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Non-Contact Measurements to Estimate the Elastic Properties of Rocks Under In Situ Conditions

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Abstract

Laboratory rock physics measurements are important for understanding how the physical properties of rocks control the behaviour of elastic waves propagating in the earth. Traditionally, ultrasonic waves are excited and detected in the laboratory with contacting transducers inside fluid-filled pressure vessels that replicate in situ subsurface confining stress.

Instead of transducers, we use laser ultrasonics (LUS) to generate and record ultrasound in rock samples, an entirely non-contact technique. This method offers several advantages over transducer measurements: mechanical coupling issues are avoided, very broadband (30 kHz to 24 MHz) waveforms can be recorded, the small footprint of the laser beams allows a single rock to be densely sampled, and group velocity is always measured. However, LUS has so far been limited to studies at atmospheric pressure, and since the elastic properties of rocks are strongly dependent on confining stress, the advantages of LUS have not yet been realised for realistic rock physics measurements under in situ conditions.

We have designed and implemented a methodology to perform non-contact LUS compressional wave measurements under in situ confining stress whereby rock samples are mounted inside a pressure vessel with two optical windows for the source and receiver laser beams. Experimental acquisition and arrival time picking are both automated.

To demonstrate the advantages of this technique, we investigated the anisotropy and pressure dependence of four rocks from the Alpine Fault in New Zealand. Due to the dense sampling, we experimentally determined the orientation of transverse isotropy for three protolith samples and showed that transverse isotropy was not a valid assumption for a highly fractured cataclasite sample. Fitting a curve to over 90 independently measured P-wave velocities for each sample significantly improved both the accuracy and precision in the estimates of the elastic constant $c_{13}$. Although the rocks had similar mineralogy, the observed differences in velocity, anisotropy, attenuation, and pressure dependence were mainly controlled by variations in microcrack density and alignment at shallow crustal pressures.

In the future, we intend to improve this methodology to exploit the advantages of non-contact LUS measurements under in situ conditions for a range of rock, ice, and material physics experiments.
To the One in whom is my greatest delight,
be all praise and honour and glory!
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Chapter 1

Introduction

Most problems in subsurface geophysics are inverse problems: the properties and characteristics of what we cannot see are inferred from a sparse number of noisy measurements. Moreover, in realistic scenarios these inverse problems are ill-posed and the data can be described by many different, equally valid, models. To form reliable interpretations of geophysical data where many models can explain the measurements, supplementary geophysical and geological datasets provide necessary constraints. Laboratory rock physics measurements form one of these important datasets by connecting the physical properties of the rocks, such as porosity, permeability, pore fluids, lithology, and saturation, with indirect seismic exploration or earthquake seismology data (Wang, 2001).

The propagation of elastic waves in rocks depends on many physical parameters, including mineralogy, porosity, pore shape, permeability, and fractures. Additionally, the subsurface environment, described by the pressure, temperature, pore fluid, and pore fluid pressure, affects the elasticity of rocks (Wang, 2001; Mavko et al., 2003). Laboratory rock physics is mainly concerned with describing how the physical rock properties and subsurface environment control the behaviour of elastic waves propagating in rocks. The insights and data gained from these experiments are used in a range of applications, including accurately processing seismic reflection and refraction data (e.g. Godfrey et al., 2002), extracting valuable information from subsurface images (e.g. Vernik and Nur, 1992), and better understanding and predicting geophysical processes, such as faulting (e.g. Carpenter et al., 2014; Allen et al., 2017). As such, laboratory studies of rock elasticity over the past 70 years have created a valuable body of knowledge in solid earth geophysics (Wang, 2001).

1.1 Laboratory Rock Physics Methods

In the laboratory, the elastic properties of rocks are commonly determined by the pulse transmission experiment using ultrasonic transducers (e.g. Birch, 1960; Winkler and
Plona, 1982; Vernik and Nur, 1992; Dellinger and Vernik, 1994; Sarout et al., 2015; Allen et al., 2017). This method is based on the premise that the propagation of linear elastic waves transmitted through a rock is governed by the elastic properties of the rock. By holding constant most parameters which influence rock elasticity and varying a single parameter, the effect of that parameter on the ultrasonic wave velocity, and thus elasticity, can be determined with this technique.

In transducer experiments, one or more ultrasonic transducers are coupled to the surface of a rock sample with glue or a liquid transducer couplant. A good mechanical coupling of the transducer to the rock is essential to minimise the acoustic impedance contrast and maximise the ultrasound energy transmitted between the transducer and the sample (Winkler and Plona, 1982; Durán et al., 2018). A high voltage pulse applied across a piezoelectric crystal within the transducer creates a mechanical pulse which propagates as an ultrasonic elastic wave through the rock sample (kHz-MHz range). Different orientations of the crystals allow for both longitudinal and shear wave polarisations. After propagation through the sample, the piezoelectric crystal within a receiving transducer converts the mechanical vibrations into an electrical signal. This signal is then displayed and recorded using an oscilloscope.

Aside from transducers, other methods exist to generate and record ultrasonic elastic waves in a range of media. One such method requires no mechanical contact with the sample through the use of lasers. This method, known as laser ultrasonics (hereafter referred to as LUS), has been applied for investigating media with acoustic waves for non-destructive testing (Cho et al., 1996; Silva et al., 2003), medical imaging (Johnson et al., 2018), studies of wave propagation in heterogeneous media (Scales and Malcolm, 2003; van Wijk et al., 2019), and rock physics (Blum et al., 2013; Xie et al., 2016, 2018), among other applications.

With LUS, transmitting and receiving transducers are replaced by a pulsed laser and laser interferometer for the generation and detection of ultrasound (figure 1.1; see Scruby and Drain (1990) for a detailed treatment of LUS techniques). The pulsed laser delivers a very short (< 100 ns) pulse of electromagnetic radiation, usually in the infrared or visible wavelengths, to the surface of the sample. This short high-intensity pulse causes thermoelastic expansion or ablation of the medium, generating ultrasonic waves which propagate through the sample. The resulting vibrations are detected at the sample surface by a laser interferometer, such as a laser Doppler vibrometer (hereafter referred to as LDV) or Fabry-Perot interferometer. A laser interferometer reflects one or more laser beams off the surface which vary in phase or wavelength as a result of the ultrasonic vibrations. The reflected beam is then combined with a reference beam, and the resulting intensity or frequency variations caused by optical interference of the two beams are proportional to the displacement or velocity of the sample surface, allowing the ultrasonic waveform to be retrieved (Scruby and Drain, 1990).
1.2 Rock Physics with Laser Ultrasonics

Using laser ultrasonics to generate and record elastic waves in laboratory rock samples offers several advantages over traditional transducer methods. One advantage is that the method is entirely non-contact. This eliminates the variability in sensitivity and bandwidth caused by the quality of the transducer coupling, allowing for more reliable and repeatable ultrasound generation and detection (Scruby and Drain, 1990). Moreover, the absence of mechanical contact ensures that the ultrasound recorded at the sample surface is not influenced by the physical properties (e.g. resonance/ringing) or loading of a transducer element; the LUS waveform displays only the response of the rock sample to the ultrasonic source. This means we can record ultrasonic waveforms with a very broad range of frequencies (30 kHz to 24 MHz), obtain reliable estimates of the absolute amplitude and attenuation of ultrasound (e.g. Blum et al., 2013), and utilise the coda waves for studying small changes in velocity (similar to Grêt et al., 2006). An additional advantage of non-contact LUS is that the alignment of the laser beams and their positioning on the sample surface can be easily adjusted, unlike transducer elements which require manual coupling to the surface for each sampling position.

Another distinct advantage of the LUS method over transducer measurements is the small area of the laser beams on the sample surface, also known as the footprint. Whereas the diameter of contacting transducers is often the same order of magnitude as the propagation distance of the ultrasound (Scruby and Drain, 1990; Dellinger and Vernik, 1994), LUS laser beams have a diameter at least an order of magnitude smaller than the length of a rock sample. Not only does this enable us to measure many independent propagation directions with different transects through a sample, but it also guarantees that we
always measure the group velocity, avoiding the ambiguity often present in transducer experiments as to whether the group or phase velocity is being measured (Dellinger and Vernik, 1994; Blum et al., 2013). In this study, we exploit the advantage of small beam footprints to estimate P-wave velocities at over 90 different angles around cylindrical rock cores, improving the accuracy of calculating the anisotropic properties of these rocks.

Several studies have demonstrated the advantages of LUS over contacting transducers for performing laboratory rock physics measurements. Blum et al. (2013) successfully used non-contact source and receiver lasers to estimate the elastic anisotropy and attenuation of reservoir shales. They obtain accurate anisotropy estimates by taking many independent measurements of the P- and S-wave velocities around cylindrical horizontal cores. Xie et al. (2018) use this same technique for determining elastic anisotropy in organic-rich shales, using slight changes in the theoretical and experimental method proposed by Xie et al. (2016) to enhance reliability where LUS data is noisy. While these studies clearly demonstrate the advantages of non-contact LUS, they all share a common disadvantage: these measurements are performed on the bench top at atmospheric pressure.

Since the elastic properties of rocks vary significantly with confining stress, it is very important to perform measurements under conditions which replicate the in situ subsurface environment of the rock samples in order to record realistic results. Ultrasonic transducer rock physics experiments are typically performed inside fluid-filled pressure vessels to provide an isotropic confining stress (e.g. Birch, 1960; Vernik and Nur, 1992; Winkler and Plona, 1982). Moreover, samples are often mounted between hydraulic pistons either inside or outside a pressure vessel to create triaxial differential stresses that replicate the in situ environment of rocks (e.g. Lebedev et al., 2011; Allen et al., 2017).

Little work has been done to realise the full benefits of LUS techniques for realistic rock physics measurements under in situ conditions. Carson and Lebedev (2014) extend the work of Lebedev et al. (2011) by using an LDV to measure ultrasonic wave polarisations and velocities in a cubic rock sample under differential stress. However, an S-wave transducer glued to the sample surface was used for the ultrasonic source, and a stress could not be applied to one axis because two opposite faces had to be exposed for ultrasound generation and detection. Adam et al. (2014) present a methodology to perform fully non-contact LUS compressional wave measurements under isotropic confining stress by mounting the rock sample inside a pressure vessel with two optical windows for the source and receiver laser beams. While measurements were only demonstrated at low (1 MPa) pressure in a single direction at a time, the methodology provided a means to combine the full benefits of LUS with realistic rock physics measurements under in situ conditions. In this study, we build significantly on the work of Adam et al. (2014), demonstrating fully non-contact measurements to estimate the elastic properties of rocks under in situ conditions.
1.3 Thesis Structure

The primary aim of this research is to demonstrate the application of the high-pressure LUS methodology for performing non-contact rock physics measurements under in situ conditions. To this end, we\(^1\) present a case study of the anisotropy and pressure dependence of four rocks from the Alpine Fault in New Zealand. In chapter 2, we cover the theory behind measuring transversely isotropic rocks and give motivation for why LUS techniques are suited for this type of measurement. Moreover, we review the dependence of the elastic properties of rocks on pressure, highlighting the importance of performing measurements under in situ conditions. Chapter 3 gives a detailed description of the high-pressure LUS methodology and data processing and analysis techniques. In chapter 4, we present the background, experimental procedures, and results of our case study of Alpine Fault rocks. In light of these results, we evaluate the design and implementation of the high-pressure LUS methodology in chapter 5, and form geophysical interpretations of the Alpine Fault rocks at shallow crustal pressures. The final chapter summarises our findings, with suggestions for improvements to the methodology and ideas for future experiments.

\(^1\)The pronoun “we” is used throughout this thesis to reflect the contribution and input from the supervisors and research group. However, unless otherwise stated or implied by citations, development of the methodology, execution of experiments, and analysis of data were performed primarily by the student with guidance from the supervisors and research group.
Chapter 2

Theory

This chapter presents the theoretical background and motivation for performing non-contact rock physics measurements with high-pressure laser ultrasonics. Since we apply our methodology to investigate transversely isotropic rocks in this study, we first give a mathematical description of rock elasticity and transverse isotropy in section 2.1. Next, we review the laboratory methods used for measuring anisotropy in section 2.2, giving motivation for why LUS is especially useful for this type of measurement. Lastly, we show the importance of performing rock physics measurements under in situ conditions in section 2.3 by describing how the elastic properties of rocks depend on pressure.

2.1 Mathematical Description of Rock Elasticity and Transverse Isotropy

Hooke’s Law describes the relationship between a force applied to an object and the resulting deformation, provided the deformation is reversible (i.e. elastic). For example, a spring with stiffness $k$ deforms by a distance $x$ in one dimension when a force $F$ is applied to the spring, stated as $F = -kx$. To describe the relationship between a force and the resulting elastic deformation in a continuous medium in three dimensions, we consider the forces acting on the faces of a small cube in the medium. For this case, Hooke’s law is stated as (Mavko et al., 2003):

$$
\sigma_{ij} = \sum_{k=1}^{3} \sum_{l=1}^{3} c_{ijkl} \epsilon_{kl}
$$

(2.1)

The force per unit area is $\sigma_{ij}$, or stress, acting on the $i$-th face in the $j$-th direction, where $i$ and $j$ take values of 1, 2, and 3. Similarly, $\epsilon_{kl}$ is the strain on face $k$ in direction $l$. $c_{ijkl}$ represents an element of the elastic stiffness tensor $c$ which fully characterises the elasticity of the medium (Thomsen, 1986). Since we aim to characterise the elastic properties of
rocks in the field of rock physics, determining the elastic stiffness tensor and associated elastic moduli are important goals.

Equation (2.1) implies that every element of the strain $\epsilon$ is linearly related to every element of the stress $\sigma$ and that the elastic stiffness tensor has a shape of $3 \times 3 \times 3 \times 3$, for a total of 81 components. However, several symmetries greatly simplify the expression of $c$. Firstly, there cannot be any rotation due to a nonzero net torque on our small cube of the medium, meaning that $c$ is symmetric and $ij = ji$, $kl = lk$. Moreover, by considering the potential energy associated with the strain, it can be shown that $ij$ is interchangeable with $kl$ (Mavko et al., 2003). These symmetries reduce $c$ to a tensor with 21 independent elements which can be represented by the following $6 \times 6$ elastic stiffness matrix:

$$
c = \begin{pmatrix}
c_{11} & c_{12} & c_{13} & c_{14} & c_{15} & c_{16} \\
c_{12} & c_{22} & c_{23} & c_{24} & c_{25} & c_{26} \\
c_{13} & c_{23} & c_{33} & c_{34} & c_{35} & c_{36} \\
c_{14} & c_{24} & c_{34} & c_{44} & c_{45} & c_{46} \\
c_{15} & c_{25} & c_{35} & c_{45} & c_{55} & c_{56} \\
c_{16} & c_{26} & c_{36} & c_{46} & c_{56} & c_{66}
\end{pmatrix}
$$

(2.2)

where the indices $ij$ and $kl$ have been substituted for the indices $I$ and $J$ according to the commonly used Voigt recipe (e.g. Thomsen, 1986; Mavko et al., 2003). Using this notation, the relevant stress and strain components are contained in special six-element column vectors $T$ and $E$, so that $T = cE$. Expression (2.2) is the most general form of the elastic stiffness matrix, but symmetry within the medium of interest allows $c$ to be further simplified.

If the elastic properties of a medium vary depending on the direction of measurement, then that medium is anisotropic. Elastic anisotropy in rocks is caused by one or a combination of three main factors (Vernik and Nur, 1992; Tsvankin, 2012): (1) bedding or foliation of isotropic minerals or rock layers, (2) preferred orientation of elongated or intrinsically anisotropic mineral grains, and (3) aligned fractures or microcracks. A common type of anisotropy produced by these factors is transverse anisotropy (hereafter referred to as TI), which has a single axis of rotational symmetry (figure 2.1). Examples include rocks with horizontal layering or a network of preferentially aligned microcracks. Transverse isotropy is often divided into horizontal and vertical transverse isotropy (HTI and VTI, respectively), depending on the orientation of the symmetry axis in the subsurface. While the mathematical descriptions of HTI and VTI are identical, we use the notation for VTI given in Tsvankin (2012).

