Seismic Polarization in the Near-surface

by

Matthew Griffiths

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Supervisory Committee

Dr. André Pugin, Research Scientist (Co-supervisor)
(Geological Survey of Canada)

Dr. Dariush Motazedian, Professor (Co-supervisor)
(Carleton University, Department of Earth Science)

Dr. Giorgio Ranalli, Distinguished Research Professor
(Carleton University, Department of Earth Science)

Dr. Glenn Milne, Professor
(University of Ottawa, Department of Earth and Environmental Sciences)

ABSTRACT

Seismic polarization is investigated for near-surface reflection data with the development of two processing tools. The first tool is a method of visualizing polarization states. Principal Component Analysis (PCA) has been applied to seismic data to study polarization before, but this is the first time a red-blue-green (RGB) colour map has been used to visualize the analysis output. The new colour map allows the direction of vectors to be displayed on a single plot, instead of decomposing the direction information into different components. This visualization scheme conveys rich information in a concise manner. The second tool is a method to estimate time-dependent time-shift from trace to trace in a shot or common-midpoint (CMP) gather. The method makes use of changes in the phase content of the two signals, exploiting a well known property of the Fourier Transform. To estimate the phase content with time, the Stockwell Transform is used which maps signals into the mixed time-frequency domain. This new method of estimating time-dependent time-shift is more robust to noise, but more vulnerable to aliasing compared to previous time-shift estimation methods such as Dynamic Time Warping. When this new calculation is incorporating into an existing velocity independent processing framework, zero-offset sections can be produced in a much less labour intensive manner compared to the Semblance-Normal Moveout approach. This also serves as the first example of velocity independent seismic imaging applied to near-surface reflection data. Combining these processing tools allows for the polarization to be visualized on the scale of survey, and the resulting polarization from different source orientation can be compared conveniently.

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DEDICATION

To my grandparents; Anna, Axel, Gord and Val

Chapter 1

Introduction

1.1 Seismic Reflection

Seismic reflection is useful for finding all kinds of subsurface goodies. An active seismic source (e.g. hammer, vibrator or explosive) at the surface generates seismic waves, which propagate into the earth. When these waves encounter a new layer of earth (specifically, a change in the seismic impedance; the product of the seismic velocity and density) some of the energy is reflected back towards the surface, where it can be detected by an array of receivers (e.g. geophones/accelerometers). By analyzing the time for waves to travel from the surface to various boundaries and back, a picture of the subsurface can be computed. The method has been used for geological mapping and hydrocarbon exploration since the early part of the last century [Barton (1929), referenced in Sheriff and Geldart (1995)]. Comparatively, the application to near surface targets ($< 250m$) is new (Hunter et al., 1984). Even more recent is the acquisition and analysis of multicomponent surveys, where the wave field is fully captured in 3 dimensions Pugin et al. (2009). For some geological settings it has been found that the highest quality seismic images were processed using orthogonal source-receiver orientations (Pugin et al., 2009). This indicates that the geology influences the polarization of the reflected waves. An analysis of near-surface multicomponent data is the focus of this thesis, specifically to study the polarization of reflected waves.

Fig 1.1 shows the different types of waves that are useful for creating seismic reflection images. These are classified by the polarization (the particle motion) associated with each

Figure 1.1: **Seismic Reflection Schematic**

Seismic Reflection Schematic. **a)** P-wave incident ray **b)** S_v -wave incident ray **c)** S_h -wave incident ray. Dashed closed arrows indicate seismic ray paths. Open double arrows indicate the ground motion direction (polarization).

wave type; i.e. whether the particles oscillate parallel or perpendicular to the direction of wave propagation. 3 component (3C) geophones capture the ground motion in 3 dimensions, allowing for polarization analysis. Seismic processing schemes often assume horizontal, homogeneous and isotropic layers. Here only waves predicted by classic theory are shown (Zoeppritz, 1919). **Fig 1.1a** shows the reflected P-wave and converted $P-S$ -wave resulting from an incident P -wave. All ground motion takes place within the plane of the survey. **Fig 1.1b** shows the reflected S_v and converted $S-P$ wave from an incident S_v wave. **Fig 1.1c** shows the ray path of a reflected S_h wave. No conversions are predicted for an incident S_h wave, as the ground motion is purely transverse to the survey plane; the motion is orthogonal to the motion of P & S_v waves.

The Geological Survey of Canada has acquired over 800 kilometers of near-surface seismic data using an in-house developed landstreamer system (Pugin et al., 2009). This system has been used in a variety of environments (Pugin et al., 2013a,b), for applications such as regional hydrostratigraphy and earthquake hazard assessment. Already 3C surveys have proven utility where single component data would have failed (Pugin et al., 2009). The 3C acquisition also allows for sophisticated analysis of the wavefield. The purpose of this research is to convey and explore the richness of these data sets. The fact that there is a vast volume of data further constrains the problem. The analysis applied and developed in this thesis should be practically reproducible on other data sets. This is done by automating processing steps when possible.

1.2 Polarization of Seismic Waves

Polarization refers to the oscillation character (degree of linearity and orientation) of ground motion. Along with seismic velocity, polarization is a characteristic that differentiates seismic wave types. The two kinds of body waves are

- P Waves (Compressional)
- S Waves (Shear)

P waves are the fastest seismic waves and their particle motion is parallel to the direction

Figure 1.2: **Particle Motion Illustration.**

a) P-wave particle motion. **b)** S-wave particle motion. **c)** Surface wave particle motion: top - Rayleigh wave particle motion. bottom - Love wave particle motion. From Everett (2013)

of wave propagation. The polarization is linear and longitudinal (see **Fig 1.2a**).

Shear waves travel slower and consequently arrive later than P-waves. The particle motion is perpendicular to the direction of wave propagation see **Fig 1.2b**. S-waves are often further categorized by the orientation of the ground shaking relative to the surface and the plane of incidence; S_v being in the plane of incidence and S_h being perpendicular to it and parallel with the surface. Because of the slower velocity, shear waves have smaller wavelengths compared to P waves of the same frequency. As a result the resolution of S-waves is higher, a fact that has been applied by Pugin et al. (2009).

These are all linear polarization states. The longitudinal and transverse nature of P and S waves (respectively) are predicted in isotropic media. The polarization of body waves in anisotropic media may have slightly different properties; the fastest wave may not have exactly longitudinal polarization, nor the slowest waves have exactly transverse polarization, though these quasi-P and quasi-S waves are orthogonal to each other in theory (Landau

and Lifshitz, 1959).

1.3 Problem

The initial challenge in understanding the polarization of seismic waves is to characterize the variation of 3D ground motion in space and time. Considering a 2D seismic survey, the data dimensions are source location, receiver locations and two way travel time. For multicomponent surveys there can be a large data volume ($48 \text{ geophones} \times 3 \text{ components} \times \# \text{ shots} \times \# \text{ source orientations}$). For single component surveys the creation of a zero-offset section has been the preferred approach to analyse and reduce the data volume. The creation of the zero-offset set is also an important piece in the polarization analysis, but because 3 zero-offset sections must be made, a more automated approach compared to the conventional semblance-nmo processing scheme was sought out and developed. While it is an aside to the initial thesis objective, the development of this technique has become a key part of this research. The product of this research is the implementation and development of two multi-component processing tools.

1. A tool to analyze and visualize seismic polarization.
2. A method to obtain zero-offset sections automatically.

1.4 PCA based Polarization

Principal Component Analysis (PCA) computes the axis that accounts for the majority of the variance in a multi-dimensional signal; so if the ground is shaking purely in the vertical direction, the first principal component (PC1) will align with the vertical axis. **Fig 1.3a** is a 2 dimensional example of PCA, most of the variance is along the orange axis (PC1). **Fig 1.3b** is an example of PCA in 3 dimensions, the computed principal components overlie the hodogram plot. The gray curve is the 'shadow' of the ground motion trajectory projected into the x-y plane to better convey the 3D nature of the trajectory. Since it was first applied to seismic data by Flinn (1965), numerous studies have also used the technique including Perelberg and Hornbostel (1994). This operation is purely in the time domain.

Figure 1.3: **PCA in Two & Three Dimensions**

a): 2d example of PCA **b)**: 3d example of PCA. The black is the x-y projection to better represent the 3 dimensional nature of the trajectory. The majority of the variation in the signal is along the orange axis.

Figure 1.4: **Mapping Vectors to Colours**

Visualizing directions in 3 dimensions using colours

The first principal component is used as a proxy for the polarization. In the case of perfectly linear polarization, PC1 is just that, a perfect characterization of the polarization state. For elliptically polarized waves the second principal component (PC2) must also be used. PC1 & PC2 then define a plane of shear strain that the particle motion is contained in. However, it is rare to have a single plane of ground motion.

Visualizing linear polarization requires that a vector quantity be displayed with respect to the dimensions of the survey. A new colour map was created that represents the components of the polarization vector by varying the amounts of red, blue and green at a point. The red-blue-green (RGB) colour map is

$$\begin{aligned}
 \text{Vertical (V)} &\propto \text{blue} \\
 \text{In-line (H1)} &\propto \text{green} \\
 \text{Cross-line (H2)} &\propto \text{red}
 \end{aligned}$$

This is illustrated for a unit vector in **Fig 1.4**. In this way the orientation of the polarization axis can be represented at a point.

1.5 Non-stationary Time-shift Estimation

A fruitful (albeit long) tangent to the polarization analysis was a method of producing a zero-offset section in an automatic fashion. The core idea is illustrated in **Fig 1.5**; by

Figure 1.5: **Local Shift Concept**

If the local shift between two traces is known for all times, this can be applied to the second trace to re-align it with the first, and make the arrival coherent.

knowing how much the second signal is delayed, it can be re-aligned with the first trace by re-interpolation. The difficulty comes from estimating the delay as it depends on time. A new approach to calculating the time-dependent time-shift is developed in Chapter 4. The new calculation is then incorporated into an alternative processing framework formalized by Fomel (2007a) called the velocity independent approach. This alternative scheme does not require velocity picking and is therefore less labour/time intensive for the user.

The study of seismic polarization is addressed in this thesis by the implementation/development of 2 distinct processing tools. First is the development of visualization scheme for 3 dimensional vector directions, using a new RGB colour map. Second is the development of a non-stationary time-shift estimation calculation, which when incorporated into the velocity independent processing framework can produce zero-offset sections in a much less labour intensive manner. Combining these processing/analysis steps allows for the polarization to be visualized on the scale of a survey, and the resulting polarization from different source orientations can be compared conveniently.

Using these processing tools, the following observations about the near-surface seismic wave-field were made:

1. Horizontal shake is most common in bedrock reflection for all source orientations.
2. Polarization analysis is more appropriately conducted on corrected gathers or zero-offset sections.

1.6 Document Structure

The rest of this document has the following structure:

Chapter 2 catalogs the data sets and summarizes the preprocessing steps used in this research. Preliminary observations on the multicomponent common-midpoint (CMP) and shot gathers are made.

Chapter 3 details the application of Principal Component Analysis (PCA) to seismic data to study the polarization. The mathematics of PCA is reviewed and the approach to visualizing the analysis is developed. This analysis is applied to the multicomponent data allowing for more quantitative observations to be made about the seismic polarization state.

Chapter 4 introduces an alternative approach to processing seismic reflection data, specifically in constructing zero-offsets sections. A new method of estimating time dependent time-shift is developed and tested. This is followed by the review of the velocity independent approach to seismic imaging in which the new time delay calculation is used. Multicomponent zero-offset sections are computed and compared to sections processed in the semblance normal-moveout manner.

Chapter 5 combines the two processing techniques presented in chapter 3 & 4 so that observations of the polarization of a multicomponent survey maybe investigated as a whole.

Chapter 6 summarizes the key findings of this research and provides some ideas for future studies.

Chapter 2

Datasets & Approach

To investigate seismic polarization in the near-surface, several multicomponent datasets were collected near Ottawa, Canada. At every point of acquisition, shots were recorded using two or three source orientations to compare the polarization of the reflected seismic waves. A schematic of the survey geometry along with the direction naming convention is shown in **Fig 2.1**. At all 5 experimental areas, data were collected with vertical (V) and cross-line (H2) horizontal sources ($2src \times 3rec = 6C$). Three of these locations also included the inline (H1) source data ($3src \times 3rec = 9C$), making them 9C survey. With the exception of the French Hill site, which was collected with a hammer source, the datasets were acquired using the Microvibe, a battery powered sweeping source developed in-house at the GSC (Pugin et al., 2013b). The Microvibe has 2 orthogonal shaking orientations, 1 vertical and 1 horizontal. The horizontal component can be oriented in any direction. The 3C geophones are mounted on steel sleds that can be towed along the road surface for the rapid collection of reflection data. This system, shown in **Fig 2.2**, allows for the rapid acquisition of multicomponent seismic data.

