090 | 5.983 | 6.700 | 5.189 | 3.746 | 3.073 | 3.077 |
| 2 | 5.596 | 6.355 | 6.918 | 5.753 | 3.909 | 3.253 | 3.286 |
| 3 | 5.721 | 6.434 | 6.968 | 5.796 | 3.949 | 3.055 | 3.332 |
| 4 | 5.784 | 6.512 | 6.997 | 6.000 | 3.972 | 3.337 | 3.353 |
| 5 | 5.831 | 6.542 | 7.018 | 6.060 | 3.990 | 3.361 | 3.369 |
| 6 | 5.869 | 6.572 | 7.035 | 6.109 | 4.006 | 3.380 | 3.382 |
| 7 | 5.902 | 6.593 | 7.050 | 6.150 | 4.017 | 3.397 | 3.392 |
| 8 | 5.931 | 6.614 | 7.063 | 6.187 | 4.028 | 3.411 | 3.402 |
| 9 | 5.957 | 6.630 | 7.074 | 6.219 | 4.037 | 3.424 | 3.410 |
| 10 | 5.980 | 6.646 | 7.084 | 6.249 | 4.046 | 3.436 | 3.417 |


| Sample A-16 (Haast schist (chlorite IV zone)), Density $=2.770$ |  |  |  |  |  |  | $\mathrm{~g} / \mathrm{cm}^{3}$ |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| 1 | 4.504 | 6.212 | 6.386 | 5.235 | 3.903 | 2.773 | 2.903 |
| 2 | 5.102 | 6.453 | 6.632 | 5.785 | 4.014 | 2.976 | 3.068 |
| 3 | 5.338 | 6.521 | 6.732 | 5.990 | 4.042 | 3.055 | 3.120 |
| 4 | 5.474 | 6.588 | 6.786 | 6.093 | 4.054 | 3.101 | 3.147 |
| 5 | 5.574 | 6.621 | 6.822 | 6.159 | 4.062 | 3.136 | 3.166 |
| 6 | 5.656 | 6.654 | 6.850 | 6.211 | 4.068 | 3.164 | 3.182 |
| 7 | 5.726 | 6.678 | 6.873 | 6.253 | 4.073 | 3.188 | 3.195 |
| 8 | 5.787 | 6.701 | 6.893 | 6.290 | 4.078 | 3.208 | 3.207 |
| 9 | 5.842 | 6.719 | 6.911 | 6.323 | 4.082 | 3.227 | 3.217 |
| 10 | 5.892 | 6.737 | 6.927 | 6.353 | 4.085 | 3.244 | 3.226 |


| Sample A-43 (Haast schist (chlorite IV zone)), Density $=2.626$ |  |  |  |  |  |  |  |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| $\mathrm{~g} / \mathrm{cm}^{3}$ |  |  |  |  |  |  |  |
| 1 | 5.454 | 6.484 | 6.484 | 5.752 | 3.736 | 3.243 | 3.377 |
| 2 | 5.919 | 6.779 | 6.779 | 6.304 | 3.885 | 3.416 | 3.550 |
| 3 | 6.031 | 6.842 | 6.862 | 6.418 | 3.927 | 3.462 | 3.592 |
| 4 | 6.090 | 6.904 | 6.904 | 6.461 | 3.949 | 3.486 | 3.608 |
| 5 | 6.135 | 6.931 | 6.934 | 6.488 | 3.965 | 3.505 | 3.618 |
| 6 | 6.171 | 6.958 | 6.958 | 6.510 | 3.978 | 3.520 | 3.626 |
| 7 | 6.203 | 6.977 | 6.978 | 6.528 | 3.989 | 3.533 | 3.633 |
| 8 | 6.230 | 6.996 | 6.996 | 6.544 | 3.999 | 3.544 | 3.639 |
| 9 | 6.255 | 7.011 | 7.110 | 6.558 | 4.008 | 3.554 | 3.644 |
| 10 | 6.277 | 7.025 | 7.025 | 6.570 | 4.015 | 3.563 | 3.649 |


| Sample A-18 (Haast schist (chlorite III zone)), Density $=2.765 \mathrm{~g}^{\prime} \mathrm{cm}^{3}$ |  |  |  |  |  |  |  |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| 1 | 4.749 | 5.854 | 6.087 | 5.173 | 3.640 | 2.892 | 2.839 |
| 2 | 5.283 | 6.163 | 6.408 | 5.610 | 3.787 | 3.113 | 3.078 |
| 3 | 5.532 | 6.279 | 6.557 | 5.799 | 3.845 | 3.215 | 3.187 |
| 4 | 5.675 | 6.395 | 6.640 | 5.923 | 3.872 | 3.274 | 3.246 |
| 5 | 5.774 | 6.447 | 6.693 | 6.020 | 3.889 | 3.317 | 3.285 |
| 6 | 5.850 | 6.498 | 6.733 | 6.100 | 3.900 | 3.350 | 3.314 |
| 7 | 5.913 | 6.532 | 6.765 | 6.169 | 3.910 | 3.378 | 3.338 |
| 8 | 5.967 | 6.566 | 6.793 | 6.229 | 3.918 | 3.403 | 3.358 |
| 9 | 6.015 | 6.593 | 6.817 | 6.283 | 3.924 | 3.424 | 3.376 |
| 10 | 6.059 | 6.619 | 6.838 | 6.332 | 3.931 | 3.444 | 3.391 |

