Seismic Anisotropy Measurements of a Fractured Crystalline Formation in Outokumpu, Finland Using High Resolution VSP

by

Heather Schijns

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Abstract

A high resolutions seismic survey is undertaken in Outokumpu, Finland using a 2.5 km deep borehole. The Outokumpu area subsurface is composed primarily of a biotite schist and is expected to display significant seismic velocity anisotropy as a result of preferentially oriented minerals and aligned fractures. The seismic velocity anisotropy of the area is measured using data from a multi-depth multi-azimuth walk-away VSP. The three component data is processed using a harmonic filter, eigenvalue based rotation, a polarization, a directional and, where necessary, a bandpass filter. Static corrections are developed and applied from traveltime inversion of seismic refraction data. Phase velocity anisotropy is then measured from plane wave decompositions of the processed VSP data sets. Significant P-wave anisotropy is observed, as well as shear-wave splitting.
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1.0 Introduction

This thesis presents seismic anisotropy measurements made using a 2.5 km deep borehole in Outokumpu, Finland. This chapter serves as a qualitative and quantitative introduction to the topic of seismic velocity anisotropy. The causes of anisotropy, methods of measuring anisotropy and previous research, particularly from geologically similar study areas, are presented. Finally, the basic geology of the Outokumpu area and the acquisition details of the seismic survey analyzed in this thesis are described.

1.1 Background

Seismic anisotropy refers to the dependence of seismic wave velocity and polarization on the direction of propagation of the seismic wave through a particular rock mass. It is prevalent throughout most of the Earth.

In an isotropic medium seismic body waves travel as compressional waves, or P-waves, and as shear waves, or S-waves (Fig. 1.1). The P-wave is the fastest wave, and is therefore the first arriving wave. As a compressional wave, it is polarized parallel to the direction of propagation. The S-wave is slower and, in the absence of mode conversions, arrives second. The S-wave oscillates perpendicular to the direction of propagation, and has particle motion confined within the vertical plane and within the horizontal plane. The two directional components, although traveling together at the same velocity, are termed SV- and SH-waves, respectively.

In an anisotropic medium the movement of these waves is more complex. Anisotropy is observed in the three kinds of seismic body waves. In P-waves the directional dependence of velocity impacts the traveltime of the P-wave when it is measured to have traveled with different directions of propagation through the rock. The
observation of P-wave anisotropy therefore requires measurements in multiple directions. Further, anisotropy causes the oscillation of the P-wave to deviate slightly away from parallel to the direction of propagation. To signify this deviation away from a pure compressional wave, P-waves in anisotropic media are termed quasi P-waves, or qP-waves (Crampin, 1989). In shear waves anisotropy results in the phenomenon of shear wave splitting, where the two orthogonal components of the shear wave travel at different velocities due to their different directions of oscillation. Shear wave anisotropy can therefore be observed from a single measurement in a single direction. Like the qP-wave, the directions of oscillation of the two shear waves deviate slightly away from orthogonal to the direction of propagation in anisotropic media. The first arriving shear wave is termed qS$_1$, and the second qS$_2$ (Crampin, 1989).

The speed of seismic wave propagation in a given direction is governed by the rock’s elastic stiffness tensor. Hooke’s Law defines the elastic stiffness tensor in terms of stress and strain; using the Einstein summation convention

$$\sigma_{ij} = C_{ijkl} \varepsilon_{kl} \quad \text{for } i,j,k,l = 1,2,3 \quad (1.1)$$

where $\sigma_{ij}$ is the stress tensor, $\varepsilon_{kl}$ the strain, and $C_{ijkl}$ the elastic stiffness tensor or elastic modulus. The reader is referred to Auld (1973) for a more in depth description. Although $C_{ijkl}$ is a fourth order tensor with 81 variables, symmetries in the stress and strain tensors result in $C_{ijkl}$ having a total of only 21 independent variables, with

$$C_{ijkl} = C_{iklj} = C_{ijlk} = C_{klij} \quad \text{for } i,j,k,l = 1,2,3 \quad (1.2)$$

The infinitesimal strain can be described by the Cauchy strain equation

$$\varepsilon_{ij} = \frac{1}{2} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \quad \text{for } i,j,k,l = 1,2,3 \quad (1.3)$$
where \( u \) is displacement and \( x \) is position. The strain equation can be combined with the equation for linear momentum

\[
\frac{\partial \sigma_{ij}}{\partial x_j} = \rho \frac{\partial^2 u_i}{\partial t^2} \quad \text{i,j}=1,2,3 \tag{1.4}
\]

where \( t \) is time, and \( \rho \) is density, to give the full elastic wave equation

\[
\rho \frac{\partial^2 u_i}{\partial t^2} - \frac{\partial}{\partial x_j} \left( C_{ijkl} \frac{\partial u_k}{\partial x_l} \right) = 0 \quad \text{i,j,k,l}=1,2,3 \tag{1.5}
\]

Assuming \( u \) can be described as a harmonic oscillation

\[
u_i = Ab_i \exp \left[ i \omega \left( \frac{n_m}{v} x_m - t \right) \right] \quad \text{i,m}=1,2,3 \tag{1.6}
\]

where \( b_i \) represents the particle displacement or oscillation direction polarization, \( \omega \) is the angular frequency, \( n \) is a unit vector parallel to the direction of wave propagation, and \( v \) is the magnitude of the phase velocity, the equations can be combined to yield a system of linear equations

\[
(C_{ijkl} n_j n_l - \rho v^2) b_k = 0 \quad \text{i,j,k,l}=1,2,3 \tag{1.7}
\]

Where \( n_j \) and \( n_l \) are the direction cosines. A non-trivial solution is achieved if and only if the determinant of the coefficient matrix is zero

\[
\det \begin{vmatrix}
\Gamma_{11} - \rho v^2 & \Gamma_{12} & \Gamma_{13} \\
\Gamma_{21} & \Gamma_{22} - \rho v^2 & \Gamma_{23} \\
\Gamma_{31} & \Gamma_{32} & \Gamma_{33} - \rho v^2
\end{vmatrix} = 0 \tag{1.8}
\]

Here \( \Gamma \) is the Christoffel symbol, and \( \Gamma_{ij} = C_{ijkl} n_j n_l \). The symmetry properties of the elastic stiffness tensor, \( C_{ijkl} \), ensure that \( \Gamma_{ij} \) is both symmetric and real. This results in three distinct eigenvalues; and therefore three distinct phase velocities. Each phase velocity
corresponds to a particular body wave, qP, qS₁ or qS₂ (Mainprice, 1990). When the elastic tensor contains 21 independent, non-zero components, the resulting velocities have a triclinic symmetry. As the number of independent parameters decreases, the symmetry of the velocity field increases, until an isotropic velocity field is reached when the elastic tensor includes only two independent parameters. The phase velocities given by the elastic tensor, which have a plane wave front, are different than the group (or ray) velocities observed in most seismic experiments, which have a curved wavefront (Figure 1.2). Moving between these two velocities is not a simple matter and is discussed in future chapters.

1.2 Causes of Seismic Anisotropy

Seismic anisotropy is produced in numerous situations that include the preferred orientation of anisotropic minerals, layering of isotropic and anisotropic lamina, pore texture, faults, and non-randomly oriented fractures within the rock (Crampin, 1981). Anisotropy can be loosely divided into anisotropy caused by the intrinsic physical properties of the rock itself, such as preferentially aligned minerals within the rock, and anisotropy caused by external factors, such as fractures. The anisotropic symmetry of rocks can be approximated by any of eight symmetries, ranging from the simplest, isotropic, to the most complex, triclinic. Most measurements of anisotropy, for simplicity, assume rocks exhibit transversely isotropic (TI) or orthorhombic symmetry. TI symmetry, with a single symmetry axis, is described by five independent elastic stiffness constants, while orthorhombic symmetry, with three symmetry axes, is described by nine.
1.2.1) Anisotropy Due to Aligned Crystals and Grains

Minerals can have eight possible different symmetries of seismic anisotropy ranging from perfectly isotropic and described by two independent elastic parameters to triclinic symmetry that is described by the full complement of twenty one independent elastic parameters (Table 1.1). Variations in symmetries and physical properties of minerals mean that anisotropy varies substantially from mineral to mineral; most of the minerals found in the crust and upper mantle are strongly anisotropic (Weiss et al., 1999). For this reason, the volumetric proportion, and particularly the orientation and alignment, of minerals can play a large role in the anisotropy of a given rock. Minerals tend to align due to the stress field in the area or during recrystallization (Drury and Urai, 1990). Some rock forming minerals are more likely to align themselves than others, causing them to play a more important role in lattice preferred orientation (LPO) anisotropy.

Phyllosilicate minerals, such as biotite and muscovite, appear to demonstrate some of the greatest anisotropy, with strong shear wave splitting and substantial P-wave anisotropy (Barruol and Mainprice, 1993). Maximum P-wave velocities are seen along the basal plane and minimum velocities along the symmetry axis (Valcke et al., 2006). These minerals, while generally exhibiting monoclinic symmetry, can be considered to be close to transversely isotropic (Vaughan and Guggenheim, 1986). Within a host rock, the phyllosilicate texture can cause the rock to have either an overall transversely isotropic symmetry or an orthorhombic symmetry (Sintubin, 1994). Within rocks, phyllosilicate minerals have a strong tendency to preferentially orient themselves, and are often responsible for delineating the texture of the rock (Cholach and Schmitt, 2006).
anisotropy of these minerals is so great that, when they are preferentially oriented, they are often the principle cause of LPO anisotropy within aggregates. Within sedimentary rocks, Valcke et al. (2006) measured the strongest LPO anisotropy to occur due to phyllosilicates, with the strongest effect in shales, followed by siltstones and then sandstones. Chlupacova et al. (2003) find high anisotropy (25-49%) in metamorphic slate with high phyllosilicate content. After an examination of igneous rock properties, Barruol and Mainprice (1993) conclude that, within deformed felsic rocks, seismic properties are “essentially controlled by mica,” while Weiss et al. (1999), based on a compilation of laboratory data on single crystals and whole rocks at high confining pressure, as well as numerical simulation, reach the more general conclusion that, for lower crustal rocks, those with high mica content demonstrate the highest anisotropy.

While phyllosilicates are undoubtedly the most important minerals in terms of LPO anisotropy, LPO is observed in rocks composed of other minerals as well. In sedimentary rocks, quartz grains are generally found to be oriented nearly randomly. Deposition in the presence of a current, however; can cause weak anisotropy due to grain alignment, e.g. imbrication in stream sediments. Weak anisotropy due to preferred orientation can also be observed to be caused by dolomite and siderite in siltstones and sandstones (Valcke et al., 2006). In igneous rocks, it has been found that pyroxene crystals are usually randomly oriented, whereas amphiboles often have high LPO and contribute substantially to anisotropy and shear wave splitting (Barrol and Mainprice, 1993).

Measurements of crustal foliated rocks such as schists, gneisses and amphibolites have shown P-wave anisotropy from 9-20% due to LPO, and it can be estimated that under appropriate circumstances (e.g. 10-20 km thick, steeply dipping schist) LPO could
contribute up to 45% of the observed shear wave splitting in the crust normally attributed to textures within the mantle (Godfrey et al., 2000, Cholach et al., 2005).

1.2.2) Anisotropy due to Layering

Layering is slightly larger scale than preferential alignment, and occurs when thin layers of rock with differing seismic velocities are inter-layered. Some examples might include inter-layered sand, silt, shale, and carbonate (Wang, 2002). If the layering is fine enough that it is significantly smaller than the seismic wavelength, then it will affect the bulk anisotropy of the region. Layering will usually result in vertically transversely isotropic symmetry (Liner and Fei, 2006). Fine layering is prevalent primarily in sedimentary rocks, and is a common cause of anisotropy here. Kebaili and Schmitt (1996) measure anisotropy in layered shale and find an anisotropy of 15%, with velocity being largest at angles oblique to the normal of the layering. Theoretical models show that layering can cause a maximum of 14.2% anisotropy and Kebaili and Schmitt hypothesize that the additional anisotropy observed in their study is due to the intrinsic anisotropy of shale. This observation is supported by Wang (2002) who finds that shale is always intrinsically anisotropic.

1.2.3) Anisotropy Due to Aligned Fractures and Microcracks

Microcracks and fractures that are randomly oriented tend to isotropically slow the velocity of seismic waves. Measurements of attenuation in addition to seismic anisotropy can aid in the distinction between anisotropy caused by LPO and anisotropy caused by fractures, as attenuation is expected to increase with increased fracturing, but not with LPO of minerals (Carter and Kendall, 2006). Attenuation measurements can also help
differentiate between different models of fractured media that could cause identical seismic anisotropy (Kelner et al., 1999).