2.1 Test Sites

Five multicomponent datasets were acquired in and around Ottawa, Canada (**Fig 2.3c**) during 2017 & 2018. The site locations are John Shaw Rd (**Fig 2.3d**) near Kinburn, ON; Kaymar Drive in Orleans (**Fig 2.3e**); French Hill, North-East of Ottawa (**Fig 2.3f**); Natural

Figure 2.1: **Survey Geometry Schematic**

Survey geometry and source-receiver orientation naming convention for V (vertical), H1 (horizontal in-line) and H2 (horizontal cross-line) components.

Figure 2.2: **Survey Equipment - Microvibe**

Microvibe and 3 component landstreamer (0.75m geophone spacing). This photo is from the John Shaw rd. survey location. Microvibe depicted in cross-line shaking mode.

Site	Source	Source Orient.	Shot Space (m)	Station #	Bedrock Depth (m)
French Hill	Hammer	V,H1,H2	-	1	~ 30
Rte 300	μ -vibe	V,H2	3	312	~ 40
NRC Site	μ -vibe	V,H1,H2	0.75	90	~ 18
John Shaw Rd	μ -vibe	V,H1,H2	0.75	110	~ 110
Kaymar Dr.	μ -vibe	V,H2	1.5	135	~ 29.5

Table 2.1: **Dataset Summary Table**

Research Council Geotechnical Soil test site (NRC site) located South-East of Ottawa (**Fig 2.3g**) and Route 300 near Embrun, ON (**Fig 2.3h**). Survey parameters for each data sets are summarized in Table 2.1.

2.1.1 French Hill Road

A single location 9C record was acquired with a hammer source specifically to investigate polarization in the near surface. The data was collected with 0.75m receiver spacing, and 5 hammer hits per shot orientation were stacked to achieve a higher signal-to-noise ratio. The data was also acquired with an extremely high temporal resolution ($2.5 \times 10^{-4} s$ sample interval or 4kHz sampling frequency). Acquired February 20th, 2017; some of the geophone sleds froze to the ground resulting in excellent coupling. Shot location is shown in **Fig 2.3f**.

Fig 2.4 shows the shot gathers for each of the 9 source - receiver combinations. The 98th percentile of the signal magnitude serves as a display limit (display saturation), each panel is annotated with this value 'sat'. Despite the different orientation in hammer blows, there is similarity within the receiver orientations (eg. Vsrc-H2rec (**Fig 2.4c**) is similar to H2src-H2rec (**Fig 2.4i**)). Two explanations of this similarity are;

1. Reflection polarization is controlled by the geology.
2. The source produces all modes in various proportion of energy.

These interpretations are not mutually exclusive, however based on this exact dataset Pugin and Yilmaz (2017) observes that sources are not pure in their generation of the various waves types (P, S_v or S_h), which leads to the second interpretation more directly.

Figure 2.3: **Location Map**

Location of test datasets collected around Ottawa, Canada. **a)** Ontario (Eastern Ontario boxed in green) **b)** Eastern Ontario (Ottawa region boxed in blue) **c)** Ottawa Region and site locations. **d)** John Shaw Road. **e)** Kaymar Drive. **f)** French Hill Road. **g)** NRC site. **h)** Route 300

Figure 2.4: **French Hill 9C shot records.**

Shot gathers for each of the 9C source - receiver combinations. **a)** Vsrc-Vrec. **b)** Vsrc-H1rec. **c)** Vsrc-H2rec. **d)** H1src-Vrec. **e)** H1src-H1rec. **f)** H1src-H2rec. **g)** H2src-Vrec. **h)** H2src-H1rec. **i)** H2src-H2rec.

Interfering arrivals are seen in the H1-H1 component, particularly at 0.3 seconds (**Fig 2.4e**). At the earliest arrival times there is a clear difference in the frequency content and speed of the arrivals, between the different components (**Fig 2.4a&c**).

2.1.2 Route 300

A 6C (V,H2 src) survey was collected along Route 300 at 1.5m shot spacing on July 13th, 2017. Route 300 is a particularly well studied area located South East of Ottawa (**Fig 2.3h**). Clays overly glacial sands and gravels (Pugin et al., 2013b). The signal-to-noise of this data set is very good at this location although a culvert running through the middle of line produces some interferences.

Fig 2.5 shows the cmp gather at 170m from Route 300. The recordings are of very high quality, particularly the H2-H2 component (**Fig 2.5e**).

2.1.3 NRC Test Site

At the Natural Research Council (NRC) geotechnical research soil site #1, a major 9C dataset was collected. At 0.75m shot spacing, this survey has very high lateral resolution. The dataset was used by Pugin and Yilmaz (2019). **Fig 2.3g** shows the acquisition line.

Fig 2.6 shows the cmp gather at 57m from the NRC test site. V-V (**Fig 2.6a**) has the highest amplitude based on the display saturation value (98th percentile magnitude), this is followed by the H2-H2 (**Fig 2.6i**), then the H1-H1 (**Fig 2.6e**).

Hyperbolic arrivals take on a linear form when gathers are mapped into $X^2 - T^2$ coordinates. This allows deviations from hyperbolic geometry to be more easily recognized. **Fig 2.7** is a schematic of 3 hyperbolic arrivals and a linear arrival in $X - T$ and $X^2 - T^2$ coordinates. Hyperbolas are linearized and lines (e.g. direct arrivals) becomes non-linear in the stretched $X^2 - T^2$ coordinate system.

A shot from the NRC site is shown in both $X - T$ and $X^2 - T^2$ coordinates in **Fig 2.8**. This clearly shows the non-hyperbolic nature of the arrivals and indicates that the idealized subsurface assumptions (homogeneous, isotropic and horizontal layers) have been violated. This means the normal-moveout correction cannot perfectly correct the arrivals for the optimal stack.

Figure 2.5: **Route 300 Raw Seismic CMP Gathers**

a) Vsrc-Vrec. b) Vsrc-H1rec. c) Vsrc-H2rec. d) H2src-Vrec. e) H2src-H1rec. f) H2src-H2rec.

Figure 2.6: **NRC Site Data**

Common Midpoint sorted data, CMP: 57m a) Vsrc-Vrec. b) Vsrc-H1rec. c) Vsrc-H2rec. d) H1src-Vrec. e) H1src-H1rec. f) H1src-H2rec. g) H2src-Vrec. h) H2src-H1rec. i) H2src-H2rec.

Figure 2.7: **Hyperbolic Arrivals in $X^2 - T^2$ Coordinates**

a) 3 synthetic hyperbolic arrival times plotted in regular $X - T$ coordinates along with a direct arrival. The bottom 2 reflections have the same velocity, while the top has a higher velocity. **b)** In $X^2 - T^2$ coordinates the hyperbolic arrival time curves are linearized.

Figure 2.8: **H2-H2 Shot in $X^2 - T^2$ Coordinates**

a) Shot gather shown in regular $X - T$ coordinates. **b)** Shot gather in $X^2 - T^2$ coordinates. **c)** Zoomed $X^2 - T^2$.

Figure 2.9: **NRC Seismic Anisotropy**

3 component shot gather using H2 source recorded at NRC site. Ellipses highlight the arrival time for the bedrock reflection. **a)** V rec **b)** H1 rec **c)** H2 rec

H2 arrival comes in at a later time relative to the H1 arrival, and V arrives even later. Another potential cause of non-hyperbolic arrivals is the presence of seismic anisotropy; which is the polarization/directional dependence of seismic velocity. This might explain the observation that the nature of the arrival time with increasing offset is different for each of component (**Fig 2.9**). In **Fig 2.9** the bedrock reflection has distinctly different arrival time curves for each of the three components, most notably different is the vertical component **Fig 2.9a**. A simple model based on Thomsen (1986) (**Fig 2.10**) predicts the arrival times in isotropic and anisotropic media with similar subsurface conditions to the NRC site. Even weak-anisotropy (Transversely Isotropic symmetry about the vertical axis [VTI]) can lead to arrival time differences between components. Elastic/seismic anisotropy theory is reviewed in Appendix C. Based on Vertical Seismic Profiling measurements, Winterstein et al. (2001) notes that the very near-surface is often the most anisotropic, leading to shear wave splitting.

Figure 2.10: **Modelled Anisotropy**

Two layer model of arrival time. **a)** isotropic layer. **b)** anisotropic layer.

2.1.4 Kaymar Drive

A 6C survey collected with the Microvibe in May 2018, East of Ottawa (**Fig 2.3e**). In the middle of the survey is a large drainage pipe (Pugin et al., 2019). **Fig 2.11** shows the cmp gather at 94m from Kaymar Drive.

2.1.5 John Shaw Road

A high lateral resolution (0.75m shot spacing) 9C survey was acquired along a segment of John Shaw Road in Summer of 2018 (**Fig 2.3d**). This is another well studied area, an existing 6C (V,H2 source) survey collected with the Minivibe was acquired in 2012. This area has borehole control (Crow et al., 2015), and there are published examples using this data set (Pugin and Yilmaz, 2019).

Figure 2.12 shows the cmp gather at 56m from the John Shaw 9C test. The recordings have a lower signal-to-noise ratio compared to the other surveys due to windy conditions during acquisition. For this gather the 3 components of the V source recording has the highest 98th percentile signal magnitude.

2.2 Summary

Five multi-component surveys were acquired to investigate the polarization of seismic waves in the near-surface. These were acquired around Ottawa using a 3C landstreamer and multiple source orientations. An overview of the datasets and basic observations about the data was discussed. Some note worthy observations in the datasets are:

1. non-hyperbolic arrivals
2. different arrival time slopes for different components.

These observations might both be results of seismic anisotropy. A vast amount of data was acquired and only a subset is presented above. Using hodogram analysis to investigate polarization would require too much time for such a large catalog and is therefore not a practical approach. With this motivation a new polarization visualization technique is developed in the following chapter.

Figure 2.11: **Kaymar Data Overview**

CMP: 94m **a)** Vsrc-Vrec. **b)** Vsrc-H1rec. **c)** Vsrc-H2rec. **d)** H2src-Vrec. **e)** H2src-H1rec. **f)** H2src-H2rec.

Figure 2.12: **John Shaw Data Overview**

CMP: 56m a) Vsrc-Vrec. b) Vsrc-H1rec. c) Vsrc-H2rec. d) H1src-Vrec. e) H1src-H1rec. f) H1src-H2rec. g) H2src-Vrec. h) H2src-H1rec. i) H2src-H2rec.

Chapter 3

Principal Component Analysis

Studying seismic polarization is not a new idea, with early examples in the 1970s and 1980s (Montalbetti and Kanasevich, 1970; Rene et al., 1986; Jurkevics, 1988). Technically, the polarization could be characterized directly using the cmp gathers as there are currently displayed by comparing the amplitude between components. In principle, this is sufficient to make quantitative observations on the nature of the polarization. Many seismic polarization studies make use of hodograms (Pugin et al., 2009; Michaels, 2002) to analyse and classify seismic waves, **Fig 1.3b** is an example of such analysis. Hodograms have proven utility in understanding fundamental ground motion in 3 dimensions. The draw back to direct amplitude measurement and hodograms is that only a small subset of data can be studied in detail (several traces for a specific reflection). Although hodograms are an intuitive method to study seismic polarization, they are not used in this thesis because a survey level understanding of the seismic polarization is sought rather than individual trace studies for certain arrivals.

In order to look at the polarization on the scale of a survey, several processing tools and visualization techniques were used and developed. Principal component analysis is used on the 3 component traces to calculate the axis that accounts for the most seismic energy at a given time. The intent is to develop an analysis tool that conveys the richness of multicomponent data, while remaining practical for wide spread application: the analysis should be repeatable on new datasets in a practical manner. This constraint rules out the traditional form of polarization analysis which is using hodograms, because it requires too

much processing time when a full survey.

3.1 PCA Mathematical Formulation

Principal Component Analysis is eigenvalue decomposition on the covariance matrix of a multicomponent signal. It is a commonly used technique in many scientific disciplines such as chemometrics (Bro and Smilde, 2014), bioinformatics (Lee et al., 2010), climate science (Hannachi et al., 2006) and of course seismology (Jahan et al., 2017). The mathematics described here closely follows that outlined in Nguyen (1989). Using index notation, the covariance matrix is

$$\sigma_{ml}^2 = \frac{1}{N-1} \sum_{i=1}^N (x_m^i - \mu_m)(x_l^i - \mu_l) \quad (3.1)$$

μ_m is the mean of the m-th component;

$$\mu_m = \frac{1}{N} \sum_{i=1}^N x_m^i \quad (3.2)$$

these equations are written in every statistics textbook, including Bethea et al. (1995). For our purposes the indices correspond to the three directions of ground motion, i.e. m,l = V,H1,H2 components. To calculate the covariance as it varies in time, the summation is constrained to smaller time intervals;

$$\sigma_{ml}^2(t) = \frac{1}{M-1} \sum_{i=t-\frac{M}{2}}^{t+\frac{M}{2}} (x_m^i - \mu_m)(x_l^i - \mu_l) \quad (3.3)$$

here t is the index of the center sample within a time window that is M-samples long. Next the eigenvalues and eigenvectors of the covariance matrix are calculated by solving the equation:

$$[\tilde{\sigma}^2 - \lambda I] \vec{u} = 0 \quad (3.4)$$

where I is the identity matrix. In three dimensions, there are usually three solutions: the eigenvalues, $\lambda_1 > \lambda_2 > \lambda_3$, and their associated eigenvectors $\vec{u}^1$, $\vec{u}^2$, $\vec{u}^3$ which are orthogonal

to each other. These eigenvectors are the principal axes of the data for the specified window. Performing this calculation for every time sample we obtain PC1 ($\vec{u}^1$), PC2 ($\vec{u}^2$) and PC3 ($\vec{u}^3$) as they depend on time.

