TWO-DIMENSIONAL WAVEFORM TOMOGRAPHY OF
THE QUEEN CHARLOTTE BASIN OF WESTERN CANADA AND THE SEATTLE FAULT ZONE

by

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Abstract

Two-dimensional frequency domain visco-acoustic waveform tomography is applied to limited-offset marine seismic reflection data from the Queen Charlotte sedimentary Basin of western Canada, and from the Seattle Fault Zone in Puget Sound, Washington. It was possible to obtain high resolution P-wave velocity and attenuation images of the subsurface, and to practically evaluate the effectiveness of the visco-acoustic waveform tomography method.

A specific data preconditioning and inversion strategy is developed to recover models to a depth of 1.2 to 1.3 km. The preconditioning of the data converts the field data to a form similar to that predicted by the acoustic waveform modelling algorithm. A multiscale inversion strategy was designed to mitigate non-linearity issues and to improve the estimation of attenuation. The starting velocity model is derived from first arrival traveltime tomography and the starting attenuation model is a homogeneous $Q_p$-value.

Four seismic lines in the Queen Charlotte Basin are imaged, and the recovered velocity models aid in interpreting shallow structures such as Quaternary strata and Pliocene faulting. The joint interpretation of the velocity and attenuation models enables the identification of siltstone, shales, the presence of hydrocarbons and seafloor pockmarks. The shallowmost basement rocks are interpreted to be volcanic.

Using a section of the seismic data across the Seattle Fault Zone, synthetic visco-acoustic and visco-elastic modelling was used to verify the effectiveness of applying visco-acoustic waveform tomography to visco-elastic data. The results show that
visco-acoustic waveform tomography of marine seismic reflection data is reliable when high velocity gradients are absent from the model.

Finally, an interpretation is provided for the inversion results across the Seattle Fault Zone. The inverted velocity and attenuation models enable the identification of glacial and post-glacial Pleistocene, Tertiary sedimentary rocks, and Eocene volcanic rocks. Several north-dipping shallow thrust faults, anticlines and a syncline are identified across the Seattle uplift and the Seattle Fault Zone. The orientation of the faults are consistent with the interpretations of the Seattle Fault Zone as either a fault propagating fold with a forelimb breakthrough, or as the leading edge of a triangle zone within a passive roof duplex.

**Keywords:** Waveform tomography; traveltime tomography; visco-acoustic modelling; visco-elastic modelling; velocity; attenuation; geological interpretation
To My Father, Mr. Jean Takougang
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Chapter 1

General Introduction

1.1 Overview

One of the most important aspects of seismic exploration is the accurate estimation of seismic reflectivity and physical properties of the subsurface that can provide useful geological information. P-wave velocity is commonly estimated from seismic data. Ideally, such a velocity should be suitable for its use in pre-stack depth migration, the goal of pre-stack depth migration being to restore to their appropriate subsurface position any reflection event, thus providing a clear image of the reflectivity. The quality of the migration improves when the reference velocity model contains both low and intermediate wavenumbers (Sirgue, 2003). In 1972 when a technique based on the use of curved ray travel times to estimate the seismic velocity distribution between two boreholes was implemented (Bois et al., 1972), several strategies and techniques were developed to derive more accurate velocity models. The earliest methods, based on information (time) from first arrivals, is called travel time tomography. The term tomography, borrowed from the Greek word “tomos” meaning slice or section, simply refers to constructing a picture of the subsurface from seismic data. The most recent method is called waveform tomography. This method is based on the inversion of the waveform itself, which carries
more information about the subsurface. Although this last method is more expensive computationally, complex to implement and subject to greater non linearity, it can provide greater resolution of the subsurface velocity variation than travel time tomography. Other subsurface physical properties such as S-wave velocity, density and attenuation, can in principle also be derived using waveform tomography. The application of waveform tomography to field seismic data constitute the core of this thesis.

In the following, I will provide a brief review of the principle of tomographic velocity estimation.

### 1.1.1 Travel Time Tomography

Travel time tomography is the process of creating a velocity model by projecting travel times from sources to receivers along the ray path. This process is a non linear inverse problem that can be solved by linearization and iteration (Aldridge and Oldenburg, 1993; Zelt and Barton, 1998). In principle, we want to reduce to an acceptable small value, the misfit function, which is related to the travel time residuals. The travel time residuals characterizes the difference between the observed travel times (often simply from the first arrivals) and the calculated ones from forward modelling in the predicted velocity model. A predicted velocity model is obtained after each iteration, starting from an initial guessed velocity model or one based on empirical geological knowledge of the area. The relation between the travel time \( t \) and the slowness in a two-dimensional medium \((x,y)\) is given by the following relation, known as the Eikonal equation (Aki and Richards, 1980):

\[
\| \nabla t \|^2 = s(x,y)^2, \quad (1.1)
\]
where \( s(x, y) \) represents the slowness at a point with spatial coordinates \((x, y)\). Knowing the ray path \( L \) with respect to a reference point, the travel time takes the form:

\[
t = \int_L s(x, y) \, dl.
\] (1.2)

Equation 1.2 represents the calculated travel time from a source to a receiver through the slowness field \( s(x, y) \). The 2D velocity model can be discretized into a set of \( N \) cells \( c_i \) with \( i = 1, \ldots, N \), each characterized by a constant slowness value \( s(x, y) \). The calculated travel time \( t \) results from the addition of the elementary travel times \( t_i \) due to the propagation through each cell intersected by the raypath. The travel time-time difference \( \Delta t \) has the form:

\[
\Delta t = t_{obs} - t_{calc},
\] (1.3)

where \( t_{obs} \) is the observed travel time and \( t_{calc} \) is the calculated time from forward modelling in the initial or updated model. \( \Delta t \) is a matrix containing a collection of travel time differences \( \Delta t \) due to wave propagation from many sources and many receivers. It is not possible to update the slowness field by reducing the traveltime differences \( \Delta t \) in a single step, because the problem is non linear. The ray path matrix is usually large and non square due to the acquisition geometry, and the problem might be inconsistent due to random noise in the observed data (Aldridge and Oldenburg, 1993). The strategy (as illustrated in figure 1.1) consists of reducing \( \Delta t \) by a small quantity \( \delta t \) at each iteration.
Figure 1.1: Illustration of the reduction of the data difference $\Delta t_i$ ($\Delta t_i$), with $\delta t_i$ ($\delta t_i$) at each iteration $i$.