Table 2. (continued)

| $P$ | $V_{P}\left(0^{\circ}\right)$ | $V_{P}\left(90^{\circ}\right) \mathrm{a}$ | $V_{P}\left(90^{\circ}\right) \mathrm{b}$ | $V_{P}\left(45^{\circ}\right)$ | $V_{S H}\left(90^{\circ}\right)$ | $V_{S V}\left(0^{\circ}\right)$ | $V_{S V}\left(90^{\circ}\right)$ |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| Sample |  |  |  |  |  |  |  |
| A-35 | $($ Haast schist (chlorite III zone)), Density $=2.740$ | $\mathrm{g} / \mathrm{cm}^{3}$ |  |  |  |  |  |
| 1 | 5.239 | 6.509 | 5.870 | 5.169 | 3.769 | 3.193 | 3.300 |
| 2 | 5.488 | 6.655 | 6.065 | 5.495 | 3.834 | 3.298 | 3.399 |
| 3 | 5.608 | 6.712 | 6.158 | 5.631 | 3.864 | 3.346 | 3.438 |
| 4 | 5.688 | 6.768 | 6.218 | 5.716 | 3.882 | 3.377 | 3.461 |
| 5 | 5.751 | 6.800 | 6.262 | 5.779 | 3.895 | 3.401 | 3.478 |
| 6 | 5.802 | 6.831 | 6.299 | 5.831 | 3.905 | 3.420 | 3.491 |
| 7 | 5.846 | 6.854 | 6.329 | 5.876 | 3.914 | 3.437 | 3.502 |
| 8 | 5.884 | 6.876 | 6.356 | 5.915 | 3.922 | 3.451 | 3.512 |
| 9 | 5.918 | 6.894 | 6.380 | 5.949 | 3.928 | 3.463 | 3.520 |
| 10 | 5.949 | 6.912 | 6.401 | 5.980 | 3.934 | 3.475 | 3.528 |

Sample A-33 (Haast schist (chlorite II zone)), Density $=2.626 \mathrm{~g} / \mathrm{cm}^{3}$

| 1 | 4.711 | 5.622 | 5.271 | 4.865 | 3.142 | 2.974 | 3.110 |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| 2 | 5.330 | 5.996 | 5.760 | 5.417 | 3.656 | 3.292 | 3.309 |
| 3 | 5.611 | 6.105 | 5.972 | 5.659 | 3.742 | 3.410 | 3.387 |
| 4 | 5.768 | 6.213 | 6.084 | 5.792 | 3.788 | 3.475 | 3.430 |
| 5 | 5.874 | 6.263 | 6.155 | 5.880 | 3.821 | 3.522 | 3.459 |
| 6 | 5.954 | 6.313 | 6.207 | 5.948 | 3.848 | 3.559 | 3.482 |
| 7 | 6.021 | 6.349 | 6.250 | 6.004 | 3.870 | 3.591 | 3.502 |
| 8 | 6.079 | 6.384 | 6.286 | 6.052 | 3.890 | 3.620 | 3.519 |
| 9 | 6.130 | 6.349 | 6.317 | 6.095 | 3.907 | 3.645 | 3.534 |
| 10 | 6.176 | 6.439 | 6.345 | 6.134 | 3.923 | 3.667 | 3.547 |

${ }^{\mathrm{a}} P$ is pressure in kbar. Velocities are in $\mathrm{km} / \mathrm{s} . V_{P}\left(90^{\circ}\right) \mathrm{a}$ and $V_{P}\left(90^{\circ}\right) \mathrm{b}$ are $P$ wave velocities for propagation perpendicular to the symmetry axis ( $V_{P}\left(90^{\circ}\right)$ a is orthogonal to $\left.V_{P}\left(90^{\circ}\right) \mathrm{b}\right)$. Note that densities are the average of four measured cores for each sample.
tensen, 1995]. Phase and group velocities are equal for propagation directions parallel and perpendicular to the symmetry axis.
In an anisotropic medium, $V_{P}$ is usually a quasi-compressional wave and $V_{S V}$ is a quasi-shear wave since particle motion is neither parallel or perpendicular to the symmetry axis for most angles of incidence [Auld, 1990; Winterstein, 1990] (Figure 7). $V_{P}$ and $V_{s v}$ are only pure waves (particle motion is parallel or perpendicular to the symmetry axis) for $\theta=0^{\circ}$ and $90^{\circ} . V_{S H}$, however, is always a pure shear wave since particle motion is always within the foliation (perpendicular to the symmetry axis) [Auld, 1990; Winterstein, 1990], and more importantly, the particle motion always remains perpendicular to the direction of propagation. We will refer to $V_{P}$ and $V_{S V}$ as $q V_{P}$ and $q V_{S V}$ for completeness.

We have calculated $q V_{P}, q V_{S V}$ and $V_{S H}$ for all angles of incidence $(\theta)$ from zero (parallel to the symmetry axis) to $90^{\circ}$ (perpendicular to the symmetry axis, parallel to foliation) for the six samples we analyzed (Figure 8).

We have estimated the maximum error resulting from the approximation of orthorhombic symmetry to transverse isotropy in our study (Figure 9). We use Haast Schist sample A-1 (the most orthorhombic sample we approximated to transverse isotropy) to calculate the percentage difference between $q V_{P}, q V_{S V}$, and $V_{S H}$ calculated from average values and those calculated from the fastest and slowest values. The percentage differences shown in Figure 9 are the maximum for this sample, and since this sample is the most orthorhombic of the samples for which the approximation was done, it should also be considered the maximum error associated with any of our samples. The $q V_{P}$ error $(< \pm 0.5 \%)$ is much smaller than the error in $q V_{S V}$ or $V_{S H}( \pm 4 \%)$.

The variation of $q V_{p}$ with angle depends on the relative value of $V_{P}\left(45^{\circ}\right)$ compared with $V_{P}\left(0^{\circ}\right)$ and $V_{P}\left(90^{\circ}\right)$ (Figures 5 and 8 ). Generally, there is little increase in $q V_{P}$ between $\theta=0$ and $\sim 40^{\circ}$ but then an increase with angle from $\sim 40^{\circ}$ to a maximum at $90^{\circ}$ (Figure 8).