Deviatoric stress through much of the Earth’s upper crust means that microcracks and fractures within the subsurface tend to be aligned, often in a horizontal direction at the near surface, and then vertically at a slightly larger depth (Crampin, 1990). Shear cracks are generated due to sufficiently high compressive stresses, and these are aligned sub-parallel (±15°) to the direction of maximum compressive stress. A single such fracture set would exhibit transversely isotropic symmetry, and it is postulated that stress-aligned fractures are the dominant cause of observed anisotropy within the upper crust (Crampin and Lovell, 1991). The cracks are generated intragranularly or between grain boundaries. Models of quartzite found that intragranular cracks were the principle mechanism of fracturing under low stress, while intergranular cracks occurred under higher stress. Cracks nucleating from pores tend to be random in orientation (Malan and Napier, 1995). Cracks are commonly approximated to be ellipsoidal in shape, and are then described by their aspect ratio $b/a$, where $a$ is the length of the two equal semi-axes of the ellipsoid, and $b$ is the length of the third axis, the axis of rotational symmetry (Douma, 1988). As cracks tend towards a more spherical shape, their aspect ratio approaches unity, while flatter cracks have an aspect ratio that tends towards zero. The aspect ratio of cracks appears to be small in hard rocks (Crampin and Lovell, 1991), and larger in sedimentary rocks (Crampin, 1991). Uniform aligned fractures can contribute different amounts to anisotropy depending on the saturation of the fractures. Near the surface, wider fractures may be drier, while smaller microcracks may retain water due to surface tension and capillary forces (Bamford and Nunn, 1979).
In a material containing aligned, oblate spheroid cracks, velocity is slowest in the direction normal to the plane the cracks lie in (Anderson et al., 1974). In parallel, vertically oriented cracks, the fastest vertically propagating shear waves are polarized in the direction of the fractures. This polarization is not static, and thus shows only the direction of the fractures near the receiver (Crampin and Lovell, 1991). This has been confirmed with laboratory measurements, which found this to be true within ±22.5° (Peacock et al., 1994). Liu et al. (2006) find that shear wave polarization varies with signal frequency. The variation in polarization depends on the angle of incidence in relation to the orientation of the different fracture sets. They speculate that signal frequency and polarization direction are correlated with fracture size, with high frequency shear waves being polarized in the direction of microcracks and low frequency shear waves being polarized in the direction of larger fractures in the case where there are multiple fracture sets.

Bakulin et al. (2000) investigate the effect of fractures on anisotropy in orthorhombic media and find that the anisotropy is governed by the weaknesses of fractures orthogonal to the vertical plane of symmetry.

### 1.3 Measuring Anisotropy

The elastic modulus of an aggregate, and thus the seismic anisotropy, can be estimated using the Voigt (1928) and Reuss (1929) estimates. The Voigt estimate, which assumes uniform strain, provides an upper bound for the elastic constants

\[
C_{upper} = \sum_{i=1}^{N} \phi_i C_i
\]  

(1.9)
where $N$ is the total number of mineral modes, $\phi$ is the volumetric proportion of the minerals, and $C_i$ is the elastic stiffness of the mineral. The Reuss estimate, which assumes uniform stress, provides a lower bound for the elastic constants:

$$C_{upper} = \left[ \sum \frac{\phi_i}{C_i} \right]^{-1}$$  \hspace{1cm} (1.10)

These bounds, while fairly effective for isotropic aggregates, and do not constrain the elastic constants of anisotropic rocks as precisely (Cholach and Schmitt, 2006). Numerous other techniques have been proposed to more accurately estimate the elastic constants (e.g. Hill, 1952, Hashin and Shtrikman, 1963), but none have perfect agreement with experimental results. In order to accurately characterize the anisotropy and elastic constants, it therefore becomes necessary to determine the anisotropy of an aggregate through measurements and/or computer modeling.

Seismic anisotropy can be measured in a variety of ways, either in situ or in a laboratory setting. Most techniques aim to measure seismic velocities in enough directions to fully describe the elastic stiffness tensor, as, once the elastic tensor is determined, the velocities in all unmeasured directions can be theoretically calculated, however; this is much simpler in a laboratory setting than in situ. Even here, however; the symmetry of the sample is usually assumed known. This greatly simplifies the calculation.

1.3.1) Laboratory methods

Laboratory measurements of seismic anisotropy and the representative elastic stiffness tensor of a rock can be made using several different techniques, including x-ray diffraction (e.g. Pavese et al., 1999, Smyth et al., 2000), ultrasonic pulse transmission
(e.g. Best et al, 2007; Kebaili and Schmitt, 1997), and Brillouin scattering (e.g. Jiang et al., 2004, Murakami et al., 2007). Most suffer from similar difficulties, including the challenge of machining appropriate samples, the assumption that measurements are made along known symmetries, and the difficulty in measuring the phase (plane) velocities which are required to invert for the elastic constants (Kebaili and Schmitt, 1997). As previously discussed, microcracks, fractures and pores can play a large role in anisotropy. The removal of the rock from its in situ environment will cause the shape, size, and possibly number, of these microcracks, fractures and pores to be altered, changing their effect on the anisotropy. If the laboratory measurements are made at sufficiently high confining pressure then this problem can be minimized.

1.3.2) In situ methods

In situ measurements of the anisotropy clearly give the best picture of the anisotropy in a given location. While earthquakes can provide valuable data, the unpredictability of the source and the lack of control over the source receiver orientations is less than ideal. Active seismic sources, such as dynamite, weight drops and seismic vibrators can allow the simultaneous acquisition of P- and S-wave data, but will then require three component receivers. Sources with a defined orientation, such as seismic vibrators, can potentially help avoid ambiguity in P- and S-wave measurements (Luschen, 1999). In situ measurements, however; are plagued by many of the same problems as laboratory measurements: it is impossible to determine the phase velocities directly from seismic velocity measurements (which measure the group or ray velocity), there are often limitations on the azimuth and zenith angles that can be covered using typical source and receiver layouts and heterogeneities within the subsurface often result in an inability to
invert for an accurate bulk elastic modulus. Finally, seismic noise and the low frequency seismic signal used, typically 100-400 Hz (Hayles et al., 1999), can cause significant difficulties in the precision of the determination of the P- and S-wave arrival times. In practice, the difficulty in picking S-wave arrival times can make it quite challenging to accurately measure shear wave splitting.

In situ measurements are often made using cross-hole surveys (e.g. Hayles et al., 1999), reflections profiles (e.g. Gretchka and Tsvenkan, 1999, Bakulin et al., 2000), or vertical seismic profiles (VSP). VSP surveys involve placing a receiver downhole and using a surface source. The source can be stationary, either near the borehole, a zero-offset VSP, or at a distance from the borehole, a far offset VSP; in these cases the receiver is moved through a range of depths. Alternatively, in a walk-away VSP, the receiver can be left at a single depth while the source is moved along an azimuth from the borehole (Figure 1.3). VSP surveys require only a single borehole and can include numerous azimuth angles. Difficulties, however, lie in that measurement through the zenith angles is more difficult and requires long offsets, and static corrections for the near surface are often required. Nonetheless, VSP surveys provide the most reliable anisotropy measurements of the in situ geology (Morozov et al., 1997)

VSP measurements first detected seismic anisotropy through shear wave splitting in sedimentary basins in the late 1980s (e.g. Johnston, 1987, Daley and McEvilly, 1990, Douma et al., 1990). Since then, more sophisticated multi-offset and multi-azimuth VSP surveys have been undertaken to observe both the P-wave anisotropy and the shear wave splitting. Kebaili and Schmitt (1996) use a multi-offset multi-depth VSP (e.g. Figure 1.3) to measure anisotropy in a sedimentary rock environment. In order to expand anisotropy
measurements from the limited directions measureable by VSP into all possible directions, it is necessary to use the measurements to model the elastic tensor of the rock. Grech et al. (2002) use a multi-offset VSP and assume TI symmetry to invert for anisotropy parameters of a dipping shale. Newrick and Lawton (2003) use qP-wave first arrivals from a multi-offset VSP in a sedimentary basin to estimate TI anisotropy parameters. Similarly, Li et al. (2004) use a three-component receiver and multi-offset VSP to invert for TI anisotropy parameters.

Anisotropy caused by layering has TI symmetry. In an unfractured sedimentary environment, where this is usually the principle cause of anisotropy, an assumption of TI symmetry is usually valid. In metamorphic and igneous rocks, where the causes of anisotropy are more varied, such an assumption is not always as accurate. Seismic methods are still only rarely used for mineral exploration and as a result most VSP research in crystalline rock environments is limited to the use of scientific boreholes. As a result very few such anisotropy measurements have been made. Li et al. (1988) study anisotropy in the Cajon Pass scientific borehole, located near the San Andreas fault. The authors use the shear wave splitting and polarization directions observed at several source offsets and azimuths to infer a possible regional stress orientation. Digranes et al. (1996) use two far offset VSPs in the Kola superdeep borehole, located in northwest Russia, to measure shear wave splitting and conduct a shear wave polarization analysis in a metamorphic terrane. Further anisotropy analysis is performed in a laboratory setting using core samples from the Kola borehole. Okaya et al. (2004) use a multi-offset multi-azimuth in combination with travel time inversion and comparisons to core samples to measure anisotropy in the KTB super-deep borehole. The KTB borehole is located in
southeast Germany in a metamorphic terrane; the authors find TI to be an adequate approximation in this case. Okaya et al. find that, if the region does not exhibit horizontal or vertically oriented symmetry, it is necessary to have >180° of azimuthal coverage in the VSP survey to be able to expand the bulk anisotropy results to every possible source-receiver orientation.

1.4 Outokumpu Area Geology

Outokumpu, the field site for this survey, is located in Southeast Finland (Figure 1.4). It is the site of a historic base metal mine, and was chosen as the location of a 2.5 km deep, fully cored, International Continental Scientific Drilling Program (ICDP) borehole. The borehole was drilled by the Russian company NEDRA; borehole logging was undertaken during the drilling process first by NEDRA and again later by the ICDP after the completion of the borehole.

The Outokumpu area is part of the Fennoscandian shield and has previously undergone significant glaciation. As a result, the bedrock is covered by an overburden tens of metres thick, composed principally of glacial till and glaciofluvial sediments (Chork and Salminen, 1993). The borehole drilling results show an overburden 33 m thick composed of glacial deposits. Below these glacial deposits is a crystalline bedrock composed of a biotite metamorphic schist (Figure 1.5). The schist extends to a depth of ~2 km in the borehole and is quite laterally continuous as well, as evidenced by regional geological surveys (Gaal et al., 1971, Huhma, 1971, Koistinen, 1981) and by crustal scale seismic reflection profiles (Figure 1.6) through the area (Sorjonen-Ward, 2006). At 1.3 km depth in the borehole, the schist encloses a ~200 m thick body composed of Outokumpu-assemblage rock (Koistinen, 1981) consisting primarily of serpentinite,
skarn, quartz and black schist. Below 2 km, the schist becomes increasingly interlayered with a pegmatitic granite.

Representative core samples of the schist, from the borehole, are composed of 22.6-47.7% quartz, 19.9-43.8% plagioclase, 16.4-34.2% biotite and 1.6-16.7% muscovite, by volume (Kern and Mengel, 2007). In some instances iron sulfide (0.1 vol.%) is present as an accessory mineral (Kern et al., 2008). The schist is expected to demonstrate significant intrinsic seismic anisotropy as a result of its high biotite content. The biotite demonstrates a lattice preferred orientation, and core samples from the borehole show that the foliation of the schist appears to be horizontal or subhorizontal (Kern and Mengel, 2007). Where the schist outcrops ~2.5 km to the northwest, it has a dip 35° to the southeast, and a strong NE-SW lineation (I.T. Kukkonen, personal communication, 2007), however; from the recent seismic reflection profiles undertaken as part of the Finnish Reflection Experiment (FIRE) (Sorjonen-Ward, 2006), the schist appears to be fairly flat lying in the vicinity to the east of the borehole (Figure 1.6). This indicates that it is likely the intrinsic anisotropy of the schist near the borehole has symmetry that can be approximated as vertically transversely isotropic (VTI)\(^1\).

As most of the Earth’s crust is expected to include aligned fractures (Crampin, 1990), it is likely that preferentially aligned fractures are present within the Outokumpu schist. The presence of fractures is confirmed by the borehole core samples. The combination of aligned fractures and intrinsic anisotropy means that the seismic anisotropy in the schist likely has an orthorhombic symmetry. Indeed, theoretical studies show that no matter how many fracture sets are present within a host rock, the symmetry is never more complex than orthorhombic (Grechka and Kachanov, 2006). As the

\(^1\) VTI refers to a transversely isotropic symmetry where the symmetry axis is in the vertical direction
symmetry of the anisotropy caused by LPO of biotite within the schist is no more complicated than that of the most complex fracture sets, it stands to reason that these results also demonstrate that this combination of LPO and fracture anisotropy cannot be more complex than orthorhombic symmetry.

1.5 Outokumpu Area Seismic Survey

In May, 2006, personnel from the University of Alberta, in conjunction with personnel from the Geological Survey of Finland (GTK) and from the University of Helsinki, undertook a high resolution seismic survey using the Outokumpu borehole. The survey included three parts: a surface reflection/refraction profile, a zero and far offset VSP and a multi-depth multi-azimuth walk-away VSP. The surveys took place principally along two ~2 km long azimuths from the borehole: one to the northeast, and one mainly to the southeast (Figure 1.7).

The University of Alberta IVI minivib\textsuperscript{TM} seismic vibrator source was used for all of these surveys and employed 8 s linear taper sweeps with frequencies 15-250 Hz. Surface receivers were single component 14 Hz OYO geophones, with a 3-component geophone employed downhole. Surface geophones were located 4 m apart. The downhole receiver was located at depths of 1000, 1750 and 2500 m for the walk-away VSP; for the zero offset VSP and the far offset VSP, the receiver was raised in 2 and 25 m increments, respectively. Shot points were located approximately 20 m apart along the NE azimuth survey line, with some few shot points missed due to local topographical variation. Shot points where located approximately 10 m apart along the SE azimuth survey line as a result of interference from local infrastructure. Infrastructure along this survey line meant that many stretches of the line could not have shot points; where shot points were
permitted they were therefore spaced closer together to allow the building of a common mid-point fold for the reflection data.