The change in time $\delta t$ along the ray path $l(x,y)$ can be used to calculate a perturbation $\delta s(x,y)$ in the slowness field such that (Aldridge and Oldenburg, 1993):

$$\delta t = \Delta t_{perturb} = \sum_{x,y} \delta s(x,y)\delta l(x,y).$$

Equation 1.4 is non linear because $\delta l(x,y)$ implicitly depends on $\delta s(x,y)$. The linearization consists of assuming that the perturbation is small enough such that the ray path in each elementary cell is unchanged in the reference model. Taking $m$ to be the slowness field ($m$ is a matrix whose elements are the slowness values at each cell), and $\Delta m$ a perturbation to the slowness field, then for a collection of many sources and receivers, equation 1.4 takes the form:

$$\Delta t_{perturb} = A(m)\Delta m,$$

where $A(m)$ is a matrix whose elements are the raypath length segments within each cell.

The data residual is expressed as

$$\Delta t_{res} = \Delta t - \Delta t_{perturb} = \Delta t - A(m)\Delta m,$$
and its norm $\Phi$ is the misfit or the cost function,

$$\Phi = \|(A\Delta m - \Delta t)\|^2. \quad (1.7)$$

The minimization of this cost function can be computed iteratively using a finite difference schema (Vidale, 1988) and following the steepest descent direction (Aldridge and Oldenburg, 1993) toward an improved slowness field $m + \Delta m$. In order to limit the non-uniqueness of the solution, a form of regularization is applied. This is particularly important when the data are noisy, and the system of equations is ill-conditioned (Aldridge and Oldenburg, 1993). The regularization can be done in several ways: Bishop et al. (1985) and Bregman et al. (1989) proposed weighting the model perturbation, whereas Lytle and Dines (1980) introduced a Laplacian smoothing operator when calculating the model perturbation. Aldridge and Oldenburg (1993) proposed the application of linear equality constraints to the slowness model rather than to a model perturbation, since the model itself is the desired result. After applying the regularization as suggested by Aldridge and Oldenburg (1993), the cost function now takes the form:

$$\Phi = \|W(A\Delta m - \Delta t)\|^2, \quad (1.8)$$

with $W$ the weighting matrix containing the regularizing parameters.

Travel time tomography is a robust method, easy to implement, and relatively cheap computationally. It has been successfully applied in a variety of geological areas at shallow and crustal depth, e.g., Calvert et al. (2003), Takahashi et al. (2007), Hayward and Calvert (2007) and Jaiswal et al. (2008), and was successfully used to monitor $CO_2$ gas injection in enhanced oil recovery (Mathisen et al., 1995).
1.1.2 Waveform Tomography

Waveform tomography is a method very similar to pre-stack migration, but more complete in the sense that the model obtained provides both qualitative and quantitative information. The similarity to pre-stack depth migration is expressed in the way the final model is constructed; however, waveform tomography employs the first arrival waveforms which include transmitted waves at wide-angle. The back propagated data residual is iteratively correlated (multiplied in the frequency domain) with the forward propagated wavefield (Tarantola, 1984; Mora, 1987; Pratt and Worthington, 1990; Crase et al., 1992) recovering both low and high wavenumbers of the model under reconstruction ($V_s$, $V_p$, $Q$, density...). This idea was first recognized in the 1980’s by Tarantola (1984), and constituted an important step in the implementation of the method. The approach was then applied numerically in the time domain by Mora (1987) and by Crase et al. (1992). Later, Pratt and Worthington (1990, 1988) showed that waveform tomography can be implemented in both the time domain and the frequency domain. The implementation in the time domain is convenient when time windowing the residual is important (Shipp and Singh, 2002). Time windowing of the residuals can also be implemented in the frequency domain with the use of complex frequencies (Mallick and Fraser, 1987). Also, preconditioning the gradient in the frequency domain is equivalent to using a single sinusoidal component in the time domain (Sirgue and Pratt, 2003). In fact, the frequency domain and the time domain inversions are equivalent when all the frequencies are inverted at the same time (Pratt and Worthington, 1988). However, the implementation in the frequency domain is less computationally expensive in 2D than the time domain method, and can be used in a multiscale approach consisting of using frequencies or groups of frequencies sequentially from the lowest frequency to the highest frequency (Pratt and Worthington, 1988; Pratt et al., 1996; Pratt, 1999), thus helping to mitigate the non-linearity of the problem. When all the available frequencies are inverted sequentially,
the method is called full waveform tomography (Pratt and Worthington, 1990; Pratt, 1999; Benders and Pratt, 2007a). Another variant of this approach, named efficient waveform tomography was investigated by Sirgue and Pratt (2004) and Yokota and Matsushima (2004), and consists of the inversion of few carefully selected frequencies, which aims to reduce the redundancy in the wavenumber coverage with the advantage of a significant reduction of computational cost. This strategy, mathematically derived from a 1D velocity model, has been successfully applied to 2D synthetic data (Sirgue and Pratt, 2004; Yokota and Matsushima, 2004; Benders and Pratt, 2007b) and field data (Takam Takougang and Calvert, 2011). However, the efficient waveform tomography approach provides images with lower resolution than the full waveform tomography approach (Benders and Pratt, 2007b; Takam Takougang and Calvert, 2011). When the data are noisy the best approach remains full waveform tomography, which suppresses the effect of noise in the data by inverting groups of frequencies sequentially.

The advantages of waveform tomography over travel time tomography are:

1. **Inclusion of more information from the recorded data**

   Since waveform tomography is based on the seismic waveform, it utilizes both the amplitude and phase of the propagating wave. This information includes reflections, refractions, diffractions and other scattered arrivals. In travel time tomography we only use the first arrivals.