Most of the samples have $V_{P}\left(45^{\circ}\right)$ values that are less than the value midway between the average $V_{P}\left(0^{\circ}\right)$ and $V_{p}\left(90^{\circ}\right)$ (Figure 5 a ). They show a maximum in the $q V_{s v}$ versus angle plot (Figure

Table 3. Modal Analyses ${ }^{\text {a }}$

|  | Sample | Biotite | Musc. | Chl. | Qtz | Plag. | Gnt | Mag. | Sph. | Opaq. Epid. Carb. |  |
| :--- | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| Single-crystal $P$ wave velocity ${ }^{\text {b }}$, km/s |  | 5.26 | 5.78 | 5.26 | 6.09 | 6.22 | 8.52 | 7.39 | 7.39 | 7.39 | N/A |
| Haast schist (garnet zone) | A-1 | 12 | 13 | 0 | 65 | 7 | 2 | 1 | 0 | 0 | 0 |
| Haast schist (garnet zone) | A-5 | 4 | 50 | 3 | 20 | 17 | 4 | 1 | 1 | 0 | 0 |
| Haast schist (garnet zone) | A-12 | 15 | 20 | 1 | 25 | 37 | 0 | 0 | 0 | 2 | 0 |
| Haast schist (biotite zone) | A-13 | 10 | 22 | 3 | 55 | 6 | 0 | 0 | 0 | 4 | 0 |
| Haast schist (chlorite IV subzone) | A-16 | 0 | 25 | 2 | 40 | 13 | 0 | 0 | 0 | 20 | 0 |
| Haast schist (chlorite IV subzone) | A-43 | 2 | 2 | 3 | 42 | 48 | 0 | 0 | 0 | 0 | 0 |
| Haast schist (chlorite III subzone) | A-18 | 0 | 30 | 2 | 50 | 16 | 0 | 0 | 0 | 0 | 2 |
| Haast schist (chlorite III subzone) | A-35 | 5 | 37 | 2 | 49 | 7 | 0 | 0 | 0 | 0 | 0 |
| Haast schist (chlorite II subzone) | A-33 | 15 | 5 | 3 | 54 | 20 | 0 | 0 | 0 | 0 | 3 |
| Pelona schist | LA-1 | 0 | 35 | 0 | 34 | 21 | 0 | 0 | 0 | 0 | 10 |
| Chugach phyllite | TA-2 | N/A | N/A | N/A | N/A | N/A | N/A | N/A | N/A | N/A | N/A |
| Chugach phyllite | TA-23 | N/A | N/A | N/A | N/A | N/A | N/A | N/A | N/A | N/A | N/A |
| N/A |  |  |  |  |  |  |  |  |  |  |  |
| Coldfoot schist | TA-80 | 0 | 42 | 3 | 38 | 16 | 0 | 1 | 0 | 0 | 0 |
| Poultney slate | VT-1 | N/A | N/A | N/A | N/A | N/A | N/A | N/A | N/A | N/A | N/A |

${ }^{\text {a }}$ N/A, sample too fine grained for mineralogical analysis.
${ }^{6}$ Data from Christensen [1965].
${ }^{\text {c }}$ Musc., muscovite; Chl., chlorite; Qtz, quartz; Plag., plagioclase; Gnt, garnet; Mag., magnetite, Sph., sphene; Opaq., opaques; Epid., epidote; Carb., carbonate.
8). We demonstrate the effect of the $V_{P}\left(45^{\circ}\right)$ value with respect to $V_{P}\left(0^{\circ}\right)$ and $V_{P}\left(90^{\circ}\right)$ on the shape of the $q V_{S V}$ versus angle of incidence curve (Figure 10) using values based on sample A-1 at 6 kbar (Table 1). We increase $V_{P}\left(45^{\circ}\right)$ at $0.1 \mathrm{~km} / \mathrm{s}$ intervals between $V_{P}\left(0^{\circ}\right)$ and $V_{P}\left(90^{\circ}\right)$ (Figure 10$)$. Our plot shows that for $V_{P}\left(45^{\circ}\right)>$
$\left[V_{P}\left(0^{\circ}\right)+V_{P}\left(90^{\circ}\right)\right] / 2$, the $q V_{S V}$ curve has a minimum. If $V_{P}\left(45^{\circ}\right)=$ $\left[V_{P}\left(0^{\circ}\right)+V_{P}\left(90^{\circ}\right)\right] / 2$, the $q V_{S V}$ curve is flat and for $V_{P}\left(45^{\circ}\right)<$ $\left[V_{P}\left(0^{\circ}\right)+V_{P}\left(90^{\circ}\right)\right] / 2$, the $q V_{S V}$ curve has a maximum. This is supported by our samples (compare sample A-16 with all the other samples in Figures 5 and 8). These observations are important


Figure 5. Range of compressional wave velocities for each sample studied. (a) Samples which we approximate to transverse isotropy (shown at 6 kbar ). Percent anisotropy is calculated from the average fast velocity (cross) minus the slowest velocity (solid circle) divided by the average of the three orthogonal velocities. (b) Samples which are significantly orthorhombic (shown at 6 kbar ). (c) Our measurements plotted with Voigt and Reuss averages (measurements shown for 2-kbar pressure appropriate for single-crystal data). The sin-gle-crystal $P$ wave velocity data that we use (Table 3) are Voigt-Reuss-Hill averages [Christensen, 1966]. The Voigt and Reuss averages calculated for our samples follow the method of Christensen [1966]. We have assumed that chlorite has the same velocity as biotite, that the carbonates are calcite, that the feldspar is oligoclase, and that the opaques and sphene have the same velocity as magnetite. There are no velocity data for epidote. We therefore recalculate the modal fractions in Table 3 assuming that no epidote is present.