In general, 3-5 shots were conducted at each shot point. In most cases, the data from these shots were automatically correlated with the seismic vibrator signal and stacked. In some cases the seismic vibrator truck was out of range of the radio trigger which linked it and the data collection computer. Ordinarily, this remote triggering of the source allows the start of the seismic vibrator sweep and the start of data collection to occur simultaneously. When the seismic vibrator truck was out of range, it had to be manually triggered, with the result that the start of geophone data collection and the seismic vibrator sweep did not occur simultaneously. In the case of the reflection/refraction profiles these source points were excluded from the data set. For the walk-away VSPs, however; they were included by calculating and removing the time difference between the start of data collection and the triggering of the seismic vibrator sweep. The first arrivals of all the data were picked based on the first extremum in amplitude of the signal. As the actual traveltime from source to receiver could not be known when the seismic vibrator failed to be triggered remotely, a linear interpolation between the nearest two source points with known start time was used to determine a first arrival time for the source points with unknown start times. All sweeps from the same shot point were then correlated with a synthetic seismic vibrator signal and stacked.

The principle goal of the seismic survey was to acquire walk-away VSP data for the purpose of measuring the anisotropy in the area. The data had quite a low signal to noise ratio, and significant static problems as a result of the varying topography and near surface composition. In order to make use of the walk-away VSP data it was necessary to
process and filter the data quite substantially, and to perform adequate static corrections
to remove the influence of the glacial deposits and topography from the data set. These
corrections are discussed in the following chapters.
Table 1.1

Form of the \( (S_3) \) and \( (C_4) \) matrices

**KEY TO NOTATION**

- zero component
- non-zero component
- equal components
- components numerically equal, but opposite in sign

For \( s \) \( \bigcirc \) twice the numerical equal of the heavy dot component to which it is joined
For \( c \) \( \bigcirc \) the numerical equal of the heavy dot component to which it is joined
For \( s \) \( \bigbullet \) \( 2(s_{11}-s_{12}) \)
For \( c \) \( \bigbullet \) \( \frac{1}{2}(c_{11}-c_{12}) \)

All the matrices are symmetrical about the leading diagonal.

---

**TRICLINIC**

*Both classes*

![TRICLINIC Diagram](image)

(21)

---

**MONOCLINIC**

*All classes*

**Diad \( \| x_3 \)**

(standard orientation)

![MONOCLINIC Diagram](image)

(13)

---

**ORTHORHOMBIC**

*All classes*

![ORTHORHOMBIC Diagram](image)

(9)

---

**CUBIC**

*All classes*

![CUBIC Diagram](image)

(3)
Elastic stiffness tensor $C_{ijkl}$ and elastic compliance tensor $S_{ijkl} = C_{ijkl}^{-1}$ represented in Voigt notation as $C_{mn}$ and $S_{mn}$ for the various symmetries. From Nye (1990), with permission.
Figure 1.1

The three seismic body waves in an isotropic media. The P-wave is a compressional wave, and oscillates parallel to its direction of propagation, the S-waves are shear waves and oscillate perpendicular to their direction of propagation.
A comparison between phase velocity and group velocity at a given point P. Note also the difference between the group angle $\phi$ and phase angle $\theta$. After Kebaili and Schmitt (1997).
A multi-depth walk-away VSP. The seismic wave generated by the multiple source points (*) are observed by downhole receivers R1 and R2. Note that for receivers located at deeper depths, sources require further offset from the borehole to achieve the same angle aperture as achieved for receivers at lesser depths.
Figure 1.4

Outokumpu is located in Southeast Finland.
A lithological cross section of the 2.5 km deep Outokumpu borehole. The upper 1.3 km is composed of a biotite rich schist (metasediments) with an overburden of glacial deposits. Below 1.3 km depth the schist is interlayered with Outokumpu assemblage rock (ophiolitic rocks) and pegmatitic granite. With permission from the Geological Survey of Finland (GTK).
Part of the FIRE profiles through Outokumpu, these three separate reflection profiles are represented three dimensionally to show the approximate location of the Outokumpu borehole and the outcropping of the Outokumpu assemblage rock on the surface. The strong reflector intercepted by the borehole is known to be Outokumpu assemblage rock from study of core samples. With permission from the Geological Survey of Finland (GTK).
Figure 1.7

A map of the survey area, the green circle shows the borehole location, while the dotted line shows the seismic survey lines and shot point numbers.
1.6 Bibliography


Voigt, W., 1928. Lehrbuch der Kristallphysik, Teubner-Verlag, Leipzig, Germany.


2.0 Data Processing

This chapter covers the processing of the Outokumpu walk-away VSP data. The data was processed using a harmonic filter, axes rotation, and the application of rectilinear polarization and directional filters. In some instances a bandpass frequency filter was additionally applied. The motivation behind the processing stream, the theoretical background of the techniques applied and the processing results are presented here.

2.1 Introduction

In May 2006 a multi-azimuth multi-depth walk-away VSP survey was conducted in Outokumpu, Finland using the 2.5 km deep International Continental Scientific Drilling Program (ICDP) borehole for the purpose of studying seismic anisotropy within the area’s fractured bedrock. A combination of the poor coupling between the downhole receiver and fractured borehole walls, attenuation of the seismic signal at the larger receiver depths, strong powerline harmonics and ambient seismic noise in the area caused the signal to noise ratio of the seismic records to be lower than desired.

A complete study of anisotropy within a host rock involves the measurement of both P- and S-wave velocities; an accurate identification of the waves and their arrival times within the seismic record is crucial to the success of any such measurement. The P-wave is the first arriving wave in a walk-away VSP, and while it is thus relatively easy to identify, a low signal to noise ratio can make accurately picking its arrival time difficult (Tronicke, 2007). The arrival times of waves lying within the coda of the seismic signal, such as the S-wave, are even more difficult to measure (Zelt et al., 2003). An increase in the signal to noise ratio in the record can be instrumental in the correct identification of
the compressional and shear wave arrivals. This study presents the results of the application of harmonic, rectilinear polarization and directional filters in increasing the signal to noise ratios of P- and S-waves within the seismic record.

2.2 Harmonic Filter

Seismic noise within a record is unavoidable, and filtering to increase the signal to noise ratio is often essential. Powerlines in Europe introduce strong 50 Hz harmonic fields into the Earth. Within North America, the third harmonic of the dominant powerline frequency contributes significantly to harmonic noise, as it is only a factor of 4-10 times smaller in amplitude than the dominant frequency (Butler and Russell, 2003). Indeed, this contribution from the 150 Hz harmonic is observed in the European study, and requires attenuation as well.

A notch filter is commonly used to remove harmonic noise, however; when the harmonics lie within the frequency band of the signal, as they do in this study, the use of a notch filter attenuates substantial parts of the signal and causes distortion of the frequencies near the notch. Additionally, multiple notches are required to remove the additional odd harmonics, further degrading the signal. Spiking deconvolution is also commonly used to suppress powerline harmonics, but can cause distortion of the seismic signal in the presence of strong spectral peaks (Linville and Meek, 1992). A third method is to subtract a sinusoid of the appropriate frequency, amplitude and phase from the record to remove the noise. Numerous techniques of calculating the appropriate sinusoid have been developed. Nyman and Gaiser (1983) propose an adaptive rejection technique to measure and remove harmonic noise. Butler and Russell (1993) built on this method and also present a computationally inexpensive block subtraction technique. Linville and

Most of these techniques involve calculation of a precise fundamental frequency for the harmonic noise since powerline noise can be expected to drift by ±0.03 Hz over time within North America (Adams et al., 1982), with similar deviations likely in Europe. This study used stacked seismic records typically summing 3-5 correlated seismic vibrator sweeps. Each sweep was 8 s in length, with 5 s listening time, and frequency drift during the data collection is quite possible. Since the calculation of a fundamental frequency would invariably be inaccurate if the frequency was not stationary, and the problem of calculation of phase is complex due to the stacked records, a modified version of the simpler block subtraction technique (Butler and Russell, 1993) is applied in this study.

This technique involves using the pre-signal noise to create a harmonic noise filter. The first breaks of the seismic signal were picked for all traces, and the pre-signal noise was divided into 20 ms sections corresponding to the fundamental 50 Hz frequency. In Butler and Russell’s (1993) method, one of these 20 ms segments is chosen to be subtracted from each 20 ms block of the record. Here, as a modification of their method, the 20 ms sections of noise were averaged prior to subtraction. Averaging the 20 ms noise segment before subtraction reduced the development of even harmonics within the record, as well as increasing the ratio of harmonic noise to random noise within the
subtraction block. Using this technique produced an overall reduction in the amplitude of the harmonic noise present within the data set (Figure 2.1a,b).

Deviation away from 50 Hz was noticed in some records, although not all, confirming the occurrence of frequency drift. This deviation from the assumed frequency caused beating at the end of some records, however; the arrivals of interest, the P- and S-waves, all occur near the beginning of the record where this beating has not yet developed (Figure 2.2). Deviations away from 50 Hz therefore have minimal effect on the noise reduction of the harmonics present.

2.3 Polarization and Directional Filter

Other sources of seismic noise were more random and therefore more difficult to attenuate. Nonetheless, in order to improve the signal to noise ratio within the records, it remained necessary to filter the data further. P- and S-waves are linearly polarized, and, along a given ray path, propagate with motion in mutually orthogonal planes. Unfortunately the more complex behaviour of quasi P- and S-waves (so named due to their only superficial resemblance to their isotropic counterparts) in anisotropic media means that the qP- and qS-waves cannot be separated from the principally elliptically or unpolarized noise as easily as might be implied.

Polarization directions (Fouch and Fischer, 1998, Liu et al., 2006) and velocities (Marson-Pidgeon and Savage, 1997, Tod and Liu, 2002, Liu et al., 2003) of qS-waves within fractured anisotropic media can be frequency dependant. Since the seismic vibrator signal in this study employed a sweep with frequencies 15-250 Hz, and the subsurface is expected to have aligned cracks, qS polarizations and velocities can be expected to show this frequency dependence. When considering the overall seismic
record, this frequency dependence results in the energy of qS1 and qS2 (the fast and slow shear waves) being smeared out slightly due to their range of velocities. Additionally, the qS1- and qS2-waves may appear to be elliptically polarized as a result of the superposition of different frequency waves being linearly polarized in different directions.

The qP-wave polarization is also frequency dependant, and can vary by up to 40° from the pure longitudinal motion for an elastic wave when propagation is in an anisotropic medium (Crampin, 1981, Bear et al., 1999). The anisotropy in the Outokumpu area is not expected to cause such extreme variations in polarization directions, but some variation is inevitable and the direction of qP-wave polarization is therefore unknown as well.

Furthermore, although qP- and qS-waves traveling along the same ray path have mutually orthogonal polarizations, within an anisotropic medium the qP- and qS- waves observed by the receiver are likely to have traveled different paths. It is therefore unlikely that the observed waves are mutually orthogonal (Crampin, 1981).

Numerous polarization filters have been developed to separate body waves; the success of a particular filter varies with the data attributes. Polarization filters are ordinarily applied in either the time (e.g. Flinn, 1965, Montalbetti and Kanasewich, 1970, Vidale, 1986, de Franco and Musacchio, 2001, De Meersman et al., 2006) or frequency domain (e.g. Samson and Olson, 1981, Park et al., 1987, Du et al., 2000), although a few methods use both (e.g. Jurkevics, 1988). Filters of both types are usually based on the application of an eigenvalue technique to a data covariance matrix (termed a spectral density matrix in the frequency domain).
Time domain based filters typically use singular value decomposition (SVD) to measure the polarization attributes within a time window. The filters can be used on real (Flinn, 1965, Montalbetti and Kanasewich, 1970) or complex (Vidale, 1986, Bataille and Chiu, 1991, De Meersman et al., 2006) time series. Filters are used to separate either linearly polarized waves (e.g. Flinn, 1965, Montalbetti and Kanasewich, 1970) or both linearly polarized body waves and elliptically polarized surface waves (e.g. de Franco and Musacchio, 2001).

Although the expected frequency dependence of the qP- and qS-waves polarizations would suggest that a frequency domain based filter would be more appropriate for anisotropic data, frequency domain filtering is most successful if superpositioned waves have distinct frequency content (Schimmel and Gallart, 2003). In this study, the body waves have a broad frequency band due to the 15-250 Hz sweep used (Figure 2.3). The waves are, however; well separated in time, and for this reason a time domain based polarization filter was applied.

The algorithm of Montalbetti and Kanasewich (1970) was modified and applied to the seismic records. This algorithm uses eigenanalysis of the covariance matrix of a sliding time window of the data, and filters both for linear polarization and directionality. Since it is in the time domain, it does not consider frequency dependence. It is more able to remove high frequency noise, since more wavelengths of this noise will be present within a given time window. Additionally, like most filters, it does not consider non-orthogonal body waves. Filters which consider non-orthogonality (e.g. Lei, 2005) perform poorly when, as in this study, the frequency dependence of the qP- and qS-waves results in a broad range of simultaneous polarization directions. Despite the numerous
difficulties associated with using a polarization filter on anisotropic data, the modified Montalbetti and Kanasewich (1970) algorithm appears to work well.

The directional filter of the algorithm performs better if the axes of the initial records are firstly rotated into the principle axes of qP and qS motion. Since the directions of the principle axes is unknown, a method similar to the Karhunen-Loève transform described by Jackson et al. (1991) is developed and applied. The method relies on SVD to rotate the axes into the correct direction, and an accurate identification of the time of two of the qP and qS arrivals is required.