2. **Greater resolution**

   The resolution $r_1$ of waveform tomography is of the order of the wavelength $\lambda$ of the propagated wavefield (Wu and Toksöz, 1987; Sirgue and Pratt, 2004):

   $$ r_1 = \lambda. \quad (1.9) $$

   On the other hand, the resolution $r_2$ of travel time tomography is limited to the width
of the first Fresnel zone (Williamson, 1991):

\[ r_2 = \sqrt{\lambda L}, \]  

(1.10)

with \( L \) the maximum offset. This implies that for a given frequency of say 15 Hz, if the velocity of our model is 3000 ms\(^{-1}\), with a maximum offset of 4000 m, then the resolution of waveform tomography is of the order \( r_1 = 200 \) m, whereas the resolution for travel time tomography is \( r_2 = 894.4 \) m. This shows that the resolution of waveform tomography is 4.5 times higher than that of travel travel tomography at 15 Hz, and the difference will become greater as the frequency increases. The reason for this difference is the infinite frequency assumption in travel time tomography, which results in its relatively low resolution (Nolet, 1992).

Due to its ability to provide high resolution images of the subsurface, waveform tomography can be used to complement subsurface interpretation using more conventional techniques such as seismic migration, especially in areas where image distortions due to gas and complex geology make the imaging process using more conventional techniques difficult. Waveform tomography can therefore have a profound impact in the hydrocarbon industry.

Unfortunately, the abundance of information used in the waveform inversion leads to its high non-linearity. A number of key points are necessary for successful application of waveform tomography to field data. These include: a good preparation of the data, a good starting model, a robust inversion strategy, and the use of very low starting frequencies if available. When low starting frequencies are not available, an efficient inversion strategy and a very good starting model must be selected to ensure sufficient convergence to the global minimum (Sirgue and Pratt, 2003; Takam Takougang and Calvert, 2010, 2011). Traveltime tomography can be used to get a good starting model (Pratt, 1999; Ravaut et al.,
2004; Brenders and Pratt, 2007a; Malinowski and Operto, 2008; Takam Takougang and Calvert, 2011) with the general requirement that it must predict the first arrival wavefield to within half a cycle (Sirgue and Pratt, 2004). Therefore, waveform tomography cannot be seen as a replacement for travel time tomography, but as an additional tool for higher resolution imaging.

A common approach to reduce the computational cost of waveform tomography consists of simplifying the visco-elastic wave-equation to acoustic or visco-acoustic approximation. The application of waveform tomography using the elastic or visco-elastic wave equation is ideal because it corresponds to the true nature of the seismic waves, but it is more complicated due to a greater non-linearity and greater computational cost (Crase et al., 1992; Minkoff and Synes, 1997; Shipp and Singh, 2002). Implementations of the acoustic wave-equation have been used successfully in a variety of synthetic studies, from crosshole tomography to seismic reflection and refraction surveys (Pratt, 1999; Wang and Rao, 2006; Song and Williamson, 1995; Pratt and Worthington, 1990; Pratt et al., 1996; Hicks and Pratt, 2001; Sirgue and Pratt, 2004; Shin and Ming, 2006; Brenders and Pratt, 2007a) as well as with field data (Ravaut et al., 2004; Pratt et al., 2004; Operto et al., 2006; Malinowski and Operto, 2008; Bleibinhaus et al., 2008; Dessa et al., 2004; Chironi et al., 2006; Hicks and Pratt, 2001; Wang and Rao, 2009; Takam Takougang and Calvert, 2011).

The equation of motion of a seismic wavefield \( u \) excited by a body force \( f \) and propagating in an elastic medium with a density \( \rho \) can be written as (Aki and Richards, 1980):

\[
\rho \ddot{u} = f + (\lambda + 2\mu) \nabla(\nabla \cdot u) - \mu \nabla \times (\nabla \times u),
\]

where \( \lambda \) and \( \mu \) are the Lame parameters with which:

\[
\nu_p^2 = \frac{\lambda + 2\mu}{\rho},
\]
and
\[ v_s^2 = \frac{\mu}{\rho}. \]  

(1.13)

The displacement field \( \mathbf{u} \) can be written in terms of vector and scalar potentials (\( \psi \) and \( \phi \)) using the Helmholtz equation such that:
\[ \mathbf{u} = \nabla \phi + \nabla \times \psi. \]  

(1.14)

\( \nabla \phi \) and \( \nabla \times \psi \) are called the P-wave and S-wave components of \( \mathbf{u} \), respectively. Taking the divergence of \( \mathbf{u} \) yields:
\[ \nabla \cdot \mathbf{u} = \nabla \cdot (\nabla \phi + \nabla \times \psi) = \nabla^2 \phi, \]  

(1.15)

and the curl of \( \mathbf{u} \) yields:
\[ \nabla \times \mathbf{u} = \nabla \times \nabla \times \psi = -\nabla^2 \psi. \]  

(1.16)

The acoustic approximation is based on the assumption that the data are dominated by unconverted P-waves. This implies that the following condition should be satisfied:
\[ \nabla \cdot \mathbf{u} \gg \nabla \times \mathbf{u}. \]  

(1.17)

This condition (equation 1.17) may not always hold because the proportion of S-wave in the data can be significant. Therefore, depending on the type of data used, and the nature of rocks encountered, the acoustic approximation can provide reliable results in some cases but can also fail in other cases. Mulder and Plessix (2008) and Barnes and Charara (2008, 2009) have shown with synthetic marine dataset that, in the case where S-waves are significant in the data, acoustic waveform tomography provide reliable results only when the S-wave velocity profile is smooth. When S-waves are absent, if the Amplitude Versus Offset (AVO) effect due to S-conversion is significant, acoustic waveform tomography
leads to reliable results only for the near offset data. This implies that in the case of limited offset marine seismic reflection surveys, if the data are dominated by P-waves, successful application of acoustic or visco-acoustic waveform tomography can be envisaged.