Figure 6. (a) All velocity measurements from sample A-1 (gar-net-oligoclase zone of the Haast schist) versus pressure. Solid line is compressional wave parallel to symmetry axis (perpendicular to foliation). Short dashed line is quasi-compressional wave at $45^{\circ}$ to symmetry axis. Long dashed line is compressional wave parallel to foliation but perpendicular to lineation. Shaded line is compressional wave parallel to lineation within the foliation. Solid line with crosses is the shear wave propagating parallel to foliation with particle motion also parallel to foliation. Solid line with diamonds is the shear wave propagating parallel to foliation with particle motion perpendicular to foliation. Solid line with squares is the shear wave propagating perpendicular to foliation. Symmetry axis is shown by the solid line in each schematic. Propagation direction is shown by the bold arrow in each schematic. (b) Velocity measurements from sample A-1 approximated to transverse isotropy. Solid line is compressional wave parallel to the symmetry axis. Short dashed line is quasicompressional wave at $45^{\circ}$ to symmetry axis. Long dashed line is compressional wave perpendicular to the symmetry axis. Solid line with crosses is the shear wave propagating parallel to the symmetry axis, which is equal to the shear wave propagating perpendicular to the symmetry axis with particle motion parallel to the symmetry axis. Solid line with diamonds is the shear wave propagating perpendicular to the symmetry axis with particle motion also perpendicular to the symmetry axis.
since they show that the $V_{P}\left(45^{\circ}\right)$ value is critical to the shape of the curve and should not be assumed if the measurement is not available. Our measurements of this parameter show that it is quite variable between samples (Figure 5).

Shear wave velocities ( $q V_{S V}$ and $V_{S H}$ ) are equal for waves propagating parallel to the symmetry axis $\left(\theta=0^{\circ}\right)$ since both waves exhibit particle motion within the foliation plane. Shear wave birefringence or splitting, where the two shear waves propagate with different velocities, is observed for shear waves propagating at an angle to the symmetry axis. The maximum difference in shear wave velocities occurs for shear waves propagating perpendicular to the symmetry axis $\left(\theta=90^{\circ}\right)$. In this case the particle motion for one shear wave is parallel to the symmetry axis (perpendicular to foliation) and for the other shear wave it is perpendicular to the symmetry axis (within the foliation). The $V_{S H}$ wave shows an increase in velocity as the angle from the symmetry axis increases (Figure 8). The $q V_{S V}$ wave, which is dependent on the $V_{P}\left(45^{\circ}\right)$ value (see equations (6) and (9)), generally shows an increase in velocity up to an angle of $40-45^{\circ}$ before decreasing again to the initial $\theta=0^{\circ}$ value at $\theta=90^{\circ}$ (Figure 8).

In most cases $q V_{s v}$ is the faster of the two shear waves for low angles of $\theta$. In these cases, as $\theta$ increases, the $q V_{s V}$ and $V_{s H}$ curves intersect, and $V_{S H}$ becomes the faster of the two shear waves (Figures $8 \mathrm{a}-8 \mathrm{~d}$ and 8 f ). In some cases the $V_{S H}$ curve rises more steeply than the $q V_{s v}$ curve at low angles of $\theta$. In these cases the two curves never cross, and $V_{S H}$ is the faster phase at all angles (Figure 8 e ). Where the two shear wave curves simply touch (e.g., $\theta=$ $0^{\circ}$ ), there is a "kiss"-type shear wave velocity singularity, and where the curves intersect, there is a line singularity [Winterstein, 1990]. More information about line and kiss-type singularities is given by Crampin [1991]. For the majority of cases in which the two shear wave curves intersect, the amount of shear wave splitting decreases to zero at the line singularity angle before rising again to a maximum at $\theta=90^{\circ}$. The angle at which the line singularity occurs varies from $\sim 30^{\circ}$ to $60^{\circ}$.

### 4.2. Variation of $V_{P} / V_{s}$ With Propagation Direction

The ratio $V_{P} / V_{s}$ is, in combination with the compressional and shear wave velocity measurements themselves, more sensitive to changes in lithology than either velocity alone. $V_{P} / V_{s}$ is often calculated from compressional and shear wave velocity models for a particular region to help determine the lithology of the subsurface in that region. In isotropic regions, Poisson's ratio, which is also calculated from compressional and shear wave velocities, may also be used to aid lithological interpretation. Poisson's ratio, however, is not meaningful for an anisotropic medium.

We have calculated $V_{P} / V_{S}$ for our samples at each propagation angle. Since there are two shear wave values, we have calculated different $V_{P} / V_{s}$ ratios for each shear wave phase (Figure 11). Our results show that there is an extremely wide range of $V_{P} / V_{s}$ with propagation direction, and the implications of this will be discussed further in section 5.4.

## 5. Discussion

We now discuss six issues that arise from our transverse isotropy measurements on samples from various schist belts.

### 5.1. The Significance of a Crustal Component of Shear Wave Splitting

As noted earlier, the two shear wave phases travel with different velocities except at singularities. For a single homogeneous anisotropic layer this phenomenon is known as shear wave birefringence. When several layers are involved, some anisotropic (symmetry axes not constrained to be in any particular orientation) and some isotropic, the cumulative effect of birefringence in the different anisotropic layers is known as shear wave splitting [Winterstein, 1990]. Shear wave splitting over a path between the core and the surface of the Earth can be measured using teleseismic $S K S$ waves (see summary on $S K S$ measurements and shear wave splitting by Savage (1999)). Generally, SKS splitting ob-


Figure 7. Schematic diagrams showing $q V_{S V}$ and $V_{S H}$ as defined for a horizontally propagating shear wave. Dashed line is symmetry axis. Arrows show $S$ wave propagation direction and "zig-zag" lines show particle motion directions. (a) Diagram showing angle of incidence of $0^{\circ}$. Particle motion for both $V_{S V}$ and $V_{S H}$ is within the foliation (perpendicular to the symmetry axis) and $V_{S V}$ equals $V_{S H}$. (b) Diagram showing angle of incidence of $45^{\circ}$. Particle motion for $q V_{S V}$ is neither parallel nor perpendicular to the symmetry axis. Particle motion for $V_{S H}$ is perpendicular to the symmetry axis. (c) Diagram showing angle of incidence of $90^{\circ}$. Particle motion for $V_{S V}$ is parallel to the symmetry axis. Particle motion for $V_{S H}$ is perpendicular to the symmetry axis.
servations are assumed to result primarily from anisotropy in the mantle, and possible crustal contributions are ignored. Previously, it has been assumed that the crust does not contribute enough to shear wave splitting to be significant [Barruol and Mainprice, 1993; Crampin, 1994; Klosko et al., 1999; Savage, 1999]. Our results suggest that the shear wave splitting from suitably oriented schist terranes of large extent and thickness may contribute a significant amount of shear wave splitting to the SKS observations (Figure 12).