Due to its large amplitude, the qP-wave arrival could be easily identified, allowing the vertical component of the receiver to be rotated into the qP arrival direction as a first step. The firstbreaks of all traces had been previously picked for the harmonic filter, and these firstbreak times were reused here. A time window that captured approximately the first cycle of the qP-wave signal was determined for each walk-away depth, and a covariance matrix $V$ was calculated for this time period

$$
Cov(X_i(t), X_j(t)) = E\{(X_i(t) - \mu_i(t))(X_j(t) - \mu_j(t))\}
$$

$$
t \in \left[ t_0 - \frac{N-1}{2}, t_0 + \frac{N-1}{2} \right]
$$

$$
V = Cov(X_i(t), X_j(t))
$$

where $\mu$ is the mean of $X$, $N$ is the length and $t_0$ is the centre of the time window and $X_i$ ($i=1,2,3$) are the three components of the seismic trace. The eigenvectors, $\xi$, of the covariance matrix are found. The principle eigenvector, $\xi_1$, is associated with the eigenvalue of greatest magnitude. Since, over the time window $t$, the amplitude of the incoming qP-wave is much greater than the amplitude of any noise present, motion
should principally be in the direction of the wave. The eigenvector $\xi_1$ can therefore be assumed to approximate the direction of the incoming qP-wave. The seismic traces are then rotated so that the axis direction of $X_i$ is rotated into the direction of the incoming qP-wave, $X_i''$, for all times $t$

\[
\theta = \tan^{-1}\left(\frac{\xi_3}{\xi_1}\right) \quad (2.3)
\]

\[
\alpha = \tan^{-1}\left(\frac{\xi_2}{\xi_1}\right) \quad (2.4)
\]

\[
\begin{bmatrix}
X_1'(t) \\
X_2'(t)
\end{bmatrix} = \begin{bmatrix}
\cos(\theta) & \sin(\theta) \\
-\sin(\theta) & \cos(\theta)
\end{bmatrix} \begin{bmatrix}
X_3(t) \\
X_1(t)
\end{bmatrix} \quad (2.5)
\]

\[
\begin{bmatrix}
X_1''(t) \\
X_2''(t)
\end{bmatrix} = \begin{bmatrix}
\cos(\alpha) & \sin(\alpha) \\
-\sin(\alpha) & \cos(\alpha)
\end{bmatrix} \begin{bmatrix}
X_1'(t) \\
X_2'(t)
\end{bmatrix} \quad (2.6)
\]

The second and third channels are then rotated through 90° about the $X_i''$ axis until a qS-wave amplitude is maximized in $X_2''$. This should, in turn, isolate the other qS-wave in $X_i''$. As discussed, seismic noise within the record, frequency dependant polarization and non-orthogonal body waves make this eigenvalue technique imperfect, however; a substantial improvement in the isolation of the seismic waves can be noticed (Figure 2.1c).

Any differences in noise amplitude between the channels would appear as polarization within the record. In order to minimize any such polarization of the noise prior to the application of the polarization filter, the amplitude of the noise within each trace is normalized using the traces pre-signal record (Samson and Olson, 1981).

After rotating the data axes into the principle directions of body wave motion and minimizing the polarization of the ambient noise, the Montalbetti and Kanasewicz’s
(1970) polarization and directional filter algorithm is applied. Body waves are polarized approximately linearly in an anisotropic media, whereas surface waves and noise are polarized elliptically. Filtering for polarization should allow the separation of the body waves from ambient noise and surface waves. Since the noise in the record is reasonably omni-directional, whereas the directions of motion of the body waves should correlated strongly with the directions of the axes after rotation, directional filtering should serve to further enhance the signal to noise ratio.

A covariance matrix $V$ (Equation 2.) is determined for a time window of length $N$ centered about $t_0$. The rectilinearity of the polarization of particle motion in the time window can then be estimated through the eigenvalues of $V$

$$F(\lambda_1, \lambda_2) = 1 - \left( \frac{\lambda_2}{\lambda_1} \right)^n \quad n>0$$

(2.7)

where $\lambda_1$ and $\lambda_2$ are the largest and second largest eigenvalues, respectively. Here, $n$ is an empirically determined weighting factor and can be varied to affect the weight of the eigenvalue term on the final value of $F(\lambda_1, \lambda_2)$. When rectilinearity is high, most of the particle motion will be concentrated in a single direction, and will result in $\lambda_1 >> \lambda_2$. Therefore, when rectilinearity is high, $F(\lambda_1, \lambda_2)$ will approach unity, and when rectilinearity is low, ie $\lambda_1 \approx \lambda_2$, $F(\lambda_1, \lambda_2)$ will approach zero.

The time $t_0$ is allowed to range over the full length of the record, and the measure of rectilinearity at time $t_0$ is given by
\[ RL(t_0) = \left[F(\lambda_1, \lambda_2)\right]^j \quad j>0 \]  

(2.8)

Similar to \( n \), the term \( j \) is a further empirically determined weighting factor. As \( j \) is increased the magnitude of rectilinear filtering on the seismic traces is increased. A measure of the direction of particle motion is given by the normalized eigenvectors of \( V \)

\[ D_{X_i}(t_0) = \left[\|\xi_i\|\right]^k \quad i=1,2,3; k>0 \]  

(2.9)

A final empirically determined exponent \( k \) is used to weight the effect of the directional filtering applied. Further smoothing to the functions \( RL(t) \) and \( D(t) \) is applied through the use of a time window \( M \) that is approximately half the length of \( N \) (Kanasewich, 1981)

\[ RL^*(t_0) = \frac{1}{M} \sum_{\ell=-L}^{L} RL(t_0 + \ell) \]

\[ D_{X_i}^*(t_0) = \frac{1}{M} \sum_{\ell=-L}^{L} D_{X_i}(t_0 + \ell) \quad i=1,2,3 \]  

(2.10)

Finally, the filters are applied to the traces to maximize the appropriate body waves within each channel of the recording (Figure 2.1d)

\[ X_{i}^*(t) = X_{i}^*(t) \cdot RL^*(t) \cdot D_{X_i}^*(t) \quad i=1,2,3 \]  

(2.11)

### 2.4 Discussion

The unknown arrival times of the qS-waves, their reduced amplitude, and the close proximity of the qS\(_1\) and qS\(_2\) waves compared to the filter window length causes poorer results than those achieved for the qP-wave. Orthogonal wave motion is implicitly assumed within the filter, and it is possible that non-orthogonal qS-waves are, in part, the
cause of the poorer results observed. To improve the observations of the qS arrivals, a bandpass filter is applied to the records emphasizing these in order to eliminate some of the high frequency residual noise (Figure 2.4).

The frequency dependence of the qP-wave polarization and velocity is evident in the filtered traces. The qP-wave continues to have substantial amplitude on all three components after rotation, however; the largest amplitudes occur at different times, as a response to a changing polarization direction (Figure 2.1c).

When the receiver is located a larger depths, the predominant frequencies of the body waves are lower as a result of attenuation. To accommodate the lower frequencies, the filter then requires a larger sliding window to assess rectilinearity and direction, resulting in a smoother filter with increased depth.

2.5 Conclusions

Despite substantial variation in the level of harmonic contamination in the seismic records, the harmonic filter performs well within the time window of interest. Evidence of beating near the ends of only select records confirms the presence of frequency drift. Since all frequency calculation techniques assume stationary noise, the presence of frequency drift within the records confirms that more sophisticated measurements of the frequency of the noise would be unlikely to perform better than the assumption of a 50 Hz harmonic.

The polarization and directional filter performs best when the signal is of large amplitude compared to the noise. The large amplitude of the qP-wave means that the filter is most successful at isolating this first arrival compared to the later qS-arrivals.
Figure 2.1

Three component trace from the 1000 m walk-away with the axis direction of each trace in the top left corner showing a) the unfiltered trace and the trace after b) harmonic filtering, c) rotation and d) polarization and directional filtering.
Seismic trace a) before and b) after harmonic filtering showing the development of beating towards the end of the trace. Note that the harmonic filtering still improves the signal-to-noise ratio at the beginning of the trace, where the qP- and qS-arrivals are present.
Superposition of Fourier transforms of all 1000 m walk-away VSP data showing the signal frequency ranges from ~25 Hz to 250 Hz, with strong 50 Hz harmonics.
Figure 2.4

The three components of the 1000 m, northeastern azimuth, walk-away VSP. Before (left) and after (right) harmonic, polarization and directional filtering to isolate and amplify a) the qP-wave and harmonic, polarization, directional and bandpass filtering to isolate and amplify b) qS₁ and c) qS₂.
2.6 Bibliography


3.0 Seismic Refraction Traveltime Inversion for Static Corrections

Topographical and overburden variation in the Outokumpu area caused the walk-away VSP data set to require static corrections. In this chapter a seismic velocity model of the near surface is developed using critically refracted wave travel times measured from the seismic reflection survey data. Static corrections are calculated from the model and applied to the walk-away VSP data set. This chapter has been submitted to the journal *Geophysical Prospecting* for publication and is currently under revision.

3.1 Introduction

The town of Outokumpu, in northern Karelia, Finland, is the site of a historical but now depleted base metal mine. Renewed exploration of the area has been spurred in part by the recent acquisition of a crustal-scale seismic reflection profile through Finland acquired as part of the FIRE (Finnish Reflection Experiment) project (Kukkonen et al., 2006, Sorjonen-Ward, 2006) which prompted a reassessment of the regional geology. The profile near Outokumpu revealed an unexpected series of horizontal reflectors that appeared related to the Outokumpu assemblage exposed at the surface to the northwest and this further motivated the drilling of the International Continental Scientific Drilling Program (ICDP) wellbore to a depth of 2.5 km (Kukkonen and the Outokumpu Deep Drilling Working Group, 2007). High resolution seismic reflection/refraction profiles and multi-azimuth multi-depth walk-away vertical seismic profiles (VSP) were obtained as part of this study. However, the analyses of these data were hampered by severe (10-70 ms) static time shifts caused by a combination of the large velocity contrast between the underlying metamorphic rock and Quaternary glacially deposited sands and gravels and the rapidly fluctuating surface topography.
Conventional static correction techniques failed to satisfactorily correct the VSP data. Here, an alternative strategy that determines the source and receiver static corrections via seismic refraction traveltime inversion is presented. The utility of these static corrections is illustrated by the improvement to the moveout curves observed in a multi-azimuth, multi-depth VSP survey.

3.2 Study Area

The study area is near the town of Outokumpu, Finland (Figure 3.1). An examination of the walk-away VSP data showed that substantial topographical variation combined with a heterogeneous subsurface to require significant static corrections (Figures 3.2, 3.4a). Geological studies of the area show the overburden is up to tens of meters thick and is principally composed of glacial till and glaciofluvial sediments (Chork and Salminen, 1993). This is confirmed by the borehole drilling results, which show a 33 m thick cover of glacial deposits overlying the bedrock that is composed of a metamorphic biotite schist. At least three ice lobes are known to have terminated in Outokumpu during the last deglaciation, and the complex structure and topography of the glacial deposits (Chork and Salminen, 1993) causes substantial traveltime anomalies. In order to apply static corrections to the walk-away data, it was necessary to create a seismic velocity model of the subsurface.

3.3 Prior Work

Near surface modeling techniques range from the more traditional and relatively simple plus-minus method (Hagedoorn, 1959, van Overmeeren, 2001, Bridle, 2007) to more complex techniques such as wavefield inversion (e.g. Pratt et al., 1996). Recently, there have been advances in using other geophysical methods to estimate near surface
seismic velocities and static corrections: Krieger et al. (2000) use gravity measurements, and Ley et al. (2006) use vibrator controller measurements of ground viscosity and stiffness to refine models and resulting static corrections. Similarly, Vesnaver et al. (2006) krig uphole data and combine satellite imagery and topography to improve static corrections. Selection of a particular near surface modeling technique involves considering the geological context and scale of the area to be modeled as well as the desired resolution. The quality, quantity and type of data available must also be considered. Where high resolution seismic data is available it typically yields static corrections of higher accuracy than is achievable with the other above mentioned methods as a result of the imperfect correlation these other methods have with seismic velocity. In this study it was possible to take advantage of the walk-away VSP offsets to collect high resolution single component reflection/refraction data which could be used for a seismic based model of the near surface.

The near surface can be difficult to image with reflection seismic profiling due to the interference of source noise, surface waves, and direct arrivals with reflected waves in this region (Miller et al., 1998). For this reason it is usually necessary to use modeling of refracted waves to examine the near surface structure. Full wavefield inversion of refracted waves can create detailed models, but involves a significant increase in time and effort over most other methods (Sheng et al., 2006). Coherence inversion (Landa et al., 1995) can yield useful results, but is ineffective when coherent noise is present. Refraction traveltime inversion has previously been used to successfully model the near surface (Lanz et al., 1998, Bergman et al., 2006, Zelt et al., 2006, Yordkayhun et al., 2007), and Bridle et al. (2006) show that it yields better results than the plus-minus
method. Traveltime inversion models are quite efficient, often requiring only a few iterations to converge, and are less likely to include unnecessary subsurface features than forward modeling of the same data (White and Boland, 1992). Due to the non-linear nature of the inversion process, an initial starting model of the seismic velocity as a function of depth and lateral location is required, and the final model is quite sensitive to these initial parameters (Kissling et al., 1994). Although occasionally both turning-ray refractions and critical refractions are used in the traveltime inversion (Bohm et al., 2006), most methods involve choosing between the gradient imaging ability of turning-ray refractions and the boundary imaging ability of critically refracted waves. In the Outokumpu area sharp velocity contrasts are present, from ~1800 m/s in the glaciofluvial sediment to ~6000 m/s in the schist, and critically refracted arrivals are necessary to accurately image these. Bridle et al. (2006) find that static corrections resulting from a traveltime inversion model using critically refracted arrivals yield less noisy results than those resulting from an inversion model using turning-ray refractions, although the authors do not expand further on the reason for this difference.