The single axis of rotational symmetry allows us to fully describe the elasticity of TI rocks with five independent $c_{IJ}$’s in the elastic stiffness matrix (Tsvankin, 2012):
Having characterised the elasticity of a TI medium with five independent elastic constants, we now describe how these constants are estimated by measuring elastic waves propagating in the medium. Applying Newton’s second law to our small cube produces a differential equation with a second-order derivative of displacement with respect to time and a first-order derivative of stress with respect to space (e.g. equation (1.1) of Tsvankin (2012)). Since both the displacement of the medium and the stress which causes that displacement are unknowns, we need to utilise a relationship between the stress and strain in order to solve the differential equation. In the elastic regime where the strain is small, we assume the strain linearly depends on stress and Hooke’s law (2.1) is the appropriate stress-strain relationship. Combining Hooke’s law with the differential equation yields the generalised homogeneous linear wave equation (Tsvankin, 2012):

$$\frac{\partial^2 u_i}{\partial t^2} - \frac{c_{ijkl}}{\rho} \frac{\partial^2 u_k}{\partial x_j \partial x_l} = 0$$

(2.4)

where \( u \) is a component of the displacement vector \( \mathbf{u} = (u_1, u_2, u_3) \), \( t \) is time, \( c_{ijkl} \) denotes an element of the elastic stiffness tensor, \( x \) denotes a Cartesian coordinate, and \( \rho \) is the density of the medium.
The coefficient $c_{ijkl}/\rho$ has dimensions of length squared over time squared, equivalent to the dimensions of velocity squared. This means that the velocity of an elastic wave described by equation (2.4) depends on the elasticity and the density of the medium. While the propagation velocity (also known as the phase velocity) of an arbitrary wave often depends on multiple elastic constants and the direction of propagation, the relationship between the wave velocity and the elastic constants demonstrates an important principle: we can estimate the elastic properties of a medium by measuring the relevant phase velocities of the elastic waves.

Because a TI medium is rotationally symmetric, the directional dependence of the velocity is a function of only the propagation angle $\theta$ from the symmetry axis $x_3$. Moreover, the description of waves propagating in any plane containing $x_3$ can be used to describe waves propagating in any direction (Tsvankin, 2012). In such a plane, TI media supports three waves with different polarisations:

1. A longitudinal wave with an in-plane polarisation in the direction of propagation, denoted as P,
2. A shear wave with an in-plane polarisation that is always perpendicular to the direction of propagation, denoted as SV,
3. A shear wave with an out-of-plane polarisation that is always parallel to the layering of the TI medium, denoted as SH.

The phase velocities $V$ of these three modes as a function of $\theta$ are:

\[
V_P(\theta) = \sqrt{\frac{(c_{11} + c_{55}) \sin^2 \theta + (c_{33} + c_{55}) \cos^2 \theta + D}{2\rho}}
\] (2.5)

\[
V_{SV}(\theta) = \sqrt{\frac{(c_{11} + c_{55}) \sin^2 \theta + (c_{33} + c_{55}) \cos^2 \theta - D}{2\rho}}
\] (2.6)

\[
V_{SH}(\theta) = \sqrt{\frac{c_{66} \sin^2 \theta + c_{55} \cos^2 \theta}{\rho}}
\] (2.7)

where

\[
D = \sqrt{[\{(c_{11} - c_{55}) \sin^2 \theta - (c_{33} - c_{55}) \cos^2 \theta\}^2 + 4(c_{13} + c_{55})^2 \sin^2 \theta \cos^2 \theta]}
\] (2.8)

While expressions (2.5) to (2.8) are reasonably complicated, the equations quickly simplify when the propagation direction is limited to parallel ($\theta = 0^\circ$) and perpendicular ($\theta = 90^\circ$) to the symmetry axis:
Thus, by measuring the P-wave velocities both parallel and perpendicular to $x_3$, along with both polarisations of the shear wave at $90^\circ$, four of the five independent elastic constants can be estimated. Since $c_{11}$ is greater than $c_{33}$ in most TI rocks (Thomsen, 1986), $V_P(0^\circ)$ is referred to as the slow P-wave and $V_P(90^\circ)$ as the fast P-wave. Similarly, $c_{66} > c_{55}$ in most cases, so $V_{SH}(0^\circ) = V_{SV}(0^\circ) = V_{SV}(90^\circ)$ are the slow S-waves and $V_{SH}(90^\circ)$ is the fast S-wave. Since this study will be concerned mostly with these four wave speeds in TI rocks, we simplify the notation according to expression (2.11). Figure 2.2 shows a visual summary of the different wave polarisations and velocities in TI media.

\[ V_{P0} = \sqrt{\frac{c_{33}}{\rho}}, \quad V_{P90} = \sqrt{\frac{c_{11}}{\rho}}, \quad V_{S0} = \sqrt{\frac{c_{55}}{\rho}}, \quad V_{S90} = \sqrt{\frac{c_{66}}{\rho}} \] (2.11)

Additionally, we quantify the P-wave anisotropy as a percentage by the difference of the fast and slow P-wave velocities over their average:

\[ \text{Anisotropy} = 100 \times \frac{V_{P90} - V_{P0}}{\frac{1}{2} [V_{P90} + V_{P0}]} = 100 \times \frac{\sqrt{c_{11}} - \sqrt{c_{33}}}{\frac{1}{2} \left( \sqrt{c_{11}} + \sqrt{c_{33}} \right)} \] (2.12)
As shown by expressions (2.5) and (2.8), estimating $c_{13}$ is more complicated than determining the other elastic constants; we must use at least one measurement of the P- (or SV-) wave velocity between $0^\circ$ and $90^\circ$. Such measurements have traditionally been difficult to accurately obtain on a single sample (see section 2.2). However, LUS allows us to obtain many individual samples of the P-wave velocity at different angles around the sample, providing an experimental estimate of $V_P(\theta)$. With this data, we can invert for $c_{13}$ by fitting a curve in the form of expression (2.5).

Because the footprint of the laser receiver on the sample surface is much smaller than the propagation distance through the sample, our LUS setup measures the group velocity as opposed to the phase velocity (Dellinger and Vernik, 1994; Blum et al., 2013). In order to use the theory for phase velocity and phase angle described above, we convert group parameters to phase parameters using the relationships given in Tsvankin (2012):

$$V_G = V \sqrt{1 + \left( \frac{1}{V} \frac{dV}{d\theta} \right)^2}$$  \hspace{1cm} (2.13)

$$\tan \psi = \frac{\tan \theta + \frac{1}{V} \frac{dV}{d\theta}}{1 - \frac{\tan \theta}{V} \frac{dV}{d\theta}}$$  \hspace{1cm} (2.14)

where $V_G$ is the group velocity and $\psi$ is the group angle. The derivative $dV/d\theta$ is zero for phase angles of $0^\circ$ and $90^\circ$, implying the group velocity and angle are the same as the phase velocity and angle in these directions. This allows us to estimate the elastic constants in expression (2.11) using the measured group velocities at $0^\circ$ and $90^\circ$. However, to estimate $c_{13}$, equations (2.13) and (2.14) are used to convert the phase velocity and angle to group velocity and angle before fitting a curve to the data.

### 2.2 Methods to Measure Transverse Isotropy

Because most rocks are not elastically isotropic but exhibit some degree of anisotropy, usually transverse isotropy, many laboratory rock physics studies are concerned with describing the anisotropy of rocks (e.g. Jones and Wang, 1981; Vernik and Nur, 1992; Godfrey et al., 2000; Allen et al., 2017). Accurately understanding the anisotropy of shales is of particular importance for the hydrocarbon industry, as anisotropic shales comprise a large percentage of sedimentary basins, reservoir seals, and source rocks (Wang, 2002a). If elastic anisotropy is not accounted for in the processing of seismic reflection data, for example, seismic reflectors will not be accurately positioned in subsurface profiles, leading to uncertainty in the location of reservoirs and subsurface structures (Godfrey et al., 2002; Tsvankin, 2012). However, reliable estimates of elastic anisotropy have traditionally been difficult to obtain in rock physics experiments. In the following paragraphs, we review
the traditional methods for estimating the elastic constants of TI rocks and the studies which have suggested improved methods to fully characterise transverse isotropy in the laboratory.

Traditionally, two measurements are made at 90° and 0° to the symmetry axis to determine the fast and slow P- and S-wave speeds using transducers, and a third off-axis measurement, usually at 45°, provides an intermediate wave speed for determining $c_{13}$ in TI rocks. In some studies, these measurements are carried out on three separate cores (e.g. Jones and Wang, 1981; Vernik and Nur, 1992; Hornby, 1998), while in other studies a single core is used for all three orientations (e.g. Wang, 2002a,b; Dewhurst and Siggins, 2006). Whether one or more cores are used, two main challenges impede the reliability of measuring transverse isotropy in the laboratory using this method: (1) the orientation of the symmetry axis is either unknown or must be assumed a priori, and (2) the single off-axis measurement results in a large degree of uncertainty in the calculated values of $c_{13}$. These measurement uncertainties have resulted in inaccuracies for published values of $c_{13}$ that can be greater than 50% (Yan et al., 2012, 2016; Sarout et al., 2015).

Several studies have focused on improving the accuracy of measuring $c_{13}$. Lu and Cheshnokov (2015) used the shear waves in an off-axis core to determine its true angle relative to the TI symmetry axis rather than assuming a 45° orientation to improve accuracy in the calculation of $c_{13}$. Other studies have obtained multiple off-axis measurements on a single sample with transducers (Yan et al., 2014) and laser ultrasonics (Blum et al., 2013; Xie et al., 2018), allowing $c_{13}$ to be determined from a least-squares fit of equation (2.5) to the data. Furthermore, theoretical constraints have been applied to guide the inversion for $c_{13}$ of shale samples (Yan et al., 2016; Xie et al., 2016). While these studies successfully report greatly improved methods for estimating $c_{13}$, they still assume these measurements are recorded in a plane perpendicular to the symmetry axis and have not yet been demonstrated under in situ stress conditions. Finally, Sarout et al. (2015) use an array of transducers around a cylindrical TI sample under confining pressure to obtain 65 P-wave velocities at several angles to the symmetry axis, determining $c_{13}$ with a novel inversion method.

In this study, we use our high-pressure LUS apparatus to acquire many estimates of the P-wave velocity as a function of angle around a single cylindrical sample. In doing so, we determine the orientation of the TI symmetry axis experimentally rather than assuming this a priori, and we fit equation (2.5) to over 90 estimates of the P-wave velocity around the sample to obtain accurate and precise values for $c_{13}$. We build on the work of Blum et al. (2013) and Xie et al. (2018) by performing these measurements under confining stress while retaining the advantages of non-contact LUS.
2.3 Dependence of Velocity on Pressure

So far in this chapter, we have described the constants that characterise the elastic properties of rocks and the relationships to velocity, but have not yet related these to the physical properties of the rocks or the subsurface environment. In this section, we review how the elastic properties of rocks, and thus the wave velocities, depend on pressure, demonstrating the importance of performing rock physics measurements under in situ conditions.

The elastic parameters which appear in expression (2.11) describe the resistance of the TI medium to the particle motion of the respective body waves. For example, $c_{11}$ can be thought of as the resistance of the medium to longitudinal motion of waves propagating in the fast direction, while $c_{66}$ is the in-plane resistance of the medium to shear motion for a transverse wave propagating parallel to the layering. These elastic constants are a function of both the physical properties of the rock—such as the mineralogy, porosity, and pore shape—as well as the subsurface environment, described by parameters such as saturating fluid, temperature, and pressure (Wang, 2001; Mavko et al., 2003).

In the earth, the overburden pressure acts to push inward on pore spaces, while the pressure of the pore fluid acts to push outward on the pores. As such, the elastic properties depend on the net difference between the overburden and pore pressures, commonly called the effective pressure. While non-isotropic stresses from tectonic forces can be responsible for angle-dependent velocity variations (Nur and Simmons, 1969; Mavko et al., 2003), the net overburden pressure is the most important stress affecting the elastic properties of rocks in the subsurface. The pore fluid pressure is usually taken to be the hydrostatic pressure from a column of fluid (usually water) at a depth $d$ by assuming interconnected pore spaces. Hence, a first-order expression of the effective pressure $P_E$ a rock experiences at a depth $d$ is:

$$P_E = (\rho_l - \rho_f) gd$$  \hspace{1cm} (2.15)

where $\rho_l$ is the density of the overburden rocks, $\rho_f$ is the pore fluid density, and $g$ is the acceleration due to gravity. For metamorphic rocks with a density of 2500 kg/m$^3$ where water is the saturating fluid, the effective pressure gradient is approximately 15 MPa/km.

Both P- and S-wave velocities increase as the effective pressure increases in most rocks. This increase occurs for several reasons: better contact between compliant grain boundaries, closure of microcracks which have significantly more compressibility in one orientation, and compression of the tips of lens-shaped pores (Walsh, 1965; Mavko et al., 2003; Ullemeyer et al., 2018). The velocity does not increase linearly with pressure but increases more rapidly at lower confining pressures when the microcracks and compressible pore spaces are not yet fully closed (Eberhart-Phillips et al., 1989; Wang, 2001).
For TI rocks, the five independent elastic constants vary independently as effective pressure increases. A common observation is that $c_{33}$ increases by a larger proportion than $c_{11}$ at low effective pressures, corresponding to a decrease in P-wave anisotropy (e.g. King, 1966; Vernik and Nur, 1992; Wang, 2002b; Ullemeyer et al., 2018). This is because the microcracks and mineral grains are preferentially aligned in TI rocks, unlike isotropic rocks where the microcrack and grain boundary alignment is more random. When pressure increases, the increased contact between mineral grains and closure of microcracks in a direction perpendicular to the layering increases the stiffness for particle motion in that direction, while stiffness in the direction parallel to the layering remains relatively unaffected by these changes.
Chapter 3

Methods

Performing non-contact measurements of rock samples under high pressure requires a unique experimental setup which combines apparatus and techniques from both conventional rock physics and laser ultrasonics. In section 3.1, the experimental setup is described both in terms of its individual components and also how these components are used together to acquire experimental data. Figure 3.1 shows a diagram of the entire experimental setup which will provide a useful context for understanding the place of each component in the experiment as it is described. We also explain the techniques and software that were used to process and analyse the LUS data in section 3.2.
Figure 3.1: Diagram showing the experimental setup. 1: High-pressure gas bottle. 2: Pressure gauge. 3: Pulsed source laser. 4: Infrared source laser beam. 5: Pressure vessel. 6: Optical window. 7: Focusing lens. 8: Brass-jacketed core sample. 9: Stepper motor. 10: Receiver laser beam. 11: Laser Doppler vibrometer. 12: Arduino motor controller and circuitry. 13: Control and acquisition PC. 14: Vibration isolated laser table.
3.1 Experimental Setup

3.1.1 Pressure Vessel

In order to perform LUS measurements under high pressure, a specialised pressure vessel was constructed. This pressure vessel serves the same purpose as those typically used to perform measurements of rock samples with transducers, providing an isotropic confining stress to replicate in situ pressure conditions. Two custom-made optical windows are positioned on opposite sides of the vessel to allow line-of-sight for the laser beams to the sample under pressure. The optical windows are made from 25 mm thick plugs of sapphire, as sapphire has high transmissivity over a wide range of wavelengths from ultraviolet to near infrared, permitting both the source and receiver laser beams to pass with minimal reflection and attenuation. To minimise the attenuation of laser light, the vessel was pressurised by nitrogen gas rather than hydraulic fluid. Using gas pressure also allowed us to operate a small electric stepper motor within the vessel for rotating the sample.

Constructed from 75 mm thick stainless steel, the pressure vessel has a design pressure of 41 MPa (6000 psi) and a maximum operating temperature of 200°C. However, the maximum pressure used in this study is limited by the nitrogen gas bottles which have a gauge pressure of 20 MPa. A series of stainless steel discs was stacked below and above the sample to adjust the sample positioning and provide a convenient mounting surface. These discs also filled unused space within vessel, minimising the volume in order to reduce the safety risk and prolong the use of the nitrogen gas bottles. A 2.5 kW heating mantle could be placed around the outside of the pressure vessel to increase the temperature of the gas.