We have calculated the amount of shear wave splitting ( $\Delta t$ ) that would result from a vertically propagating teleseismic shear wave traveling through a $10-\mathrm{km}$-thick region of transversely isotropic anisotropic material with bulk orientation varying from a vertical symmetry axis ( $\phi=0^{\circ}$, horizontal foliation) to a horizontal symmetry axis ( $\phi=90^{\circ}$, vertical foliation). We used samples A-1 (Haast schist) and TA-80 (Coldfoot schist) to demonstrate the different curve shapes that result from the relative behavior of $V_{S H}$ and $q V_{S V}$ (Figure 12a). Some samples have no shear wave


Figure 8. Plots of velocity versus angle from the symmetry axis (perpendicular to foliation) for each sample. Lines with no symbols are quasi-compressional $\left(q V_{P}\right)$ wave velocities, lines with crosses are the quasi-shear wave velocities ( $q V_{S V}$ ) and lines with open squares are true shear wave velocities ( $V_{S H}$ ). Grayscale is consistent between compressional and shear waves and represents measurements and calculations at a particular pressure. Solid represents lowest pressure ( 1 kbar ) and lightest shading represents highest pressure ( 9 kbar ). Intermediate lines increment by 2 kbar each.


Figure 9. Plot of percentage difference versus angle from the symmetry axis for sample A-1 at 6 kbar . Three calculations of $q V_{P}, q V_{S V}$, and $V_{S H}$ are done using the fast and slow measured values of $V_{P}\left(90^{\circ}\right)$ and $V_{S H}\left(0^{\circ}\right)$ and the average of these values. The same values for $V_{P}\left(0^{\circ}\right), V_{P}\left(45^{\circ}\right)$, and $V_{S H}\left(90^{\circ}\right)$ are used in each calculation. The difference between the calculation using the fastest velocities ( $P$ and $S$ wave) and that using the average velocities is shown by a solid line. The difference between the calculation using the slowest velocities ( $P$ and $S$ wave) and that using the average velocities is shown by a dashed line.
splitting at a specific angle due to the shear wave line singularity (e.g., sample A-1, Figures 8a and 12a). Others show shear wave splitting at all angles except $0^{\circ}$ (e.g., $50^{\circ}$ for sample TA-80, Figures 8 e and 12a).

We have also calculated the amount of shear wave splitting $(\Delta t)$ that would result from a vertically propagating teleseismic shear wave traveling through a $5-, 10-$, and $20-\mathrm{km}$-thick region of anisotropic material with bulk orientation varying from a vertical symmetry axis ( $\phi=0^{\circ}$, horizontal foliation) to a horizontal symmetry axis ( $\phi=90^{\circ}$, vertical foliation) (Figure 12b). We used samples A-1 (garnet-oligoclase zone of the Haast schist terrane), TA-2 (Chugach phyllite), and LA-1 (Pelona schist) for our examples (Figure 12b). The schists in the Chugach terrane are known to have steeply dipping foliation. The Haast schist has nearvertical foliation in outcrop. The subsurface orientation of the Pelona schist foliation is not known. Our calculations show that for a $5-\mathrm{km}$-thick anisotropic layer as much as 0.2 s of splitting could occur for a schistose body with vertical foliation (horizontal symmetry axis, $\phi=90^{\circ}$ ). If the thickness of the anisotropic body increases to 20 km , shear wave splitting of $>0.2 \mathrm{~s}$ could occur for schist orientations where the symmetry axis is at only $30^{\circ}$ from vertical (foliation dipping at $30^{\circ}$ ). For these three schists, there would be no shear wave splitting if the bulk orientation of the schistose body was such that the symmetry axis was tilted at $\sim 50^{\circ}$ from vertical (foliation dipping at $\sim 40^{\circ}$ ). For a more steeply dipping symmetry axis the degree of splitting would increase reaching a maximum of $0.7-0.9 \mathrm{~s}$ (samples A-1 and TA-2) for a horizontal symmetry axis (vertical foliation). These measured values suggest larger amounts of shear wave splitting than estimated from quantitative modeling based on single-crystal velocities
(0.1-0.2 s over 10 km thickness with vertical foliation) [Barruol and Mainprice, 1993].

A global compilation of shear wave splitting observations shows up to 2.4 s of shear wave splitting has been measured between the core and the Earth's surface beneath continental regions [Silver, 1996; Savage, 1999]. Shear wave splitting observations have been made at Landers, California (open circle, Figure 3d), College, Alaska (open circle, Figure 3b), and on South Island, New Zealand (open circle, Figure 2).

Our results suggest that large, thick anisotropic regions with regionally dipping foliation of $30^{\circ}$ or more could contribute between a quarter and a fifth of the observed shear wave splitting measurements (Table 4). The values we calculate will most likely be a maximum since we are assuming that the petrofabric of the entire terrane is consistently oriented. There may be significant local variations at the scale of the seismic wavelength due to rheological inhomogeneities that would reduce the bulk anisotropy. Our results affect the interpretation of shear wave splitting results from teleseismic earthquakes where the anisotropy is generally assumed to come from the mantle and suggest a significant amount of shear wave splitting could come from the crust. Our data can be used in shear wave splitting models to remove the assumed crustal contribution. If an appropriate thickness and orientation of anisotropic material are assumed, the contribution to the