3.4 Field Program

The reflection/refraction and walk-away VSP surveys were simultaneously acquired along the same surface lines. Indeed, although the VSP surveys were the primary motivation for the work with the surface profiling secondary, the use of the latter here in static corrections impacts significantly the validity of the VSP results. The surveys consisted of two seismic lines, one along a northeastern azimuth, and one principally along a southeastern azimuth from the borehole (Figure 3.1); hereafter these will be referred to as the NE and SE lines, respectively. Both lines extend approximately 2 km in
The NE survey line had fairly consistent source coverage, with only a small segment of the survey line inaccessible due to local topography, and shot point spacing was typically 20 m. The SE line had several substantial gaps in source coverage due to local infrastructure. In order to attempt to build common midpoint fold, the shot points on the SE line were therefore spaced only 10 m apart in the places where acquisition was allowed. The geophone singles (14 Hz OYO™) used on the surface were at a nominal 4 m spacing; unfortunately there is no coverage along the most eastern section of the SE seismic line due to survey time constraints. The same three component geophone receiver was used in the borehole as has been earlier employed at the KTB scientific wellbore (e.g. Luschen et al. 1996).

Seismic data was acquired with a 1 ms sampling period on 216 channels using a semi-distributed system (Geode®, Geometrics, California). A high frequency vertical seismic vibrator (IVI Minivib™, Industrial Vehicles International, Oklahoma) source employed an 8 s linear taper sweep with frequencies 15-250 Hz with a nominal force of 25 KN (~5500 lbs). Good coupling of the vibrator plate to the frozen ground and ice produced uniform force levels across the sweep. In general, 3-5 sweeps were stacked after correlation to improve the signal to noise ratio. An automatic gain control (AGC) was applied to the seismic traces for visualization, but the data were otherwise unprocessed prior to picking first breaks for the traveltime inversion model.

Although seismic waves which turn within a layer offer important velocity gradient information, the shallow depth of the model (~40 m) and strong velocity contrasts meant that only critically refracted waves could be confidently identified, limiting each layer of the model to a constant vertical velocity. Furthermore, since traveltime picks within the
coda of the seismic signal are considered to be less accurate (Zelt et al., 2003), only the first breaks were used as input to the model. The first breaks were picked manually by picking the first amplitude extremum to 1 ms accuracy using a commercial software package (Vista® software, GEDCO, Calgary). Correct identification of the phase and time of the first breaks is crucial to the successful modeling of the velocity structure of the subsurface through traveltime inversion. In some cases identification of the first break or assignation of the correct phase to a first break was not possible due to noise present in the record, and these seismic traces were not used in the modeling.

In order to be able to compare synthetic traveltimes to first break observations it was necessary to determine the error in the first break picks. As part of this, the reciprocal times for the 549 source-receiver pairs lying within the bounds of the proposed models were calculated. The maximum difference in reciprocal times was 17 ms, with 73% of the pairs having a difference of \( \leq 2 \) ms. The few larger reciprocal times may have been caused by a misidentification of the first break due to a phase change or low signal to noise ratio, or by incorrect triggering of the vibroseis source. In assessing the typical error in first break picks, signal to noise ratio and the frequency of the first arrivals were also considered. From these and the reciprocal time information, an error of \( \pm 2 \) ms was assessed for all first break picks.

### 3.5 Application of Refraction Traveltime Inversion

Zelt and Smith’s (1992) algorithm, initially developed for crustal scale studies, was chosen for the purpose of traveltime inversion with input parameters modified to compensate for the much reduced scale. The algorithm uses ray tracing to calculate traveltimes through the model and an iterative damped least squares method to solve the
linearized inverse problem. Although the algorithm is intended principally for large scale wide angle crustal data, it remains effective when applied to smaller scale surveys (e.g. Domes, 2004) such as the Outokumpu survey. The algorithm can invert multiple phases simultaneously, cope with irregular node spacing and hold parameters fixed during inversion as required.

The algorithm requires an initial starting model as input. A three layer initial model was chosen to correspond to the expected near surface structure. Evidence from the shot gathers, which showed two distinct critical refractions (Figure 3.2) support this interpretation. From previous published research in the area (Chork and Salminen, 1993), borehole drilling and seismic results, the layers can be expected to correspond to a topmost layer of glaciofluvial sand and gravel and a middle layer of more competent glacial till all underlain by the thick mica schist. For simplicity, this initial model was laterally homogeneous. Refracted ray path geometry calculations based on the intercept-time method were used on several shot gathers to obtain an appropriate initial value for the thickness and seismic velocity of each layer. This technique is limited in that it assumes homogeneous planar layers and does not consider the errors in picking the traveltimes, crossover offsets, and zero offset intercepts of refractions. The initial values can therefore be expected to have substantial error, however; the calculated values for layer thickness and velocity nonetheless suffice for an initial model.

The NE and SE surveys were modeled separately, however; the crookedness of the survey lines required both of these surveys to be further segmented prior to modeling (Figure 3.1). The algorithm requires a two dimensional input and output, and hence, in each case, the crooked survey line needed to be projected onto a two dimensional plane.
For each segment, a line of best fit was determined based on all of shot points within the bounds of the segment. As the walk-away VSP shot points occasionally differed somewhat from the reflection/refraction survey shot points, using both sets of shot points in the determination of a line of best fit ensured the most general traveltime inversion model possible. The shot points from the reflection/refraction survey were projected perpendicularly on the line of best fit for each segment, and the receivers were then placed along the line at their true source-receiver offset. The maximum offset between the shot point and geophone locations and the line of best fit was 91 m, and the average offset was 21 m. As a result of these offsets, and of the dense geophone and shot spacing, a particular geophone could differ considerably in its projected location on the line of best fit depending on which shot point it was associated with. As the seismic line was split into segments, each segment demonstrated increased linearity, however; each splitting of the seismic line was accompanied by a reduction in the amount of data available for the model, since only data with both shots and geophones within the segment boundaries could be included. The length of each model was determined by weighing this loss in data quantity against the gain in data quality and reduction in artefacts caused by the crookedness of the original survey line. Regardless, the number of observations traced in a given inversion over a segment was never less than 867, still a respectable number (Table 3.1).

The NE survey line was particularly crooked and, to combat this problem, was divided into four separate segments for the purposes of modeling (Figure 3a). The models varied in length from 360-600 m. All of the NE line had coverage by both source and geophones, but between the four segments there remained some shot points excluded
from the modeling. These shot points were not included within any segment model due to the large discrepancy between their location and those of the neighbouring lines of best fit. In these cases, after the segment models were complete, a linear interpolation between the nearest modeled shot points was employed to determine an appropriate static correction for these unmodeled locations. While all models were inverted using a fine grid parameterization (Table 3.1), the complexity of the subsurface meant that some models or layers were less resolved than others and required larger gridding to avoid oscillations in the layer boundaries and velocities.

While the SE survey line was straighter, the line could not be completely covered by geophones, with the result that traveltime picks from the first refraction were not available in some areas. Additionally, the shot points were more limited in this line because of local infrastructure. The SE line was divided into two sections for the purposes of modeling (Figure 3.3b). For the most part, the SE model was parameterized using a fine grid, but shot spacing was not as consistent in this model as in the NE model; where there was a substantial gap in shot spacing and geophone coverage, the model nodes were reduced to one lateral node between shots.

The first segment of the SE line had geophone coverage for all but the last 80 m of the model, and only three small gaps in source coverage throughout the segment. There were no geophones past this segment of the seismic line, and so modeling the second segment of the SE line proved challenging. To this end, the last 260 m of the model of the first segment was rotated onto the line of best fit of the second segment to comprise the first part of this segments model. This part of the model included the only receivers present in this segment, and was held fixed from inversion during the subsequent
modeling of this segment. This allowed a reduction in the model uncertainties, since only the ray path outside of this section was considered unknown. As a further complication in this segment, there existed a 440 m gap between the last shot point of the first segment of the SE line and the first shot point of the second segment; this large gap, combined with the lack of geophone coverage in this model, resulted in a particularly high dependency of the final model on the initial model parameters, particularly the initial velocity of the third layer. Since direct arrivals and first refractions were only observed at shot-receiver offsets of <100 m (Figure 3.2), the geophones, more than 440 m away, recorded only the second refraction for these shot points. The critically refracted first breaks recorded from the first shot point of this segment show an initial velocity of 6000 m/s for the third layer in this area, and this initial velocity is used in the model. The model of this second segment of the SE line is necessarily underdetermined, and substantial depth/velocity tradeoffs are evident due to the measurement of only one phase in the refraction data (Docherty, 1992), however; it yields enough information to cause some improvement in the model and resulting static corrections. Despite part of the model for this segment having been created from a different data set, the final model of this segment manages to trace over 95% of the 1309 traveltime observations. The model has an RMS traveltime residual of only 3 ms, and can therefore be expected to be an effective representation of the subsurface traveltimes, if not necessarily of the subsurface geology.

Past the end of the second segment, at station 3290, there remained only a few shot points for the refraction survey, with the final shot point at station 3315 (Figure 3.1). These were increasingly broadside to the receivers with a very low signal to noise ratio,
and it was judged that the two dimensional model would not yield useful results if these were included. The walk-away VSP survey, however, included shot points to station 3330, and static corrections were required for these outlying shot points. The model at the final point, station 3290, was extended laterally for all points past this one. The true topography of all points was used in the calculation of static corrections.

3.6 Refraction Traveltime Inversion Results

The models required 4-6 iterations to converge, with the final models having RMS traveltime residuals of 2-4 ms for both phases modeled (Table 3.1). Normalized $\chi^2$ values were calculated as an additional measure of the accuracy of the traveltimes through the model, and these values ranged from 1.072-2.515. They could not be reduced closer to the ideal value of 1, likely as a result of projecting a crooked survey onto a 2-D plane (Zelt and Smith, 1992). The larger traveltime errors occurred in the SE model due the poorer parameterization of the surface topography resulting from an attempt to minimize the total parameters of this larger model.

After modeling was complete, the resolution of the models was tested using the parameter perturbation technique described by Zelt and Smith (1992), where node parameters are perturbed and the model is inverted to see how many other nodes the perturbation affects. Representative nodes are tested in each model, and the models were found to be well resolved to a lateral precision of 20 m where node resolution was $\leq 20$ m. Since previous research using this algorithm indicates the lateral resolution is limited by the shot spacing (Zelt, 1999), this seems reasonable. Absolute uncertainty of representative nodes was measured, again using parameter perturbation. In this case a parameter is perturbed and held fixed during inversion; this is repeated with increasingly
greater perturbations until the recovered model is no longer able to trace rays to all traveltime picks observed in the original model or the normalized $\chi^2$ value of the recovered model fails an F-test when compared to the $\chi^2$ value of the original model (Zelt and Smith, 1992). Using this technique, vertical boundary uncertainty averaged ±2 m, and velocity uncertainty averaged ±60 m/s, ±260 m/s and ±520 m/s for the first second and third layer velocities, respectively. This increased resolution near the surface is expected due to the greater angle aperture provided by multiple phases through the top layers (White, 1989).

Once the models were completed, the vertical traveltime through the model from the surface to the datum elevation of 57 m above sea level was calculated for each shot point within the walk-away VSP data sets. The velocity anisotropy of the schist was ignored here since the distance traveled by the ray through the schist is small. If, as expected in this setting, the ray is assumed to travel 10% slower in the vertical direction than in the modeled direction, the anisotropy is less than the average 11% velocity uncertainty calculated using parameter perturbation, and the maximum effect observed by such anisotropy would be only a 0.6 ms increase in the traveltime to the datum elevation. The P-wave static corrections were then performed by correcting each seismic trace of the walk-away VSPs by the calculated traveltime to the datum elevation (Figure 3.4).

The analysis of the walk-away VSP data also required S-wave static corrections. The S-wave traveltimes, however, could not be consistently picked from the refraction data due to the very low signal to noise ratio, and it was not possible to directly create a model using traveltime inversion. A strong S-wave is observed to originate at the glacial till-schist boundary in the refraction data set (Figure 3.2), and it is likely that S-wave
arrivals observed in the walk-away VSP data are converted from the down-going P wave at this boundary. No S-wave is observed to originate at the glaciofluvial sediment-glacial till boundary, although Carr et al. (1998) measure an average $V_p/V_s$ ratio of 3.6 for glacial till, at which point it is possible the refraction, if it exists, might be hidden within the source generated noise. Similarly, direct S-waves generated by the source would likely travel slower than air waves and not be observable. In the absence of direct evidence, it is assumed the S-wave arrival observed in the walk-away VSP has been converted at the schist boundary. Sonic logging of the P- and S-wave velocities in the Outokumpu borehole by both the drilling contractor NEDRA initially, and repeated by the ICDP Operational Support Group, gives an average $V_p/V_s$ ratio of 1.76±0.05 through the schist, and this ratio is assumed throughout the model to determine values for the S-wave velocities from the modeled P-wave velocities. Although velocity measurements by Christensen (1996) and Cholach et al. (2005) in similar rocks yield comparable $V_p/V_s$ ratios, Godfrey et al. (2000) finds $V_p/V_s$ ratios in anisotropic schists are highly dependent on direction of propagation. Any anisotropic traveltime effects cannot be accounted for within the static corrections, but, as in the P-wave corrections, are expected to be negligible over the distance to the datum elevation. S-wave static corrections are therefore determined by calculating the P-wave traveltime through the upper two layers, and the S-wave traveltime through the schist to the datum elevation using the estimated $V_p/V_s$ ratio of the schist (Figure 3.5).