3.1.2 Rotational Stage

A distinct advantage of non-contact LUS measurements over transducer measurements is the ability to easily adjust the position of the source and receiver beams which have a much smaller footprint than typical transducers. In order to fully exploit the benefits of this flexible beam positioning and higher resolution, the position of the rock must be easily adjustable to record waves at different locations. Rock cores are commonly rotated in order to study angle-dependent variations with LUS, especially in TI rocks. However, rotating the sample inside the pressure vessel provided a unique challenge, as the electric motor had to be robust enough to remain accurate under high pressure, but inexpensive enough so that the cost of experimenting with different arrangements was not prohibitive.

To rotate the sample inside the pressure vessel, we used an inexpensive stepper motor (an electric DC motor that moves the magnet in small angular steps when current is provided to the motor coils in sequence). Due to its small number of robust components and capability to rotate 0° to 360°, a stepper motor was favoured over a servo which has more components and is often limited in rotation angle. The particular stepper motor
used (RS Pro brand) moved at increments of $0.90 \pm 0.05^\circ$ and positioned the sample accurately and reliably even at the highest pressures. It was powered via four wires that were connected through a cable gland in the pressure vessel cap (figure 3.1).

An Arduino micro-controller was used to control the position of the stepper motor (see https://www.arduino.cc/). The Arduino was programmed with customised code which received position commands from the control PC through serial communication and then relayed this position to a stepper motor driver circuit through four digital output pins. These digital signals controlled a series of switches in the stepper motor driver circuit which supplied power to the motor coils in the correct sequence, moving the stepper motor the appropriate number of steps. A rotating potentiometer knob connected to the Arduino enabled manual fine-tuning of the stepper motor position to correctly align the laser beams with the sample.

### 3.1.3 Pulsed Source Laser

Elastic waves were excited on the surface of the rock using a pulsed laser (SpectraPhysics QuantaRay INDI Nd:YAG laser). The pulsed laser uses neodymium-doped yttrium aluminium garnet (Nd:YAG) as the lasing medium, which is optically pumped with a flash-lamp to excite the Nd atoms. $Q$-switching produced a short-duration (6-9 ns) pulse of 1064 nm light by suddenly changing the quality factor $Q$ of the laser optical cavity from low to high when the number of Nd atoms in the excited state is greatest (Silfvast, 2004). This generated an infrared laser beam with an energy of up to 450 mJ per pulse at a rate of 10 pulses per second. The variable output intensity of the flashlamp determined the actual energy of the laser pulse, which was usually set to create the largest amplitude elastic waves without damaging the surface of the sample (typically between 20 mJ per pulse and 150 mJ per pulse, depending on the degree of source beam focussing). Absorption of the laser pulse at the surface of the sample causes very rapid localised heating which results in rapid thermoelastic expansion. This creates stresses and strains within the sample that propagate as ultrasonic waves (Scruby and Drain, 1990).

The source laser beam had an unfocussed diameter of 10 mm. However, it was often desirable to decrease the diameter of the source laser beam both to decrease the area over which waves are generated and increase the frequency content of the elastic waves (Scruby and Drain, 1990). A lens fixed inside the pressure vessel was used to reduce the beam diameter to approximately 3 mm on the sample surface.

### 3.1.4 Laser Doppler Vibrometer

Following propagation through and around the sample, vibrations from elastic waves were recorded on the surface of the sample with a laser Doppler vibrometer (LDV). As
opposed to the high-power pulsed source laser, the vibrometer emitted a continuous low-power helium-neon laser and received the reflections from the sample surface. We used the Polytec OFV-505 sensor head with the OFV-5000 vibrometer controller, transforming the optical output into an electrical signal with the DD-300 displacement decoder. This decoder recovered the displacement of the sample surface as a function of time from the phase modulation of the Doppler signal. It was particularly suited to detecting ultrasonic waves as it is capable of recording vibrations with frequencies between 30 kHz and 24 MHz with a maximum peak-to-peak amplitude of 75 nm.

Since the laser Doppler vibrometer measured the displacement of the sample by the laser light reflected from the surface, the highest signal-to-noise ratio (S/N) was attained when the most light was reflected from the sample back to the receiver. To achieve this, focussing the laser beam onto the sample was essential, as the small footprint of a focussed beam allows reliable reflection back to the detector even when a surface is not optically finished or oriented exactly perpendicular to the incident beam (Scruby and Drain, 1990). The OFV-505 sensor head automatically found the focus point by adjusting the position of the internal lenses until the reflected signal was a maximum. At optimum focus, the diameter of the vibrometer laser beam on the sample surface was approximately 20 µm.

The range of focus positions over which the sensor automatically scanned was limited so that the beam never focussed on the surface of the optical window. Moreover, to further increase S/N, the sensor head was positioned at the minimum stand-off distance (53 cm) so that the beam diameter through the window was large, reducing direct reflections from the window surface. Also, retro-reflective tape, which reflects the laser beam in the incident direction regardless of the incident angle, was applied to the surface of samples to maximise the S/N ratio. The tape was 0.14 mm thick and has a flat frequency response with negligible impact on the amplitude of the recorded ultrasound. For particle motion perpendicular to the surface, the tape introduced a constant time offset of 0.06 ± 0.02 µs, which was corrected for in the analysis of waveform data.

The source and receiver laser beams were aligned on diametrically opposite points of the cylindrical sample. It is important to note that only particle motion towards and away from the incident receiver beam could be detected by the laser Doppler vibrometer, such as those caused by P-waves travelling through the sample and the component of Rayleigh waves perpendicular to the sample surface.

3.1.5 Data Acquisition

Data acquisition was handled by a personal computer with an AlazarTech ATS660 PCI oscilloscope card, digitising to 16 bits at 50 million samples per second. The 50 MHz sampling rate allowed for a maximum signal frequency (Nyquist) of 25 MHz, well above the highest frequencies generated in our rock samples. Electrical signals from the vibrom-
eter decoder were digitised at a scale of 16 bits to 200 mV, implying that the minimum resolvable surface displacement change is 0.15 pm with the 50 nm/V sensitivity of the decoder. Recording was triggered by the source laser firing, and an optimum signal-to-noise ratio was achieved by averaging 200 realisations to form a single waveform at each sample position.

All components of the experimental setup were arranged on a vibration-isolated laser table to avoid recording unwanted noise. Figure 3.2 shows a photograph of the setup.

![Figure 3.2: Photograph showing the experimental setup on the laser table. The green and red lines denote the paths of the source and receiver laser beams, respectively.](image)

3.1.6 Experiment Automation and Control

So far in this section, we have described the hardware used to perform LUS measurements under pressure. However, controlling these individual hardware components in a safe, convenient, and coordinated fashion to produce high-quality and well-structured data is primarily a task for software. The software which performed this task was PLACE: an open-source Python package for Laboratory Automation, Control, and Experimentation developed in-house at the Physical Acoustics Laboratory (Johnson et al., 2015; see https://github.com/palab/place for source code and documentation).

PLACE is modular and open-source in design, allowing new laboratory equipment to be easily incorporated with support from Python’s large number of powerful open-source libraries for scientific applications (Johnson et al., 2015). A new PLACE module was created for the Arduino stepper motor controller to easily control the start position,
increment and/or end position of the rotation. This new module was included with the
other instrument modules to configure and automate the LUS experiments.

An experiment was performed by first configuring the instruments and recording ex-
periment metadata in the PLACE web interface. Once an experiment was started, PLACE
proceeded through the set number of updates (in our case, the number of positions the
core is rotated to), acquiring the data from each instrument. At completion of the experi-
ment, PLACE saved two files—a configuration file that recorded all metadata and allowed
the experiment to be easily repeated, and a file that stored all recorded data. This au-
tomated workflow using PLACE ensured high-quality and reliable data acquisition for each
experiment.

3.2 Processing and Analysis

Typical scans recorded with the high-pressure LUS setup in this study contained up to 100
waveforms recorded at different positions, each with thousands of data points. These scans
were recorded at multiple pressures for several rock samples, inevitably leading to large
datasets. In order to efficiently and reliably analyse and interpret this data, specialised
code was written. The following sections describe details of the software written for data
processing and analysis, as well as the methodology for automatic arrival time picking.

3.2.1 Processing Software and Workflow

All processing, analysis, and visualisation of experimental data was performed using the
Python programming language. Python is a diverse language with a user-friendly syntax
and a large number of powerful libraries for scientific computing (Oliphant, 2007; see also
https://www.python.org/). The NumPy library (Oliphant, 2006) was used extensively for
handling the large amounts of recorded time series data, while all plots were created using
the Matplotlib library (Hunter, 2007). Functions from SciPy (Jones et al., 2001-) were
utilised for filtering and fitting data, and all propagation of errors was facilitated using
the Python uncertainties package (Lebigot, 2010-).

In order to efficiently manage and analyse large datasets, a software class was defined
to represent a scan as a Python object. This object, called a PlaceScan, contained all the
meta information and recorded data for a scan in a single Python variable, allowing for
convenient interaction with all experimental data.

A PlaceScan object was created by retrieving the scan configuration and data from file.
This information was stored in a Python variable, and plotting, processing, and arrival
time picking could then be performed by calling the relevant methods of the PlaceS-
can object (figure 3.3). The various routines for analysing and displaying the data were
written in separate Python modules which the PlaceScan object referred to. Structuring
the code in this way meant that simple control scripts could be used for performing customised workflows without interacting with backend code. Moreover, multiple scans could be conveniently analysed together, and the common PlaceScan object ensured consistency throughout data analysis and plotting.

All PlaceScan code and control scripts used for data processing and analysis can be viewed and downloaded from the GitHub repository at https://github.com/jsimpsonUoA/PlaceScan.

3.2.2 Arrival Time Picking

As demonstrated in chapter 2, the elastic properties of a rock are experimentally determined by estimating the velocity of P- and S-waves. Therefore, accurately picking the arrival time of waves in the measured waveforms is essential for obtaining reliable elasticity results. With hundreds of waveforms recorded at several different pressure steps, a semi-automated picking algorithm was necessary to pick arrival times both reliably and rapidly.

Many automatic picking algorithms have been successfully applied to detect the arrival times of waves in various fields of geophysics (Vassallo et al., 2012; Akram and Eaton, 2016). The most widely used class of algorithm detects when a certain threshold is exceeded, either in the signal itself or in the ratio of a short term average and long term average of the signal (Allen, 1982; Baer and Kradolfer, 1987). Autoregressive algorithms comprise a second class, which use statistical considerations to detect the boundary between noise and signal (Sleeman and van Eck, 1999; Kurz et al., 2005; Sarout et al., 2009). Moreover, techniques such as cross-correlation have been used to determine the time difference between two waveforms, which is particularly relevant in laboratory rock physics experiments where changing parameters (pressure, temperature, fluids, etc.) cause small changes in the elasticity of the rock, leading to differences in arrival time of the wave (Molyneux and Schmitt, 1999; Sarout et al., 2009; Durán et al., 2018).

Recently, a dynamic time warping (DTW) algorithm has been applied to assist in the arrival time picking of ultrasonic rock physics data (Durán et al., 2018). This algorithm maps all the points of a query waveform onto a reference waveform via a warping function, thereby allowing the arrival time of the query to be determined from the arrival time of the reference. DTW is considered more robust than cross-correlation for determining time shifts, as it accounts for waveform changes caused by dispersion due to attenuation and suppresses cycle-skipping (Venstad, 2014; Mikesell et al., 2015; Durán et al., 2018).

DTW was applied to the LUS scan data recorded in this study to pick arrival times. However, since DTW relies on matching a query waveform to a reference waveform, it is not particularly suited for LUS data. This is because amplitudes and frequency content can vary appreciably between adjacent waveforms due to the localised sampling of the
Figure 3.3: Diagram showing the modular structure of the Python code used to analyse and plot scan data. A PlaceScan object was created with the experiment configuration and data, and the scan referred to individual modules which controlled analysis and plotting. An example of a simple control script with a typical workflow is given.

small beam footprints and variation in the absorption and conversion of the source laser energy into elastic waves. Despite this, we still used DTW to pick the arrivals of S-waves in waveforms recorded at different pressures with transducers (see section 4.4.2).

As an alternative automatic picking algorithm to DTW, we used the Aikaike Information Criterion (AIC) method. This autoregressive method considers the statistical variance of the time series data to determine the boundary between the noise preceding the onset of the P-wave phase and the signal afterwards (Sleeman and van Eck, 1999). It has been
successfully applied at both seismic and ultrasonic frequencies (Sleeman and van Eck, 1999; Zhang et al., 2003; Kurz et al., 2005; Akram and Eaton, 2016), including in laboratory experiments of anisotropic rocks (Sarout et al., 2009). For a discrete time series \( u_i(t) \) of \( N \) total samples, Maeda and Naoki (1985) showed that the AIC at the \( k \)th sample point is:

\[
\text{AIC}(k) = k \log \left[ \text{var}(u_i[1,k]) \right] + (N - k - 1) \log \left[ \text{var}(u_i[k + 1, N]) \right]
\]  

(3.1)

where \( \text{var} \) is the statistical variance, evaluated over the samples on the interval \([1,k]\) preceding and including sample \( k \), or the interval \([k + 1, N]\) after sample \( k \). The onset of the P-wave is the point at which expression (3.1) reaches its global minimum (figure 3.4).

While AIC is often combined with other techniques such as the wavelet transform and cross-correlation (Zhang et al., 2003; Kurz et al., 2005; Akram and Eaton, 2016; Sarout et al., 2009), we chose a 5-10 \( \mu \)s window around the expected P-wave arrival for each rock sample and allowed the AIC algorithm to suggest the arrival time. A visual inspection was then performed for each scan, and where the AIC picker was unable to accurately determine the arrival time due to excessively noisy waveforms or ambiguous first breaks, the pick was adjusted manually. For most of the data, the AIC picker accurately picked the first break of the P-wave, and since it is independent of adjacent waveforms, simple in application, and efficient, it was favoured over DTW for automatic arrival time picking.
Figure 3.4: AIC picking example, showing an LUS waveform $u_i(t)$ recorded on a schist sample (top) with the AIC function of this waveform (bottom). The global minimum of the AIC function indicates the onset of the P-wave, indicated by the red line.
Chapter 4

Case Study: Anisotropy and Pressure Dependence of Alpine Fault Rocks

In this chapter, we demonstrate the application of the high-pressure laser ultrasonics methodology by performing non-contact rock physics measurements under in situ conditions. We investigate four samples from the Alpine Fault in New Zealand with the aim of demonstrating how the methodology can produce valuable data which leads to reliable geophysical interpretations. In particular, we focus on anisotropy and the dependence of wave speed on pressure, making use of the high-resolution LUS measurements under pressure to fully characterise the elastic properties of the rocks. First, we introduce the Alpine Fault and the four samples, before describing the experimental procedure and then presenting the laboratory results.

4.1 The Alpine Fault

The Alpine Fault is a major plate boundary transform fault and geomorphological structure in the South Island of New Zealand. The fault extends for 600 km on land in a northeast-southwest orientation, accommodating slip between the Pacific Plate to the east and the Australian Plate to the west (figure 4.1; Wellman, 1955; Norris and Cooper, 2001). Motion on the fault is predominantly dextral strike-slip with a component of reverse faulting towards the northwest. This oblique slip is responsible for the uplift of the Southern Alps to the east of the Alpine Fault (Norris et al., 1990). The fault slips at an average rate of 27 ± 5 mm/yr, generating earthquakes around magnitude $M_W8.0$ at an average recurrence interval of 300 years with the last such event occurring in 1717 AD (Norris and Cooper, 2001; Sutherland et al., 2007).

From a geoscience standpoint the Alpine Fault is of great interest, as it represents a globally significant example for understanding continental plate boundary environments and processes. Moreover, knowledge of the structure and processes of the Alpine Fault is
important for understanding the natural hazard posed by a large magnitude earthquake on the fault which is late in its earthquake cycle (Townend et al., 2009). In a major endeavour to better understand the current state and geological evolution of the Alpine Fault as well as gain insight into a major plate boundary fault before a significant rupture, three boreholes were drilled transecting the central Alpine Fault during two phases in 2011 and 2014-15 as part of the Deep Fault Drilling Project (DFDP; Townend et al., 2009; Sutherland et al., 2017).

Figure 4.1: Locations of outcrops along creeks and the DFDP-1 site where two boreholes (DFDP-1A and -1B) were drilled along the Alpine Fault. (A) Regional tectonic setting, with the Alpine Fault running through the west of the South Island. (B) Zoomed map of the white box in (A), showing the Alpine Fault trace in red and the locations of the DFDP-1 boreholes and outcrops along creeks where rocks in this study were collected. (C) Schematic cross section through a typical fault segment. Notice the progression of rock metamorphism and shear deformation with distance to the principal slip zone (PSZ) in the hanging wall. Figure adapted from Toy et al. (2015).