Figure 10. Plot of velocity versus angle from the symmetry axis (perpendicular to foliation) for different $V_{P}\left(45^{\circ}\right)$ values (with respect to $V_{P}\left(0^{\circ}\right)$ and $V_{P}\left(90^{\circ}\right)$ ). The velocity values used in this example are based on sample A-1 at 6 kbar . In this example density is $2.718 \mathrm{~g} / \mathrm{cm}^{3}, V_{P}\left(90^{\circ}\right)$ is $6.5 \mathrm{~km} / \mathrm{s}, V_{P}\left(0^{\circ}\right)$ is $5.75 \mathrm{~km} / \mathrm{s}, V_{S H}\left(0^{\circ}\right)$ is $3.332 \mathrm{~km} / \mathrm{s}$ and $V_{S H}\left(90^{\circ}\right)$ is $3.931 \mathrm{~km} / \mathrm{s}$. The value of $V_{P}\left(45^{\circ}\right)$ was varied to examine the effect that it has on the shape of the $q V_{S V}$ curve. The grayscale is consistent from $q V_{S V}$ to $q V_{P}\left(V_{S H}\right.$ (dashed line with crosses) is independent of $V_{P}\left(45^{\circ}\right)$ and is the same for all calculations). Light shading represents fast $V_{P}\left(45^{\circ}\right)(6.5,6.4,6.3$, and $6.2 \mathrm{~km} / \mathrm{s}$ ) while dark shading and solid represent slow $V_{P}\left(45^{\circ}\right)$ values $(6.1,6.0,5.9,5.8$ and $5.75 \mathrm{~km} / \mathrm{s})$. The flat $q V_{S V}$ curve is for a $V_{P}\left(45^{\circ}\right)$ value of $6.125 \mathrm{~km} / \mathrm{s}$, which is midway between $V_{P}\left(90^{\circ}\right)$ and $V_{P}\left(0^{\circ}\right)$.


Figure 11. Plots of $V_{P} / V_{S}$ ratio versus angle from the symmetry axis (perpendicular to foliation) for each sample. Lines with open squares are $V_{P} / V_{S}$ calculated from $q V_{P}$ and $q V_{S V}$, lines with no symbols are $V_{P} / V_{S}$ calculated from $q V_{P}$ and $V_{S H}$. Grayscale is consistent between the two determinations of $V_{P} / V_{S}$ and also between the velocities used to calculate them in Figure 8. Grayscale represents measurements and calculations at a particular pressure. Solid represents the lowest pressure ( 1 kbar ), and lightest shading represents the highest pressure ( 9 kbar ). Intermediate lines increment by 2 kbar each.
shear wave splitting can be estimated and removed from the observed splitting value. The remaining amount of splitting can then be modeled and attributed to the mantle.

### 5.2. The Implication of Two Shear Wave Phases for Active Source Seismic Data

Earlier, we observed that in most cases $q V_{S V}$ is the faster of the two shear wave phases for low angles of $\theta$ and that as $\theta$ increases, the $q V_{S V}$ and $V_{S H}$ curves intersect at a line singularity and $V_{S H}$ becomes the faster of the two shear waves (Figures 8a-8d and 8f). In some cases the $V_{S H}$ curve rises more steeply than the $q V_{S V}$ curve at low angles of $\theta$, and in this case the $V_{S H}$ curve is the faster phase at all angles (Figure 8e).

This phenomenon affects the shear wave travel times picked from active source seismic data. Typically, only the first arriving shear wave is picked and modeled to produce a shear wave velocity model for a region. Our data show that the shear wave phase picked will probably not be the same phase for all propagation directions. In most cases the first arriving phase will be $q V_{S V}$ for propagation directions close to the symmetry axis (low values of $\theta$ ) and $V_{s H}$ for higher angles of $\theta$ (Figures 8 and 13a). For the cases where the two shear wave curves do not cross, $V_{S H}$ is the first arriving phase.

### 5.3. Variation in $V_{P} / V_{s}$ Using First Arriving Shear Wave Phases

The $V_{P} / V_{s}$ ratio that is used in lithological identification is generally calculated from compressional and shear wave velocity models. As discussed above, the shear wave velocity model is most likely to be derived from the first arriving shear wave phase. The $V_{P} / V_{S}$ ratio will correspondingly be biased toward a lower value than might be true if the $V_{\rho} / V_{S}$ calculation were based on a
single shear wave phase (Figure 13b). On the basis of the plots shown in Figures 11 and 13 the variation in $V_{P} / V_{S}$ with propagation direction will generally be reduced from the maximum variation seen in $V_{P} / V_{S}$ derived solely from $q V_{S V}$. The variation in $V_{P} / V_{S}$ based on the first arriving shear wave phase, however, still varies significantly from the value determined for propagation parallel to the symmetry axis, which is the angle of propagation for which published $V_{p} / V_{s}$ data are generally given [Christensen, 1996]. For some samples, $V_{P} / V_{s}$ (calculated from the first arriving shear wave phase) decreases with increasing propagation angle, while for other samples, there is a net increase in $V_{P} / V_{S}$ with propagation angle (Figure 11).

### 5.4. Implications for Lithological Identification

In field experiments we measure the group velocity (for both compressional and shear waves) since the waves are generated from point sources [Johnston and Christensen, 1995]. In the laboratory we measure phase velocities (for both compressional and shear waves) since the sources are plane waves [Dellinger and Vernik, 1994; Johnston and Christensen, 1995]. The phase and group velocities are equal for propagation directions parallel and perpendicular to the symmetry axis. In general, we are comparing group velocities in velocity models derived from active source seismic field data with phase velocities measured in the laboratory. Since the phase velocities are different from the group velocities, our lithological interpretation based on velocity correlation may be inaccurate. Johnson and Christensen [1995] show, however, that shales with up to $\sim 24 \% P$ wave anisotropy show very little difference between phase and group velocity. The group velocity only becomes significantly different from the phase velocity for shales with more extreme anisotropy. Our slate sample, which is most similar to shale, has a $P$ wave anisotropy of $\sim 21 \%$. The difference between phase and group velocity in this


Figure 12. (a) Plot of amount of shear wave splitting (seconds) versus angle of the transversely isotropic symmetry axis from vertical (bulk orientation of the anisotropic body) for a vertically traveling teleseismic wave traveling through a $10-\mathrm{km}$-thick anisotropic layer based on samples A-1 and TA-80. This plot demonstrates different curve shapes resulting from the relative behavior of $q V_{S V}$ and $V_{S H}$ (see Figure 8). (b) Plot of amount of shear wave splitting (seconds) versus angle of the symmetry axis from vertical (bulk orientation of the anisotropic body) for a vertically traveling teleseismic wave traveling through a $5-, 10$-, and $20-$ km-thick anisotropic layer based on samples TA-2, A-1, and LA1.
sample is unlikely to be significant. The other schist samples are even less anisotropic. Teleseismic arrivals may be considered as plane waves and are thus likely to yield phase velocities.