3.7 Discussion

The subsurface model was created in order to apply static corrections to the walk-away VSP data set, but has the added benefit of allowing the study of the near surface
conditions at Outokumpu as well. In some areas it is possible to see where the surface of the schist has likely been reworked by subsequent glacial events. Variations in the thickness and proportions of the glacial deposits are evident, and some velocity variation can also be observed. The model yields important information about the area geology, but, as in all traveltime inversion models based on critical angle refractions, is unable to resolve heterogeneities smaller than the sampling wavelength, and cannot detect low velocity layers hidden under higher velocity layers.

P-wave static corrections from the model yield substantial improvements in the walk-away VSP data set. The smoothness of the move-out increases dramatically, indicating that the effects of the overburden and surface topography have largely been accounted for. While the confidence in the static corrections from the second segment of the SE model is lower than in the other segments due to the high dependence of the results on the initial velocity chosen for the third layer of the model, numerous shot gathers were examined and care was taken in selecting an initial velocity for this layer so the static corrections from the resulting model are expected to be reasonable representations.

The S-wave arrival in the walk-away VSP also shows improvement after the application of S-wave static corrections. The method appears to be quite successful even though the results are less good than those observed in the P-wave arrivals due to the necessity of approximating the S-wave velocity in the schist using the V_p/V_s ratio measured by sonic logging. Interference from other waves and noise within the data make the S-wave quite difficult to pick; after the static corrections the S-wave arrival becomes substantially clearer.
Although some traveltimes through the models disagree with observations by slightly more than the ±2 ms error expected by the error analysis of the first break pick observations, and normalized $\chi^2$ values are >1 in most cases, these small errors can be attributed to the crooked seismic line, and the inability of the high precision data to sample small-scale heterogeneities within the subsurface (Zelt and Smith, 1992). Two dimensional models are not able to perfectly model crooked line geometry, particularly if three dimensional heterogeneities exist (Zelt and Zelt, 1998). Traveltime models are non-unique, however; using a grid parameterization and inversion yields a minimum structure model, where the structure present can therefore be expected to be required by the model. The small travelt ime errors show that the inversion is successful at creating a model that is the traveltime equivalent to the true subsurface, and therefore successful at determining effective static corrections.

The thickness of the glacial deposit cover shows quantitative agreement with unpublished electromagnetic surveys in the area and with the borehole drilling results. The velocities of the weathered glaciofluvial sediments and glacial till calculated by the model (averaging 420±100 m/s and 1800±160 m/s, respectively) also appear reasonable when compared to published literature (Stumpel et al., 1984, Carr and Hajnal, 1999, Juhlin et al., 2002), and are achieved relatively independently of the initial parameters used. While the uppermost layer of the model includes seismic velocities up to 10 m/s below the air wave velocity (approximately 330 m/s), it is important to recall that the velocities in this region of the model have an average resolution of only ±60 m/s. It is therefore not possible to definitively exclude these velocity values from the model. The velocities in the upper layers of the model appear more uniform than those modeled in the
schist, partly due to the lower lateral resolution resulting from the measurement of fewer first refraction than second refraction arrivals, but also as a result of the difference in magnitudes of the velocities. The percent variation of the modeled velocities is quite similar in each layer of the model, but as the magnitude of the velocities increased with increased layer depth, the magnitude of the variation also increases and therefore becomes more apparent.

Luosto et al. (1990), using traveltime inversion, measure a P-wave velocity of 6050-6100 m/s for the upper ~1 km of the crust in the Outokumpu area, as part of a 400 km long cross section. Although the resolution of the model by Luosto et al. is orders of magnitude larger, these velocities agree well with the velocities of 4900-6580 m/s measured in the schist by this near surface model, when heterogeneity in the crust and near surface effects are considered. Borehole sonic logs confirm the plausibility of the modeled results, with P-wave velocities measured from 0-60 m above sea level varying, with no discernable trend, from 4860-5830 m/s.

3.8 Conclusions

A high resolution seismic refraction survey yields an important asset with which to conduct static corrections and examine subsurface geology. Refraction traveltime inversion can provide sufficient resolution to calculate accurate source static corrections, and can resolve sharp velocity contrast boundaries well. Good static corrections can be achieved even when receivers and sources are not coincident if an initial model is constructed which includes good seismic velocity estimates and the seismic structure under the receivers is well constrained.
This analysis was critical to the processing of the walk-away VSP as conventional static correction techniques failed to satisfactorily correct either data set. The near surface traveltime effects obscured many features of the data, and the removal of these was crucial to the successful processing and interpretation of the data. In particular, these static corrections allowed for the careful analysis of the seismic anisotropy of the area as discussed in later chapters.
The final parameters for each model, showing the number of rays successfully traced by each model, the traveltime residuals of the 1\textsuperscript{st} and 2\textsuperscript{nd} refractions, the normalized chi-squared of the model, the degrees of freedom of the model, and finally the lateral parameterization of the model. While held laterally fixed, the boundary nodes were allowed to vary vertically.

Within each layer the velocity was constant in the vertical direction, but varied laterally across the layer.
Figure 3.1.

Map of the survey area, including map on Finland (inset). The locations of the borehole (green), the two seismic lines (dotted) and the model segments (red) are plotted. Model segments and select station numbers along the seismic lines are labeled. Each station is 10 m apart.
Figure 3.2.

Shot gather from the northeastern seismic line plotted as time vs. offset (bottom), with corresponding topography (top). Note the first (see inset) and second refraction visible as first breaks. Some of the more severe static time shift problems are emphasized with white arrows. Also of note is the strong S-wave refraction originating with the second refraction of the P-wave.
Figure 3.3.
a) The four models of the northeastern seismic line and b) the two models of the southeastern seismic line (vertically exaggerated by a factor of 10) showing P-wave velocities in m/s and resulting static corrections to a datum elevation of 57 m, with distance along the model segments (bottom) and distance from the borehole (top). Note the 260 m overlap between the models of the SE line is not exact due to sampling differences between the models.
Figure 3.4.

Seismic traces from the walk-away VSP data set along the northeastern azimuth, with receiver at a depth of 1000 m, rotated and filtered using a harmonic and polarization filter. a) Showing the P-wave arrival before and b) after P-wave static corrections.
Figure 3.5.

a) Seismic traces from the walk-away VSP data set along the northeastern azimuth, with receiver at a depth of 1000 m, rotated and filtered with a harmonic, polarization and Ormsby filter. a) Showing one of the S-wave arrivals before and b) after S-wave static corrections.
3.9 Bibliography


4.0 Anisotropy Measurements

The anisotropic velocity measurements made using the Outokumpu data set are covered in this chapter. The group (ray) velocities are measured directly from the processed walk-away VSP, and a plane-wave decomposition method is presented and applied to the walk-away VSP data to allow the measurement of phase velocities. Several gaps exist in source coverage along the seismic lines, and the effect of these gaps on the phase velocity measurements is investigated using synthetic seismic data.

4.1 Introduction

In-situ measurements of seismic anisotropy are challenging in part because of the limited range of angles that can be achieved using surface-to-borehole seismic geometries. Measuring horizontal propagation would require an infinite source-receiver offset. Measurements of azimuthal angles are often limited by time and terrain. In-situ measurements are nonetheless very important, as accurate measurements of the anisotropy within a host rock allow the depth and lateral position of subsurface features to be more accurately delineated (Banik, 1984, Alkhalifah and Larner, 1994, Grech et al., 2002).

A 2.5 km deep borehole drilled in Outokumpu, Finland by the ICDP (International Continental Scientific Drilling Program) revealed a subsurface composed of a mica-rich schist to a depth of 1.3 km, and provided an opportunity to study in-situ anisotropy through a relatively homogeneous formation. Seismic anisotropy can have a variety of causes, ranging from the presence of preferred orientations of minerals, to layering, pore texture, faults, and non-randomly oriented fractures (Crampin, 1981). The Outokumpu
area schist is expected to be anisotropic as a result of LPO (lattice preferred orientation) of biotite within the rocks and the presence of aligned fractures.

Phyllosilicate minerals such as biotite appear to demonstrate some of the greatest LPO anisotropy, with strong shear wave splitting and substantial P-wave anisotropy (Barruol and Mainprice, 1993). The anisotropy of these minerals is so great that, when they are preferentially oriented, they are often the principle cause of LPO anisotropy within aggregates; they are often also responsible for delineating the texture of the rock as a result of their strong tendency to preferentially orient themselves (Cholach and Schmitt, 2006). Measurements of crustal foliated rocks such as schists, gneisses and amphibolites have shown P-wave anisotropy ranges from 9-20% due to LPO (Godfrey et al., 2000). Although phyllosilicate minerals themselves have close to transversely isotropic elastic symmetry (Vaughan and Guggenheim, 1986), the phyllosilicate texture can cause the aggregate rock to have either an overall transversely isotropic symmetry or an orthorhombic symmetry (Sintubin, 1994). The maximum P-wave velocities in phyllosilicates are seen along the basal plane and minimum velocities along the symmetry axis (Valcke et al., 2006). For biotite, these velocities range from 4000 to 7800 m/s for the qP-wave (Cholach and Schmitt, 2006).

Aligned fractures are also expected to cause anisotropy in the Outokumpu area schist: deviatoric stress through much of the Earth means that microcracks and fractures within the subsurface tend to be aligned. Shear cracks result from compressive stress and are aligned sub-parallel (±15°) to the direction of maximum stress. This causes the seismic velocity to have transversely isotropic symmetry (Crampin and Lovell, 1991).
Uniform aligned fractures can contribute different amounts to anisotropy depending on the fracture fluid saturation.

For a material containing aligned, oblate spheroid cracks, the seismic velocity is slowest in the direction normal to the plane of the cracks (Anderson et al., 1974). In parallel, vertically oriented cracks, vertically propagating fast shear waves are typically polarized in the direction of the fractures. This polarization is not static, and thus shows only the direction of the fractures near the receiver (Crampin and Lovell, 1991). This has been confirmed with laboratory measurements, which found this to be true within ±22.5° (Peacock et al., 1994). Liu et al. (2006) find that shear wave polarization varies with signal frequency. The variation in polarization depends on the angle of incidence in relation to the orientation of the different fracture sets. They speculate that signal frequency and polarization direction are correlated with fracture size, with high frequency shear waves being polarized in the direction of microcracks and low frequency shear waves being polarized in the direction of larger fractures in the case where there are multiple fracture sets.

4.2 Previous Work

Previously, Kern et al. (2008) have made measurements on a core sample of biotite schist from the Outokumpu borehole. At 200 MPa confining pressure, they measure the maximum qP-wave velocity to be 6.49 km/s and minimum qP-wave velocity to be 5.57 km/s, and observe the schist to demonstrate an orthorhombic symmetry. Shear-wave splitting is also measured to vary from 0.01-0.30 km/s in the axial directions, and qS velocities to vary from 3.4-3.9 km/s along the axial directions if the x-axis is taken lie in the foliation plane and point in the direction of lineation. Laboratory measurements at
lower confining pressure can be expected to vary greatly from in-situ results as a result of
decomposition of the core during retrieval, while measurements at higher confining
pressure, such as these, are at greater pressure than is necessarily present in the borehole.
Nonetheless, the laboratory measurements remain a useful guide for the expected in-situ
results.

In-situ anisotropy can be measured using either reflection or VSP methods. Seismic
methods are still only rarely used in mineral exploration, and thus most anisotropy
measurements are made for the purpose of petroleum exploration and focus on the
properties of sedimentary rocks. This is despite the fact that many ore bodies are hosted
in deformed metamorphic rocks that are expected to be anisotropic. The petroleum
industry commonly uses reflection methods, such as amplitude versus offset (AVO), to
characterize TI or orthorhombic symmetry anisotropy (e.g. Ruger, 1998, Perez et al.,
1999, Behura and Tsvankin, 2006). AVO is the method of measuring the amplitude of
reflected waves as a function of source-receiver offset. Aligned fractures cause a distinct
variation in AVO, particularly in qS-waves, when compared to AVO measurements in a
non-fractured environment. Since aligned fractures are also a cause of anisotropy, AVO
measurements allow anisotropy measurements to be approximated. While AVO can
yield good results, phase changes at far offsets can cause measurement inaccuracies.
Other reflection methods include measurements of non-hyperbolic move-out. A plot of
arrival time of a reflected seismic wave vs. source-receiver offset in an isotropic media
will have an approximately hyperbolic shape; in an anisotropic material the shape
deviates from hyperbolic, and the amount of deviation can be equated to the degree of
anisotropy. These equations become less valid at wide angles where isotropic move-out
can no longer be approximated as hyperbolic. Additionally, analysis of non-hyperbolic move-out involves stacking coherent data, and is therefore also hampered by any correlated noise present in the data (Grechka and Tsvankin, 1999). A final method of measuring anisotropy from reflection seismic is the method of seismic critical-angle reflectometry (Landro and Tsvankin, 2007). Landro and Tsvankin find that the critical angle of a qP-wave seismic reflection varies azimuthally by 5-6° in orthorhombic media. If this variation can be measured, it can allow the calculation of anisotropy, however; the authors state that it cannot be considered a stand alone method.