The rocks surrounding the Alpine Fault consist of country rock protoliths that exhibit increasing degrees of metamorphism with proximity to the principal slip zone (PSZ), with hydrothermally altered cataclasite and fault gouge within the PSZ itself (see Toy et al. (2015) for detailed descriptions of the fault rock lithologies). Several studies have investigated Alpine Fault rocks to understand the relationship between the rock physical
properties and the processes of the fault, especially in relation to earthquake rupture characteristics (e.g. Christensen and Okaya, 2007; Carpenter et al., 2014; Jeppson, 2017). Seismic anisotropy is a defining feature of these rocks, as metamorphism and the high shear stress deformation environment around the fault produce foliation due to preferred mineral orientations, mineral segregation, and aligned microcracks (Okaya et al., 1995; Godfrey et al., 2002). As such, laboratory rock physics studies have characterised the elastic anisotropy of the fault rocks and protoliths to improve velocity models for seismic imaging and understand the fault zone environment (Okaya et al., 1995; Godfrey et al., 2002; Christensen and Okaya, 2007; Guerin-Marthe, 2015; Allen et al., 2017).

4.2 Description of Samples

We investigate four rock samples representative of the different lithologies surrounding the Alpine Fault, focusing in particular on their anisotropy and the dependence of wave velocity on pressure. Figure 4.2 shows photographs of these samples. The schist comes from a surface outcrop at Smithy Creek and is part of the Alpine Schist tectonostratigraphic unit, while the protomylonite and mylonite are taken from a surface outcrop at Stoney Creek (figure 4.1). The cataclasite sample is from the drill core obtained from the DFDP-1A borehole at a depth of 84.2 m below the surface, less than 10 m from the PSZ which was intercepted at 91 m (Carpenter et al., 2014). All rocks are from the hanging wall of the fault.

The protomylonite and mylonite are derived from the schist and share approximately the same mineralogy as the schist. Quartz and feldspar comprise 50-75% of the minerals, with the remainder dominated by micas (15-30%) and smaller amounts of other minerals such as calcite (Toy et al., 2015; Adam et al., 2019). We have a finely-banded schist, where foliation is characterised by sub-millimetre quartz-feldspar and mica bands. Mineral bands are wider (0.5-5 mm) and more defined in the protomylonite, while the mylonite has much finer mineral grains with a preferred orientation (Guerin-Marthe, 2015). The cataclasite is from the DFDP-1A upper foliated unit, with a mineralogy determined from XRD analysis of approximately 30% each of quartz and mica, 15% plagioclase feldspar, and lesser amounts of chlorite and calcite from hydrothermal alteration (Adam et al., 2017). Some protolith foliation is present among comminuted grains and fragments (< 1 cm in size). Hereafter, the schist, protomylonite, and mylonite will be referred to as the protoliths—i.e., these rocks are the base lithologies from which the cataclasite is derived.
4.3 Experimental Methods

To prepare the rocks for ultrasonic measurements, 25 mm cylindrical samples were cored from larger outcrop specimens for the protoliths and hand-sculpted from a section of the drill core from DFDP-1A for the cataclasite. Cores were drilled such that the axis of the cylinder lay in the foliation plane, while the cataclasite cylinder axis is aligned with the borehole axis. Samples were dried in an oven at 70°C to remove moisture from the pore spaces before being weighed and measured; the densities of the samples were then determined by dividing the mass by the cylindrical volume. Small chips at the edges of the protomylonite and cataclasite samples cause only small errors in the density estimates which fall within the stated uncertainties. Table 4.1 lists the dimensions, masses, and densities of the samples.

For high-pressure LUS measurements, all four samples were surrounded in a 38 µm thick brass jacket, applied to the surface with a thin layer of epoxy, to prevent nitrogen entering the pore spaces. Appendix A discusses the effect of this jacket on the generation and detection of ultrasound, showing that it does not significantly affect comparisons between different confining pressures. Samples were dry and thus no fluid pressure was applied. Therefore, all values for pressure quoted hereafter are assumed to be effective pressures. Samples were mounted upright on the rotational stage inside the pressure vessel and the laser beams were aligned at diametrically opposite points halfway along the cylinders. LUS waveforms were recorded at 1.8° increments over a range of 165.6° for a total of 93 waveforms in each scan; data were not recorded to 180° in order to prevent the source laser burning the edge of the reflective tape or the epoxy covering the jacket.
Table 4.1: Dimensions, masses, and densities of the four Alpine Fault samples. The uncertainties of the diameters and heights are standard deviations of 20 and 10 measurements at different locations, respectively. The densities are calculated from the dimensions and masses by assuming the cores are cylinders, with uncertainties propagated through the calculations.

<table>
<thead>
<tr>
<th></th>
<th>Diameter (mm)</th>
<th>Height (mm)</th>
<th>Mass (g)</th>
<th>Density (g/cm³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Schist</td>
<td>25.72 ± 0.02</td>
<td>26.84 ± 0.19</td>
<td>37.16 ± 0.01</td>
<td>2.665 ± 0.019</td>
</tr>
<tr>
<td>Protomylonite</td>
<td>25.75 ± 0.03</td>
<td>25.46 ± 0.25</td>
<td>35.50 ± 0.01</td>
<td>2.677 ± 0.027</td>
</tr>
<tr>
<td>Mylonite</td>
<td>25.62 ± 0.03</td>
<td>41.24 ± 0.07</td>
<td>58.45 ± 0.01</td>
<td>2.749 ± 0.008</td>
</tr>
<tr>
<td>Cataclasite</td>
<td>24.56 ± 0.26</td>
<td>19.60 ± 0.06</td>
<td>22.62 ± 0.01</td>
<td>2.44 ± 0.05</td>
</tr>
</tbody>
</table>

Seam. The seam of the jacket was positioned away from the plane of any visible foliation, ensuring that the fast and slow directions were measured within the angular range of the scan. During data processing, the angle with the fastest P-wave arrival was set to 90°. Due to the geometry of the scans, the slow P-wave direction occurs around 180°. This is consistent with the notation introduced in chapter 2, as TI media are periodic over 180°.

Scans were performed at nine different confining pressures between 1 MPa and 16 MPa. To ensure velocities were not underestimated due to rock hysteresis from plastic deformation, which has previously been observed for Alpine Fault rocks (Guerin-Marthe, 2015; Jeppson, 2017), scans were recorded on the second decrease of pressure from 16 MPa after the pressure had first been cycled to 16 MPa and then back to 2 MPa. Data acquisition at each pressure step took approximately 40 minutes, governed mostly by the 10 Hz repetition rate of the pulsed laser.

Since the high-pressure LUS setup measures only longitudinal motion, S-waves were recorded after the LUS scans using conventional transducers inside a vessel pressurised with hydraulic fluid. Two custom-built 25 mm diameter transducers with a central frequency of 1 MHz were coupled to opposite ends of the rock with shear couplant (Magnaflux® Shear Gel), and this assembly was housed inside a sealed rubber tube to prevent the pressuring fluid filling the pore spaces. A 300 V square wave activation pulse excited the transmitting transducer, and the electrical signal from the receiving transducer was analogue bandpass filtered between 1 Hz and 300 kHz using an SR560 pre-amplifier before being recorded with a Tektronix DPO 3014 digital oscilloscope. The analogue bandpass filter enhanced the visibility of the lower frequency S-wave from the higher frequency converted P-waves which arrived before the S-wave. Results were recorded on the second decrease of pressure from 16 MPa at the same pressure steps as the LUS scans.

Following the pressure cycles, the fast and slow S-waves were recorded on the bench
with 0.5 MHz V151 Olympus transducers. The fast S-wave had a particle motion aligned parallel to the plane of visible foliation, while the slow S-wave was recorded with the transducers generating a particle motion perpendicular to the foliation. These benchtop S-wave measurements provided a clean reference for picking the S-wave arrivals in the waveforms recorded under pressure. S-waves were not recorded for the cataclasite sample.

The transducers used inside the pressure vessel contained two S-wave piezoelectric crystals oriented perpendicularly, allowing both the fast and slow polarisations to be recorded by switching wiring. However, imperfections in the transducer elements meant that both the fast and slow polarisations were generated at the same time, causing the arrival of the slow S-wave to be obscured by the fast S-wave. This prevented an independent measurement of the slow S-wave velocity from being obtained. Instead, we used the absolute velocity difference between the fast and slow S-waves from the Olympus transducers to calculate the slow S-wave velocities at each pressure step. Results from previous S-wave transducer measurements of these lithologies (Jeppson, 2017) and samples (Guerin-Marthe, 2015) have reported little to no change in the difference between the fast and slow S-wave velocities at these confining pressures.

4.4 Results

4.4.1 Schist Waveforms

In this section, we present the ultrasonic waveforms recorded on the schist sample. To begin, we compare the fast and slow LUS waveforms recorded at 1 MPa confining stress and then compare these to the fast and slow directions recorded at 16 MPa confining stress. We highlight the changes in the waveforms caused by increasing pressure at the end of subsection 4.4.1.1. In subsection 4.4.1.2, we present LUS waveforms recorded at every orientation of the sample at 1 MPa and 16 MPa. Finally, the waveforms of the S-waves recorded with transducers are presented in subsection 4.4.1.3. The step-by-step results are shown only for the schist here as it is considered representative of the other samples. However, all LUS waveforms recorded for all four samples are available online as wavefield plots (Simpson, 2019).

4.4.1.1 Fast and Slow P-Wave Directions

LUS waveforms recorded in the fast and slow P-wave directions for the schist sample at a confining pressure of 1 MPa are shown in figure 4.3. For this sample, the fast P-wave direction was parallel to the visible foliation, and the slow P-wave direction was perpendicular to the visible foliation. The $x$-axis measures the elapsed time since the source laser firing, and the $y$-axis shows the true displacement of the sample surface in nanometres detected by the LDV. To remove high-frequency noise unrelated to the
ultrasonic waves, a zero-phase bandpass filter has been applied with corner frequencies of 100 Hz and 3 MHz (this same filter was applied to all waveforms presented in this chapter). The first break of the P-wave occurs at $5.26 \pm 0.14 \mu s$ in the fast direction, while in the slow direction the P-wave arrives at $6.76 \pm 0.18 \mu s$. In the fast direction, the P-wave displays a more impulsive onset and an amplitude about twice as large as the first P-wave in the slow direction, which has a more gradual onset.

![Figure 4.3: Waveforms recorded at 1 MPa for the schist sample. The data has been bandpass filtered between 100 Hz and 3 MHz, and P-wave arrival picks are indicated by the vertical black bars. Notice the time difference in the first breaks of the fast ($90^\circ$) and slow ($180^\circ$) directions. The first break of the P-wave in the slow direction also has an amplitude about half that of the first break in the fast direction.](image)

To examine the frequency content of the waveforms, we plot the power spectral densities (PSDs) in the fast and slow directions (figure 4.4). The PSDs were calculated using the multitaper spectrum methods in the mtspec Python module (Krischer, 2016), which average many independent estimates of the PSD function with different weights (tapers) to reduce random noise and increase the confidence of the overall estimate of the frequency content (Prieto et al., 2009). Figure 4.4 shows that frequencies in the low (100-400) kilohertz range dominate the signal in both directions. However, the fast direction contains one to two orders of magnitude more power in the 750-2000 kHz range than the slow direction. This higher frequency content in the fast direction is clearly visible in the waveform plot (figure 4.3) and is responsible for the more impulsive onset of the P-wave.

For the scan of the schist sample recorded at the highest pressure (16 MPa), the difference between the first break of the P-wave in the fast and slow directions is significantly
less than that observed for the 1 MPa scan (figure 4.5). In the fast direction, the P-wave arrives at $4.93 \pm 0.14 \mu s$, while in the slow direction, the first break occurs at $5.8 \pm 0.16 \mu s$. Thus, the difference between the fast and slow P-waves at 16 MPa ($0.87 \mu s$) is 42% less than the difference at 1 MPa ($1.50 \mu s$). Additionally, the first break of the P-wave is impulsive for both the fast and slow directions, but similar to the 1 MPa waveforms, we observe that the amplitude of the slow direction first break is approximately half that of the first break in the fast direction. It is interesting to note that the maximum amplitude in the 16 MPa scan, representing a surface wave with an amplitude of $\pm 0.75 \text{ nm}$, is 25% less than the maximum amplitude of the surface wave observed in the 1 MPa scan (see Appendix A for an explanation of this effect).

Power spectral density plots of the 16 MPa waveforms (figure 4.6) show that the peak frequencies are very similar for the fast and slow directions (140-160 kHz). While the fast direction generally contains more power in the frequencies above 750 kHz compared to the slow direction, the difference in the frequency content of the two directions is noticeably less than the difference observed for the 1 MPa scan.
Figure 4.5: Waveforms recorded at 16 MPa for the schist sample. The time difference between the first breaks of the fast (90°) and slow (180°) directions is less than the difference in the 1 MPa scan (figure 4.3), as indicated by the vertical black bars.

Figure 4.6: Power spectral densities of the waveforms in figure 4.5. Shaded regions denote the 90% confidence intervals for the multitaper spectral estimates. The peak powers occur at 140 kHz and 160 kHz for the slow and fast directions, respectively. Similar to the 1 MPa scan, the fast direction contains higher frequency content overall than the slow direction, although the difference in frequency content is smaller.
To emphasise the changes in the waveforms caused by increasing confining stress, we plot the waveforms in the slow direction of the schist at all pressures (figure 4.7). We use the slow direction, as greater changes are visible in the characteristics of the waveforms than for the fast direction. As observed previously, the first break of the P-wave decreases noticeably in arrival time with increasing confining stress, indicating that the velocity of the sample in the slow direction increases with increasing pressure. Moreover, the onset of the P-wave becomes more impulsive and higher frequencies are superposed onto the lower frequency trend as pressure increases.

These frequency content differences are highlighted in figure 4.8. The power in frequencies below 750 kHz decreases monotonically with increasing pressure, while the power in frequencies above 1.3 MHz increases monotonically with increasing pressure. A transition zone exists around 750 kHz to 1.3 MHz where these two trends overlap.
Figure 4.7: LUS waveform comparisons in the slow P-wave direction of the schist sample at different pressures. Notice the clear decrease in the first break arrival time of the P-wave along with an increase in higher frequencies as pressure increases. The waveforms have been normalised by the greatest amplitude.

Figure 4.8: Power spectral density comparisons of the waveforms in the slow P-wave direction of the schist sample at different pressures. In general, scans at higher pressure have more power at frequencies above 1 MHz than scans at lower pressure, and vice versa for power at frequencies below 1 MHz.
4.4.1.2 Full Scan Wavefields

Next, we examine the waveforms recorded at all orientations of the schist. Figure 4.9 shows all waveforms plotted as a function of group angle for scans at 1 MPa and 16 MPa confining pressure for the schist sample. These wavefield plots reveal a succession of wave arrivals. The P-wave arrives first, between 5 $\mu$s and 7 $\mu$s, with the first break arrival displaying a clear sinusoidal variation with angle. As shown in the previous section, the fast and slow P-wave directions are indeed separated by 90°. These observations confirm the clear transversely isotropic nature of this rock. The sinusoidal variation in the arrival time of the P-wave is noticeably flatter for the 16 MPa scan which again indicates that the relative difference between the fast and slow directions is lower at higher pressure.

Following a series of scattered P-waves after the first break, a surface wave arrival is visible as a positive peak between 14$\mu$s and 16$\mu$s for both scans. This wave is a Rayleigh wave which propagates around the circumference of the rock. For these scans, the amplitude of the Rayleigh wave is up to five times greater than the amplitude of the P-wave since the energy of the Rayleigh wave spreads out in only two dimensions as opposed to three for the volumetric P-waves. Since the surface wave samples all orientations of the rock each time it propagates from source to receiver, the Rayleigh wave arrival is invariant with the group angle. However, the peak of the wave decreases from 16.0 ± 0.4 $\mu$s for the 1 MPa scan to 14.8 ± 0.3 $\mu$s for the 16 MPa scan, showing that the surface wave velocity does depend on pressure.