1.1.3 Theory of Acoustic Waveform Inversion

The following describes the principle of non-linear waveform inversion in the frequency domain, based on acoustic wave propagation using the iterative gradient method (Wu and Toksöz, 1987; Pratt and Worthington, 1990; Pratt et al., 1998). Consider a wave propagating in a medium with a constant density $\rho$. If the source is an impulse, located at a vector position $s$, then the propagation of a wave $u_o(s, x, \omega)$ from the position $s$ to a position $x$ in the medium can be expressed as:

$$\nabla^2 u_o(x, s, \omega) + \frac{\omega^2}{c_o^2(x)} u_o(x, s, \omega) = -\delta(x - s),$$

(1.18)

where $c_o(x)$ is the velocity of the medium and $\omega$ is the angular frequency. Since the source is a delta function, the solution of equation 1.18 is a Green’s function. We can write,

$$u_o(x, s, \omega) = G(x, s, \omega).$$

(1.19)

For a more general source function $f(x, \omega)$ representing the spatial distribution of source energy in the model, equation 1.18 becomes:

$$\nabla^2 u_o(x, \omega) + \frac{\omega^2}{c_o^2(x)} u_o(x, \omega) = -f(x, \omega),$$

(1.20)
and the solution of equation 1.20 is therefore the convolution of the source function and the Green function:

$$u_o(r, s, \omega) = \omega^2 \int d^3 x G(r, x, \omega) f(x, \omega).$$  (1.21)

Now from equation 1.18 suppose that we want to upgrade the background velocity from $c_o(x)$ to $c(x)$ with a perturbation velocity $o(x)$ such that,

$$c^{-2}(x) = c_o^{-2}(x) + o(x).$$  (1.22)

The wavefield will also be upgraded from the initial one $G(x, s, \omega)$ to $u(x, s, \omega)$ due to a scattering field $u_{sc}(s, x, \omega)$ such that:

$$u(x, s, \omega) = G(x, s, \omega) + u_{sc}(x, s, \omega).$$  (1.23)

The wave equation for the upgraded wavefield $u(x, s, \omega)$ will take the form

$$\nabla^2 u(x, s, \omega) + \frac{\omega^2}{c^{-2}(x)} u(x, s, \omega) = -\delta(x - s).$$  (1.24)

Plugging in equations 1.22 and 1.23 into 1.24 yields:

$$\nabla^2 u_{sc}(x, s, \omega) + \frac{\omega^2}{c_o^2(x)} u_{sc}(x, s, \omega) = -\omega^2 o(x) u(x, s, \omega).$$  (1.25)

The solution of equation 1.25 is the convolution of the Green function $G(r, x, \omega)$, with the source function $-\omega^2 o(x) u(x, s, \omega)$ and has the form:

$$u_{sc}(r, s, \omega) = -\omega^2 \int d^3 x G(r, x, \omega) o(x) u(x, s, \omega).$$  (1.26)
where the introduction of the variable $\mathbf{r}$ represents the receiver locations at which the wavefield is recorded. This solution is non linear as $G(\mathbf{r}, \mathbf{x}, \omega)$ implicitly depends on $o(\mathbf{x})u(\mathbf{x}, \mathbf{s}, \omega)$. Linearizing the solution consists of using the first Born approximation, which approximates the wave field $u(\mathbf{x}, \mathbf{s}, \omega)$ by the background field $G(\mathbf{x}, \mathbf{s}, \omega)$, similar to the linearization for travel time tomography when assuming a constant ray path. The solution now takes the form:

$$u_{sc}(\mathbf{r}, \mathbf{s}, \omega) = -\omega^2 \int d^3x G(\mathbf{r}, \mathbf{x}, \omega) o(\mathbf{x}) G(\mathbf{x}, \mathbf{s}, \omega).$$

(1.27)

Equation 1.27 shows that the scattered field is obtained by multiplying the forward propagating wavefield or incident field $G(\mathbf{x}, \mathbf{s}, \omega)$ from the source to the scattering point, by the backward propagating wavefield $G(\mathbf{r}, \mathbf{x}, \omega)$ from the scattering point to the receiver, after interacting with the scattering potential $o(\mathbf{x})$. $o(\mathbf{x}) = \delta c^2(\mathbf{x})$ is obtained from equation 1.22. The expression of the wave equation (equation 1.20) can be simplified in matrix form (after Pratt, 1999) as:

$$Su = f,$$

(1.28)

where $S$ represents the wave equation operator or impedance matrix, $f$ the source wavefield and $u$ the pressure wavefield. Equation 1.28 implicitly depends on the acoustic model parameters $m$. The pressure field, $u$, can thus be calculated by solving this system of linear equations:

$$u = S^{-1}f.$$

(1.29)

The data residual $\delta \mathbf{d}$ is defined as the difference between the calculated wavefield $u$ and the observed wavefield $\mathbf{d}$:

$$\delta \mathbf{d} = u - \mathbf{d}.$$  

(1.30)
The objective of the inversion is to find \( m \) such that \( \delta d \) is a minimum. This is equivalent to minimizing the misfit \( E \) which is defined using the \( L_2 \) norm as:

\[
E = \frac{1}{2} \delta d^T \delta d, \tag{1.31}
\]

where \( d^T \) is the conjugate transpose of the data residual.

The minimization of the misfit function can be achieved by computing the gradient of the misfit function with respect to a perturbation to the model parameter \( m \), following the steepest descent direction. The gradient has the form:

\[
G_m = \frac{\partial E}{\partial m} = \nabla_m E = J^T \delta d, \tag{1.32}
\]

where \( J \) is the matrix of Fréchet derivatives,

\[
J = \frac{\partial S}{\partial m} = S^{-1} F. \tag{1.33}
\]

\( F \) is a matrix with all virtual sources in its columns as described in (Pratt et al., 1998). Substituting equation 1.33 into equation 1.32 yields

\[
G_m = F^T \left[ S^{-1} \right]^T \delta d = F^T v, \tag{1.34}
\]

where \( v = \left[ S^{-1} \right]^T \delta d = S^{-1} \delta d^* \). Expanding equation 1.34 gives

\[
G_m = - \left\{ u^T \left[ \frac{\partial S}{\partial m} \right] v \right\} = - \left\{ u^T \left[ \frac{\partial S}{\partial m} \right] S^{-1} \delta d^* \right\}, \tag{1.35}
\]

which is similar to the “adjoint state” method. Details of this method can be found in (Pratt et al., 1998). In summary, the gradient is calculated in two steps:

1. The backpropagated wavefield is computed using the impedance matrix \( S \) of
equation 1.28 and the complex conjugate of the data residual is used as the source.