Our results show how the interpretation of lithology from compressional wave velocities alone could lead to a misinterpretation of lithology. If velocities are determined in a region with a uniformly dipping anisotropic layer, the velocities may be significantly different from those measured in the laboratory at $0^{\circ}$ or $90^{\circ}$ to the symmetry axis. The slow and fast velocities of a sample may each be interpreted as different lithologies based on published data (e.g., samples A-1 and VT-1, Table 5).
$V_{\rho} / V_{s}$ ratios determined from velocity models are generally compared with laboratory data. Published data may only be determined for the $\theta=0^{\circ}$ propagation direction or may consist of average values for anisotropic rocks. Published $V_{\rho} / V_{s}$ values of common crustal rocks fall in the range of 1.707-1.863 (determined from velocities measured at 6 kbar) [Christensen, 1996]. Significant outliers to this range include 1.492 for quartzite and 2.094 for serpentinite (determined from velocities measured at 6 kbar) [Christensen, 1996]. As discussed above, we have noted considerable variation in $V_{P} / V_{s}$ with propagation direction. For sample A-1 (at 6 kbar ), $V_{P} / V_{S}$ is 1.75 at $\theta=0^{\circ}$, reaches a minimum of 1.56 at $\theta=37^{\circ}$, and increases again to 1.65 at $\theta=90^{\circ}$
(based on first arriving shear waves) (Figure 13). This range of $V_{P} / V_{s}$ values encompasses the following lithologies based on published data: quartzite, granite, metagraywacke, granite gneiss, biotite (tonalite) gneiss, hornblendite, pyroxenite, and dunite, although most of these rocks have values between 1.73 and 1.75 [Christensen, 1996]. Values of $V_{P} / V_{s}$ between 1.5 and 1.7 , which result from our data, are not represented in the published data [Christensen, 1996]. The variation of $V_{P} / V_{s}$ values with propagation angle might make it difficult to determine lithology at all based on published values of $V_{P} / V_{s}$. However, if anisotropy in field-based studies is recognized, observations of $V_{P} / V_{s}$ variations with propagation direction will provide powerful constraints in crustal lithology interpretations.

### 5.5. Effect of Anisotropy on Active Source Seismic Experiments

Data from active source seismic experiments may include both near-vertical and wide-angle arrivals. Refraction/wide-angle reflection experiment geometries generate wide-angle ray paths,


Figure 13. (a) Plot of velocity versus angle from the symmetry axis for sample A-1 at 6 kbar . Solid line shows the first arriving shear wave phase which for this sample is $q V_{s v}$ for $\theta<48^{\circ}$ and $V_{s r^{\prime}}$ for $\theta>48^{\circ}$. (b) Plot of $V_{P} / V_{s}$ ratio versus angle from the symmetry axis for sample A-1 at 6 kbar. Solid line shows $V_{P} / V_{s}$ that would be calculated using the first arriving shear wave phase (Figure 13a).

Table 4. Shear Wave Splitting Observations

|  | Observed Shear Wave Splitting (s) | Foliation Orientation in Schist Belt | Fast Direction Determined From Observed Shear Wave Splitting | Contribution to Splitting From 10-km-Thick Crustal Layer With Vertical Foliation, s (Observed Splitting, \%) | Contribution to Splitting From 20-km-Thick Crustal Layer With Vertical Foliation, s (Observed Splitting, \%) |
| :---: | :---: | :---: | :---: | :---: | :---: |
| College, Alaska | $1.55{ }^{\text {a }}$ | Coldfoot schist, east-west (parallel to terrane) Chugach phyllite, eastwest, but arcuate (parallel to arcuate terrane) | east-west | 0.35 (23\%) | 0.7 (45\%) |
| Landers, California | $1.2{ }^{\text {b }}$ | unknown | upper layer, parallel to San Andreas fault (NW-SE) <br> lower layer, eastwest | 0.14 (12\%) | 0.28 (23\%) |
| South Island, New Zealand | $2.2{ }^{\text {c }}$ | parallel to Alpine fault (NE-SW) | parallel to Alpine fault (NE-SW) | 0.46 (21\%) | 0.92 (42\%) |

${ }^{\text {a }}$ Silver and Chan [1991].
${ }^{\mathrm{b}}$ Savage et al. [1990].
${ }^{\text {c }}$ Klosko et al. [1999].
which encompass a range of propagation orientations ranging from the takeoff angle to horizontal. Rays traveling with different orientations through a material with a given bulk anisotropy sample different velocities depending on whether rays travel in slow, fast, or intermediate directions, or most likely a sequential, and thus cumulative combination of these. A set of rays with a range of ray paths traveling through an anisotropic region will arrive at the surface with different travel times compared to a similar ray set passing through an equivalent isotropic region. If these travel time differences are significant, modeling techniques that assume isotropy, which are generally used to model crustal-scale seismic data, may not achieve a good fit to the data. Even if a good fit to the data is achieved, the model will not be correct since it is underparameterized.

In passive seismic experiments where teleseismic incoming waves are near vertical, we have already shown that crustal anisotropy could be significant for large terranes with uniformly steeply dipping foliation. Active source experiments, however, particularly those recording wide-angle energy, will be susceptible to horizontally oriented foliation as well as more steeply dipping foliation since a wide range of ray angles are involved.