Several arrivals which are not continuous over the full azimuthal range are prominent in the wavefield plots. Since arrivals related to the cylindrical geometry of the core would be continuous over the entire angular range, these discontinuous arrivals reveal information about the heterogeneity and anisotropy of the rock sample. For example, the foliation of the rock may be acting as a waveguide, preserving large P-wave amplitudes through several reflections between edges of the sample for azimuths close to the fast direction. These multiple reflections may be responsible for the large positive amplitudes between 13-15 $\mu$s and 20-25 $\mu$s in both scans.

Full scan wavefields of all four samples at each confining stress are available online, along with animated comparisons of the wavefields at different pressures (Simpson, 2019).
Figure 4.9: Wavefields of scans recorded at 1 MPa (top) and 16 MPa (bottom) for the schist sample. Waveforms are plotted as a function of group angle, with filled regions denoting positive polarity. The amplitudes of the waveforms are normalised by the largest amplitude in the scan. The red lines show the P-wave arrival times, with thin shaded regions denoting the picking uncertainty. The peaks of the Rayleigh waves are denoted by the green lines.
4.4.1.3 S-Waves

Figure 4.10 shows the shear wave transducer waveforms for the schist sample. The waveforms recorded under pressure correspond to the fast S-wave polarisation, which has a particle motion aligned parallel to the plane of foliation. Both the fast and slow S-wave polarisations were recorded with 0.5 MHz Olympus transducers on the bench, and the S-wave arrival times on these waveforms were used to estimate the difference between the fast and slow S-wave velocities. As opposed to the broadband LUS waveforms, the S-wave waveforms do not vary significantly in appearance at different pressures. However, the arrival time of the fast S-wave decreases slightly as the confining stress increases, indicating an increase in shear wave velocity with increasing pressure.

![S-wave waveforms](image)

Figure 4.10: S-wave transducer waveforms for the schist sample. Waveforms were recorded under pressure using the 1 MHz transducers. These waveforms correspond to the fast S-wave polarisation. The waveforms at atmospheric pressure were recorded on the bench with 0.5 MHz Olympus transducers in both the fast and slow polarisations. The vertical red bars indicate the S-wave pick on all waveforms. Varying delay times between the 1 MHz and 0.5 MHz transducer elements cause converted P-waves to be misaligned.

4.4.2 Wave Velocities

The following figures display the P- and S-wave velocities for all four Alpine Fault samples at all pressures. For the in situ LUS P-wave measurements, the arrival times of the P-waves were picked automatically with the AIC algorithm described in section 3.2.2. Since
the algorithm does not provide an estimate of the picking uncertainty, the uncertainty was fixed at ±2% of the arrival time for all waveforms. This uncertainty was considered representative of the overall picking confidence considering the S/N of the waveforms. Also, using a percentage rather than an absolute time reflects the greater uncertainty of the picking at later arrival times, where less impulsive onsets in the slow directions rendered first break picking more difficult. Velocities were calculated by dividing the propagation distance (i.e. the diameter of the samples) by the arrival times. Corrections were applied to the arrival times of the hand-sculpted cataclasite sample to account for slight irregularities in the diameter.

![Graph showing P-wave velocities as a function of pressure for the fast (solid lines and circles) and slow (dashed lines and crosses) directions.](image)

Figure 4.11: P-wave velocities as a function of pressure for the fast (solid lines and circles) and slow (dashed lines and crosses) directions.

Figure 4.11 plots the fast and slow P-wave velocities as a function of pressure. As for the schist, the fast and slow P-wave directions of the protomylonite and mylonite were aligned parallel and perpendicular to the direction of visible foliation, respectively. For the schist, protomylonite, and mylonite, the fast direction P-wave velocities are all similar within the 4800-5800 m/s range. Within error margins, these velocities increase linearly with increasing pressure by 300-400 m/s. More variation between these samples is observed in the slow direction velocities, both in terms of the absolute values and the rate at which the velocity increases with pressure. The schist varies the least, from 3800±100 m/s at 1 MPa to 4430±120 m/s at 16 MPa, while the protomylonite varies the most, from 3640±90 m/s at 1 MPa to 4800±130 m/s at 16 MPa. Unlike the fast direction velocities, the slow direction velocities do not increase linearly for all pressures. Instead,
the increase follows a logarithmic trend from lower pressures until 8-10 MPa, at which point the increase appears to become approximately linear for the schist, protomylonite, and mylonite.

The cataclasite P-wave velocities are noticeably different from the other three samples. Aside from the mylonite slow direction, both the fast and slow velocities are lower than the velocities of the schist and mylonites, ranging from 2200-3800 m/s. This P-wave velocity range agrees closely with those reported by Carpenter et al. (2014) and Jeppson (2017) for DFDP-1A cataclasites. Moreover, the fast and slow velocities each increase by approximately the same amount for each pressure step, following a logarithmic trend with pressure.

In figure 4.12, we plot all the P-wave velocities obtained from the in situ LUS scans for all four samples, as functions of both angle and pressure. For the schist, protomylonite, and mylonite, we observe an elliptical variation of the P-wave velocity with angle, representing the TI nature of these rocks. The schist displays the most regular variation, with clearly defined fast (90°) and slow (180°) directions. The protomylonite and mylonite also have fast and slow directions at approximately 90° and 180°, although the slowest and fastest P-waves do not always occur at exactly the same angle for every pressure. Of these three samples, the mylonite displays the strongest change in velocity with angle. As observed previously, greater variation is present between the velocities at different pressures in the slow direction compared to the fast direction. For all three protolith samples, the fast P-wave direction (90°) is aligned with the plane of visible foliation.

Again, the cataclasite velocities do not follow the same trends as the other samples. Variation in velocity with angle is irregular, with two fast directions displaying similar velocities at 90° and 150°, and two slow directions with similar velocities at 130° and 200°. None of these orientations align with the plane of weak foliation visible in the sample, which is aligned approximately 110°-290°. These observations suggest that the cataclasite is not accurately described as a TI rock, but exhibits a more complex dependence of velocity on angle. Moreover, as observed for the cataclasite fast and slow directions, the velocity increases by the same amount for all angles between consecutive pressure steps.

Fast and slow S-wave velocities determined from the transducer experiments are plotted as a function of pressure in figure 4.13. S-wave arrival times were picked by first identifying the S-wave in the 16 MPa scan using the Olympus transducer reference. Once the arrival time was selected, the dynamic time warping algorithm matched all points in the 16 MPa waveform to those in the 14 MPa waveform, thus identifying the S-wave arrival in that waveform. This process was continued until the lowest pressure scan to consistently pick all the S-waves. The uncertainty in all S-wave picks is ±2.5%, or approximately ±0.2 µs.

For the fast S-wave, the velocities increase approximately linearly with pressure. The fast S-wave velocity of the schist, protomylonite, and mylonite increases by 75-100 m/s
between 1 MPa and 16 MPa, or 3% on average. This is less than the fast P-wave which increased by 7% on average for these samples. The slow S-wave velocities were estimated by subtracting the velocity difference between the fast and slow polarisations measured
Figure 4.13: S-wave velocities as a function of pressure for the fast (solid lines and circles) and slow (dashed lines and crosses) polarisations. Fast S-wave velocities were calculated from the transducer waveforms recorded under pressure. The slow S-wave velocities were determined by subtracting from the fast S-waves the velocity difference between the fast and slow polarisations measured on the bench.

with the Olympus transducers from the fast S-waves measured under pressure. For the schist, protomylonite, and mylonite, these velocity differences were 430 m/s, 250 m/s, and 600 m/s, respectively, giving slow S-wave velocities that were 80-90% of the fast S-waves. These observations agree with the differences previously reported by Guerin-Marthe (2015) and Jeppson (2017), who both observe slow S-wave velocities that are between 75% and 95% of the fast S-waves for these rocks.

4.4.3 Elasticities

As described in section 2.1, the elasticity of TI rocks is fully characterised by five independent elastic constants. Four of these constants ($c_{11}$, $c_{33}$, $c_{55}$, and $c_{66}$) are calculated with the simple expressions given in equation (2.11). We used the fast and slow P- and S-wave velocities presented in the previous section to calculate these elastic constants for all four Alpine Fault samples at each pressure step. To reliably determine the fifth elastic constant $c_{13}$, we performed a nonlinear least-squares inversion using the dense array of P-wave velocities measured around the samples under confining stress. Since the inversion assumed the rocks were transversely isotropic, it was not performed for the cataclasite sample, as the variation of the P-wave velocity with angle did not display TI symmetry for this sample.
Nonlinear inversion was performed by first specifying an initial guess for \( c_{13} \). An initial value which worked well for these samples was 1.2 times the minimum constraint on \( c_{13} \) deduced by Yan et al. (2016) for hydrocarbon source rocks \( (c_{13,\text{min}} = \sqrt{c_{33}(c_{11} - 2c_{66}) + c_{66}^2 - c_{66}}) \). Using the other four elastic constants with this initial guess and equations (2.5), (2.13), and (2.14), the theoretical P-wave group velocity as a function of angle was calculated and compared to the measured group velocities. The orthogonal distance regression library in the Python SciPy module was used to execute the inversion, determining the cost of the fit for the given \( c_{13} \) value and iteratively refining the estimate until the model fitted the measured data within the noise threshold. Improved results were obtained by allowing \( c_{11} \) and \( c_{33} \) to vary within their measured uncertainties in the inversion process, along with the exact angle of the TI symmetry axis. Moreover, the best estimate for \( c_{13} \) was not permitted to exceed the theoretical maximum of \( c_{13,\text{max}} = \sqrt{c_{11}c_{33}} \) (Tsvankin, 2012). Variability between the measured velocities and the best fit models determined the standard error for the best fit \( c_{13} \) values.

Figure 4.14 shows the results of the \( c_{13} \) inversion for the schist sample at confining stresses of 1 MPa and 16 MPa. The best estimate \( c_{13} \) values corresponding to the best fit curves are 8.9 \( \pm \) 1.3 GPa and 15.5 \( \pm \) 0.7 GPa, respectively, an overall increase of 6.6 GPa. For the most part, the best fit curves match the measured velocities, rarely lying outside the uncertainty of the velocity values.

All fast and slow P- and S-wave velocities, along with the calculated elastic constants for the four Alpine Fault samples at each of the nine pressure steps, are reported in Appendix B. For reference, we also report the three dimensionless Thomsen parameters for the TI samples. Figure 4.15 summarises the elastic constants for each sample at the highest and lowest confining pressures. This plot shows that every elastic constant increases with pressure. On average, \( c_{11} \) increases by 41\% and \( c_{33} \) by 86\%, while \( c_{55} \) and \( c_{66} \) increase substantially less (7.3\% and 6.3\%, on average). \( c_{13} \) increases the most on average (150\%) between 1 MPa and 16 MPa, although the uncertainties in the estimates of \( c_{13} \) at 1 MPa are large. While \( c_{13} \) did not increase monotonically for each sample, the best estimates follow an increasing trend overall while the uncertainty in the estimates generally decreases with increasing pressure (see Tables B.1 through B.4).
Figure 4.14: Measured P-wave group velocities for the schist sample as a function of group angle (crosses) with the best fit curves at 1 MPa and 16 MPa. The best fit curves were found using the inversion process described in the text. The best estimate of $c_{13}$ for these scans is the value which was used in the equation of the best fit line. Shaded regions denote the errors in the measured velocity estimates.
Figure 4.15: Comparisons of the five independent elastic constants for all samples, calculated for 1 MPa and 16 MPa.
Chapter 5

Discussion

We have designed and developed a new methodology to perform entirely non-contact rock physics measurements under in situ conditions and applied this methodology to study anisotropic rocks. In this chapter, we discuss the advantages of this methodology and suggest improvements that would expand the suite of rock physics data we can acquire. Moreover, we interpret the data recorded on the Alpine Fault samples, showing that the high-pressure LUS methodology leads to valuable geophysical interpretations.

5.1 Evaluation of the High-Pressure LUS Design and Implementation

The high-pressure non-contact LUS methodology offers many advantages over traditional transducer techniques for acquiring laboratory rock physics data. Waveforms can be recorded at many closely spaced points around a single sample due to the small footprint of the receiver laser, and the absence of mechanical coupling allows broadband ultrasonic waveforms to be recorded without ringing from transducer elements. Moreover, data acquisition is fully automated. By making use of these advantages, we did not have to assume the orientation of anisotropy in our samples, and we could characterise in situ rock anisotropy more accurately by fitting curves to densely spaced estimates of the group velocity around a sample. Additionally, the dependence of the frequency content on direction and pressure could be observed. We discuss these advantages in more detail and identify how the methodology could be improved to expand the range of non-contact rock physics measurements that can be acquired under in situ conditions.

The high spatial density of LUS measurements allowed us to experimentally determine the orientation of the fast and slow directions of the TI rocks without assuming this beforehand. Thus, we could obtain the most accurate estimates of $c_{11}$, $c_{33}$, and $c_{13}$. This is particularly important for rocks where the fast and slow directions may not align with the visible layering, or where visible layering is difficult to identify when the sample is cored.
Moreover, the presence of TI symmetry could be tested rather than assumed. Aligned fractures that are not parallel to the bedding or foliation can cause a rock to exhibit orthorhombic or monoclinic symmetry (Tsvankin, 2012), or randomly oriented fractures may mean a rock does not exhibit any regular anisotropic symmetry. This advantage was highlighted for the case of the cataclasite, as although some foliation was visible, our measurements revealed that transverse isotropy was not a valid assumption for this rock (figure 4.12).

However, while the rotational scans do not assume the orientation of the fast and slow directions, we do assume that the TI symmetry plane is aligned with the long axis of the sample cylinders. The photograph of the schist sample (figure 4.2) suggests that the plane of foliation may not in fact be aligned with the cylinder axis which would lead to an overestimate in all velocities except in the fast direction. The possibility exists to use a methodology similar to Sarout et al. (2015) who record ray paths with transducers through a cylindrical TI sample in more than one plane. While they assume vertical TI, the inversion for the elastic constants could account for tilted TI where the orientation of the layering was clearly tilted or uncertain. To achieve this within the pressure vessel, measurements could be recorded at different vertical positions by using an automated translational stage, or by adjusting the vertical alignment of the laser beams.

Of the five independent elastic constants describing TI media, $c_{13}$ is the most sensitive to measurement error. Yan et al. (2012) showed that if $c_{13}$ is calculated using only one off-axis P-wave velocity estimate at 45° to the symmetry axis, a 1% error in that velocity can translate into a 40% error in the estimate of $c_{13}$. This is because, for only a slight variation in the 45° estimate of the P-wave velocity, the shape of the function describing the variation of group velocity with angle can vary a large amount. In figure 5.1, we compare the $c_{13}$ values of the schist calculated using only one off-axis P-wave velocity at 45° with the $c_{13}$ estimates determined using the curve fitting method. The $c_{13}$ estimates from the curve fitting method are considered more accurate because they show a regular increase with pressure, unlike the $V_P(45^\circ)$ estimates which show an overall decrease with pressure and greater variation. Moreover, the average standard error in $c_{13}$ values from curve fitting is 8% while the average uncertainty in the $V_P(45^\circ)$ estimates is 100%, showing that the curve fitting method can constrain the estimates of $c_{13}$ much more precisely. Thus, by acquiring many velocity estimates between the fast and slow directions and fitting a curve, we significantly reduce the uncertainty and increase the accuracy in the estimates of $c_{13}$. While several studies have demonstrated similar improvements in accuracy and precision by taking several off-axis measurements (see section 2.2), we improve on most of these methods by demonstrating improved $c_{13}$ estimates under confining pressure.