2. The backpropagated wavefield is then multiplied by the virtual sources generated by the estimated wavefield \( u \).

The calculation of the gradient therefore involved forward propagation of the wavefield, backpropagation of the residual wavefield, and a multiplication of both wavefields. The gradient is used to update the model such that if at iteration \( n \) the model is \( m^n \), then at iteration \( n + 1 \) the model would be:

\[
m^{n+1} = m^n - \alpha^n G_m^n, \tag{1.36}
\]

where the step length \( \alpha \) can be computed using the line search method or the linear estimate method (Tarantola, 1984; Mora, 1987),

\[
\alpha = \frac{|| \nabla mE ||^2}{|| J \nabla mE ||^2}, \tag{1.37}
\]

where \( || \) represents the Euclidean length of the vectors. The iterations are then repeated until a chosen convergence criteria is achieved.

For visco-acoustic forward modelling, attenuation can be introduced using complex velocities (Song and Williamson, 1995; Hicks and Pratt, 2001). In this case, dispersion is required to keep the propagation causal. The complex phase velocity takes the following expression (Aki and Richards, 1980):

\[
v(f) = v_o \left[ 1 + \frac{1}{\pi Q} \ln \left( \frac{f}{f_o} \right) - \frac{i}{2Q} \right], \tag{1.38}
\]

with \( f \) the modelling frequency, \( f_o \) a reference frequency, \( v_o \) the velocity at the reference frequency and \( Q \) the quality factor, with \( Q^{-1} \) the attenuation.
1.2 Thesis Objectives

Although the Queen Charlotte Basin (QCB) and the Seattle Fault zone (SFZ) have been thoroughly studied before with seismic experiments (Rohr and Dietrich, 1992; Dietrich, 1995; Woodsworth, 1991; Johnson et al., 1994, 1996; Brocher et al., 2004; Calvert et al., 2003), none of these studies have used full waveform tomography. For instance the works of Rohr and Dietrich (1992) and Dietrich (1995) on the QCB were based on migrated seismic reflection data and those from Johnson et al. (1994, 1996), Brocher et al. (2004) and Calvert et al. (2003) on the SFZ were based on migrated seismic data, magnetics data, gravity data and travel time tomography.

The objectives of this research project are the application of the 2D frequency domain visco-acoustic waveform tomography (Pratt, 1999) to limited offset marine seismic reflection data from the Queen Charlotte Basin (QCB) offshore British Columbia, and the Seattle Fault Zone (SFZ) in Washington to obtain high resolution P-wave velocity and attenuation images, which can be used for a better understanding of the subsurface geology, especially in areas were unambiguous interpretation of conventional migrated data is difficult. More specifically, the elements of this project comprise:

1. Study of the QCB

The QCB is the largest Tertiary basin on the west coast of Canada, with an area of approximately 80,000 km$^2$ (Whiticar et al., 2003), and has the greatest hydrocarbon potential in westernmost Canada, when compared to other sedimentary basins in the area such as the Tofino basin and the Nechako basin (Higgs, 1991; Dietrich, 1995). The geological structure of the basin offshore has been intensively documented (Rohr and Dietrich, 1992; Woodsworth, 1991; Whiticar et al., 2003; Lyatsky, 1993); the geological information is mainly derived from petroleum-exploration wells drilled in the 1960s and from seismic reflection surveys. However, much of the geological
character of the basin, especially that related to hydrcarbon, remains poorly known. Waveform tomography can be particularly useful in providing valuable additional information.

I used four seismic lines 88-04, 88-05, 88-06 and 88-07 from the seismic reflection data collected in 1988 by the Geological Survey of Canada (GSC) to image the basin to a depth of 1.2 km from Hecate Strait to Dixon Entrance using the velocity and attenuation models derived from waveform tomography. The top 1.5 km of the QCB corresponds to the Tertiary rocks of the Skonun formation made up of interbeded sandstones, shales, conglomerates and lignites and the Masset formation of basalt and volcanicleastics (Dietrich, 1995). The higher resolving power of waveform tomography enables a more precise structural interpretation within the Skonun formation, which may constitute in places an excellent cap rock for hydrocarbon reservoirs, due to its high impermeability (Lyatsky, 1993).

2. Study of the SFZ

The SFZ is a zone of multiple thrust faults located beneath the greater Seattle metropolitan and other densely populated areas. It constitutes one of the greatest seismic hazards along the Pacific coast with a M7 earthquake that occurred approximately 1,100 years ago (e.g., Bucknam et al., 1992, 1999). In 1998, marine seismic reflection data were collected beneath Puget Sound as part of the Seismic Hazard Investigation Program in Puget Sound (SHIPS) to better constrain the structural interpretation of the fault zone. Oil industry seismic data (Johnson et al., 1994; Pratt et al., 1997) as well as high resolution lines collected by the U.S. Geological Survey (USGS) were also used for the same purpose (Johnson et al., 1999). The interpretation of structures within the SFZ has, however, remained ambiguous due to the absence of clear reflections in the area and the complexity of the geology. Consequently, different interpretations of the fault zone have been
proposed (Johnson et al., 1994; Calvert et al., 2003; Pratt et al., 1997; Brocher et al., 2004; Liberty and Pratt, 2008).