VSP experiments are much more direct measurements of wave propagation and hence are less sensitive to these issues, but are constrained by borehole location. Shear wave splitting can be measured using zero offset VSP, but characterization of qP- and qS-wave anisotropy requires walk-away VSP. In sedimentary rocks, anisotropy is often governed by the layered structure of the rock or aligned fractures and is often approximated as pure TI or orthorhombic symmetry; most research has focused on measuring these anisotropy parameters (e.g. Deparscau, 1991, Newrick and Lawton, 2003, Li et al., 2004, Grechka and Mateeva, 2007). Anisotropy in metamorphic and igneous rocks often has more varied and numerous causes, including a much stronger contribution from LPO effects, and is thus more complex. Previously, Okaya et al. (2004) and Rabbel et al. (2004) have used walk away VSPs to measure anisotropy in the crystalline rock of the KTB borehole in Germany, while Digranes et al. (1996) used offset VSP in the Kola superdeep borehole to measure shear-wave splitting in the surrounding metamorphic rock. Okaya et al. (2004) and Rabbel et al. (2004) use traveltime inversion of multi-azimuth multi-depth walk-away VSP data to measure
anisotropy and shear-wave splitting. Further, Okaya et al. compare the inversion results to predictions from borehole petrophysics. Digranes et al. (1996) uses two far offset VSPs to measure shear-wave splitting and polarization directions. The in-situ anisotropy analysis is supplemented with laboratory measurements on core samples.

Velocities can be measured directly from the VSP records in the time-offset domain by picking arrival times from the seismic record and assuming path lengths. This measures the velocity of the direct wavefront from the source to the receiver; such velocities are group (ray) velocities. While group velocities yield important anisotropy measurements, the corresponding phase velocities that describe the speed of a hypothetical plane wave are preferable for several reasons. First, theoretical models yield phase velocities directly from the Christoffel equations; if experimental phase velocities are determined this allows direct comparison between theoretical and experimental measurements. Furthermore, calculations done with the plane waves are mathematically much simpler. Velocity measurements over a range of depths within the subsurface, called interval velocities, allow seismic velocities to be better correlated with lithological changes. Plane waves allows interval velocities to be easily calculated using layer-stripping techniques (Kebaili and Schmitt, 1996). Finding interval velocities is a difficult task if one uses only the direct arriving wavefront times.

In order to measure the phase velocity of seismic data, it is necessary to conduct a plane-wave decomposition. This is commonly done using a τ-p transform (Stoffa et al., 1981, Kebaili and Schmitt, 1997, Mah and Schmitt, 2003, van der Baan, 2004)
4.3 Method

The survey included a multi-azimuth, multi-depth walk away VSP and a reflection/refraction survey. A vertical seismic vibrator source employed 8 s linear taper sweeps with frequencies 15-250 Hz. Two seismic lines approximately 2 km long were used, along azimuths of approximately 74° and 140°, hereafter referred to as the Northeastern and Southeastern azimuths. The shots were typically spaced 10-20 m apart, although in some places shots were not possible as a result of topography or infrastructure. A three component receiver was placed downhole at depths of 1000, 1750 and 2500 m, and a 1 ms sampling rate was employed.

The walk away VSPs required source static corrections. These were calculated from a model of the near surface created from traveltime inversion of the refraction survey data. Quasi-shear waves were assumed to convert from P-waves at the overburden-schist boundary. As this boundary lay above the datum elevation, unique static corrections were therefore made for qP- and qS-waves. Strong 50 Hz powerline harmonics were removed through a block subtraction technique (Butler and Russell, 1993). Using eigenanalysis in the time domain, the data was further processed by rotating the components of the receiver as closely as possible into the principle directions of body wave particle motion (e.g. Jackson et al., 1991), and by the application of a polarization filter and directional filter (Montalbetti and Kanasewich, 1970). Where necessary, the frequency difference between the qP- and qS-waves was taken advantage of through the application of a bandpass filter (Figure 4.1). Despite the extensive processing, qS-wave arrivals remained unidentifiable in much of the data.
All traveltime picks were made manually using the Vista\textsuperscript{\textregistered} software (GEDCO, Calgary). Group (ray) velocities were calculated after static corrections, assuming a straight ray path between the projection of the source location onto the datum elevation and the receiver. As velocity is expected to increase slightly with depth near the surface due to the increase in overburden pressure, some unquantifiable amount of ray path bending will have occurred. These velocities, in particular the velocities from the 1000 m walk-away VSP where this effect is greatest, should therefore be considered to be minimum interval velocities. Group angles were calculated based on the assumption that the ray path was vertical through the overburden to the datum elevation.

In order to obtain phase velocities, a plane-wave decomposition in the form of a $\tau$-p transform was applied to the processed walk-away VSP data. Robinson (1982) defines the transform as:

$$F(\tau, p) = \int_{-\infty}^{+\infty} f(x, \tau + px) dx$$

(4.1)

where $x$ is the horizontal offset between the source and the borehole, $p$ is the horizontal phase slowness and $\tau$ is the delay intercept time, equivalent to the zero offset arrival time (Figure 4.2). The function $f(x, t = \tau + px)$ is seismic signal in offset-time space, and $F(\tau, p)$ is the signal after transformation to intercept time-slowness space. The horizontal phase slowness is defined as

$$p = \frac{\sin \theta}{v}$$

(4.2)

where $\theta$ is the incident angle, and $v$ is the velocity, and $\tau$ is defined as

$$\tau = t - px$$

(4.3)

where $t$ is the traveltime. The discrete $\tau$-p transform is then defined as:
The discrete transform is applied to the walk-away VSP records (Figure 4.7). The calculation of phase velocities and angles requires a calculation of the vertical slowness as well. In an anisotropic medium, the vertical slowness is dependent on the horizontal slowness, \( p \), and therefore, by default, on the propagation angle, \( \theta \), as well (Mah, 1999). The vertical slowness, \( q \), is calculated over each interval by assuming a homogeneous medium (Kebaili and Schmitt, 1997)

\[
q(p) = \frac{\tau_2(p) - \tau_1(p)}{z_2 - z_1}
\]

(4.5)

Here \( \tau \) is the intercept time at a constant horizontal slowness, \( p \), and \( z \) is the receiver depth. While \( z_1 \) can be allowed to represent the datum elevation, with a corresponding intercept time of zero, to allow a comparison between phase and group velocities, this formula also allows the phase velocity over a receiver interval to be calculated (e.g. Kebaili and Schmitt, 1996), revealing the contribution of each depth interval to the overall anisotropy observations. Phase velocity and angle measurements (Figure 4.9, 4.10, 4.11) are then made by manually picking the intercept time from the transformed data for each slowness value

\[
v(\theta) = \frac{1}{\sqrt{q^2(\theta) + p^2(\theta)}}
\]

(4.6)

\[
\theta = \arctan\left(\frac{p}{q}\right)
\]

(4.7)
The accuracy of the phase velocity measurements is limited by the temporal sampling rate of the data (1 ms), by the signal to noise ratio and frequency content of the input walk-away VSP and by the spatial offset of the walk-away VSP.

The spatial sampling of the data, particularly along the Southeastern walk-away VSP, was quite irregular as a result of infrastructure in the study area. The plane-wave decomposition has some advantages in this case as it is able to average through results, and allows phase velocities to be calculated over a more continuous range of angles than is possible for the group velocities. In order to evaluate whether the calculated phase velocity curve is reasonable, given the irregular spatial sampling, a synthetic walk-away VSP was created from the measured phase velocity curve. A comparison between the synthetic and experimental VSPs, their τ-p transforms and phase velocities, will yield information about the accuracy, repeatability and backwards compatibility of the technique. This will help assess whether the experimentally measured results are reasonable and will help delineate signal and noise within the τ-p transformed data.

The measured phase velocities were converted back to group velocities and traveltimes using Thomsen’s (1986) parameters for TI anisotropy. Since the observed anisotropy is not, in fact, TI, both walk-away azimuths were considered separately and given a unique set of anisotropy parameters. Thomsen (1986) gives the group velocity, \( V \), as

\[
V^2(\phi(\theta)) = v^2(\theta) + \left( \frac{dv}{d\theta} \right)^2
\]

where \( \phi \) is the group angle (Figure 1.1) and is defined as
\[
\phi(\theta) = \arctan \left( \frac{\tan\theta + \frac{1}{v} \frac{dv}{d\theta}}{1 - \frac{\tan\theta}{v} \frac{dv}{d\theta}} \right)
\]  

(4.9)

To facilitate the calculation of \(dv/d\theta\), a polynomial curve was fit to the walk-away VSP qP-wave phase velocity curves. Once the group velocities and angles are calculated, the calculation of qP-wave traveltimes is trivial if the datum elevation and shot locations are known.

A Ricker wavelet (Ricker, 1940) is used to approximate a seismic wave

\[
Y(t) = \frac{1}{\sqrt{2\pi\sigma^3}} \left( 1 - \frac{t^2}{\sigma^2} \right) e^{-\frac{t^2}{2\sigma^2}}
\]  

(4.10)

where \(Y(t)\) is the amplitude of the wavelet. A value of \(\sigma=1.25\) ms, chosen to approximate the walk-away VSP P-wave frequency, was used in conjunction with the traveltime information to create synthetic walk-away VSPs. This synthetic walk-away was transformed using the \(\tau\)-\(p\) transform (Figure 4.8) and phase velocities of the synthetic \(\tau\)-\(p\) transforms were calculated and compared to the experimentally measured ones (Figure 4.12).

4.4 Conclusions

The Outokumpu area schist is observed to be quite anisotropic in the measured directions. The sampling rate of 1 ms gives the measured phase velocities an error on the order of ±10-30 m/s. The observed anisotropy substantially exceeds this range and can taken to be significant. Between 50-1000 m depth, the qP-wave anisotropy is observed to vary between ~5.35-5.95 km/s over the range of angles measured, which agrees well with the measurements by Kern et al. (2008) on a core sample from 818 m depth, which has
qP-wave velocities varying from 5.57-6.49 km/s. The maximum qP-wave velocity measured in the laboratory is measured in the plane of schistosity: in situ this would require a horizontal measurement, which is not achievable with walk-away VSP. The discrepancy between the minimum velocities measured can be explained by the use of a high confining pressure in the laboratory (200 MPa) which is observed to close fractures in the rock. These fractures remain open in situ, are likely to lower the minimum velocity observed in the walk-away VSPs. A similar effect is observed in the qS-wave velocities, where, in situ they are observed to vary from 3.05 to 3.50 km/s, while the laboratory measurements of Kern et al. measure qS-wave velocities to range between 3.4-3.9 km/s.

The anisotropy is quite azimuthally variable, and must have symmetry more complex than TI symmetry. Since a single fracture set would be expected to create a TI symmetry within the rock, and LPO anisotropy of biotite is known to cause either TI or slightly orthorhombic symmetry, it is unlikely, due to the substantial deviation from TI symmetry, that the observed anisotropy can be described by a single one of these causes. The observed anisotropy more probably results from the interaction of both aligned fractures and biotite within the schist, and can be described by an orthorhombic symmetry. This is supported by the measurements of Kern et al. (2008)

The velocity minimum observed to occur at about ~20° from vertical along the SE profile (Figure 4.11) seems to indicate that the fracture set is dipping; a horizontal fracture set would yield a minimum velocity for vertically propagating waves. While it is possible that this minimum could be explained by a dip in the LPO of the biotite, reflection profiles (Sorjonen-Ward, 2006) through the area and core samples from the
borehole (Kern and Mengel, 2007) appear to indicate that the schist is fairly flat lying, making this scenario unlikely.

The depth interval phase velocities show that, while anisotropy appears to increase from the surface to a depth of 1750 m, between 1750-2500 m the anisotropy is less. This is most likely as a result of a lithological change; the borehole lithology shows that, below a depth of 2000 m, the schist becomes interlayered with increasing amounts of pegmatitic granite. As pegmatitic granite is unlikely to have the same intrinsic anisotropy as the biotite-rich schist, this reduction in anisotropy appears reasonable.

The synthetic $\tau$-$p$ transforms and phase velocities show that the transform is, in fact, backward compatible, and that the phase velocities measured are therefore reasonable, and are not unduly influenced by spatial sampling gaps. The main effect of the gaps in sampling appears to be an increase in noise outside of the area of interest. This increase is clearly evident in both the synthetic and the experimental $\tau$-$p$ transforms in the form of lines tangent to the qP-wave arrival, however; appears to have little effect on the shape of the qP-wave arrival itself.
The three components of the 1000 m Northeastern (left) and Southeastern (right) walk-away VSPs, after static corrections and filtering to isolate a) the qP-wave, b) qS₁ and c) qS₂.
The relationship between $\tau$-p and x-t space when the source is at the surface and the receiver is located downhole. The amplitudes lying along the line tangent to the qP-wave arrival curve are summed to create a single point in $\tau$-p space. Figure from Kebaili and Schmitt (1997).
The qP-wave group velocities measured from the filtered 1000 m walk-away a) Northeastern and b) Southeastern VSPs.
The qS-wave group velocities measured from the filtered 1000 m walk-away a) Northeastern and b) Southeastern VSPs.
The qP-wave group velocities measured from the filtered 1750 m walk-away a) Northeastern and b) Southeastern VSPs.
Figure 4.6

The qP-wave group velocities measured from the filtered 2500 m walk-away a) Northeastern and b) Southeastern VSPs.
Figure 4.7

The τ-p transforms of the filtered Northeastern 1000 m walk-away VSP, showing the τ-p of a) the qP-wave, b) qS₁ and c) qS₂ arrivals.
Figure 4.8:

The \( \tau \)-\( \rho \) transforms of the qP-wave arrival in the a) processed, experimental and b) synthetic Southeastern 1000 m walk-away VSP.
The qP-wave phase velocities measured from the Northeastern walk-away VSPs.
Figure 4.10

Phase Velocity vs Phase Angle for Northeastern Walk-Away VSP

The qS-wave phase velocities measured from the 1000 m Northeastern walk-away VSP.
Figure 4.11

Phase Velocity vs Phase Angle for Southeastern Walk-Away VSPs

The qP-wave phase velocities measured from the Southeastern walk-away VSPs.
The qP-wave phase velocities measured from the experimental and synthetic Northeastern (blue) and Southeastern (black) walk-away VSPs.
4.5 Bibliography


Mah, M., 1999. Experimental determination of the elastic coefficients of anisotropic materials with the slant-stack method, MSc, University of Alberta, Edmonton.