A drawback to PCA is that elliptical polarization states are easily overlooked, but there are parameters that give an indication of how linear the signal is based on the relative size of the eigenvalues. Nguyen (1989) defines a linearity parameter using the eigenvalues of PCA. The linearity is defined as

$$L = 1 - \left(\frac{\lambda_2 + \lambda_3}{\lambda_1} \right)^n \quad (3.5)$$

n is found empirically, though Nguyen suggests values between 0.5-1. For pure linear polarization $L = 1$, L becomes smaller for other polarization states. In the rare circumstance that the 3C signal variance has almost no preferred orientation ($\lambda_1 \approx \lambda_2 \approx \lambda_3$) the linearity $L \rightarrow -1$.

3.2 Visualizing Principal Components

The basic idea of mapping directions to colours was shown in **Fig 1.4**. This can be done for each of the principal components, PC1, PC2 and PC3. Take PC1 as an example. The cartesian components of PC1 are scaled to calculate the amount of red, green and blue.

$$\begin{aligned} u_z^1 &= PC1_V \propto B \\ u_y^1 &= PC1_{H1} \propto G \\ u_x^1 &= PC1_{H2} \propto R \end{aligned}$$

For a colour pixel, the amount of red, green and blue ranges from on a scale from $0 \rightarrow 255$; $[R \ G \ B] = [0 \ 0 \ 0]$ is black and $[R \ G \ B] = [255 \ 255 \ 255]$ is white. The cartesian components of PC1 are uniformly scaled such that the largest cartesian component will be represented by full colour: $[u_x \ u_y \ u_z] = [0.5 \ 0.25 \ 0.25] \rightarrow [R \ G \ B] = [255 \ 128 \ 128]$. This way the directions are represented by the brightest possible colours, providing the most contrast.

The colour map is shown in 3 dimensions in **Fig 3.1a**. The colour representation of the vector is determined by where the vector intersects the surface of **Fig 3.1a**. **Fig 3.1b** is

Figure 3.1: **Colour Map for Linear Polarization Axis**

a): Colour surface in 3 dimensions. **b)**: Lower-hemisphere stereonet projection of colourmap.

Figure 3.2: **PCA Example 1: Route 300 - H2src**

a) Vertical (V) component data **b)** Inline (H1) data **c)** Cross-line (H2) data **d)** 1st principal component vector (PV1) **e)** Data projected along PV1.

the stereonet representation of the colourmap, which will be used in the rest of the thesis. The colour map allows the direction of linear polarization to be represented in a single plot. Previous studies decompose the direction information into different components, for example dip and strike values (Labonne et al., 2016; Li and Crampin, 1991; Pinnegar, 2006) in doing so the polarization required many plots. A similar approach to represent polarization direction was used by Li and Crampin (1991), but for 2 component recordings (purely horizontal; no vertical), and so was not a complete characterization of the wavefield in 3D.

As an example, a shot gather from Route 300 was used, the results of which are shown in **Fig 3.2**. In **Fig 3.2d** most of the polarization is in the H2 direction (red), significant vertical polarization (blue) is situated in a shallow and far offset window. The fact that H2 dominates the arrivals is consistent with a strong shear wave reflection from the primarily H2 source. This visualization allows for convenient comparison of the relative amplitudes of the components.

3.2.1 PCA Optimization

While the concept of performing principal component analysis on the data is relatively straight forward, the method must be tuned and modified for optimal performance; both for accurate characterization of the wavefield & efficient computation.

Parameter Tuning

The main parameter in the algorithm is the window size, both in time and offset. Jackson et al. (1991) notes the main factors that should be considered in picking the window length for this kind of analysis;

- the window should contain one arrival.
- the window should yield the maximum signal-to-noise ratio.
- the window must be long enough that noise and signal are clearly distinct.

Using the data shown in **Fig 3.2**, different window lengths were tested to investigate how the analysis responds. **Fig 3.3** displays the results of this test. **Fig 3.3a,e** correspond to the principal component, and the data projected along this principal component with analysis window of 5ms respectively. Much of the power in the signal is below 200Hz ($F = 1/T$), so the window does not span an entire wavelength and consequently yields poorly projected data; arrivals frequently reverses polarity making the signal incoherent, especially around 0.2s. The principal axis jumps around within a single arrival as the oscillation goes through zero. In **Fig 3.3b** the polarization is more stable, due to the longer window. Here the 99th percentile of the signal magnitude is used for the display saturation. This value can serve as a metric for the optimal window length: The larger the saturation value, the better the 3 components are combined. Based on the dominant frequency & the 99th percentile amplitude the window length of 30 ms was selected. 40Hz is the lower end of the frequency content (due to filtering), which has a period of 0.025 seconds. This is consistent with an optimal window being between 10 - 30ms.

Another modification to the method is to expand the window along the offset axis, so that variance includes adjacent traces, to increase consistency in the offset dimension. **Fig**

Figure 3.3: **Effect of Window Length on PCA**

PCA using window lengths from 5ms to 100ms. **a)** PC1 vector time window: 5ms **b)** PC1 vector 10ms **c)** PC1 vector 30ms **d)** PC1 vector 100ms **e)** PC1 projected data 5ms **f)** PC1 projected data 10ms **g)** PC1 projected data 30ms **h)** PC1 projected data 100ms.

Figure 3.4: **Effect of Expanding Trace Window on PCA**

The number of traces in the window increases from left to right, the time window is fixed at 30ms. **a)** 1 trace. **b)** 3 traces. **c)** 5 traces. **d)** 7 traces. **e)** 9 traces. **f-j)** are the corresponding projections of the data along the principal axis shown in **a-e)**

3.4 shows the results of expanding the number of traces used in the PCA window. As expected the principal component direction becomes more consistent with offset. Using the same procedure as before for finding the optimal time window, we use the saturation amplitude noted above the projected data (**Fig 3.4f-j**) as a metric to find the optimal PCA window in offset. Since this has only a small effect on the 99th percentile amplitude, we use a single trace because it requires the least computation time.

When the data is project along the principal component axis, the fact that there is a sign ambiguity becomes an issue: The vector representing the principal component could be represented by a vector in the opposite direction (ie. the negative of the vector). In representing the vector in a colourmap, this is not an issue because the absolute value of

the components are used. To investigate this sign ambiguity, the direction of the largest component has been encoded in the transparency of the colourmap; when a negative value of the component was shown, the transparency is set to 75%. **Fig 3.5** compares the visualization of a shot gather principal component vector using the two colourmaps. Since the sign is shown by transparency, the signed colourmap is comparatively washed-out implying that the largest component of the max energy axis (principal component 1) is often negative. This becomes important when projecting the data along the PC1 axis.

Fig 3.6 tests different ways of addressing the ambiguity in the direction. **Fig 3.6e** shows the data projected along the principal component vector, ignoring the ambiguous sign direction of the vector; which results in polarity inconsistency along a given arrival, and therefore decreases the coherency of the signal. In **Fig 3.6f** the absolute value of the principal vector components are used; this results in an increased coherency in the signal. **Fig 3.6g** shows the projected data along the principal vector that has been corrected such that the largest component is always positive. It is difficult to distinguish which projection is better between **Fig 3.6f** & **Fig 3.6g**, so **Fig 3.6h** shows the absolute difference between the magnitude of the signals. Red indicates where **Fig 3.6g** has a higher signal magnitude, suggesting that the corrected PC1 vector combines the components more constructively overall than the vector used in **Fig 3.6f**.

Computation Speed

Repeating the principal component analysis for every time sample is computationally demanding. However with optimization of the algorithm, the polarization can be calculated much more quickly. The covariance can be reformulated from equation 3.3 as a convolution with an array the same length as the window.

$$\sigma_{ij}^2(t) = \frac{1}{n-1} \left(q_{ij} - \frac{s_i \cdot s_j}{n} \right) \quad (3.6)$$

where

Figure 3.5: **Comparing PCA Visualization with and without Signed Transparency**

a) Regular PCA visualization b) Signed value of principal components used.

Figure 3.6: **Comparing PCA Projected Data**

a) V **b)** H1 **c)** H2 **d)** PC1 vector **e)** data projected along PC1 **f)** sum of V, H1 and H2 weighted by the absolute value of the components of PC1 **g)** The principal vector is corrected such that the largest cartesian component is always positive. **h)** Comparison of the project data in **f** and **g**. Red patches show where the 3C data is combined more constructively for the projecting in **g** than the projection in **f**. Blue indicates the opposite.

$$\begin{aligned}
s_i(t) &= x_i * w \\
q_{ij}(t) &= (x_i \cdot x_j) * w
\end{aligned}
\tag{3.7}$$

See Appendix E for details of the reformation. Using the original calculation scheme (calculating the covariance matrix for the window within the time loop) a single shot with 48 traces and 1000 time samples would take ~ 6.9 s on a single processor. This computation time was reduced to ~ 1.2 s using the new calculation method (computing the windowed covariance matrices up front). For a survey this scales to a substantial time savings.

3.3 Application

3.3.1 French Hill

Fig 3.7a-l shows the results from applying PCA on the French Hill 9C dataset. The shear-wave bedrock reflection arrives ~ 0.3 seconds, and the multiple (raypath: surface $\rightarrow$ bedrock $\rightarrow$ surface $\rightarrow$ bedrock $\rightarrow$ surface) is also clear ~ 0.6 seconds. It is interesting how the polarization of the bedrock reflection compares with the polarization of the multiple; the amount of cross-line energy increases with time/propagation distance. For all source orientations there is vertically polarized energy in the main reflection and an increased amount of shear horizontal (H2) in the multiple. There are notably different arrival curves depending on the direction of the receiver. Looking at plots **Fig 3.7a** and **Fig 3.7c** this is particularly noticeable between 0.3-0.4 seconds at far offsets (≥ 30 m).

Fig 3.7m-o shows the analysis on common receivers. Previously, PCA has been conducted on physical 3C signals (with common sources). By instead conducting PCA on abstract 3 component signals, signals that share a common receiver orientation (e.g. a signal composed of Vsrc-Vrec, H1src-Vrec, H2src-Vrec), we get a comparison of the source energy for each receiver; what source generates the strongest reflector with a specific polarization direction. This is an abstract physical quantity and depends on the relative strength of the hammer strikes, and the coupling of the source to the ground. Simple logic suggests

Figure 3.7: French Hill Principal Component Analysis

Components are denoted in SRC-REC pairs a) V-V b) V-H1 c) V-H2 d) V source PC1 e) H1-V f) H1-H1 g) H1-H2 h) H1 source PC1 i) H2-V j) H2-H1 k) H2-H2 l) H2 source PC1 m) V receiver PC1 n) H1 receiver PC1 o) H2 receiver PC1 p) (V,V)-(H1,H1)-(H2,H2) PC1

that changing the orientation of the source would change the orientation of the particle motion of the reflected seismic waves; a vertical source resulting in a primarily vertically polarized reflection for example. These plots show which source is contributing the most energy for a particular arrival. While the results are not at all surprising for the H2 component, where the H2 source indeed contributes the most energy (the plot is dominated by the red, the colour of H2), the amount of energy from the returning in V and H1 from the H2 strike is quite surprising. The cross-line strike generates a tremendous amount of energy to all 3 receiver components. Some potential explanations for this observation are:

1. The H2 strike is stronger than the V and H1 strikes (which was not perceived during the acquisition)
2. The coupling of the H2 strike is better than V and H1 strikes putting more energy into the ground, although the strike plate is the same for H1 & H2 shots.
3. S-wave energy preferentially vibrates in H2.
4. Streamer design preferentially picks up H2.

The fact that the same strike plate was used for both H1 & H2, does not support the first two explanations. To get a more complete view of the polarization **Fig 3.8** shows all 3 principal components, as well as the rectilinearity (defined in equation 3.5). The principal components have a similar character for all 3 source orientations. Broadly, the rectilinearity is quite similar as well, with what appears to be an increasing degree of rectilinearity with time, however the bedrock reflection from the vertical source (**Fig 3.9d**) has the highest degree of linearity.

For a quantitative estimate of how similar the polarization is between different source orientations, the dot product is used to compare PC1 from each of the 3 sources. **Fig 3.9a** shows the distribution of the absolute value of the dot products between the first principal components for each source. A dot product of 0 indicates that the two PC1s are perfectly perpendicular to each other (zero similarity), and a dot product of 1 (or -1) indicates the two PC1s are parallel. **Fig 3.9b-d** breaks up these distributions and plots them with respect to time. There is no obvious relationship between polarization state similarity and time. The

number of time samples that have a dot product near 1, meaning the PC1 from each source are similar, suggests that the hammer source is not perfectly polarized in the directions of the strike.