I applied visco-acoustic waveform tomography to a section of line PS-2 from the SHIPS survey, across the SFZ, to obtain high resolution velocity and attenuation images to a depth of 1.2-1.3 km. By combining the derived velocity and attenuation images with the migrated seismic section, I provide a joint interpretation, and discuss possible geometries for the fault zone.

3. Inversion strategy and Resolution Analysis

The seismic data used in this study are characterized by a relatively high starting frequency (4.4 Hz for the SFZ and 7 Hz for the QCB) and limited offset (2575 m for the SFZ and 3770 m for the QCB). In my project, I aim to find a strategy for the successful application of waveform inversion to such data. The absence of low frequencies increases the non-linearity of the inversion problem, and the field survey also has difficulties illuminating deeper structures due to its limited offset. This type of data is very common in seismic exploration, including many marine surveys. Low frequencies are often removed during acquisition, because they can be contaminated by various types of noise. I developed a specific inversion strategy to mitigate the non-linearity and boost the contribution of deeper structures to the inversion.

Due to its complexity, the SFZ constitutes an excellent area to test the applicability of visco-acoustic waveform tomography to the visco-elastic field data. I performed both visco-acoustic and visco-elastic modelling using velocity and attenuation models derived from the SFZ data by waveform tomography, in order to compute a set of checkerboard tests to practically check the validity of the application of visco-acoustic waveform tomography to visco-elastic field data.
1.3 Thesis Outline

This thesis can be divided into three main parts, each of which is an independent manuscript published or submitted to peer-reviewed journals.

The first manuscript focuses on developing a methodology to apply successfully visco-acoustic waveform tomography to offset-limited seismic reflection data. In this part, I only used a section of line 88-06 from the QCB to test and develop my inversion methodology. I performed the inversion using two frequency selection approaches: efficient waveform tomography and full waveform tomography. Both approaches converged to similar results showing the robustness of the inversion strategy used, but the efficient waveform tomography approach generated images with lower resolution. For this reason, only the full waveform tomography approach was used in the second and third manuscripts. The first manuscript was published in the journal “Geophysics” in April 2011.

The second manuscript consists of applying the methodology derived in the first part of this thesis to four seismic lines from the QCB dataset: 88-04, 88-05, 88-06 and 88-07. These four lines cover the basin from Hecate Strait to Dixon Entrance. The inversion strategy developed in the first manuscript was refined to image more adequately shadow zones (or zones of relatively weak seismic amplitude) encountered on some of the lines, and to better estimate attenuation values. This second manuscript focuses mainly on the geological interpretation of the velocity and attenuation models, and has been accepted for publication to “Geophysics” with minor revision in August 2011.

The last manuscript can be divided into two parts. First, I applied the refined methodology and inversion strategy developed in the first and second manuscripts to image the SFZ using a section of line PS-2, and with the recovered velocity and attenuation models, I performed a set of visco-acoustic and visco-elastic checkerboard
tests to test the reliability of visco-acoustic waveform tomography applied to elastic or visco-elastic data. Finally, I provide a geological interpretation of the result by combining the velocity, attenuation and migrated seismic sections. This last manuscript was submitted to “Journal of Geophysical Research”.

1.4 Methodology

The methodology for acoustic or visco-acoustic waveform inversion can be divided into two major parts:

1. Data preconditioning

2. Inversion

1.4.1 Data Preconditioning

A 2D acoustic or visco-acoustic propagation assumption is used for forward modelling in the waveform tomography inversion process. Since the propagation of seismic waves is in reality visco-elastic, in 3D, and always contaminated by various types of noise, a proper conditioning of the field data is necessary prior to its use in the inversion. The preconditioning of the data consists of:

1. Quality control and trace editing

This includes a careful inspection of the traces for the detection and removal of those contaminated by noise. What I mean by noise is all events that do not form part of the acoustic propagation scheme. These events include the ground roll or interface waves, converted S waves, plus coherent and incoherent noise related to the instruments and the recording process.
2. **Amplitude correction**

Due to 3D geometric spreading and attenuation, the amplitudes of the field data decrease more rapidly with time and offset than those modelled during the inversion. It is therefore necessary to apply a correction to the field data amplitudes to account for this difference. One way to correct for 3D geometric spreading, is to multiply the field data by \( \sqrt{t} \) (where \( t \) is the time) so that the decrease of amplitudes with time will be similar to that in 2D. Also, Ravaut et al. (2004) and Operto et al. (2006) have suggested a normalization of amplitudes to compensate for shot to shot energy variation during the acquisition, giving the same weight to every trace. However, this method will also remove the AVO information in the data. I applied an artificial match of the AVO behaviour of the real data to the modelled data obtained after forward modelling in 2D, using the starting model, following the methodology explained in Brenders and Pratt (2007a,b).

3. **Bandpass filtering and muting**

A bandpass filter of the data is designed to incorporate only the frequencies that will be used during the inversion. The mute operator is applied below the first break with a length that is equivalent to a few cycles of the dominant frequency. The data are muted to eliminate all the reflections that originate outside the limit of the model. A small window of data is used at the beginning of the inversion process, to encourage stability and ensure convergence to the global minimum. The window is increased in later stages to incorporate more detail in the model (Pratt et al., 2004; Brenders and Pratt, 2007a; Takam Takougang and Calvert, 2011).

Once the time domain data are well preconditioned, they are converted by Fourier transform to the frequency domain for use in the inversion.
1.4.2 The Inversion

An initial velocity model is necessary to start the inversion process in waveform tomography. This velocity model is critical for the success of the inversion. A poor initial model will lead to large errors in modelling first arrival times, and a failure of the inversion due to cycle skipping. A good initial model should contain enough low wavenumbers to ensure a continuous coverage, leading to convergence toward the global minimum. This initial model is specially critical when the starting frequency is relatively high, as it should contain all the wavenumbers corresponding to the missing low frequencies. I used travel-time tomography to get a good initial velocity model and I verified each time that it met the convergence criteria. The source function is obtained in the frequency domain following the method described by Pratt (1999), and consists of scaling with a complex value $s$ the initial wrong estimate of the source. The scaling factor $s$ is expressed as:

$$s = \frac{u'd^*}{u'u^*}, \tag{1.39}$$

with $u$, $u^*$ and $u'$ respectively the forward modelled wavefield, its complex conjugate and its autocorrelation; $d^*$ is the conjugate of the observed wavefield. The method converges in one iteration and provides a good estimate for both the amplitude and the phase of the source function. For visco-acoustic modelling, there can be a trade-off between the estimated attenuation value and the estimate of the source. A starting attenuation value very different from the real subsurface seismic attenuation may lead to some of the effects of attenuation being included in the source signature estimate.