We have so far discussed the clear advantage of the small LUS beam footprint—namely, the ability to obtain many estimates of P-wave velocity around a single sample which allows us to determine the TI symmetry axis orientation experimentally and obtain
Figure 5.1: Comparison of $c_{13}$ estimates calculated using only one off-axis P-wave velocity at 45° ($V_P(45°)$ method) with those estimated by fitting a curve to all 91 off-axis P-wave velocities for the schist sample. The $V_P(45°)$ estimates are calculated using the equation $c_{13} = \sqrt{c_{11} + c_{55} - 2\rho V_P^2(45°)[c_{33} + c_{55} - 2\rho V_P^2(45°)] - c_{55}}$ (Sarout et al., 2015) and the uncertainty is found by propagating errors through the calculation. The curve fitting estimates are those obtained using the method described in section 4.4.3, and the errorbars represent the standard error in $c_{13}$ derived from the fit.

accurate estimates of $c_{13}$. However, the scale of mineral banding and heterogeneities in relation to the small receiver footprint and probing wavelength is an important consideration for LUS rock physics measurements. Adam et al. (2016) showed that the measured P-wave anisotropy on an Alpine Fault protomylonite using LUS at atmospheric pressure varied by 39% between two different scan locations on the same 25 mm core. The small footprint of the LUS receiver beam meant they could resolve the elastic properties of mineral layers with thicknesses greater than the wavelength, while transducer measurements provided effective media averages. For the 100-400 kHz dominant frequencies of the protoliths, the wavelengths vary between about 7 mm in the slow direction to 5 cm in the fast direction. While these wavelengths are greater than most of the observable mineral layering, higher frequency waves were probably resolving the difference between elastic properties of $< 5$ mm mineral bands or inclusions. Thus, while the small footprint of the lasers allows for improved estimates of anisotropy and $c_{13}$ from dense sampling, the scale and resolution of the measurements should be considered if applying these values to field-scale studies.

The broadband recording spectrum of the LDV and its capability to record absolute
displacements allowed us to extract detailed information about the frequency content and amplitude of the ultrasound from the waveforms. Clear differences in the frequency content were observed as a function of pressure in the slow P-wave direction of the schist (figure 4.8), along with a 25% decrease in overall amplitude of the ultrasound between 1 MPa and 16 MPa. These accurate amplitude and frequency content estimates can be used to study rock attenuation—an important tool for understanding the subsurface, especially in relation to reservoir fluids and seismic imaging (Deng et al., 2009; Adam et al., 2009). The range of ultrasonic frequencies which our detector can record (30 kHz to 24 MHz) allows us to study rock attenuation and velocity dispersion over a much greater frequency range than is possible with transducers (e.g. Winkler and Plona, 1982; Deng et al., 2009; Adam et al., 2009).

We note that although the amplitude and spectra of the waveforms are not affected by mechanical coupling, the brass jacket decreases both the overall amplitude and power of higher frequencies in the ultrasound (figure A.2). However, this effect would not prohibit obtaining quantitative estimates of rock attenuation from the amplitude and spectral information. This is because many laboratory methods use relative spectral ratios to quantify attenuation in rocks (e.g. Johnston and Toksoz, 1980; Deng et al., 2009), dividing the amplitude spectrum recorded for the rock by that of a reference with very low attenuation, such as aluminium. Previous LUS attenuation studies have not been able to employ such methods because the characteristics of the laser ultrasonic source varies between the reference and the rock (Blum et al., 2013). Since our ultrasonic source is contained within the brass, we avoid this variability and open up the possibility for broadband LUS spectral ratio studies of rock attenuation.

To expand the range of rock physics measurements that can make use of the advantages of high-pressure LUS, we intend to make several improvements to the methodology. The first of these is to increase the amplitude of the P-wave first break. Like previous studies of rock anisotropy using LUS (Blum et al., 2013; Xie et al., 2018), we encountered low S/N in the first break of some traces as a result of high attenuation and low efficiency in the conversion of laser energy to P-wave energy via thermoelastic expansion. This effect diminished as the P-wave onsets became more impulsive as confining stress increased, particularly in the slow direction, but increasing the amplitude of the first break in LUS waveforms would further increase the accuracy of automatic arrival time picking, leading to reduced uncertainty in the values of the elastic parameters. Xie et al. (2018) report significantly improved P-wave amplitudes in shale samples when the surface was modified with various films or substances to increase the thermoelastic conversion efficiency. For high-pressure LUS, this could be achieved by selecting a jacket material which more efficiently converts the infrared laser pulse to elastic energy, such as steel (Scruby and Drain, 1990).
Several other improvements are related to truly recreating the in situ environment and investigating the effects of a range of parameters on rock elasticity beyond confining stress alone. These include controlling the pore fluid, pore fluid pressure, and temperature, as well as increasing the maximum confining pressure to recreate a greater range of crustal environments. These improvements are either ready to test (in the case of temperature), or are expected to require minimal upgrades to the experimental apparatus.

Developing a capability to record shear waves is a future improvement that addresses one of the more significant limitations of the current high-pressure LUS apparatus. If particle motion parallel to the surface could be recorded with the laser Doppler vibrometer or a similar sensor, then the entire elastic stiffness tensor could be determined with non-contact measurements under in situ conditions, eliminating the need for shear wave transducer measurements. Shear waves have previously been recorded with LDVs: Nishizawa et al. (1998) use a single LDV aimed at three different orientations to retrieve the shear wave motion, while Lebedev et al. (2011) and Carson and Lebedev (2014) use a similar method to record three-dimensional particle motion and polarisation in rocks. Blum et al. (2013) measure both the perpendicular component and one off-normal component with a single LDV beam. The challenges to employing these methods for high-pressure LUS are the limited space inside the pressure vessel and the small field of view through the window which prevents viewing the sample from a significant off-normal angle. These challenges could be overcome by using mirrors inside the pressure vessel or an optical sensor that can be positioned close to the window.

5.2 P-Wave Anisotropy of Alpine Fault Rocks at Shallow Crustal Pressures

Laboratory studies of rock anisotropy are performed in order to understand how the rock physical properties control the directional dependence of wave speed, and how this directional dependence changes with depth in the crust. Quantitative estimates of anisotropy are important for developing accurate velocity models that are used for locating earthquakes and processing seismic reflection and refraction data for imaging the subsurface. Much work has been done by the hydrocarbon industry to account for shale anisotropy in active source seismic surveys for imaging sedimentary basins (e.g. Alkhalifah and Tsvankin, 1995; Tsvankin, 2012). However, seismic anisotropy is rarely included in crustal-scale velocity models even though some regions of the crust are significantly anisotropic (Godfrey et al., 2000).

Godfrey et al. (2002) investigated the effect of anisotropy on imaging subsurface structures within and around the Haast schist, a prominent terrane in the South Island of New Zealand to the east of the Alpine Fault. Using full-wavefield modelling, they found that
the significant seismic anisotropy of these rocks (up to 20%) and variable orientation of the foliation within the terrane from horizontal to near-vertical significantly influenced the depth estimates of subsurface reflectors. If P-wave anisotropy was not accounted for, errors of 10-15% in reflector depths within and beneath a 10 km thick layer were observed. This demonstrates that the anisotropy of metamorphic fault rocks, which often exhibit greater anisotropy than sedimentary rocks, must be accounted for when imaging fault structures and locating earthquakes. Thus, laboratory rock physics studies that quantify wave speeds and anisotropy of Alpine Fault rocks are important to reliably process and interpret field data.

We observed strong directional dependence of wave speed in the four Alpine Fault samples we studied using the high-pressure LUS methodology. The schist, protomylonite, and mylonite displayed orthogonal fast and slow P-wave directions, and the overall sinusoidal variation of P-wave velocity with angle was indicative of TI symmetry (figure 4.12). Since the fast P-wave direction was aligned with the plane of foliation for these samples, we infer that the anisotropy symmetry is controlled by a single preferred alignment of minerals and microcracks. Structures related to lineation within Alpine Fault protoliths could potentially break the TI symmetry if they are not aligned with the foliation (Toy et al., 2013). Since we do not observe significant variations from TI trends, we conclude that any lineation present in these samples does not significantly affect the azimuthal variation of P-wave velocity.

For the three protolith samples, the fast direction P-wave velocity increased by 7% on average between 1 MPa and 16 MPa, while the slow direction wave speed increased by 29% on average. The observed pressure dependence of wave speed at these low confining pressures is related to the closure of microcracks and stiffening of grain boundary contacts, as described in section 2.3. However, the greater sensitivity of velocity to pressure in the slow direction suggests that most of the microcracks are preferentially aligned in a plane perpendicular to the propagation of the slow P-wave. This is confirmed by micro-CT images of the mylonite (figure 5.2), which show a network of microcracks that exhibit a preferential alignment. For the protoliths, the slow P-wave direction was perpendicular to visible foliation, implying that the microcracks are preferentially aligned in a plane parallel to the foliation.

The greater sensitivity of the slow P-wave velocity to pressure can be summarised by the change in P-wave anisotropy with pressure (figure 5.3). The anisotropy of the schist and protomylonite decreases most rapidly between 1 MPa and 4 MPa before decreasing more slowly below 20%. The mylonite shows strong anisotropy that decreases rapidly until 6 MPa before steadily decreasing below 50%. Overall, these values for anisotropy are greater than those previously reported for these lithologies at similar pressures. Jeppson (2017) reported little variation (2-5%) in the P-wave anisotropy between effective pressures of 1 MPa to 60 MPa for schist, protomylonite, and mylonite samples collected from
Discussion

Figure 5.2: Synchrotron micro-CT images of the mylonite sample. The image on the left shows the network of microcracks and pores (black regions) within the minerals (grey regions). The pore spaces are rendered blue in 3D using the Avizo software at the University of Otago (right). The orientation of microcracks varies, but a preferential alignment is visible in a plane that is approximately vertical and perpendicular to the page. Images courtesy of Virginia Toy and Katrina Sauer.

Stoney Creek and nearby outcrops. Maximum anisotropies of 15-16% were observed for saturated mylonites at 60 MPa confining pressure and 1 MPa pore fluid pressure. We observe higher anisotropy for several reasons: dry samples display higher anisotropy than saturated samples; measuring the fast and slow P-waves with transducers on separate samples will always yield an anisotropy equal to or less than high-density LUS measurements; and differences in the fracture density (induced or natural) between samples influence velocities the most at lower pressures.

At high pressures (>200 MPa) where most porosity is closed, there is little difference between the velocities of Alpine Fault protoliths of different metamorphic grades due to their similar mineralogy (Christensen and Okaya, 2007). Anisotropy at these pressures is controlled by mineral banding and the high intrinsic anisotropy of micas which mostly align parallel to the foliation during shear deformation. Figure 5.4 shows the relationship between P-wave anisotropy and the volume of mica in the quartzo-feldspathic matrix of Alpine Fault rocks. This plot is derived from data presented in Dempsey et al. (2011), who modelled the static elastic anisotropy using electron backscatter diffraction data. For the 15-30% mica content in Alpine Fault rocks, we expect to see P-wave anisotropy values between 11% and 18%; this is consistent with the range of anisotropy values reported by Okaya et al. (1995) and Jeppson (2017) at pressures above 50 MPa. While our anisotropy estimates decrease towards this range, our observed anisotropy values for the schist and protomylonite lie near the upper bound of 18% at 16 MPa. This indicates that open pore spaces are still likely contributing to the velocity anisotropy at this pressure. The high anisotropy values (>50%) of the mylonite probably indicate that this rock is more fractured than the other samples.
Figure 5.3: P-wave Anisotropy as a function of pressure for the Alpine Fault samples.

Figure 5.4: P-wave velocity anisotropy as a function of mica content in the quartzofeldspathic matrix of Alpine Fault rocks. The vertical orange lines denote the range of mica content present in Alpine Fault rocks and the horizontal lines show the corresponding range of anisotropies. Figure plotted using data from Dempsey et al. (2011).
The wave speeds of the cataclasite displayed much stronger dependence on pressure than the host rocks. The slowest P-wave velocity increased by 56% between 1 MPa and 16 MPa, and the fastest P-wave velocity increased by 48%. Moreover, the velocity of the cataclasite is 36% slower in the fast direction than that of the three host rock samples on average. These observations are consistent with Jeppson (2017), who observed a 20-40% overall decrease in the P-wave velocity of rocks within the damage zone close to the fault compared with the surrounding protoliths due to an increase in brittle deformation and clay content. Our cataclasite has a mineralogy that is similar to the protoliths with additional calcite and chlorite from hydrothermal alteration. The P-wave velocity of an effective medium composed of only the forming minerals is 6100-6400 m/s (Adam et al., 2017), suggesting that the stronger pressure dependence and lower velocities of the cataclasite are caused primarily by differences in the pore structure and microcracks. CT images of the cataclasite reveal that the pore network is composed of microcracks with no obvious preferential alignment and comminuted fragments, where more rounded porosity resides (figure 5.5). The closure of the microcracks at low pressure accounts for the strong pressure dependence, while much of the rounded porosity remains open at 16 MPa, accounting for the slower overall velocity compared to a pore-free effective medium made of minerals. While the dependence of the P-wave velocity with angle does not appear random, it displays less symmetry than TI rocks (figure 4.12). This could be caused by the tilted and weakly foliated mineral bands visible in the CT images not all being parallel, or by fractures which are aligned in a different direction to the weak foliation.

![Figure 5.5: CT images of the cataclasite sample, showing two side views and a top-down view. The black regions show open microcracks and some regions of rounded porosity between comminuted grains. The light regions are mineral alteration (calcite and chlorite). Notice also the faint tilted foliation of mineral bands. The diameter of the sample is 25 mm. Images adapted from Adam et al. (2017).](image-url)

The power spectral density plots of LUS waveforms showed that the fast direction contained more power at higher frequencies than the slow direction at both 1 MPa and 16 MPa. Since higher frequencies are recorded when there is less attenuation (Mavko et al.,...
this observation suggests that the direction parallel to foliation was less attenuative than the direction perpendicular to foliation. However, the power of frequencies above 1 MHz increased by up to two orders of magnitude in the slow direction as pressure increased (figure 4.8).

While attenuation mechanisms are variable in rocks and depend on parameters such as the size of scatterers, pore geometry, and fluid content (Mavko et al., 2003; Deng et al., 2009), it is likely that for these dry rocks the decreasing attenuation in the slow direction is related to the closure of microcracks. Impedance contrasts between the matrix and pore spaces decrease as better contact between crack edges and grain boundaries is established, resulting in less energy losses due to elastic scattering. Sliding friction between microcrack and grain boundaries is also reduced with stiffer contacts, although this decrease in friction is unlikely to contribute significantly to the decrease in attenuation as strains from elastic waves are very small (Deng et al., 2009). Like for velocity anisotropy, the closure of microcracks decreases attenuation perpendicular to the foliation while the direction parallel to the foliation remains relatively unaffected. Similar results have been reported for the directional and pressure dependence of attenuation in shales, where the percentage decrease in attenuation with pressure in the slow direction is often significantly greater and more sensitive to pressure than the increase in velocity (Best et al., 2007; Deng et al., 2009; Blum et al., 2013).

We note that the pressures used in this study correspond to depths less than 2 km, representative of the cataclasite in situ pressure of approximately 2 MPa and other shallow crustal environments. At the Alpine Fault, the brittle-ductile transition is between 6-10 km (Menzies et al., 2014), so representing in situ conditions of deeper processes will require measurements up to 150 MPa. Additionally, fluid-rock interactions (Carpenter et al., 2014; Allen et al., 2017), differential stresses, high pore fluid pressure, and an anomalously large geothermal gradient (Sutherland et al., 2017) are important parameters of the subsurface environment that affect the elastic properties of rocks in and around the Alpine Fault. Thus, the proposed improvements to the high-pressure LUS apparatus suggested in the previous section are important for studying Alpine Fault rocks in a greater range of in situ conditions.
Chapter 6

Conclusions

Laboratory rock physics measurements are important for understanding how the physical properties of rocks control the behaviour of elastic waves propagating in the earth. To obtain reliable estimates of the elastic properties of rocks, measurements must be performed under in situ subsurface conditions. We have designed and implemented the first ever entirely non-contact laboratory methodology to generate and record ultrasonic waves in rocks under pressure using laser ultrasonics (LUS) inside a pressure vessel. Our methodology offers several advantages over traditional transducer rock physics measurements. Firstly, by using two lasers to generate and record elastic waves, we avoid coupling and ringing issues associated with transducers, allowing us to record the true response of the rocks to the ultrasonic source over a broader frequency range. Moreover, a single rock can be densely sampled due to the small footprint of the lasers. The small footprint also guarantees we measure P-wave group velocity and avoid the ambiguity as to whether phase or group velocity is being measured. Finally, data acquisition and arrival time picking are mostly automated.