3.3.2 NRC site

Fig 3.10 displays the results of PCA applied to the averaged data from the NRC survey. Reflection modes are highlighted in their strongest component with red ellipses. The bedrock reflection is obvious in every component, but is strongest in H2-H2 (**Fig 3.10k**). We see some reflected energy in H2 for every source orientation (red in bedrock reflection **Fig 3.10d&h&l**). H2 source produces the most linearly polarized reflections. This analysis also highlights the converted phases, especially $P - S$ in **Fig 3.10d**. The presence of H2 energy for every source is also not predicted by classic seismic theory assuming horizontal layers.

3.3.3 Route 300

Fig 3.12 is the principal component analysis of a cmp gather from the Rte300 6C survey. The bedrock reflection is dominated by H2 energy for both source orientations. The Vsrc PC1 (**Fig 3.12a**) has relatively low energy near-offset arrivals. The presence of vertically polarized energy is much more prevalent in the far-offsets. In **Fig 3.12d** there is a sharp transition from H2 polarized energy to vertically polarized energy from the cross-line source around 20m offset and 0.25s travel time. This is also present, though not as pronounced in the V source (**Fig 3.12c**). Both sources have a similar degree of linearity (**Fig 3.12e+f**).

3.3.4 John Shaw

In **Fig 3.13** the 9 components of the John Shaw survey are shown with their associated PC1 vectors for a single cmp gather. The H2 source has some energy reflected in the V component, not explained by classic reflection theory assuming the layers are horizontal. This is observed in many of the cmp gathers. Other gathers, some gathers have a surprising amount of vertical energy, even with a horizontal source. There is also similarity between the V source and H1 source. Low amplitudes have been masked due to the relatively high

Figure 3.8: **Complete PCA Frenchhill**

a) Vsrc-PC1 b) Vsrc-PC2 c) Vsrc-PC3 d) Vsrc - Linearity e) H1src-PC1
 f) H1src-PC2 g) H1src-PC3 h) H1src - Linearity i) H2src-PC1 j) H2src-
 PC2 k) H2src-PC3 l) H2src - Linearity
 Linearity is calculated with equation 3.5 with $n = 1$.

Figure 3.9: **Principal Component Similarity Estimation**

a) Histograms of dot product between V_{src} - $H1_{src}$ PC1; V_{src} - $H2_{src}$ PC1; and $H1_{src}$ - $H2_{src}$ PC1 **b)** Time dependent dot product distribution for V-H1 1st principal component. **c)** Time dependent dot product distribution for the V-H2 1st principal component. **d)** Time dependent dot product distribution for the H1-H2 1st principal component.

Figure 3.10: NRC 9C PC1 Comparison

PCA for the NRC site. This is the average of the shot gathers. Components are denoted in SRC-REC pairs **a)** V-V **b)** V-H1 **c)** V-H2 **d)** V source PC1 **e)** H1-V **f)** H1-H1 **g)** H1-H2 **h)** H1 source PC1 **i)** H2-V **j)** H2-H1 **k)** H2-H2 **l)** H2 source PC1 **m)** V receiver PC1 **n)** H1 receiver PC1 **o)** H2 receiver PC1 **p)** (V,V)-(H1,H1)-(H2,H2) PC1.

Reflection modes are highlighted in their strongest component using red ellipses.

Figure 3.11: NRC 9C PCA & Linearity

PCA for the NRC site. This is the average of the shot gathers. **a)** V-PC1 **b)** V-PC2 **c)** V-PC3 **d)** V source linearity **e)** H1-PC1 **f)** H1-PC2 **g)** H1-PC3 **h)** H1 source linearity **i)** H2-PC1 **j)** H2-PC2 **k)** H2-PC3 **l)** H2 source linearity.

Figure 3.12: **Route 300 6C PCA**

6C data PCA. **a)** V-PC1 projected data **b)** H2-PC1 projected data **c)** Vsrc - PV1 **d)** H2src - PV1 **e)** Vsrc - Linearity **f)** H2src - Linearity

noise of the data. The high amplitude reflections remaining show a high degree of linearity. Polarization at depth is primary horizontal, even for the vertical source.

3.3.5 Kaymar

Fig 3.14 is the principal component analysis of a cmp gathers from the Kaymar 6C survey. The bedrock reflection is dominated by H2 energy for both source orientations, and we note the multiple ~ 0.6 seconds returns with an increasing amount of H2 energy (**Fig 3.14c&d**). The gathers projected along PC1 (**Fig 3.14a&b**) have a similar character. Reflection from both source exhibit a high degree of linearity for the early arrivals, this is an increase in the linearity with offset for the bedrock reflection.

3.4 Summary

Using PCA, the axis that accounts for most of the variance can be visualized, and for linear polarization, this is sufficient for characterizing the polarization state.

While an efficient method for characterizing and visualizing the polarization has been developed and applied to the seismic data, the output is still immense, especially when considering every shot gather in the survey. A draw back to this type of analysis/visualization is only linear states of polarization are truly represented by the first principal component. Elliptical polarization states can be easily overlooked. The relative eigenvalues of the principal components can be evaluated to address the degree to which the polarization is linear, specifically the linearity as defined in Nguyen (1989), equation 3.5.

The prevalence of the H2 energy in the bedrock reflection warrants some attention, the question being what makes this direction distinct? One might expect a difference between the strength of the vertical and horizontal polarization because the geology is mostly flat in these studies. But the systematic dominance of the cross-line horizontal energy over the inline horizontal energy suggests that it might be related to the streamer design.

A simple follow-up experiment could be carried out to test the hypothesis that the geology determines the dominant polarization direction (by polarized attenuation). Even though this explanation is unlikely because it implies that every survey just happened to

Figure 3.13: John Shaw 9C PCA

9C data PCA. a) V-PC1 projected data b) H1-PC1 projected data c) H2-PC1 projected data d) Vsrc - PV1 e) H1src - PV1 f) H2src - PV1 g) Vsrc - Linearity h) H1src - Linearity i) H2src - Linearity

Figure 3.14: **Kaymar 6C PCA**

6C data with PCA. **a)** V-PC1 projected data **b)** H2-PC1 projected data
c) Vsrc - PV1 **d)** H2src - PV1 **e)** Vsrc - Linearity **f)** H2src - Linearity

be in alignment with this preferred direction. It could be tested by acquiring two sets of 9C test shots, where the streamer is oriented at 90° to each other. Comparing these two datasets would show whether the direction is specific to the geology or if it is specific to the survey equipment.

Currently the geophone sleds are mounted to a low-stretch strap in series with one another. Residual tension in the streamer maybe the reason for the observed difference between H1 & H2 polarization. A small modification to the streamer design could remove this difference by attaching the sleds to the tow strap by a short tether. In this way the near-offset sleds would be unaffected by the tension caused by the rest of the streamer. A drawback to this streamer design would be sacrificing some accuracy in the position of the sleds. These experiments are saved for future work.

Chapter 4

Time-shift Analysis

Zero-offset sections are an important step in interpreting the subsurface structure because they

- increase the signal to noise ratio by stacking.
- correct for the added travel time due to offset.

Producing a zero-offset section usually requires a velocity model of the subsurface. In the conventional approach, a hyperbolic travel time is assumed, and velocities are picked from semblance panels (Taner and Koehler, 1969), the objective being to maximize coherency. These velocities are then used to perform the normal-moveout (NMO) correction, which align the arrivals for stacking. Manually picking velocities is a time consuming process and depends on human skill and experience; an experienced geophysicist can achieve high quality results. Producing a zero-offset section for a single component survey can take time, but doing all 3 components is very time consuming, particularly if each component is processed independently. For this reason an alternative method of making a zero-offset section was sought, one that would be less labour intensive. Ideally the approach could accommodate non-hyperbolic travel times, motivated by the observation of non-hyperbolic reflections in the NRC 9C survey (see **Fig 2.8**). This is not a new problem; there are number of approaches introduced to address non-hyperbolic reflections. For example Xu et al. (2017) extend the semblance approach by including a cubic term in the moveout. Although this can lead to a better stack, the processing is much more difficult because of

the trade-off between picking velocities and picking the cubic term coefficients; instead of semblance panels, semblance cubes are needed for a single cmp gather, since this approach increases processing time even more, it was not considered.

4.1 Time Shift Method

As mentioned, traces can be aligned if the underlying velocity structure is known, but this requires a great deal of analysis. If instead the time-shift between traces is known for every time sample, the traces can be re-aligned directly without the need for a velocity model of the subsurface as illustrated in **Fig 1.5**. There are several existing methods for estimating the time-dependent time-shift: Chen et al. (2018), Hale (2013) and Jin et al. (2017) use Dynamic Time Warping (DTW), an algorithm developed for comparing recordings in speech recognition problems (Sakoe and Chiba, 1978). Fomel (2007a) uses Plane-wave Destruction to estimate the local event slope, p , the local slope being the local time-shift divided by the change in offset ($p = \Delta t / \Delta x$).

A new method of calculating the time-dependent time-shift is developed here. The idea is based on a property of Fourier transforms:

$$\begin{aligned}
 x_2(t) &= x_1(t - \Delta t) \\
 \updownarrow \\
 X_2(f) &= X_1(f) \cdot e^{-i2\pi f \Delta t}
 \end{aligned} \tag{4.1}$$

where x_1 is an arbitrary signal, x_2 is the same signal delayed by an amount Δt , and X_1 and X_2 are the corresponding Fourier transforms of the signal. **Fig 4.1** illustrates this property with a ricker wavelet, a similar example in Yilmaz (2001) demonstrates this property. Only the phase content differentiates the 2 signals; differing by a linear ramp with slope proportional to the magnitude of Δt . Rearranging the equation for Δt in the frequency domain gives

Figure 4.1: **Motivating Property of the Fourier Transform**

a) A ricker wavelet (solid line) and the same signal delayed by Δt (dashed line). **b)** The amplitude spectrum each signal. **c)** The phase spectrum of each signal.

$$\Delta t = -\frac{1}{i2\pi f} \log\left(\frac{X_2(f)}{X_1(f)}\right) \quad (4.2)$$

The transformed signal in polar coordinates ($z = x + iy = r \cdot e^{i\phi}$) is insightful,

$$\Delta t = -\frac{1}{i2\pi f} \log\left(\frac{R_2(f) \cdot e^{i\phi_2(f)}}{R_1(f) \cdot e^{i\phi_1(f)}}\right) = -\frac{1}{2\pi f} \log\left(\frac{R_2}{R_1} \cdot e^{i(\phi_2 - \phi_1)}\right) \quad (4.3)$$

since it simplifies to

$$\Delta t = -\frac{1}{i2\pi f} \left[\log\left(\frac{R_2}{R_1}\right) + i(\phi_2 - \phi_1) \right] \quad (4.4)$$

The real part corresponds to the delay, while the imaginary part is related to the attenuation of signal. As is, the calculation is useful if the signal is delayed by a constant value for the whole recording, but for seismic, the shift varies with time. To accommodate time-dependence, the Fourier transforms are replaced with Stockwell transforms (S-transform) (Stockwell et al., 1996), which give an estimate of the frequency content of the signal as it depends on time. **Fig 4.2** shows the S-transform of a sweeping signal to highlight the changing nature of the frequency content with time, and how the time-frequency domain is insightful for non-stationary signals. More on the S-transform is available in Appendix B, but the definition is

$$X(\tau, f) = \int_{-\infty}^{+\infty} x(t) \left(\frac{|f|}{\sqrt{2\pi}} \exp\left(\frac{-f^2(\tau - t)^2}{2}\right) \right) \exp(-2\pi i f t) dt \quad (4.5)$$

The S-transform is similar to the Fourier Transform, but has a Gaussian sliding window scaled by the frequency (the window is larger for lower frequencies).

Replacing the Fourier transforms in equation 4.3 with S-transforms gives:

$$\Delta t(\tau, f) = -\mathbf{Re} \left\{ \frac{1}{i2\pi f} \log\left(\frac{X_2(\tau, f)}{X_1(\tau, f)}\right) \right\} \quad (4.6)$$

The calculated delay is now a function of the time and frequency. Lastly, a weighted average over frequency yields a local static shift depending only on time;

Figure 4.2: **S-transform Representation of a Sweep**

A 20-350Hz sweeping signal (Pilot trace from the Microvibe). **a)** Time domain representation. **b)** Signal amplitude in the time-frequency domain calculated using the S-transform. **c)** Associate S-transform phase spectrum. **d)** Fourier amplitude spectrum. **e)** Fourier phase spectrum.

$$\Delta t(\tau) = \sum_{f=0}^{f=f_n} \Delta t(\tau, f) \cdot W(\tau, f) \quad (4.7)$$

where $W(\tau, f)$ is the weighting function. The result is a time dependent local static shift. This calculation is referred to as the phase-delay calculation, since the time-shift / delay is derived from local phase content of the signal. Computing this for every trace, gives the local static shift with offset. From this the local event slope is estimated by dividing by the geophone spacing:

$$p(t, x) = \frac{\Delta t(t, x)}{\Delta x} \quad (4.8)$$

To test this calculation we use two adjacent traces, shown in **Fig 4.3**. This uses a weighting function proportional to the product of the two signals S-transform magnitude:

$$W(\tau, f) \propto |X_1(\tau, f)| |X_2(\tau, f)| \quad (4.9)$$

After re-interpolating the second trace based on the calculated static shift, there is good alignment between the two signals. The calculation accommodates changes in the wave-form, for example the arrival at time $\sim 0.16s$.