Having completed these steps, the inversion of the data can be started. The algorithm of the inversion can be summarized as follows:

1. Computation of the Green function for each source location

2. Estimate of the source function
3. Computation of the propagated wavefield by convolving the source and the Green function

4. Data residual and cost function calculation

5. Back-propagation of the residual

6. Computation of the cost function gradient by scaled multiplication of the forward and back-propagated wavefield. The function of the scale factor is to convert the migration-like image into a true model update, by among other things, correcting the units.

7. Search for the step length $\alpha$ and model perturbation

8. Model update

The validation of the final velocity and attenuation models is done by comparing the match between the velocity and attenuation models with the available well log information. It can also be done by comparison of the synthetic data obtained by forward modelling through the velocity and attenuation models with the field data, and by comparing the velocity model with the migrated section.
Reference List


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Chapter 2

Application of waveform tomography to marine seismic reflection data from the Queen Charlotte Basin of western Canada

A version of this manuscript was published as:

2.1 Abstract

To obtain a higher resolution quantitative P-wave velocity model, two-dimensional waveform tomography was applied to seismic reflection data from the Queen Charlotte sedimentary Basin off the west coast of Canada. The forward modeling and inversion were implemented in the frequency domain using the visco-acoustic wave equation. Field data preconditioning consisted of F-K filtering, 2D amplitude scaling, shot-to-shot amplitude balancing, and time windowing. The field data were inverted between 7 Hz and 13.66 Hz, with attenuation introduced for frequencies $\geq 10.5$ Hz to improve the final velocity model; two different approaches to sampling the frequencies were evaluated. The
limited maximum offset of the marine data (3770 m) and the relatively high starting frequency (7 Hz) were the main challenges encountered during the inversion. An inversion strategy that successively recovered shallow to deep structures was designed to mitigate these issues. The inclusion of later arrivals in the waveform tomography resulted in a velocity model that extends to a depth of approximately 1200 m, twice the maximum depth of ray coverage in the ray-based tomography. Overall, there is a good agreement between the velocity model and a sonic log from a well on the seismic line, as well as between modeled shot gathers and field data. Anomalous zones of low velocity in the model correspond to previously identified faults, or their upward continuation into the shallow Pliocene section where they are not readily identifiable in the conventional migration.

2.2 Introduction

Waveform tomography aims to provide higher resolution images of the subsurface than ray-based tomography. These images are mostly expressed in terms of P-wave velocity and sometimes attenuation. In principle, density, S-wave velocity and anisotropy, if well constrained can also be estimated. Essentially, waveform tomography focuses on the extraction of information from first arrivals and later phases in the recorded seismic wavefield. The resolution of waveform tomography is of the order of the wavelength of the propagated energy (Wu and Toksöz, 1987; Sirgue and Pratt, 2004), whereas that of a more conventional imaging technique such as curved ray travel time tomography is limited to the width of the first Fresnel zone (Williamson, 1991). This means that waveform tomography has the potential to provide a higher resolution image as the frequency of the propagated signal increases. Image reconstruction is carried out by correlating the forward propagated wavefield with the back propagated data residual in a
fashion very similar to prestack depth migration (Tarantola, 1984; Mora, 1987; Crase et al., 1992; Pratt and Worthington, 1990), recovering both low and high wavenumbers of a velocity model. The method can be viewed as a combination of tomographic and migration-like imaging techniques; low wavenumbers are provided by the tomographic-like inversion which is controlled by interval velocity and high wavenumbers are derived from a migration-like reconstruction of velocity contrasts (Mora, 1989).

Apart from the computational cost, one difficulty in the implementation of waveform tomography is its greater non-linearity than ray-based tomography methods due to the abundance of information in the recorded waveform used. This non-linearity is in part a consequence of the inaccurate recovery of low wavenumbers, as their convergence rate is slower than that of high wavenumbers (Sirgue, 2003). Low wavenumbers are essential for the linearisation of the problem and they contribute to the stabilization of the inversion. The Born approximation, which forms the basis of many waveform inversion methods, works best when the low wavenumbers are recovered first (Pratt et al., 1996). One way to take advantage of this issue is to implement the algorithm in the frequency domain, which allows the inversion to proceed frequency by frequency, from low to high values, resulting in the recovery of initial wavenumbers first (Pratt, 1999; Pratt and Worthington, 1988). Frequency domain inversion also has the advantage of being less computationally expensive in 2D than time domain methods. Dispersive phenomena and associated frequency dependence can be easily implemented in the frequency domain at little extra computational cost, with attenuation and dispersion introduced with complex velocity (Song and Williamson, 1995).

The starting model is critical to reducing the non-linearity of the inversion, because it provides the low wavenumbers corresponding to low frequencies missing from the data. This means that if the starting frequency of the inversion is relatively large, then the initial model needs to be more accurate to ensure continuity in the wavenumber spectral coverage.
and to facilitate the tomographic-like image reconstruction (Mora, 1989) at an early stage of the inversion. In other words, the starting model has to be close enough to the global minimum, i.e. it must predict the first arrival to within half a cycle (Sirgue and Pratt, 2004); to some extent, travel time tomography can be used to obtain such a starting model (e.g., Ravaut et al., 2004; Brenders and Pratt, 2007a; Pratt, 1999; Malinowski and Operto, 2008). The presence of low frequencies, an accurate starting model and good data quality are all critical to the success of waveform tomography.