5.0 Conclusions

In this chapter the final seismic anisotropy measurement results are presented and discussed. Future work may involve modeling the anisotropy measurements. Initial work on this problem is presented, and the difficulties associated with theoretical modeling of this data are investigated.

5.1 Results

This thesis presents seismic velocity anisotropy measurements made using the ICDP 2.5 km deep, fully cored, Outokumpu borehole. Anisotropy measurements were made from walk-away VSPs at two azimuths with a three component receiver placed at depths of 1000, 1750 and 2500 m. A seismic vibrator source was used along two, approximately 2 km long, survey lines. The survey lines covered two azimuths from the borehole, one to the northeast and one to the southeast.

The data was processed in several steps. A block subtraction technique was used to remove strong 50 Hz harmonics caused by nearby powerlines. The three component data was then normalized using the pre-signal noise and rotated into the directions of the arriving qP- and qS-waves. A directional and linear polarization filter based on an eigenvalue technique was then applied. The qS-waves were further isolated within the data using a bandpass filter. Extensive static corrections were developed and applied using a near surface model calculated from traveltime inversion of seismic refraction data. After the above processing, the group velocities were measured as a function of group angle where possible. Finally, a τ-p transform was applied to the data set, allowing the measurement of phase velocities as functions of phase angles. Synthetic walk-away
VSPs and τ-p transforms were calculated and compared to the experimental data to investigate the effect of uneven spatial sampling on the observed phase velocities.

5.2 Discussion

Anisotropy measurements were successfully made for the qP-waves at all depths for both azimuths. The qP-waves were observed to display less anisotropy along the NE azimuth, but quite significant anisotropy along the SE azimuth. Less overall anisotropy is observed between 1750-2500 m depth as a result of the geological shift from intrinsically anisotropic schist to a more isotropic pegmatitic granite in this region.

Measurements of the qS-waves were quite difficult due to their low signal to noise ratio resulting from their location within the coda of the seismic signal. Measurements of the qS-waves were successful only for the 1000 m walk-away VSP. Shear wave splitting was successfully observed along the NE azimuth of the VSP.

The seismic anisotropy measurements made using the 2006 walk-away VSPs are invaluable to the understanding of the Outokumpu area anisotropy. While the walk-away VSP measurements agree well with the laboratory measurements made by Kern et al. (2008), in situ measurements such as these provide information on the bulk anisotropy of the area that laboratory measurements cannot. While the work presented here has focused on measuring the anisotropy itself, the research has further reaching implications. To date, processing of the reflection profile along the SE azimuth has not been successful. The anisotropy measurements will allow the anisotropic depth migration of the reflection profile results, which may achieve a clearer image of the subsurface. Anisotropic depth migration will certainly result in a more accurate delineation of the
locations of subsurface features, and the magnitude of anisotropy observed means that these improvements may be quite substantial.

Additionally, the calculation of static corrections has yielded a model of the near surface. This model gives insight into some of the glacial processes and history of the area and may be a valuable tool for further such study in the area.

The anisotropy measurements have a broader value to the scientific community than their intrinsic value to Outokumpu area research. As seismic measurements, and in particular in situ anisotropy measurements in crystalline rocks, continue to be rare, this work highlights both some of the problems and profits of this type of research. Several processing techniques have been developed and applied to this data set that may be of value to other researchers dealing with the unique problems of processing seismic in hard rock and of making qP- and qS-wave anisotropy measurements.

5.3 Future work

Future work on the anisotropy measurements would involve extending the measurements into a three dimensional model. This could be done using either laboratory measurements on core samples or by using theoretical modeling. Such a model would be useful for a planned micro-seismic study due to take place in the area, as well as any future seismic profiles through the area or through similar rocks. Additionally, construction of such a model would give an indication of the fracture density in the rock and the orientation of aligned fractures, as well as an indication of the strength and orientation of LPO anisotropy.

Construction of a model is not a simple task, particularly for the Outokumpu rocks. Ideally, a model would be created to fit the 1000 m walk-away VSP data, as seismic
waves for this VSP travel through uniform schist. The deeper walk-away VSPs have a more complex geology and are thus less well suited for modeling. The upper 1 km of the crust, however, is under significantly less pressure than deeper rocks as a result of the decrease in overburden with increased proximity to the surface. This low pressure means that fractures in this area are more likely to be open and therefore quite likely to affect anisotropy. The 1000 m walk-away VSP therefore requires a model able to account for both LPO and fracture induced anisotropy. Research on creating and comparing such models to VSP results in crystalline rocks appears to be limited to the work of Kern et al. (2001) and Okaya et al. (2004).

Kern et al. (2001), studying the Kola superdeep borehole area, compares zero-offset VSP measured anisotropy with laboratory measurements on core samples. Good agreement is found between them when laboratory measurements are conducted at high pressure and compared to rocks at deeper depths, however; the authors discover that laboratory measurements made at low pressures do not adequately reproduce the VSP results, as the core samples have extensive drilling induced microcracks. The authors attempt to create a theoretical model by calculating, based on the elastic modulus of the minerals in the core samples and their modal proportions within the rock, theoretical elastic moduli. These results are useful at high pressures where fractures are mostly closed, but again fail to satisfactorily predict seismic results obtained for rocks nearer the surface at lower pressures.

Okaya et al. (2004) study the KTB borehole in southeast Germany. The authors calculate an effective elastic modulus of the area by assuming the walk-away VSP measurements describe a tilted TI symmetry, and attempt to compare the VSP anisotropy
measurements to a theoretical elastic modulus. They use foliation tilt information from downhole dip meters, as well as sonic logs, lithological logs and a geological cross section of the area to create a discretized model of the area. Each cell of the area was assigned an estimated elastic modulus based on its depth and laboratory measurements on rocks similar to that found in the cell. The elastic tensors along each walk-away VSP ray path was then volumetrically Voigt averaged to determine a bulk effective elastic modulus. The authors believe this to be a valid method to model anisotropy, however; they conclude that the observed anisotropy is intrinsic. Since the method is an extension of the method used by Kern (2001), it seems likely this method would not be able to accurately predict the results for more fractured rock.

Laboratory studies do not appear to necessarily be valid methods with which to model fractured crystalline rock anisotropy. Theoretical models may be more suited to accurately modeling the 1000 m walk-away VSP. The most popular model by far, Hudson’s (1981) model, uses Eshelby’s (1957) strain equations for penny-shaped cracks to calculate the effect of cracks on seismic velocities through the modification of the effective elastic modulus.

Intrinsic anisotropy with a TI symmetry can be approximated with the addition of an aligned set of fractures (Hudson, 1994). The overall model then yields a second order approximation for the effect of cracks on the effective elastic modulus (Hudson, 1986). Hudson’s model is applicable for crack densities, $\varepsilon$, $<0.1$ (Crampin, 1984). Crack density is measured as $\varepsilon = a^3 \nu$, where $a$ is the radius of the penny shaped crack and $\nu$ is the number density. While crystalline rocks such as the Outokumpu schist tend to have cracks that have a low aspect ratio (Crampin and Lovell, 1991), and thus are satisfied by
a penny-shaped crack approximation, it is less likely that the crack density in the schist is
<0.1.

Grad and Luosto (1994) study a seismic profile traversing 325 km, part of which is
trough the same orogenic domain as that of the Outokumpu area. The profile itself
passes ~100 km from Outokumpu, however; passes through many schist zones similar to
the Outokumpu area zone. From seismic attenuation measurements the authors estimate
fluid filled fractures with a crack density of 0.35-0.54 in the uppermost 200 m of the
crust, and 0.1-0.32 at depths of 0.2-1 km. Crack density does not fall below 0.1, the
threshold at which it can be modeled using Hudson’s model, until a depth of 1 km.

In fact, this high crack density is found when an attempt is made to model
Outokumpu results from the 1000 m walk-away VSP using Hudson’s model. The model
with the best fit (Table 5.2, Figure 5.1) requires ε=0.114 for the fluid filled fractures, with
fractures dip 30° and strike N80°W. The schist is found to be essentially flat lying, and is
adequately represented by dry cracks with ε=0.037 within an isotropic matrix (Table 5.1).
Excellent agreement is found between the model and the walk-away VSP results, but,
unfortunately the high crack density required to satisfy the data results in an incorrect
estimation of the effect of the aligned fractures on the elastic stiffness tensor parameters.
Hudson’s model is therefore not applicable for this data set, and accurately modeling the
Outokumpu 1000 m walk-away VSP results will require a different technique.
Nonetheless, while the quantitative estimation of the effects of the cracks is inaccurate as
a result of the high crack density required to model the VSP results, the more qualitative
aspects of the model appear to be in good agreement with the known geology. The model
requires an essentially flat lying schist, which agrees with the core sample results, which
show a horizontal to subhorizontal foliation in the schist (Kern and Mengel, 2007). The modeled cracks lie in a plane dipping in the direction of a strong NE-SW lineation in the schist (I. T. Kukkonen, personal communication, 2007). They dip at 30° to the foliation, and it is possible they are located along the bedding plane.

While it is not possible to accurately model the schist using Hudson’s model, it may be possible to use a different model which is valid for the higher crack densities observed. Schoenberg’s (1980) model is valid for higher crack densities (Grechka and Kachanov, 2006), however; fewer experimental results are available using this model and it is more complicated to use than Hudson’s model. Schoenberg’s model modifies the elastic compliance instead of the elastic stiffness tensor. Future work may concentrate on using Schoenberg’s work to model the 1000 m walk-away, or using Hudson’s model to model the deeper walk-away VSPs. The laboratory measurements on core samples from 818 m depth by Kern et al. (2008) may be used to calculate a representative elastic stiffness or compliance tensor for the uncracked schist. If more core samples from the Outokumpu borehole become available for anisotropy studies, it may be useful to compare measurements on the core samples from a larger range of depths to measurements made with VSP to both extend the Outokumpu anisotropy measurements and to further confirm the results of Kern et al. (2001) and Okaya et al. (2004) in regards to the validity of laboratory measurements in comparison with VSP measurements.
Table 5.1

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Hudson model elastic stiffness parameters (in Voigt notation) for the TI approximation of the uncracked schist.

Table 5.2

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Hudson model elastic stiffness parameters (in Voigt notation) for the cracked schist.
Figure 5.1

Comparison between theoretical Hudson-based model and walk-away VSP results.

qP-Wave Results: 1000 m Walk-Away VSP, NE and SE azimuths

qS-Wave Results: 1000 m Walk-Away VSP, NE azimuth
5.4 Bibliography


### 6.0 Glossary

<table>
<thead>
<tr>
<th>Term</th>
<th>Definition</th>
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<tr>
<td>AGC</td>
<td>automatic gain control, applied to change the amplitude of seismic signals over a given range</td>
</tr>
<tr>
<td>AVO</td>
<td>amplitude versus offset, a seismic method where the amplitude of a reflected wave is measured as a function of offset</td>
</tr>
<tr>
<td>first breaks</td>
<td>the first arrival of the seismic signal, usually the P-wave</td>
</tr>
<tr>
<td>geophone</td>
<td>a seismic receiver</td>
</tr>
<tr>
<td>HTI</td>
<td>horizontally transversely isotropic; a TI symmetry with the symmetry axis aligned horizontally</td>
</tr>
<tr>
<td>ICDP</td>
<td>International Continental Scientific Drilling Program</td>
</tr>
<tr>
<td>LPO</td>
<td>lattice preferred orientation, when a particular mineral forms bonds in a preferentially aligned direction</td>
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<tr>
<td>orthorhombic</td>
<td>a symmetry of the elastic stiffness tensor with three axes of symmetry. Requires 9 independent elastic parameters to define it.</td>
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<tr>
<td>P-wave</td>
<td>the first arriving body wave, the P-wave is a compressional wave.</td>
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<tr>
<td>qP-wave</td>
<td>quasi P-wave, the P-wave in an anisotropic media. Motion is no longer purely compressional but is subparallel to the direction of propagation.</td>
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</table>
| qS
subscript 1   | quasi S-wave, qS
subscript 1 is the first arriving shear wave in an anisotropic media. Oscillation is no longer purely orthogonal but rather is near to perpendicular to the direction of propagation |
| qS
subscript 2   | quasi S-wave, qS2 is the second arriving shear wave in an anisotropic media |
<p>| S-wave        | the second arriving body wave, the S-wave is polarized orthogonally to the direction of propagation |
| shear wave splitting | when the shear wave is split into two components with different velocities as it travels through an anisotropic media |
| slowness      | slowness, p, is the inverse of velocity, and is typically given as ms/m |
| static corrections | the correction of seismic data for the effect of near surface traveltime variation caused by topography or overburden |</p>
<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
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<tbody>
<tr>
<td>TI</td>
<td>Transversely isotropic, a possible symmetry of the elastic stiffness tensor where this is a single axis of symmetry. Requires 5 independent elastic parameters to define it.</td>
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<tr>
<td>VTI</td>
<td>Vertically transversely isotropic; TI symmetry with the symmetry axis aligned vertically</td>
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<tr>
<td>VSP</td>
<td>Vertical seismic profile, where the source is located on the surface and the receiver is located at depth</td>
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