The S-transform has been applied to seismic data before, and even for the analysis of polarization (Pinnegar, 2006). However using the S-transform to compare relative time shifts between two signals is a new application.

Fig 4.4 illustrates the delay calculation on a cmp gather from the NRC site. Here the time delay between traces is small ($< 3ms$). This figure also highlights the issue of interfering arrivals (20.25m offset & $\sim 0.45s$). Some time shift estimation methods limit their analysis to non-interfering arrivals, including Chen et al. (2018).

Initially the method only shifted samples by an integer number of time samples, which introduces a rounding error upon shifting. Instead an interpolation scheme was adopted so that the time delay calculation was not rounded to the nearest integer. Using efficient interpolation schemes included in the open source library Scipy (Jones et al., 2001) the calculation is faster than using the sample indexing method used previously.

Figure 4.3: **Example Time Shift Calculation**

a) Adjacent traces **b)** Calculated time shift using equation 4.7 where $W(\tau, f) \propto |X_1(\tau, f)||X_2(\tau, f)|$. **c)** Aligned traces applying the time shift from **(b)** to the second signal.

Figure 4.4: **Time Shift Method Illustration**

Illustration of the iterative approach of shifting the traces based on the time delay.
a-e) Shifting processing, noisy traces remain - **a)** 0 shifted (raw record) **b)** 5 far-offset traces shifted **c)** 15 far-offset traces shifted **d)** 25 far-offset traces shifted **e)** all traces shifted
 Averaged data from NRC site (H2,H2).

Figure 4.5: **Rounding Error in Trace Shifting**

a) Two adjacent traces (Rte300 H2,H2 cmp 20 traces 10-11). **b)** Calculated time delay, rounded to nearest sample. **c)** Shifted traces rounded vs interpolated delay comparison.

Figure 4.6: **Testing the Limit of the Delay Calculation**

a) Pilot trace. **b)** Set of traces to which the pilot trace is compared to test the time shift calculation. **c)** Estimate time shift between the traces in **(b)** and the pilot trace **(a)**. **d)** Traces after applying the shift correction calculated in **(c)**.

In **Fig 4.6** a two wavelet model was made to test the new time-shift estimation method, on a range arrival time differences. The arrival time of the first wavelet was varied from 0.25-0.35s (**Fig 4.6b**) and compared with the pilot trace in **Fig 4.6a** where the first wavelet arrives at 0.3s, this corresponds to a shift between -50 & 50ms. The second arrival is fixed at 0.75s to test the time-dependent nature of the calculation. **Fig 4.6c** is the estimated time shift using a weighting function $W(\tau, f) \propto |X_1(\tau, f)| \cdot |X_2(\tau, f)|$. To evaluate the efficacy of the calculation, all the traces are shifted to align the arrivals with the pilot trace. If the calculation performed perfectly, all the first arrivals would sit at 0.3s and all the second arrivals would remain at 0.75s. The calculation provides an accurate estimation of the time shift between -16ms to 16ms but for larger shifts the calculation fails. The fact that there are time-shift estimates present around 50ms in **Fig 4.6c** (although far from the energetic part of the arrival), suggests that calculation has the potential to produce an accurate estimation of the time-shift. It is hypothesized that the weighting function plays an important role in getting an accurate estimate of time shift and consequently extend the effectiveness of the calculation to larger time-shifts.

Several weighting functions (Table 4.1) were tested using the same model as **Fig 4.6**, only the shifted traces are shown to evaluate their performance (**Fig 4.7**). There is no

Table 4.1: **Weighting Function Legend for Fig 4.7**

Label	$W(\tau, f) \sim$	notes
a)	$ X_1(\tau, f) $	
b)	$ X_2(\tau, f) $	
c)	A_{12}	†
d)	$\exp(A_{12}) - 1$	
e)	$\begin{cases} 0, & \text{if } \Delta t/T > 0.4 \\ A_{12}, & \text{if } \Delta t/T < 0.4 \end{cases}$	†‡
f)	$\begin{cases} 0, & \text{if } f < f_1 \\ A_{12}, & \text{if } f_1 \leq f \leq f_2 \\ 0, & \text{if } f > f_2 \end{cases}$	†⊙

†) $A_{12} = |X_1(\tau, f)| |X_2(\tau, f)|$

‡) $T = 1/f$

⊙) f_1 and f_2 define a band containing the 5 most powerful frequencies in the Fourier transform of $x_1(t)$.

Figure 4.7: **Weighting Function Experimentation**

Resulting corrected gathers using different weighting functions tested on the same model as in **Fig 4.6**. The weighting functions associated with each panel are tabulated in Table 4.1.

significant difference between using the amplitude of the 1st signal S-transform (**Fig 4.7a**), using the S-transform amplitude of the second signal (**Fig 4.7b**), or the product of the two (**Fig 4.7c**). Using the exponential of the amplitude spectrum as a weighting function does not seem to improve the estimation either (**Fig 4.7d**). It was hypothesized that the poor shift estimates resulted from large delays relative to the frequency, so a filtering criteria was tested. If $\Delta t(\tau, f)$ estimate exceeds the $0.4 * T$, it is removed (**Fig 4.7e**), however it does not give a significant improvement over the previous weighting functions. The weighting function of panel **Fig 4.7f** makes use of a frequency band defined by the interval that includes the 5 most powerful frequencies in the overall signal (regular Fourier Transform), this is illustrated in **Fig 4.8**. This weighting function produced the most accurate estimation for the largest range of time shifts. It also requires the least computational time because only a subset of the S-transform frequencies are calculated. This weighting function is used for all following time-shift calculations. The estimated time delay is accurate for time shifts $< 20\text{ms}$. The calculation is no longer accurate when there is insufficient overlap of the 2 wavelets ($> 22\text{ms}$). At these large shifts, the signal begins aliasing: there is aliasing between the troughs of the first arrival, resulting in an incorrect shift calculation. For real data this means that the geophone spacing must be sufficiently short. For near-surface applications where the ground velocity is low in particular, it is recommended to have a short geophone spacing to apply the correction.

A synthetic example from Chen et al. (2018) in **Fig 4.9** shows how Dynamic Time Warping (DTW) can be applied to iteratively correct a shot gather. The noise free example **Fig 4.9a-c** yields a perfect correction without stretching at far offsets. With noise (**Fig 4.9d**) however, the arrivals are no longer corrected perfectly (**Fig 4.9e**). In Chen et al. (2018) a multi-trace averaging scheme was adopted to improve the method in the presence of noise (**Fig 4.9f**).

This model was replicated and used to test the new phase-delay calculation, the results are shown in **Fig 4.10**. The proposed method handles the noise without having to implement a multi-trace averaging scheme.

While DTW is intrinsically more robust to aliasing of the signal and can work for larger time shifts, the proposed method is demonstrably more robust to noise. Both methods

Figure 4.8: **Band-limited Time-shift Weighting Function**

Weighting function for **Fig 4.7f**, an interval including the 5 most powerful frequencies is used to define a band over which the weighted average is taken.

$W(\tau, f) \propto |X_1(\tau, f)| |X_2(\tau, f)|$ over this frequency band.

Table 4.2: Model Parameters for **Fig 4.10**

Reflector		1	2	3	4	5	6	7	8
Depth	(m)	500	1000	1400	1700	2000	2300	2700	3000
Velocity	(m/s)	2100	2300	2600	2900	3300	3700	4000	4300

From Chen et al. (2018).

Figure 4.9: **Dynamic Time Warping Results (Chen et al., 2018)**

Iterative moveout correction based on local static shift calculated using Dynamic Time Warping, from Chen et al. (2018). **a)** Synthetic gather (model in table 4.2). **b)** Local static shift from DTW. **c)** Corrected gather. **d)** Modelled data as in (a) with a band-limited Gaussian noise, SNR of 3. **e)** Corrected gather using results from direct application of DTW. **f)** Corrected gather by using DTW with a multi-trace averaging scheme

Figure 4.10: **Synthetic Time Shift Test with and without Noise**

Replicated model from Chen et al. (2018) to directly compare phase-delay method and DTW. Model parameters are found in Table 4.2. **a)** Noise-free shot gather. **b)** the calculated time delay. **c)** Corrected gather after iterative correction. **d-f)** Analogous plots with a noise amplitude of 50% of the peak amplitude in **a)**

are limited to non interfering arrivals.

The shifting process can be applied iteratively from the furthest offset to the nearest offset as demonstrated in **Fig 4.4**, **Fig 4.9**, and **Fig 4.10**. The last trace is shifted by the delay time, flattening the arrival with the second last trace. Next, the last two traces are shifted by the delay time between the third last trace and the second last trace; flattening the arrivals of the last 3 traces. This process is repeated until all the traces have been shifted, ideally flattening the arrivals with the nearest offset trace. Another field data example is shown in **Fig 4.11**. The method is practical for coherent reflectors (**Fig 4.11f-j**). A problem is errors in the time-shift estimation at near offsets effect on all further offset traces. In **Fig 4.11e** the arrival at $t \approx 0.5s$, is distorted beyond recognition after shifting due to 5 noisy traces at 14.25, 19.5, 23.25 26.25, and 33m offsets. The difficulty with this method is that the phase of the arrival is important for calculating the delay. Unlike in conventional semblance calculations, suppressing the noise by notch filter or muting doesn't address the problem. Noisy traces must either be interpolated over or removed entirely. **Fig 4.11f-j** illustrates the same procedure, but the problematic traces have been removed. Using the recalculated time delay, the shifted traces are more coherent, in particular the arrival at $t \approx 0.5s$. However, there still is significant stretching of the wavefield. The stretching is not apparent until all the traces have been shifted (The far-offset traces of **Fig 4.11i** do not show signs of stretching), which indicates the stretching is due to the time-shift estimate of nearer offset traces. In its current state, the method has two inherent limitations:

1. Imperfections in the time-shift estimation for near-offset traces effect all the further offset traces.
2. Shifting of the traces results in a finite-offset equivalent trace, not a zero-offset trace.

Even with a perfect estimation of the non-stationary time-shift and subsequent perfect stack of the traces, the result is not a zero-offset trace. To achieve the zero-offset equivalent stack, the time-delay of the nearest offset trace must be estimated. To do this one must assume a travel time model. The velocity-independent approach as formalized by Fomel (2007a) addresses these limitations, providing an existing framework to make use of the new

Figure 4.11: **Time Shift with Noisy Traces**

Illustration of the iterative approach of shifting the traces using the phase-delay method. **a-e)** Shifting process, noisy traces remain - **a)** 0 shifted (raw record) **b)** 5 far-offset traces shifted **c)** 15 far-offset traces shifted **d)** 25 far-offset traces shifted **e)** all traces shifted **f-j)**: Shifting process, noise traces removed (offsets: 14.25, 19.5, 23.25, 26.25, 33m). **f)** 0 shifted (raw record) **g)** 5 far-offset traces shifted **h)** 15 far-offset traces shifted **i)** 25 far-offset traces shifted **j)** all traces shifted. Data from Rte300 (H2,H2) cmp 20

time-delay calculation. In the velocity independent framework traces are not corrected in an iterative fashion, instead they are corrected individually based on the local slope (local time shift). As a result, noise/ inaccurate time-shift estimates only affect specific traces; not all the all farther offset traces.

4.2 Velocity Independent Seismic Imaging

Although the phase-delay method is new, using the time-dependent time-shift is not, so there is an existing framework in which this calculation can be used. Fomel (2007a) formalizes the velocity-independent approach for specific travel time geometries; both hyperbolic

Figure 4.12: **Processing Workflow Comparison**

a) conventional semblance-nmo workflow. b) velocity independent workflow.

and non-hyperbolic arrivals. This formalization is reviewed here, the overall workflow for velocity independent approach is compared with the Semblance-NMO workflow in **Fig 4.12**. The key difference is that the velocity structure is not required to correct the arrivals, hence the name velocity independent.

4.2.1 Normal Move-out

Though the arrivals are not always observed to be hyperbolic, it is useful to start with the classic and simplest case. The arrival time as a function of offset for hyperbolic reflections is

$$t(x) = \sqrt{t_0^2 + \frac{x^2}{v_n^2(t_0)}} \quad (4.10)$$

where $t(x)$ is the arrival time, x is the offset, t_0 is the zero-offset travel time, and $v_n(t_0)$ is the stacking velocity. The local slope of an arrival, $p = dt/dx$, is just the derivative of this equation.

$$p = \frac{x}{tv_n^2(t_0)} \quad (4.11)$$

Rearranging equation 4.10 for t_0 and eliminating $v_n(t_0)$ with equation 4.11 gives

$$t_0 = \sqrt{t^2 - txp(t, x)} \quad (4.12)$$

Knowing the local slope, the associated zero-offset travel time can be calculated for every sample in the gather and corrected directly.