Anisotropy is generally not accounted for in the modeling of active source seismic data. Laboratory data such as ours, however, can be used to help identify anisotropic regions in the subsurface from field seismic experiments with appropriate geometry. One case where an attempt was made to determine anisotropy
from an active source experiment was for the upper crust of the Trans-Alaska Crustal Transect (TACT). Given a favorable dipping bulk orientation of an anisotropic terrane, a refraction profile in a direction parallel to the fast direction of the anisotropy will have advanced crustal refraction arrival times in contrast to an orthogonal profile. Perpendicular refraction/wide-angle reflection profiles and rock samples for petrophysical analysis (including samples TA-2 and TA-23 used in this study) were collected as part of the TACT experiment in Alaska [Brocher et al., 1989; Brocher and Christensen, 1990; Fuis et al., 1991]. The orthogonal profiles within the schist terrane found mismatches in the $P$ wave velocity models in the upper 7 km , which suggested $10 \% P$ wave anisotropy due to preferred mineral orientation in the schists. This is consistent with the petrophysical measurements (samples TA-2 and TA-23, Figure 5). A shallow, high-resolution seismic reflection/refraction survey carried out in the same region was used to demonstrate shallow crustal anisotropy attributed to the phyllites of the Valdez group [Brocher et al., 1989; Brocher and Christensen, 1990]. In this case, a kink in the surface geometry of the seismic line allowed one subsurface region to be sampled at two azimuths. Parts of the line with different azimuths were modeled with different crustal velocities, which were also consistent with the anisotropic velocities [Brocher et al., 1989; Brocher and Christensen, 1990] measured in the laboratory for rocks at the surface in this region.
$P$ wave refraction studies across steeply dipping anisotropic

Table 5. Velocities Measured at 6-kbar Pressure and 20-km Depth ${ }^{\text {a }}$

| $\begin{gathered} \text { Velocity, } \\ \mathrm{km} / \mathrm{s} \end{gathered}$ | Candidate Lithology | Velocity, km/s | Difference Between Sample and Candidate Velocities, \% |
| :---: | :---: | :---: | :---: |
| Haast Schist (A-1) |  |  |  |
| Fast 6.5 | diorite | 6.5 | 0\% |
| Fast 6.5 | prehnite-pumpellyite facies basalt | 6.4 | 1.5\% |
| Fast 6.5 | diabase | 6.6 | 1.5\% |
| Fast 6.5 | mica quartz schist | 6.3 | 3\% |
| Fast 6.5 | zeolite facies basalt | 6.3 | 3\% |
| Slow 5.75 | metagraywacke | 5.6 | 2.6\% |
| Slow 5.75 | andesite | 5.6 | 2.6\% |
| Poultney Slate (VT-1) |  |  |  |
| Slow 5.75 | basalt | 5.9 | 2.6\% |
| Fast 6.6 | diabase | 6.6 | 0\% |
| Fast 6.6 | prehnite-pumpellyite facies basalt | 6.4 | 3\% |
| Fast 6.6 | amphibolite | 6.8 | 3\% |
| Fast 6.6 | greenschist facies basalt | 6.8 | 3\% |
| Fast 6.6 | mafic granulite | 6.8 | 3\% |
| Slow 5.4 | serpentinite | 5.3 | 1.9\% |

${ }^{\text {a Data are from Christensen and Mooney [1995]. Isotropic candidate lithologies }}$
within $3 \%$ of the sample velocity are shown.
shales were carried out to determine the insitu anisotropic parameters of these shales [Leslie and Lawton, 1999]. Head wave velocities along strike of the shales were faster than those perpendicular to strike. The results from this field study are consi stent with measurements made on shales in the laboratory.

Identifying an anisotropic medium in the subsurface from field data will not allow us to distinguish between a truly anisotropic medium and an effective anisotropic medium (a medium consisting of thin isotropic layers that behaves as a bulk anisotropic medium). Additional information from samples collected in outcrop or from boreholes will also be required.

## 6. Conclusions

Low- to medium-grade metamorphic rocks with strong fabrics produced by lattice preferred orientations of highly anisotropic minerals such as mica are common constituents of many orogenic belts throughout the world. Owing to plate collisions, foliations are often near vertical, resulting in azimuthal compressional wave anisotropy and shear wave splitting of teleseismic shear waves. In this study we present data that show how significant such bodies might be with regard to active and passive source seismic experiments. Although some readers may consider this an academic issue, it is an ignored source of error in many seismic studies, and it will have increasingly practical applications as seismologists begin to use high-resolution seismic methods to explore for ores in metamorphic terranes.

Our data suggest that a significant thickness ( $10-20 \mathrm{~km}$ ) of appropriately oriented (steeply dipping foliation) schist in the crust could contribute as much as $45 \%$ of observed shear wave splitting and that crustal contributions are therefore not negligible. Our data can be used to model such anisotropic bodies if it is suspected, or known, that they make up a significant part of the surface or subsurface in the region of a seismic experiment in order to determine how their presence might affect the modeling and interpretation of seismic data. Our results also show how lithological interpretation based on conventional velocity and $V_{P} / V_{S}$ ratio analysis may be misleading in regions of significant and favorably oriented bulk anisotropy.

It is clear from the data presented here that future work must include obtaining the additional two $V_{P}\left(45^{\circ}\right)$ measurements required to describe orthorhombic schist, since our Haast schist samples in particular show significant orthorhombic symmetry. More work also needs to be done to investigate the role of specific minerals within the schists, particularly micas. This will require measuring the orientation of such minerals within a schist sample and tying the single-crystal properties to the bulk rock properties presented here.

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N. I. Christensen, Department of Geology and Geophysics, University
of Wisconsin-Madison, Madison, WI 53706.
of Wisconsin-Madison, Madison, WI 53706.
N. J. Godfrey, 5 Delta Road, Chobham, Surrey GU24 8PY, England, United Kingdom. (nicola@terra.usc.edu).
D. A. Okaya, Earth Sciences Department, University of Southern California, Los Angeles, CA 90089.
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[^0]:    ${ }^{1}$ Now at Chobham, England, United Kingdom.