We demonstrated the advantages of our methodology by investigating rock anisotropy. The ability to record over 90 waveforms at different angular orientations around a single cylindrical rock sample allowed us to experimentally determine the orientation of transverse isotropy and obtain accurate estimates of the fast and slow P-wave velocities. Moreover, fitting a curve to the P-wave group velocity as a function of angle significantly improved the accuracy of the estimate for $c_{13}$, which has traditionally been prone to large measurement error. For the schist sample, we observed a decrease in the average uncertainty of $c_{13}$ from 100% using the traditional three P-wave velocity method to only 8% using our curve-fitting method at nine different confining pressures. Moreover, we obtained qualitative estimates of the dependence of rock attenuation on direction and pressure by observing changes in the frequency content of the broadband LUS waveforms.

Studying the anisotropy of Alpine Fault rocks is important for understanding fault processes and building reliable velocity models for locating earthquakes and imaging subsurface structures. We found that although our four Alpine Fault rocks had similar min-
eralogy, the observed differences in velocity, anisotropy, and attenuation were mainly controlled by differences in microcrack density and alignment at these low pressures. Our measurements showed that anisotropy decreased by up to 20% in the protoliths as pressure increased from 1 MPa to 16 MPa due to the closure of microcracks which are mostly aligned with the foliation. The cataclasite had lower velocity overall than the protoliths and did not display regular TI symmetry because of a more complex pore and microcrack network resulting from brittle damage closer to the fault plane. We also observed greater attenuation in the slow direction of the protoliths compared to the fast direction due to scattering from open microcracks. As pressure increased and microcracks closed, the attenuation in the slow direction decreased. Our case study of Alpine Fault rocks demonstrated the capability of our high-pressure LUS methodology to produce high-quality data leading to valuable rock physics interpretations.

In the future, we will improve our methodology to expand the range of rock physics measurements that can be acquired under in situ conditions. These improvements include increasing the maximum confining pressure, controlling the pore fluid and pore fluid pressure, and performing measurements at high temperatures. Since the current apparatus can only measure motion perpendicular to the sample surface, we also intend to develop the capability to record shear waves. This will enable the entire elastic stiffness matrix to be estimated with non-contact measurements under in situ conditions. Our methodology also opens the door to different types of experiments, such as analysing coda waves to detect very small changes in the velocity of rocks caused by microscale healing processes (e.g. Grêt et al., 2006). Additionally, our methodology is not limited to rock physics, but could be applied for investigating the elastic properties of a range of materials under high isotropic confining stress at different temperatures, such as glacial ice inside a refrigerated vessel. Finally, we intend to investigate how the elastic properties of rocks at ultrasonic frequencies relate to the elastic properties at seismic frequencies by studying dispersion. Our non-contact measurements could be complimented with resonant ultrasound spectroscopy measurements at kilohertz frequencies recorded with a laser Doppler vibrometer inside the pressure vessel as well as low frequency strain measurements closer to seismic frequencies.

In summary, we have demonstrated that high-pressure LUS is a robust new methodology for acquiring valuable laboratory rock physics data. We intend to further upgrade the methodology and perform a range of experiments utilising non-contact measurements acquired under in situ conditions.
Appendix A

Understanding the Effects of the Sample Jacket

All rock samples measured inside the pressure vessel are encapsulated with a thin brass jacket to prevent the ingress of the nitrogen gas into the pores. This brass jacket is applied to the rock surfaces with a uniform thin layer of epoxy. Since both the generation and detection of laser ultrasound occurs on the surface of the sample, the effects of the jacket on the measurement of elastic waves need to be understood and accounted for. In this appendix, these effects are investigated with LUS scans performed on a homogeneous cylinder of aluminium both with and without the brass and epoxy jacket. The results will show how the frequency, amplitude, and arrival time of the waves are influenced by the jacket.

Absorption of laser radiation produces a thermoelastic source for ultrasonic waves. For infrared radiation, the amplitude of the electromagnetic wave decays to \( 1/e \) of its original value over a distance \( \delta = (\pi \sigma \mu_r \mu_0 \nu)^{-1/2} \) due to absorption in a metal, where \( \sigma \) is the conductivity of the metal, \( \mu_r \) is the relative permeability, \( \mu_0 \) is the permeability of free space, and \( \nu \) is the frequency of the radiation (Scruby and Drain, 1990). This quantity is known as the skin depth, and has a value of \( \delta = 8 \) nm for brass, indicating that the amplitude of the incident laser radiation decays almost entirely to zero within the 38\( \mu \)m thick brass jacket. Absorption of the radiation in the brass causes rapid and localised heating, leading to thermal expansion of the metal which produces stresses and strains that couple into the sample and propagate as ultrasonic waves. Even when accounting for the thermal conductivity of brass, it is expected that for the energy density used in this study, all heating resulting from absorption of the incident radiation is contained within the brass jacket (Scruby and Drain, 1990), and no ultrasound is generated within the sample itself.
A.1 Aim and Methods

A homogeneous aluminium cylinder with a diameter of 44.26 mm was used to experimentally test any effects of the brass and epoxy jacket on the generation and detection of laser ultrasound under pressure. The specific aims of these scans were to determine the effects of the brass jacket on the frequency content, amplitude, and arrival time, as well as to test how the jacket is affected by different confining pressures.

The aluminium was first scanned at atmospheric pressure without a jacket, then after a jacket was applied, the same scan was performed at atmospheric pressure. The jacket consisted of brass shim with a thickness of 38 µm that was cut to size and applied to the aluminium cylinder with a thin layer of two-part epoxy (Loctite® Hysol® 9455). The seams of the jacket were then sealed with more epoxy to ensure the casing was airtight. The jacket increased the diameter of the cylinder by 120 µm, implying the epoxy layer had an average thickness of 22 µm. Following the scans at atmospheric pressure, the cylinder was scanned at a number of different confining pressures between 2 MPa and 14 MPa.

Ultrasonic waveforms were recorded at 17 positions separated by 10° increments around the sample for all scans. All other experimental parameters (e.g. laser beam alignment and source power) were chosen to accurately reproduce a typical scan of a rock sample (see section 3.1).

A.2 Results

A.2.1 Scan without Jacket

Figure A.1 shows the waveforms recorded at all positions on the aluminium cylinder without the jacket. Due to the relatively long interval plotted, most of the wave train represents surface waves and scattered body waves. A visual comparison shows few differences in the phase and arrival times of the waveforms, with only small variations in amplitude between positions, confirming our aluminium cylinder is homogeneous. The amplitude variations may be caused by impurities on the surface of the aluminium affecting the power of the generated ultrasound, or small fluctuations in the power of the source laser.

The first break of the average of all waveforms occurred at a time of $7.30 \pm 0.04$ µs, giving an average ultrasonic velocity of $6060 \pm 30$ m/s. This velocity is several hundred metres per second slower than quoted values for aluminium (e.g. Ledbetter and Moulder, 1979; Scruby and Drain, 1990), indicating that our aluminium cylinder may contain some impurities which reduce the velocity. However, this difference is not significant for this investigation, since these data will be used only for comparison to the scans with the jacket. Moreover, the absolute velocity of aluminium can be considered constant over the range of confining stresses used in this test (Bergman and Shahbender, 1958).
A.2.2 Scan with Jacket

Averages of the waveforms recorded at all positions for the scans with and without the jacket at atmospheric pressure are shown in figure A.2, along with the frequency content of the ultrasound. At first glance, it is obvious that the maximum displacement in the scan with the jacket is approximately three times less than the maximum displacement of the scan without the jacket. Moreover, the waveform of the scan without the jacket clearly has a higher frequency content than the scan with the jacket, and the phases of the surface wave train after 20 $\mu$s do not align. These amplitude and frequency differences are highlighted in the power spectral density plot, which shows that the jacketed scan has less power overall and up to three orders of magnitude less power at frequencies greater than 750 kHz.

A.2.3 Scans under Confining Stress

Comparisons of the average waveforms and frequency content of the aluminium scans with the jacket at different confining stresses are shown in figure A.3. The phase of the scan at atmospheric pressure (the same scan as in the previous section) lags the other scans by a small margin, and the amplitude of the waveform is greater overall. However, for scans under pressure, the phases of the waveforms are near identical with only a slight decrease in amplitude with pressure. The power spectral density plot shows that the greater power of the atmospheric pressure scan is contained in the peak frequency around 300 kHz,
Figure A.2: Comparison of waveforms (top) and power spectral densities (bottom) of the aluminium at atmospheric pressure with and without the brass and epoxy jacket. The data for each scan is the average of the waveforms recorded at all positions. Shaded regions in the lower plot denote the 90% confidence intervals for the multitaper spectral estimates.

while significantly less power is present in frequencies above 500 kHz compared to the scans under pressure. For the scans under pressure, the power contained in the dominant frequencies below 1 MHz is nearly invariant with pressure. Interestingly, scans at higher pressure have less power in the 1-2 MHz range but more power at frequencies above 2 MHz than those at lower pressure.

The difference between the scan at atmospheric pressure and those under pressure may be due to small confining stresses optimally coupling the jacket to the sample. This would decrease attenuation at the sample-jacket interface and thus increase the frequency content of the surface waves. Also, compression of the jacket to the sample where small pockets of air are trapped will decrease the propagation distance of the surface waves, causing the
waveforms recorded under pressure to lead the waveform recorded at atmospheric pressure. The regular decrease in overall amplitude with increasing pressure for frequencies below 2 MHz may be due to the higher acoustic impedance of nitrogen at higher pressures. Between atmospheric pressure and 14 MPa, the acoustic impedance of nitrogen increases by a factor of over 130. This leads to a lower acoustic impedance contrast at the boundary of the sample, allowing energy from the surface waves or body waves scattered at the surface to more readily radiate away from the sample into the nitrogen, decreasing the wave amplitudes recorded at the receiver location.

Figure A.3: Comparison of waveforms (top) and power spectral densities (bottom) of the aluminium with the jacket at different confining stresses.

Since we are primarily concerned with the P-wave velocities when determining the elastic properties of rock samples, careful consideration is given here to the effect of the jacket on the first break. Figure A.4 shows the first 20 µs of all the averaged waveforms,
Understanding the Effects of the Sample Jacket

capturing the first break and scattered arrivals of the body waves. Similar to the longer
time period investigated above, the waveform recorded without the jacket has greater
amplitudes and leads the other waveforms, whereas the waveform at atmospheric pressure
with the jacket has the lowest amplitudes and lags the other waveforms by several tenths
of a microsecond. Again, there are few differences between the waveforms recorded under
confining stress. However, the inset in figure A.4 shows that the amplitude of the first
break increases as the confining stress increases, opposite to the overall surface wave
amplitude decrease previously observed. This amplitude increase is consistent with the
previous explanation of a lower impedance contrast between the sample and the nitrogen,
as more of the direct P-wave energy is transmitted from the sample through the surface
when the acoustic impedance contrast is lower at higher pressures.

Figure A.4: Comparison of the body wave arrivals at different confining stresses for the
aluminium with the jacket. The blue waveform shows the scan without the jacket for
reference, and the red waveform is the scan at atmospheric pressure with the jacket. The
data has been bandpass filtered between 10 kHz and 5 MHz.

For the P-wave first break, the jacket introduces a small time offset of 0.04 $\mu$s. Calculating
the time for waves to propagate through the jacket with published values for the
velocity of ultrasound in brass (Kaye and Laby, 1995) and epoxy (Rokhlin et al., 1986;
Winkler and Plona, 1982) also gives a travel time of 0.04 $\mu$s, agreeing with the observed
offset.
A.3 Conclusion

The brass and epoxy jacket applied to samples for high-pressure LUS has an effect on the frequency content, amplitude, and arrival time of the recorded ultrasound. Both the amplitude and frequency content decreased after the jacket was applied, and the phase of the waveforms did not match. Most importantly, however, no significant changes in phase or frequency were observed between scans at different confining pressures above 2 MPa, implying that comparisons of the ultrasound at different pressures can be made without influence from the jacket. The regular amplitude differences between waveforms at different pressures are probably a result of energy radiating into the nitrogen and not an effect of the jacket. Finally, the jacket introduces a delay of 0.04 µs in the P-wave arrival at all pressures which can be easily corrected for to obtain estimates of absolute P-wave velocity.
Appendix B

Tables of Results

This appendix contains tables for each of the Alpine Fault rock samples studied, giving all the measured velocities and calculated elastic parameters at each confining stress. All uncertainties are calculated with the Python uncertainties package (Lebigot, 2010–), with rounding and significant digits following the conventions of the Particle Data Group (http://pdg.lbl.gov/). Although the elastic constants ($c_{ij}$’s) are sufficient to characterise the elasticity of TI media, for reference we also report the dimensionless Thomsen parameters which are commonly used to describe the strength of the velocity anisotropy and simplify the analysis of seismic wave propagation in anisotropic media (Thomsen, 1986; Tsvankin, 2012). The three dimensionless Thomsen parameters are (Thomsen, 1986):

$$\varepsilon \equiv \frac{c_{11} - c_{33}}{2c_{33}} \quad (B.1)$$

$$\delta \equiv \frac{(c_{13} + c_{55})^2 - (c_{33} - c_{55})^2}{2c_{33} (c_{33} - c_{55})} \quad (B.2)$$

$$\gamma \equiv \frac{c_{66} - c_{55}}{2c_{55}} \quad (B.3)$$

The parameter $\varepsilon$, often called the P-wave anisotropy, describes the difference between the fast and slow P-wave velocities and is usually positive. $\gamma$ has the same role for S-waves. The parameter $\delta$ is less intuitive, but is important for describing how the P-wave phase velocity changes near the symmetry axis. A positive value indicates the velocity is increasing either side of the symmetry axis, while a negative value indicates the velocity is decreasing (Tsvankin, 2012). Since it depends on $c_{13}$, $\delta$ is traditionally the parameter most susceptible to measurement error. While our percentage errors for $\delta$ are often large due to propagation of uncertainty through equation (B.2), better accuracy could be achieved by fitting a curve parameterised with $\varepsilon$, $\delta$, and $\gamma$ to the P-wave velocity as a function of angle. Because the slow S-wave $V_{S0}$ was determined from the fast S-wave $V_{S90}$ in this study, the values of $c_{55}$ and $c_{66}$ used to calculate $\gamma$ are not independent.
Table B.1: Measured wave speeds and calculated elastic parameters for the schist sample at all pressures.