4.2.2 Nonhyperbolic Move-out

Using the non-hyperbolic travel time equation from Malovichko and Sword (1978)

$$t(x) = t_0 \left(1 - \frac{1}{S(t_0)} \right) + \frac{1}{S(t_0)} \sqrt{t_0^2 + S(t_0) \frac{x^2}{v_n^2(t_0)}} \quad (4.13)$$

where S is a non-hyperbolic term, the rest of the parameters are identical as the ones used in equation 4.10. In this case the first and second derivatives are needed; $p = dt/dx$, and let $q = \partial p / \partial x = \partial^2 t / \partial x^2$ yielding.

$$p = \frac{x}{[t_0 + S(t_0)(t - t_0)] v_n^2} \quad (4.14)$$

$$q = \frac{t_0^2}{[t_0 + S(t_0)(t - t_0)]^3 v_n^2} \quad (4.15)$$

Rearranging for t_0 and using the analogous substitutions gives

$$t_0 = t - px \left[1 + \sqrt{\frac{qx}{p}} \right]^{-1} \quad (4.16)$$

Fig 4.13 illustrates the velocity independent workflow using a sample CMP gather from Route 300. The local slope is calculated from the cmp gather, this calculation can be used directly to reorient (align) the arrivals using an assumed travel time curve. The reoriented gather, **Fig 4.13d**, can be stacked to get the zero-offset trace with the highest signal to noise ratio (**Fig 4.13e**). This is done without any knowledge of the velocity structure, however the velocity can be calculated based on the local slope. **Fig 4.13c** shows the calculated velocity using the phase-delay method, it overlies the semblance as a sanity check. More

Figure 4.13: **Velocity Independent Workflow Example Shot**

a) CMP gather. **b)** Estimated local slope. **c)** Velocity profile. The velocity is overlaying the semblance as a quality check. **d)** The reoriented gather based on the local slope **e)** The stacked trace. The inset dashed box (top-right) shows the workflow. [script]

detail is provided in calculating the velocity from the local slope in a later section.

4.2.3 Local Slope Regularization

An additional processing step was used to smooth the local slope field of **Fig 4.13b** in order to obtain an optimal stack. This step, called shaping regularization was also applied by Fomel (2007a) but developed in a previous paper (Fomel, 2007b). Instead of smoothing along the axes of a pre-existing grid, shaping smooths the data with respect to an arbitrary geometry, in this case the most relevant geometry being a hyperbolic one. Here hyperbolic shaping is applied to regularize the local slope value calculated with the phase-delay.

Assuming a hyperbolic arrival time, the local slope of the next trace is calculated based on the local slope of the previous trace. This is projected iteratively for every trace in the gather. Based on the slope for a single trace, the local slope is estimated for the whole gather. Doing this for every trace then averaging them together, an averaged local slope is computed accounting for the hyperbolic geometry of the arrivals. The key piece is to predict the local slope of a trace based on the local slope of the trace before. Taking the Taylor expansion of the local slope expression with respect to the offset,

$$p(x_o + \Delta x) = p(x_o) + \left. \frac{\partial p}{\partial x} \right|_{x=x_o} \Delta x + \left. \frac{\partial^2 p}{\partial x^2} \right|_{x=x_o} \Delta x^2 + \dots \quad (4.17)$$

Using a linear approximation (truncating the expansion of the Taylor series to first order term), only the first derivative is needed;

$$\frac{\partial p}{\partial x} = \frac{\partial}{\partial x} \left(\frac{x}{tv^2} \right) = \frac{1}{tv^2} = \frac{p}{x} \quad (4.18)$$

assuming a hyperbolic geometry. To check, consider the limit as $x_o \rightarrow \infty$

$$\lim_{x_o \rightarrow \infty} p(x_o + \Delta x) = \lim_{x_o \rightarrow \infty} p(x_o) + \frac{p(x_o)}{x_o} \Delta x = p(x_o) \quad (4.19)$$

the slope approaches a constant value, which is consistent with a hyperbolic geometry.

Fig 4.14 is an example of this regularization on the local slope field. This smoothing of the slope by shaping regularization proves to be an important step in achieving a reliable

moveout using the velocity independent approach.

The next question is which travel time model is most appropriate/effective for the near surface data sets. **Fig 4.15** uses the H2-H2 components to compare the correction assuming a hyperbolic travel time curve (equation 4.12), with the correction using a non-hyperbolic travel time model (equation 4.16). Gathers from all 4 Microvibe datasets are shown. For John Shaw (**Fig 4.15a&b**) and Kaymar (**Fig 4.15c&d**) there is no difference in the 98th percentile absolute magnitude (sat value), though the corrected gathers assuming a hyperbolic travel time **Fig 4.15a&c** are more continuous over offset for the different reflections. Particularly for the bedrock reflection in Kaymar ($\sim 0.3s$ **Fig 4.15c**). The NRC site (**Fig 4.15e&f**) and Route 300 (**Fig 4.15g&h**) both have a larger saturation amplitude when assuming a non-hyperbolic arrival time. However, the hyperbolic assumption yields more continuity with offset for both surveys. It seems that the second order difference of the slope field is not particularly stable, resulting in an unstable correction.

4.2.4 Testing

Fig 4.16 compares sections for Route 300 (H2,H2) processed using different methods. **Fig 4.16a** was processed using semblance analysis to get the stacking velocity, which was then used to perform the normal-moveout correction. **Fig 4.16b** was processed in an automated fashion using the phase-delay method incorporated into the velocity independent approach assuming a hyperbolic travel-time model. The sections are of comparable quality. Semblance analysis and velocity picking in **Fig 4.16a** was conducted every 10 cmp gathers, and the stacking velocity was interpolated, resulting in a smoother reflections. In **Fig 4.16b** the velocity independent analysis was done for every CMP, which is why the section is not quite as smooth, however the continuity of the reflections suggest that the reduced smoothness reflects the actual nature of the subsurface. By applying the analysis to every cmp gather the lateral resolution is higher.

In addition to a zero-offset section, the velocity can be calculated from the local slope information. Rearranging equation 4.11, we get the velocity assuming a hyperbolic geometry:

Figure 4.14: **Hyperbolic Regularization**

a) Uncorrected CMP gather from Route 300. **b)** Corrected gather based on local slope of plot **(d)** **c)** Corrected gather using regularized local slope of plot **(e)** **d)** Un-regularized local slope field. **e)** Regularized local slope field.

Figure 4.15: **H2-H2 Hyperbolic vs. Non-hyperbolic VI correction**

CMP gathers are re-oriented using the velocity independent approach. Hyperbolic use equation 4.12 to calculate $t_o(x, t)$ and non-hyperbolic use equation 4.16. **a)** John Shaw - hyperbolic **b)** John Shaw - non-hyperbolic **c)** Kaymar - hyperbolic **d)** Kaymar - non-hyperbolic **e)** NRC - hyperbolic **f)** NRC - non-hyperbolic **g)** Route 300 - hyperbolic **h)** Route 300 - non-hyperbolic

Figure 4.16: NMO-Semb vs VI Comparison

Here are the processed section using Route 300 (H2,H2) data. **a)** section processed with semblance-nmo. **b)** section processed in velocity-independent manner assuming hyperbolic travel-time.

$$v_n = \sqrt{\frac{x}{t \cdot p(x, t)}} \quad (4.20)$$

The calculated velocity is a function of time, offset and slope. To get the velocity as a function of t_o only, the velocity has to be reoriented the same way the cmp gathers are reoriented. The output is still a function of offset, so a weighted average over offset is taken to get a representative velocity value. **Fig 4.17** compares the velocity section from Route 300 (H2-H2) attained using the semblance-nmo method (**Fig 4.17a**), and the velocity independent (VI) approach assuming a hyperbolic geometry (**Fig 4.17b**). There is good overall agreement between the two sections. Using the semblance-nmo approach gives more control and is therefore more robust to some types of interferences, such as the presence of the culvert in the middle of section. Because the velocity is calculated for every CMP in **Fig 4.17b** compared to every 10 CMPs in **Fig 4.17a**, there is more lateral variation in the velocity using the velocity independent approach.

The weighted average of the calculated velocity over offset is the most straight forward method of attaining the velocity from the local slope, however it is likely that the velocity associated with every offset could be used in a more sophisticated manner, to get a distribution of velocities or estimate the uncertainty on the velocity. This is saved for future work.

A primary motivation for using the velocity independent approach is that the processing is automated, so multiple components for a survey can be processed with minimal additional effort (only more computation time). **Fig 4.18** shows the zero-offset sections for every component recorded in the Route 300 survey. The H2-H2 component (**Fig 4.18f**) yields the highest quality section.

Fig 4.19 shows the processed sections for Kaymar Drive. The tunnel is most obvious in H2-H2 **Fig 4.19f**, in some of the components it doesn't appear. For both sources, V and H2, the bedrock reflection is strongest in H2. The V-H2 section (**Fig 4.19e**) is of good quality; the bedrock reflection is sharper in comparison to the H2-H2, though the tunnel

Figure 4.17: NMO-Semb vs VI Comparison with Velocity Overlay

a) Semblance derived velocities over a zero-offset section. b) Velocities derived from local slope values using a velocity independent approach assuming a hyperbolic geometry.

Figure 4.18: **Route 300 Multicomponent Velocity Independent Imaging**

a) V-V b) H2-V c) V-H1 d) H2-H1 e) V-H2 f) H2-H2

Figure 4.19: **Kaymar Multicomponent Velocity Independent Imaging**

a) V-V b) H2-V c) V-H1 d) H2-H1 e) V-H2 f) H2-H2

reflection is not nearly as strong.

4.3 Summary

After calculating the local time delay, two methods of applying the correction were used, the first being an iterative correction with no underlying assumptions made about the travel time curve, and the second being the velocity independent approach assuming hyperbolic and non-hyperbolic travel time models. The value of the velocity independent approach is that it allows the rapid and automatic processing of multicomponent data. Using this approach it is conceivable to process the data during acquisition.

The advantage of the iterative approach is that no assumptions are made about the arrival geometry, and the travel time curves can be directly inverted for subsurface parameters. The disadvantage is that error accumulates over the shifting process effecting all traces of greater offset.

Even though there is an ideal to process the data with no underlying assumptions about the travel time, we find the results of the velocity independent approach (assuming a hyperbolic travel time) are far superior to the iterative correction. Additionally, in comparing the corrected gathers between hyperbolic and non-hyperbolic arrivals, the hyperbolic assumption yielded more continuous move-out with offset.

Chapter 5

Synthesis and Evaluation

5.1 Application and Evaluation

A method of analyzing seismic polarization is applied here by combining the analysis tools developed in previous chapters. Even with the colour mapping technique to display the polarization of a 3C survey, the data can be cumbersome to interpret because of the volume. Particularly if one wants to analyze sections with more than one source orientation. The most obvious way to condense the information is to consider the polarization of the zero-offset stacked section. Now that a reliable automated approach to deriving the zero offset section has been developed and tested on near-surface data, PCA can be conducted on the zero-offset sections in an efficient manner. The automated approach is an important piece in this processing scheme because picking velocities for 9C data is time consuming.

A motivating idea for this polarization study was being able to continually process along the highest energy axis. One question in combining the polarization and velocity independent processing techniques is the order that they are applied:

1. Compute PC1 and create a zero-offset section from PC1 projected data cmp gathers.
2. Process the zero-offset sections for each component, then apply PCA to get PC1 of the section.

These two approaches are compared using the H2 source from the Route 300 dataset, the results are shown in **Fig 5.1**. **Fig 5.1a** was obtained by first processing the 3 zero

offset sections (H2-V, H2-H1, & H2-H2) using the phase-delay method in velocity independent framework. **Fig 5.1b** was obtained by computing the PC1 for every cmp gather, and then applying the phase-delay calculation and using a hyperbolic geometry to get the zero-offset section. **Fig 5.1a** is more continuous, especially the bedrock reflection. We also get an overall higher amplitude as indicated by the saturation value (98th percentile magnitude) suggesting the 3 components are more constructively combined by applying PCA to zero-offset sections, rather than constructing a zero-offset section from PC1 projected cmp gathers. Considering that one of the reasons zero-offset sections are produced is to enhance the signal-to-noise ratio, the fact that the PCA analysis is superior on the stacked data makes sense. So based on this test, the method used in **Fig 5.1a** is applied to the other surveys.

5.1.1 Route 300

It was hypothesized that projecting the data along PC1 would result in a higher quality section. To test whether the PC1 section has an overall higher quality, **Fig 5.2** compares PC1 zero-offset sections with V-V and H2-H2 sections as well as with sections made by a simple stack of the 3C data directly. It is interesting that the vertical source has the highest amplitude by adding the 3 zero offset sections together, although the amplitude is quite similar to the main component (V,V). In contrast, for the H2 source the amplitude is increased dramatically compared to the main component (H2,H2). 98th percentile amplitude is nearly doubled by projecting the zero-offset data along PC1 (**Fig 5.2f**). Even though the overall amplitude is higher for the PC1 data from the H2 source some of the earliest arriving reflections are less defined. For the H2 source the PC1 projected data does not have the same obvious benefit over H2-H2.

PCA was conducted on the zero-offset sections of Route 300 (**Fig 4.18**), the results of which are shown in **Fig 5.3**. From **Fig 5.3e&f** reflected wave polarization generated from the H2 source is more linearly polarized than that generated from the V source.

Figure 5.1: **Comparison for PC1 Zero-offset Sections**

a) PC1 zero offset section obtained from computing the zero-offset sections from each component (V,H1,H2) using the velocity independent technique, then performing PCA on the resulting sections. **b)** PC1 zero offset section obtained by first computing the PC1 for every cmp gather, then carrying out the velocity independent analysis to get a zero offset section.

Figure 5.2: **Route 300 3C Stack.**

Results obtained by adding up zero offset sections of all 3 receiver components for each source orientation, and comparing it with the zero offset section of the primary receiver component (V,V & H2,H2) **a)** V src, V rec. **b)** V src, V+H1+H2. **c)** V src, PC1. **d)** H2 src, H2 rec. **e)** H2 src, V+H1+H2. **f)** H2 src, PC1.

Figure 5.3: Route 300 Zero-offset Projected PC1 with Linearity V-H2 Source Comparison

a) Vsrc PC1 projected data. b) H2src PC1 projected data. c) Vsrc PC1 direction. d) H2src PC1 direction. e) Vsrc Linearity. f) H2src Linearity.

5.1.2 French Hill

In **Fig 5.4** the polarization is calculated for the raw and corrected gathers from the French Hill test.

Fig 5.5 shows how the eigenvalues are distributed before and after shifting the traces. The shifted gather has a stronger 1st principal component eigenvalue and an overall increase in linearity. If anisotropy is considered the primary mechanism for elliptically polarized body waves, applying the time shift correction, will effectively remove this phase shift, converting the elliptical polarization to linear polarization. This is one way to interpret the overall increase in linear polarization.

5.1.3 NRC 9C

CMP gathers from the NRC site are corrected using the velocity independent approach and shown in **Fig 5.6**. This also serves as a quality check for the velocity independent approach. The correction produces coherent gathers, although the earliest arrivals in the vertical receiver are slightly under corrected. From **Fig 5.6d&h&l**, we see the bedrock reflection returns primarily in the horizontal for all source orientations.

Stacking the gathers (using the same processing steps and mute) was done to all the cmp gathers to create zero-offset sections, which were then used for PCA. The results of this analysis are shown in **Fig 5.7**. These sections highlight the influence of the source orientation of the orientation of the reflected seismic waves. Again the bedrock reflection is dominated by horizontal seismic energy, both H1 and H2.

5.1.4 Kaymar

Fig 5.8 shows the section polarization for a Kaymar Drive, from both the V source (**Fig 5.8a&c&e**) and H2 source (**Fig 5.8b&d&f**). Data projected along PC1 yields high quality sections from both sources, but the drainage pipe is much more obvious in the H2 section **Fig 5.8b**. The overall quality of the section is superior from the H2 source, the reflections being more continuous. However an interesting note is the resolution of the bedrock reflection; **Fig**

Figure 5.4: **French Hill Raw-Shifted Polarization Comparison**

French Hill (H2 hammer hit). a) V b) H1 c) H2 d) PC1 e) Linearity f) V corrected g) H1 corrected h) H2 corrected i) PC1 corrected j) Linearity corrected

Figure 5.5: **French Hill Eigenvalues Before & After Shifting**

Comparison of PCA Eigenvalues before (raw) and after shifting **a)** Distribution of Normalized Eigenvalues **b)** Distribution of linearity

Figure 5.6: NRC site PCA on Corrected Gathers

All gathers have been corrected using the time shift calculation within the velocity independent approach. **a)** V-V **b)** V-H1 **c)** V-H2 **d)** Vsrc PC1 **e)** H1-V **f)** H1-H1 **g)** H1-H2 **h)** H1src PC1 **i)** H2-V **j)** H2-H1 **k)** H2-H2 **l)** H2src PC1

Figure 5.7: Section PCA of NRC 9C Survey

The sections are derived from the velocity independent imaging procedure as outlined above. **a)** V-V **b)** V-H1 **c)** V-H2 **d)** Vsrc PC1 **e)** H1-V **f)** H1-H1 **g)** H1-H2 **h)** H1src PC1 **i)** H2-V **j)** H2-H1 **k)** H2-H2 **l)** H2src PC1

5.8a derived from the V source has a much sharper reflection. Like the previous surveys, PC1 is primarily horizontal for the bedrock reflection **Fig 5.8c&d** for both sources. The linearity from the H2 source (**Fig 5.8f**) is higher than the linearity from the V source (**Fig 5.8e**).

To test this idea that shifting the traces aligns the arrivals between components and accounts for the effects of anisotropy, we investigate the linearity distribution on sample gathers from the microvibe sources in **Fig 5.9**. For each site & source the linearity distribution is calculated before and after applying the a hyperbolic correction. For every tested cmp gather, there is a substantial decrease in the linearity in contrast to the test at French Hill. This contradicts the hypothesis that correcting the gathers would align the arrivals between different components and subsequently lead to an increase in the polarization linearity.

5.2 Summary

In this chapter, tools developed over the course of the thesis have been combined to study the polarization of multicomponent surveys on the scale of the survey itself. We find that applying PCA to zero-offset sections is more effective than calculating the zero-offset section from PC1 projected data, the resulting sections from the former method are more constructively combined and more continuous.

Looking at the direction of PC1, we consistently observe the higher prevalence of horizontal energy, particularly cross-line (H2) energy in the bedrock reflection, no matter what source orientation was used.

The sample shot at French Hill showed an increase in the linearity after correcting the gathers assuming a hyperbolic travel time, based on this it was hypothesized that this correction accounted for the effects of anisotropy. However we see the exact opposite effect applying the same analysis to the other surveys, for all the source orientations contradicting this interpretation.

Figure 5.8: Kaymar 6C Zero-offset & PC1

a) Vsrc PC1 projected data b) H2src PC1 projected data c) Vsrc PC1 direction d) H2src PC1 direction e) Vsrc Linearity f) H2src Linearity

Figure 5.9: **Linearity Raw vs Corrected Gather**

Linearity distribution comparison for cmp gathers from each survey (relative abundance in % of samples). **a)** NRC Vsrc **b)** NRC H1src **c)** NRC H2src **d)** Rte300 Vsrc **e)** Rte300 H2src **f)** Kaymar Vsrc **g)** Kaymar H2src **h)** John Shaw Vsrc **i)** John Shaw H1src **j)** John Shaw H2src

Chapter 6

Conclusions

There are two distinct seismic processing tools resulting from this research.

1. Visualization of principal component analysis to investigate the polarization of multicomponent data sets.
2. New time-shift estimation method calculation and velocity independent approach for an alternative and quick processing approach of seismic data.

Principal component analysis allows the polarization states to be investigated on the scale of an entire survey. By mapping the principal components to colours via a stereonet projection, the orientation of the polarization can be viewed, without the use of hodograms. Using this technique the prevalence of the cross-line energy in the bedrock reflection was noted. This observation invites follow-up experiments reviewed in the future work section.

The new method of calculating the local shift (or local slope), the phase-delay calculation, is a useful alternative to existing methods. It was found to be more robust to noise, but more vulnerable to aliasing when compared with Dynamic Time Warping. Based on this observation it is important to have short geophone spacing to avoid temporal aliasing effect. Using this calculation within the velocity independent approach formalized by Fomel (2007a) yielded comparable results to sections processed using the semblance-nmo correction method. The phase-delay benefited and often required shaping regularization to stabilize the calculation.

While it is conceivable to have conducted the polarization analysis without the implementation of the velocity independent approach in the near-surface, this greatly reduces the processing time. Particularly for comparing the effects of source orientation on multi-component sections. With the development of the new local shift calculation, it was found that using the first principal component of the zero-offset section produced a high-quality sections. Specifically the construction a zero-offset section for each component, then conducting PCA yielded a superior quality section compared to creating a zero-offset section from the first principal component projected data. By projecting the data along the first principal component, there is an increase in the section quality for sections derived from the vertical source data, particularly imaging of the bedrock reflection. However for the H2 source data, the advantage of projecting the data along the principal component data is not obvious since the H2-H2 sections are already of such superior quality.

6.1 Future Work

There are many areas left for future work.

1. One intriguing avenue left unexplored in this research is the imaginary part of the phase-delay calculation. The imaginary part is related to change in magnitude of the signal. It is hypothesized that the imaginary part could be useful for characterizing signal attenuation and subsurface properties such as the Q-factor.
2. Only preliminary work was done in deriving the seismic velocities via the velocity independent framework. This warrants further investigation in the near-surface context, specifically in comparing the velocities derived from each of the 3 components. Potentially this could be used to characterize the seismic anisotropy.
3. The new time-shift method might be directly applied to study shear-wave splitting directly between components. This could be done on the exploration scale or on the global seismology scale. Since the time-shift method is general for time-series it seems that there maybe many other fields that the method could be applied.

4. The observation of the prevalence of H2 energy in the bedrock reflections also requires further investigation. Two 9C shots with the streamer oriented in orthogonal directions could be acquired without any modification to the equipment. This would determine whether this observation is related to the subsurface or if it is related to the survey equipment. If the H2 prevalence is related to the equipment, a small modification to how the sleds are mounted on the streamer could be made. Specifically using a short tether to attach each sled to the tow strap. This would mitigate the effects of tension between the sleds, but potentially decrease the positional accuracy of the receivers.
5. As-is the time-shift method is not robust to interfering arrivals, which is also the case for Dynamic Time-warping. Making the method robust to crossing arrivals would be a significant improvement.

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Appendices

Appendix A

Preprocessing

The initial processing steps are summarized in Table A.1 for each data set. Processing has been done with in-house developed software implemented in Python. The software makes use of open source scientific computing libraries; Numpy (Oliphant (2015)) for number crunching routines, Matplotlib (Hunter (2007)) for figures and Obspy (Beyreuther et al. (2010)) for SEG-Y reading and writing.

site	Route 300	French Hill	Kaymar	John Shaw	NRC site
3C Seperation	✓	✓	✓	✓	✓
Geometry Edition	✓	✓	✓	✓	✓
3C AGC	1 second				
Correlation	✓		✓	✓	✓
CMP sorting (shot-spacing binning)	✓		✓	✓	✓
Filter					
Source	[V]	V	[V]	[V]	[V]
Orientations		H1		[H1]	[H1]
	[H2]	H2	[H2]	[H2]	[H2]

Table A.1: **Shear Wave Pre-processing Summary**

Appendix B

S-transform

The s-transform as developed by Stockwell et al. (1996), is useful for analysing frequency content for non-stationary signals. The Fourier Transform can be defined as

$$X(f) = \int_{-\infty}^{+\infty} x(t) \exp(-2\pi i f t) dt \quad (\text{B.1})$$

The s-transform is a modification of this, allowing for the frequency content to be analysed as it changes with time,

$$X(\tau, f) = \int_{-\infty}^{+\infty} x(t) \left(\frac{|f|}{\sqrt{2\pi}} \exp\left(\frac{-f^2(\tau - t)^2}{2}\right) \right) \exp(-2\pi i f t) dt \quad (\text{B.2})$$

Pinnegar (2006) notes that by convolution it may be written in terms of the usual Fourier transform:

$$X(\tau, f) = \int_{-\infty}^{+\infty} X(\alpha + f) \exp\left(\frac{-2\pi^2 \alpha^2}{f^2}\right) \exp(2\pi i \alpha \tau) d\alpha \quad (\text{B.3})$$

Integrating over all τ reduces equation B.2 to equation B.1

$$\int_{-\infty}^{+\infty} X(\tau, f) d\tau = X(f) \quad (\text{B.4})$$

Appendix C

Seismic Anisotropy

Seismic anisotropy stems from the directional dependence of a medium's elastic properties. Instead of a single bulk and shear modulus to characterize a material, more parameters are required. In general every component of stress can be a linear combination of every component of strain:

$$\sigma_{ij} = C_{ijmn}\epsilon_{mn} \quad (\text{C.1})$$

where σ_{ij} is the stress tensor, ϵ_{mn} is the strain tensor, and C_{ijmn} is an 81 component ($3 \times 3 \times 3 \times 3$) tensor of elastic parameters. However only 21 of the components are independent Thomsen (2002). Reindexing the tensors allows the relationship to be written on regular flat paper (the Voigt recipe - (Thomsen, 1986)),

$$\begin{array}{cc|cccccc} \text{ij} & \text{mn} & 11 & 22 & 33 & 32=23 & 31=13 & 12=21 \\ \downarrow & \downarrow & \downarrow & \downarrow & \downarrow & \downarrow & \downarrow & \downarrow \\ \alpha & \beta & 1 & 2 & 3 & 4 & 5 & 6 \end{array}$$

Thomsen (2002) notes that this is no longer a tensor in the reindexed form, but a matrix. Therefore one must specify the coordinate system (which must be aligned with principal stresses) before writing it in the 6×6 form.

$$\sigma_{\alpha} = C_{\alpha\beta}\epsilon_{\beta} \quad (\text{C.2})$$

written explicitly in matrix form,

$$\begin{bmatrix} \sigma_1 \\ \sigma_2 \\ \sigma_3 \\ \sigma_4 \\ \sigma_5 \\ \sigma_6 \end{bmatrix} = \begin{bmatrix} C_{11} & C_{12} & C_{13} & C_{14} & C_{15} & C_{16} \\ & C_{22} & C_{23} & C_{24} & C_{25} & C_{26} \\ & & C_{33} & C_{34} & C_{35} & C_{36} \\ & & & C_{44} & C_{45} & C_{46} \\ & & & & C_{55} & C_{56} \\ & & & & & C_{66} \end{bmatrix} \begin{bmatrix} \epsilon_1 \\ \epsilon_2 \\ \epsilon_3 \\ \epsilon_4 \\ \epsilon_5 \\ \epsilon_6 \end{bmatrix} \quad (\text{C.3})$$

21 independent material parameters are needed to characterize a completely anisotropic material. ($C_{\alpha\beta}$ is symmetric and only the upper triangle is shown). This is the most general case of anisotropy and is related to the triclinic crystal class. There are no physical symmetries within the material. These elastic parameters are referred to throughout the literature on seismic anisotropy.