<table>
<thead>
<tr>
<th>P (MPa)</th>
<th>$V_{P0}$ (m/s)</th>
<th>$V_{P90}$ (m/s)</th>
<th>$V_{S0}$ (m/s)</th>
<th>$V_{S90}$ (m/s)</th>
<th>$c_{11}$ (GPa)</th>
<th>$c_{33}$ (GPa)</th>
<th>$c_{55}$ (GPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 MPa</td>
<td>3810 ± 100</td>
<td>4890 ± 140</td>
<td>2640 ± 60</td>
<td>3070 ± 70</td>
<td>63.7 ± 3.5</td>
<td>38.6 ± 2.0</td>
<td>18.5 ± 0.7</td>
</tr>
<tr>
<td>2 MPa</td>
<td>3940 ± 100</td>
<td>4920 ± 140</td>
<td>2640 ± 60</td>
<td>3070 ± 70</td>
<td>64 ± 4</td>
<td>41.3 ± 2.2</td>
<td>18.6 ± 0.7</td>
</tr>
<tr>
<td>4 MPa</td>
<td>4200 ± 110</td>
<td>5040 ± 140</td>
<td>2680 ± 60</td>
<td>3110 ± 80</td>
<td>68 ± 4</td>
<td>46.9 ± 2.5</td>
<td>19.1 ± 0.8</td>
</tr>
<tr>
<td>6 MPa</td>
<td>4280 ± 110</td>
<td>5130 ± 140</td>
<td>2690 ± 60</td>
<td>3120 ± 80</td>
<td>70 ± 4</td>
<td>48.8 ± 2.6</td>
<td>19.3 ± 0.8</td>
</tr>
<tr>
<td>8 MPa</td>
<td>4360 ± 120</td>
<td>5170 ± 150</td>
<td>2710 ± 60</td>
<td>3140 ± 80</td>
<td>71 ± 4</td>
<td>50.6 ± 2.7</td>
<td>19.5 ± 0.8</td>
</tr>
<tr>
<td>10 MPa</td>
<td>4380 ± 120</td>
<td>5170 ± 150</td>
<td>2720 ± 60</td>
<td>3150 ± 80</td>
<td>71 ± 4</td>
<td>51.2 ± 2.8</td>
<td>19.7 ± 0.8</td>
</tr>
<tr>
<td>12 MPa</td>
<td>4400 ± 120</td>
<td>5190 ± 150</td>
<td>2730 ± 60</td>
<td>3160 ± 80</td>
<td>72 ± 4</td>
<td>51.7 ± 2.8</td>
<td>19.8 ± 0.8</td>
</tr>
<tr>
<td>14 MPa</td>
<td>4400 ± 120</td>
<td>5210 ± 150</td>
<td>2730 ± 60</td>
<td>3160 ± 80</td>
<td>72 ± 4</td>
<td>51.5 ± 2.8</td>
<td>19.9 ± 0.8</td>
</tr>
<tr>
<td>16 MPa</td>
<td>4430 ± 120</td>
<td>5220 ± 150</td>
<td>2740 ± 60</td>
<td>3170 ± 80</td>
<td>73 ± 4</td>
<td>52.4 ± 2.8</td>
<td>20.0 ± 0.8</td>
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Table B.1: Continued

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<th>P (MPa)</th>
<th>$c_{66}$ (GPa)</th>
<th>$c_{13}$ (GPa)</th>
<th>Anisotropy (%)</th>
<th>$\varepsilon$</th>
<th>$\gamma$</th>
<th>$\delta$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 MPa</td>
<td>25.1 ± 1.2</td>
<td>8.9 ± 1.3</td>
<td>25 ± 4</td>
<td>0.33 ± 0.06</td>
<td>0.18 ± 0.04</td>
<td>0.06 ± 0.10</td>
</tr>
<tr>
<td>2 MPa</td>
<td>25.2 ± 1.2</td>
<td>12.6 ± 1.6</td>
<td>22 ± 4</td>
<td>0.28 ± 0.06</td>
<td>0.18 ± 0.04</td>
<td>0.12 ± 0.11</td>
</tr>
<tr>
<td>4 MPa</td>
<td>25.7 ± 1.2</td>
<td>11.0 ± 1.2</td>
<td>18 ± 4</td>
<td>0.22 ± 0.06</td>
<td>0.17 ± 0.04</td>
<td>−0.07 ± 0.08</td>
</tr>
<tr>
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<td>25.9 ± 1.2</td>
<td>13.5 ± 1.1</td>
<td>18 ± 4</td>
<td>0.22 ± 0.06</td>
<td>0.17 ± 0.04</td>
<td>−0.05 ± 0.08</td>
</tr>
<tr>
<td>8 MPa</td>
<td>26.2 ± 1.2</td>
<td>15.6 ± 1.2</td>
<td>17 ± 4</td>
<td>0.20 ± 0.05</td>
<td>0.17 ± 0.04</td>
<td>−0.02 ± 0.09</td>
</tr>
<tr>
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<td>26.4 ± 1.3</td>
<td>16.2 ± 1.1</td>
<td>16 ± 4</td>
<td>0.19 ± 0.05</td>
<td>0.17 ± 0.04</td>
<td>−0.00 ± 0.09</td>
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<td>15.8 ± 0.8</td>
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<td>0.19 ± 0.05</td>
<td>0.17 ± 0.04</td>
<td>−0.02 ± 0.08</td>
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<td>17.1 ± 0.8</td>
<td>17 ± 4</td>
<td>0.20 ± 0.05</td>
<td>0.17 ± 0.04</td>
<td>0.02 ± 0.09</td>
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<tr>
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<td>26.8 ± 1.3</td>
<td>15.5 ± 0.7</td>
<td>16 ± 4</td>
<td>0.19 ± 0.05</td>
<td>0.17 ± 0.04</td>
<td>−0.04 ± 0.08</td>
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</table>
Table B.2: Measured wave speeds and calculated elastic parameters for the protomylonite sample at all pressures.

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<th>V\textsubscript{P0} (m/s)</th>
<th>V\textsubscript{P90} (m/s)</th>
<th>V\textsubscript{S0} (m/s)</th>
<th>V\textsubscript{S90} (m/s)</th>
<th>c\textsubscript{11} (GPa)</th>
<th>c\textsubscript{33} (GPa)</th>
<th>c\textsubscript{55} (GPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 MPa</td>
<td>3640 ± 90</td>
<td>5190 ± 150</td>
<td>3080 ± 80</td>
<td>3330 ± 90</td>
<td>72 ± 4</td>
<td>35.4 ± 1.8</td>
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<tr>
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<td>3730 ± 100</td>
<td>5220 ± 150</td>
<td>3090 ± 80</td>
<td>3340 ± 90</td>
<td>73 ± 4</td>
<td>37.3 ± 2.0</td>
</tr>
<tr>
<td>4 MPa</td>
<td>4360 ± 120</td>
<td>5340 ± 150</td>
<td>3130 ± 80</td>
<td>3380 ± 100</td>
<td>76 ± 4</td>
<td>51.0 ± 2.8</td>
</tr>
<tr>
<td>6 MPa</td>
<td>4480 ± 120</td>
<td>5390 ± 150</td>
<td>3140 ± 80</td>
<td>3390 ± 100</td>
<td>78 ± 4</td>
<td>53.7 ± 2.9</td>
</tr>
<tr>
<td>8 MPa</td>
<td>4620 ± 130</td>
<td>5460 ± 160</td>
<td>3130 ± 80</td>
<td>3380 ± 100</td>
<td>80 ± 5</td>
<td>57.2 ± 3.2</td>
</tr>
<tr>
<td>10 MPa</td>
<td>4740 ± 130</td>
<td>5510 ± 160</td>
<td>3140 ± 80</td>
<td>3390 ± 100</td>
<td>81 ± 5</td>
<td>60.2 ± 3.3</td>
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<tr>
<td>12 MPa</td>
<td>4760 ± 130</td>
<td>5570 ± 160</td>
<td>3140 ± 80</td>
<td>3390 ± 100</td>
<td>83 ± 5</td>
<td>60.7 ± 3.4</td>
</tr>
<tr>
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<td>4790 ± 130</td>
<td>5660 ± 160</td>
<td>3150 ± 80</td>
<td>3400 ± 100</td>
<td>86 ± 5</td>
<td>61.3 ± 3.4</td>
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<tr>
<td>16 MPa</td>
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<td>5660 ± 160</td>
<td>3150 ± 80</td>
<td>3400 ± 100</td>
<td>86 ± 5</td>
<td>61.6 ± 3.4</td>
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Table B.2: *Continued*

<table>
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<tr>
<th>c\textsubscript{66} (GPa)</th>
<th>c\textsubscript{13} (GPa)</th>
<th>Anisotropy (%)</th>
<th>ε</th>
<th>γ</th>
<th>δ</th>
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</thead>
<tbody>
<tr>
<td>1 MPa</td>
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<td>5 ± 4</td>
<td>35 ± 4</td>
<td>0.52 ± 0.08</td>
<td>0.08 ± 0.04</td>
</tr>
<tr>
<td>2 MPa</td>
<td>29.9 ± 1.6</td>
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<td>33 ± 4</td>
<td>0.48 ± 0.07</td>
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<td>4 MPa</td>
<td>30.5 ± 1.6</td>
<td>8.4 ± 1.7</td>
<td>20 ± 4</td>
<td>0.25 ± 0.06</td>
<td>0.08 ± 0.04</td>
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<tr>
<td>6 MPa</td>
<td>30.7 ± 1.7</td>
<td>11.5 ± 1.7</td>
<td>18 ± 4</td>
<td>0.22 ± 0.06</td>
<td>0.08 ± 0.04</td>
</tr>
<tr>
<td>8 MPa</td>
<td>30.6 ± 1.7</td>
<td>13.9 ± 1.5</td>
<td>17 ± 4</td>
<td>0.20 ± 0.05</td>
<td>0.08 ± 0.04</td>
</tr>
<tr>
<td>10 MPa</td>
<td>30.7 ± 1.7</td>
<td>15.1 ± 1.3</td>
<td>15 ± 4</td>
<td>0.18 ± 0.05</td>
<td>0.08 ± 0.04</td>
</tr>
<tr>
<td>12 MPa</td>
<td>30.7 ± 1.7</td>
<td>16.3 ± 1.3</td>
<td>16 ± 4</td>
<td>0.19 ± 0.05</td>
<td>0.08 ± 0.04</td>
</tr>
<tr>
<td>14 MPa</td>
<td>30.9 ± 1.7</td>
<td>17.6 ± 1.6</td>
<td>17 ± 4</td>
<td>0.20 ± 0.06</td>
<td>0.08 ± 0.04</td>
</tr>
<tr>
<td>16 MPa</td>
<td>31.0 ± 1.7</td>
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<td>17 ± 4</td>
<td>0.20 ± 0.06</td>
<td>0.08 ± 0.04</td>
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Table B.3: Measured wave speeds and calculated elastic parameters for the mylonite sample at all pressures.

<table>
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<tr>
<th>$P_0$ (MPa)</th>
<th>$V_{P0}$ (m/s)</th>
<th>$V_{P90}$ (m/s)</th>
<th>$V_{S0}$ (m/s)</th>
<th>$V_{S90}$ (m/s)</th>
<th>$c_{11}$ (GPa)</th>
<th>$c_{33}$ (GPa)</th>
<th>$c_{55}$ (GPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 MPa</td>
<td>2600 ± 60</td>
<td>5160 ± 150</td>
<td>2820 ± 40</td>
<td>3420 ± 60</td>
<td>73 ± 4</td>
<td>18.6 ± 0.9</td>
<td>21.8 ± 0.6</td>
</tr>
<tr>
<td>2 MPa</td>
<td>2700 ± 70</td>
<td>5210 ± 150</td>
<td>2820 ± 40</td>
<td>3420 ± 60</td>
<td>75 ± 4</td>
<td>20.0 ± 1.0</td>
<td>21.9 ± 0.6</td>
</tr>
<tr>
<td>4 MPa</td>
<td>2900 ± 70</td>
<td>5250 ± 150</td>
<td>2840 ± 40</td>
<td>3440 ± 60</td>
<td>76 ± 4</td>
<td>23.1 ± 1.1</td>
<td>22.1 ± 0.6</td>
</tr>
<tr>
<td>6 MPa</td>
<td>3180 ± 80</td>
<td>5320 ± 150</td>
<td>2860 ± 40</td>
<td>3460 ± 60</td>
<td>78 ± 4</td>
<td>27.8 ± 1.4</td>
<td>22.5 ± 0.6</td>
</tr>
<tr>
<td>8 MPa</td>
<td>3210 ± 80</td>
<td>5340 ± 150</td>
<td>2870 ± 40</td>
<td>3470 ± 60</td>
<td>78 ± 4</td>
<td>28.3 ± 1.4</td>
<td>22.7 ± 0.6</td>
</tr>
<tr>
<td>10 MPa</td>
<td>3280 ± 80</td>
<td>5360 ± 150</td>
<td>2910 ± 40</td>
<td>3510 ± 60</td>
<td>79 ± 4</td>
<td>29.6 ± 1.5</td>
<td>23.2 ± 0.7</td>
</tr>
<tr>
<td>12 MPa</td>
<td>3340 ± 80</td>
<td>5420 ± 150</td>
<td>2930 ± 40</td>
<td>3530 ± 60</td>
<td>81 ± 5</td>
<td>30.6 ± 1.5</td>
<td>23.7 ± 0.7</td>
</tr>
<tr>
<td>14 MPa</td>
<td>3380 ± 90</td>
<td>5430 ± 160</td>
<td>2930 ± 40</td>
<td>3530 ± 60</td>
<td>81 ± 5</td>
<td>31.3 ± 1.6</td>
<td>23.6 ± 0.7</td>
</tr>
<tr>
<td>16 MPa</td>
<td>3600 ± 90</td>
<td>5450 ± 160</td>
<td>2940 ± 40</td>
<td>3540 ± 60</td>
<td>82 ± 5</td>
<td>35.7 ± 1.8</td>
<td>23.7 ± 0.7</td>
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Table B.3: Continued

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<tr>
<th>$c_{66}$ (GPa)</th>
<th>$c_{13}$ (GPa)</th>
<th>Anisotropy (%)</th>
<th>$\varepsilon$</th>
<th>$\gamma$</th>
<th>$\delta$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 MPa</td>
<td>32.1 ± 1.1</td>
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<td>65.9 ± 3.3</td>
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</tr>
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<td>2 MPa</td>
<td>32.2 ± 1.1</td>
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<td>0.235 ± 0.032</td>
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<td>4 MPa</td>
<td>32.4 ± 1.1</td>
<td>3.1 ± 3.3</td>
<td>57.7 ± 3.4</td>
<td>1.14 ± 0.12</td>
<td>0.234 ± 0.032</td>
</tr>
<tr>
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<td>32.9 ± 1.1</td>
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<td>50.2 ± 3.5</td>
<td>0.89 ± 0.11</td>
<td>0.232 ± 0.032</td>
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<td>33.1 ± 1.1</td>
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<td>0.231 ± 0.032</td>
</tr>
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<td>0.228 ± 0.032</td>
</tr>
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<td>34.3 ± 1.2</td>
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<td>0.82 ± 0.10</td>
<td>0.225 ± 0.032</td>
</tr>
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<td>47 ± 4</td>
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<td>0.226 ± 0.032</td>
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Table B.4: Measured wave speeds and calculated elastic parameters for the cataclasite sample at all pressures.

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<th>$V_{P90}$ (m/s)</th>
<th>$V_{S0}$ (m/s)</th>
<th>$V_{S90}$ (m/s)</th>
<th>$c_{11}$ (GPa)</th>
<th>$c_{33}$ (GPa)</th>
<th>$c_{55}$ (GPa)</th>
</tr>
</thead>
<tbody>
<tr>
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<td>2170 ± 60</td>
<td>2580 ± 70</td>
<td>−</td>
<td>−</td>
<td>16.2 ± 0.8</td>
<td>11.4 ± 0.5</td>
<td>−</td>
</tr>
<tr>
<td>2 MPa</td>
<td>2490 ± 70</td>
<td>2900 ± 80</td>
<td>−</td>
<td>−</td>
<td>20.5 ± 1.0</td>
<td>15.1 ± 0.7</td>
<td>−</td>
</tr>
<tr>
<td>4 MPa</td>
<td>2840 ± 80</td>
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<td>−</td>
<td>24.9 ± 1.3</td>
<td>19.6 ± 1.0</td>
<td>−</td>
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<tr>
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<td>3010 ± 80</td>
<td>3430 ± 100</td>
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<td>−</td>
<td>28.6 ± 1.5</td>
<td>22.0 ± 1.1</td>
<td>−</td>
</tr>
<tr>
<td>8 MPa</td>
<td>3160 ± 90</td>
<td>3570 ± 100</td>
<td>−</td>
<td>−</td>
<td>31.0 ± 1.6</td>
<td>24.4 ± 1.2</td>
<td>−</td>
</tr>
<tr>
<td>10 MPa</td>
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<td>3600 ± 100</td>
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<td>−</td>
<td>31.6 ± 1.6</td>
<td>26.1 ± 1.3</td>
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<td>32.8 ± 1.7</td>
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Table B.4: Continued

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<th>$c_{66}$ (GPa)</th>
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<th>Anisotropy (%)</th>
<th>$\varepsilon$</th>
<th>$\gamma$</th>
<th>$\delta$</th>
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<td>17.3 ± 3.3</td>
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<td>−</td>
</tr>
<tr>
<td>2 MPa</td>
<td>−</td>
<td>−</td>
<td>15.1 ± 3.4</td>
<td>0.18 ± 0.05</td>
<td>−</td>
<td>−</td>
</tr>
<tr>
<td>4 MPa</td>
<td>−</td>
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<td>11.9 ± 3.5</td>
<td>0.13 ± 0.04</td>
<td>−</td>
<td>−</td>
</tr>
<tr>
<td>6 MPa</td>
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<td>−</td>
<td>13 ± 4</td>
<td>0.15 ± 0.05</td>
<td>−</td>
<td>−</td>
</tr>
<tr>
<td>8 MPa</td>
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<td>0.14 ± 0.05</td>
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<td>0.12 ± 0.04</td>
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</tr>
<tr>
<td>14 MPa</td>
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<td>−</td>
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<td>0.11 ± 0.04</td>
<td>−</td>
<td>−</td>
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<tr>
<td>16 MPa</td>
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<td>−</td>
<td>12 ± 4</td>
<td>0.14 ± 0.05</td>
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