Fortunately materials often have some physical symmetry, this leads to a simpler stress-strain relation. Any material symmetry results in fewer independent components in the stiffness tensor.

The monoclinic system possesses a symmetry about a plane; mathematically the system must not change under the transformation

$$x_1 \rightarrow x_1, x_2 \rightarrow x_2, x_3 \rightarrow -x_3.$$

Any component possessing an odd number of 3s in the subscripts of C_{ijmn} must change sign Landau and Lifshitz (1959). In the 2 subscript notation this means

$C_{42} \rightarrow -C_{42}$, $C_{51} \rightarrow -C_{51}$, ... etc. Writing this out for all of the components before and after the transformation gives:

$$\begin{bmatrix} C_{11} & C_{12} & C_{13} & C_{14} & C_{15} & C_{16} \\ & C_{22} & C_{23} & C_{24} & C_{25} & C_{26} \\ & & C_{33} & C_{34} & C_{35} & C_{36} \\ & & & C_{44} & C_{45} & C_{46} \\ & & & & C_{55} & C_{56} \\ & & & & & C_{66} \end{bmatrix} = \begin{bmatrix} C_{11} & C_{12} & C_{13} & -C_{14} & -C_{15} & C_{16} \\ & C_{22} & C_{23} & -C_{24} & -C_{25} & C_{26} \\ & & C_{33} & -C_{34} & -C_{35} & C_{36} \\ & & & C_{44} & C_{45} & -C_{46} \\ & & & & C_{55} & -C_{56} \\ & & & & & C_{66} \end{bmatrix} \quad (\text{C.4})$$

So any components where $C_{\alpha\beta} = -C_{\alpha\beta}$ are zero, which leaves 13 independent components to characterize the monoclinic system McNeil and Grimsditch (1993);

$$C_{\alpha\beta} = \begin{bmatrix} C_{11} & C_{12} & C_{13} & & C_{15} \\ & C_{22} & C_{23} & & C_{25} \\ & & C_{33} & & C_{35} \\ & & & C_{44} & C_{46} \\ & & & & C_{55} \\ & & & & & C_{66} \end{bmatrix} \quad (\text{C.5})$$

The hexagonal system has complete rotational symmetry about a single axis, like a cylinder. This popular kind of anisotropy is known as transverse isotropy (TI). In the case of TI, the media is isotropic with respect to a specific plane and has 5 independent elastic parameters. As Thomsen (1986) puts it, "there is one distinct direction that is faster or slower, while the other 2 directions have the same seismic velocity". This type of anisotropy is thought to apply in horizontally layered media. For TI the elastic parameter matrix is

$$C_{\alpha\beta} = \begin{bmatrix} C_{11} & (C_{11} - 2C_{66}) & C_{13} & & & \\ & C_{11} & C_{13} & & & \\ & & C_{33} & & & \\ & & & C_{44} & & \\ & & & & C_{44} & \\ & & & & & C_{66} \end{bmatrix} \quad (\text{C.6})$$

And finally with the most strict sense of symmetry applied, that being where materials parameters must be invariant under any rotation, it yields,

$$C_{\alpha\beta} = \begin{bmatrix} C_{33} & (C_{33} - 2C_{44}) & (C_{33} - 2C_{44}) & & & \\ & C_{33} & (C_{33} - 2C_{44}) & & & \\ & & C_{33} & & & \\ & & & C_{44} & & \\ & & & & C_{44} & \\ & & & & & C_{44} \end{bmatrix} \quad (\text{C.7})$$

which the isotropic case. Going back to index notation and simplifying the expression leads to the familiar form of Hooke's law

$$\sigma_{ij} = \lambda \Delta \delta_{ij} + 2\mu \epsilon_{ij} \quad (\text{C.8})$$

Elastic anisotropy gives rise to seismic anisotropy. A ubiquitous parametrization of seismic anisotropy in exploration seismology was developed by Thomsen (1986). He designed his anisotropy parameterization to meet the following criteria;

- equations should be simpler.
- anisotropic parameters should reduce to zero in the isotropic case.

Assuming the media is isotropic in the horizontal plane, which is expected to be relevant for horizontal media, Thomsen reparameterized the 5 elastic constants to the vertical p-wave

and s-wave velocities and 3 new parameters of anisotropy;

$$\alpha_o = \sqrt{\frac{C_{33}}{\rho}} \quad (\text{C.9a})$$

$$\beta_o = \sqrt{\frac{C_{44}}{\rho}} \quad (\text{C.9b})$$

$$\epsilon = \frac{C_{11} - C_{33}}{2C_{33}} \quad (\text{C.9c})$$

$$\gamma = \frac{C_{66} - C_{44}}{2C_{44}} \quad (\text{C.9d})$$

$$\delta = \frac{(C_{13} + C_{44})^2 - (C_{33} - C_{44})^2}{2C_{33}(C_{33} - C_{44})} \quad (\text{C.9e})$$

He observed that seismic anisotropy is common, but weak; typically the seismic velocity of the fast direction is not that different from the seismic velocity of the slow direction. When the anisotropic parameters are small (in the case of weak anisotropy) the angular dependence of velocities are;

$$V_p(\theta) = \alpha_o (1 + \delta \sin^2 \theta \cos^2 \theta + \epsilon \sin^4 \theta) \quad (\text{C.10a})$$

$$V_{sv}(\theta) = \beta_o \left[1 + \frac{\alpha_o^2}{\beta_o^2} (\epsilon - \delta) \sin^2 \theta \cos^2 \theta \right] \quad (\text{C.10b})$$

$$V_{sh}(\theta) = \beta_o [1 + \gamma \sin^2 \theta] \quad (\text{C.10c})$$

where θ is the angle between the direction of wave propagation and the vertical axis, and V_p, V_{sv}, V_{sh} are the p-wave, vertically polarized shear wave, and horizontally polarized shear wave velocities, respectively.

Appendix D

Amplitude Correction

Applying and visualizing principal component analysis highlighted a group of traces that had a fixed polarization in the H1 direction **Fig 3.2d**. Initially this was interpreted as settling of streamer; just happening to be apart of the example chosen to illustrate the visualization technique. **Fig D.1** shows common offset sections of the polarization. Here it is clear that there are distinct channels that have this H1 dominance in the signal of the whole recording **Fig D.1e** at 27m offset for example. The sum of the signal squared was used as an indication of the energy at each receiver over a 1 second time window, this value is plotted for each shot in the Route 300 H2src survey. Each component has 4 clearly outlying channels. For the V & H2 components, **Fig D.2a** & **Fig D.2c** respectively, the signal is consistently lower for geophones 32-35 (0 starting index). The H1 component, **Fig D.2b**, is consistently higher for these same geophones. **Fig D.3a** shows the averaged value for all the shots. It is clear geophones 32-35 have anomalous sensitivities; V & H2 being insensitive and H1 being over-sensitive compared to the other geophones around this group. A logarithmic fit was used to correct the geophones, adjusting the amplitudes to match the trend from geophones 10-43. **Fig D.4** shows the newly calculated polarization after correcting the amplitudes of the problem geophones. The polarization at offset 27m **Fig D.4e** is corrected.

Figure D.1: **Increasing offset Polarization Uncorrected**

Common offset polarization sections from Route 300 H2 source data. **a)** 6m offset. **b)** 11.25m offset. **c)** 16.5m offset. **d)** 21.75m offset. **e)** 27m offset. **f)** 33m offset. Offset position is marked by a vertical line on the sample shot shown on the right.

Figure D.2: **Route 300 Receiver Energy Map**

Sum of traces squared gives a value proportional to the energy. a) V receiver b) H1 receiver c) H2 receiver.

Figure D.3: **Route 300 Average Receiver Energy**

Sum of traces squared gives a value proportional to the energy. a) Average before correcting receivers b) Average after correcting receivers.

Figure D.4: **Increasing offset Polarization Corrected Amplitude**

Common offset polarization sections from Route 300 H2 source data. Amplitudes from geophones 32-35 have by a scaling factor as in **Fig D.3b**.

a) 6m offset. b) 11.25m offset. c) 16.5m offset. d) 21.75m offset. e) 27m offset. f) 33m offset. Offset position is marked by a vertical line on the sample shot shown on the right.

Appendix E

Sliding Window Covariance

In Chapter 3 time dependent PCA was reformulated for more efficient computation. The mathematics to go from equation 3.3 is reviewed in this appendix. Equation 3.1 is the definition of covariance in index notation:

$$\sigma_{kl}^2 = \frac{1}{N-1} \sum_{i=1}^N (x_k^i - \mu_k)(x_l^i - \mu_l) \quad (\text{E.1})$$

μ_k is the mean of the k-th component;

$$\mu_k = \frac{1}{N} \sum_{i=1}^N x_k^i \quad (\text{E.2})$$

First we expand the term within the summation (**NOTE:** superscripts are indices, **not** exponents):

$$\begin{aligned} \sum_{i=1}^N (x_k^i - \mu_k)(x_l^i - \mu_l) &= (x_k^1 - \mu_k)(x_l^1 - \mu_l) + (x_k^2 - \mu_k)(x_l^2 - \mu_l) + \dots \\ &= x_k^1 x_l^1 - x_k^1 \mu_l - x_l^1 \mu_k - \mu_k \mu_l + x_k^2 x_l^2 - x_k^2 \mu_l - x_l^2 \mu_k - \mu_k \mu_l + \dots \\ &= (x_k^1 x_l^1 + x_k^2 x_l^2 + \dots + x_k^N x_l^N) \\ &\quad - \mu_k (x_l^1 + x_l^2 + \dots + x_l^N) \\ &\quad - \mu_l (x_k^1 + x_k^2 + \dots + x_k^N) \\ &\quad + N \mu_k \mu_l \end{aligned} \quad (\text{E.3})$$

Now if we let

$$q_{kl} = x_k^1 x_l^1 + x_k^2 x_l^2 + \dots x_k^N x_l^N \quad (\text{E.4})$$

and

$$s_k = x_k^1 + x_k^2 + \dots x_k^N \quad (\text{E.5})$$

then

$$\sum_{i=1}^N (x_k^i - \mu_k)(x_l^i - \mu_l) = q_{kl} - \mu_k s_l - \mu_l s_k + N \mu_k \mu_l \quad (\text{E.6})$$

and because $\mu_k = \frac{s_k}{N}$, the equation can be simplified further:

$$\begin{aligned} \sum_{i=1}^N (x_k^i - \mu_k)(x_l^i - \mu_l) &= q_{kl} - 2 \frac{s_k s_l}{N} + \frac{s_k s_l}{N} \\ &= q_{kl} - \frac{s_k s_l}{N} \end{aligned} \quad (\text{E.7})$$

Substituting this result back in equation E.1 we get

$$\sigma_{kl}^2 = \frac{1}{N-1} \left(q_{kl} - \frac{s_k \cdot s_l}{N} \right) \quad (\text{E.8})$$

This is often referred to as one-pass formulation, or the textbook algorithm (Chan et al., 1979). Because q_{kl} and s_k are just summations, it is straightforward to see how a convolution an array of ones the length of the window leads to a time dependent calculation of these parameters:

$$\begin{aligned} s_k(t) &= x_k(t) * w \\ q_{kl}(t) &= [x_k(t) \cdot x_l(t)] * w \end{aligned} \quad (\text{E.9})$$

and so we arrive at equation 3.6:

$$\sigma_{kl}^2(t) = \frac{1}{n-1} \left(q_{kl} - \frac{s_k \cdot s_l}{n} \right) \quad (\text{E.10})$$