The acoustic or visco-acoustic approximations provide the cheapest computational implementation of waveform tomography. Application to synthetic data has demonstrated the method’s great potential to recover detailed velocity and attenuation models of the subsurface with geometries ranging from crosshole surveys (e.g., Pratt, 1999; Wang and Rao, 2006; Song and Williamson, 1995; Pratt and Worthington, 1990) to surface reflection and refraction data (e.g., Pratt et al., 1996; Hicks and Pratt, 2001; Sirgue and Pratt, 2004; Shin and Ming, 2006; Brenders and Pratt, 2007a). Application of the method to field data, however, is more challenging due to the presence of noise, missing low frequencies, shot-to-shot energy variations and many other factors related to the acquisition conditions. In such situations, careful data preconditioning and a specific inversion strategy are required. Waveform tomography has been applied to field data from crosshole surveys (e.g., Wang and Rao, 2006; Pratt and Shipp, 1999; Pratt et al., 2004; Zhou and Greenhalgh, 2003) and wide-angle data (e.g., Ravaut et al., 2004; Operto et al., 2006; Malinowski and Operto, 2008; Bleibinhaus et al., 2008; Dessa et al., 2004; Chironi et al., 2006), but there are relatively few examples of results from limited-offset seismic reflection surveys (e.g., Hicks and Pratt, 2001; Wang and Rao, 2009). However, seismic reflection surveys are commonly used in the oil industry and the successful application of waveform tomography to these can potentially improve refraction statics solutions and subsurface imaging in hydrocarbon exploration.
This paper presents the application of frequency domain waveform tomography (Pratt, 1999) to seismic reflection data from the Queen Charlotte Basin offshore western Canada. The abundance of faults, grabens and subbasins makes the area a good candidate for the application of waveform tomography; however the characteristics of the field data necessitate careful pre-processing. After a review of the mathematical foundation of waveform tomography in the first section of the paper, the next section presents the study area and the characteristics of the input data; the acquisition geometry and associated aliasing issues are described, as well as the starting model requirement in relation to the minimum frequency of the data. In the third section, the preconditioning of the data prior to their use in the inversion is presented, followed by the description of the inversion strategy. The last section of the paper focuses on the presentation and discussion of the results.

### 2.3 Mathematical Background

We used the 2D frequency domain waveform tomography approach as described by Pratt et al. (1998); Pratt (1999); Pratt and Shipp (1999). Following is an overview of the method; further details may be found in the above papers. The forward problem is based on the acoustic wave equation, defined in the case of a constant density as:

\[
\nabla^2 u(x, \omega) + \frac{\omega^2}{c^2(x)} u(x, \omega) = -f(x, \omega),
\]

(2.1)

which includes the model parameter \( m(x) = 1/c^2(x) \), the slowness squared. The term \( f(x, \omega) \) represents the distribution of the source energy in the model. The forward solution \( u(x, \omega) \) is the pressure wavefield at the spatial location \( x \). In matrix form (after Pratt, 1999), equation 2.1 has the form

\[
Su = f,
\]

(2.2)
where $S$ represents the wave equation operator, $f$ the source wavefield and $u$ the pressure wavefield. Equation 2.2 implicitly depends on the acoustic model parameters $m$. The data residual $u_{res}$ is defined as:

$$u_{res} = u - d,$$  \hspace{1cm} (2.3)

the difference between the calculated wavefield $u$ and the observed wavefield $d$. The objective of the inversion is to find $m$ such that $u_{res}$ is a minimum. This is equivalent to minimizing the misfit $E$ which is defined using the $L_2$ norm as:

$$E = \frac{1}{2} u_{res}^T u_{res},$$  \hspace{1cm} (2.4)

where $u_{res}^T$ is the conjugate transpose of the data residual. The minimization of the misfit is computed iteratively in the frequency domain using the negative gradient $G_m = -\nabla E$ with respect to the model parameters $m$. At any given iteration $n$ of the computation, the gradient is computed by multiplying the forward modelled wavefield by the backward propagation data residual (Lailly, 1983), without calculating any partial derivative of the data. The model estimate is then updated at iteration $n + 1$ with an appropriate step length $\beta$ such that:

$$m^{(n+1)} = m^{(n)} + \beta^{(n)} G_m^{(n)}.$$  \hspace{1cm} (2.5)

The step length $\beta$ is usually computed using a linear estimate of the forward problem (Mora, 1987) or a line search method.

The inversion procedure requires an estimate of the source signature $f$, which can be calculated in the frequency domain using the method described in Pratt (1999). From the input data $d$, an initial wavefield $u$ is computed using the velocity model and an initial estimate of the source function. The extracted source is obtained by scaling the initial
source estimate by a complex scalar \( s \) which has the following expression:

\[
    s = \frac{u\,d^*}{u'\,u^*},
\]

where subscript \( \,^t \) and \( * \) respectively represent the transpose and the complex conjugate operators. This method can be applied on a shot-to-shot basis or to the entire dataset simultaneously to estimate a single best fit. This last approach was used for this study. During the inversion, the source function is reestimated as higher resolution velocity models are obtained.

In practical terms, the inversion is run sequentially from the lowest to higher frequencies. A single or a group of frequencies are used each time. For every frequency or group of frequencies, an updated velocity model is obtained following a user-defined number of iterations in the data residual minimization. Several parameters which will be discussed later are critical for the success of the inversion.

### 2.4 Seismic Survey

The seismic reflection survey is from the Queen Charlotte Basin (QCB), which is located between the British Columbia mainland and the Queen Charlotte Islands. It is the largest Tertiary basin on the west coast of Canada with an area of approximately 80,000 km\(^2\) (Whiticar et al., 2003). The basin is bounded to the south and to the north by Vancouver Island and Alaska respectively, and is terminated to the east by the Coast Plutonic Complex and to the west by the Queen Charlotte fault, which separates the North American plate from the Pacific plate (Figure 2.1). Exploration of the basin has been encouraged by numerous oil seeps on the Queen Charlotte Islands. From 1950 to 1960, 18 exploration wells were drilled with 8 offshore in Hecate Strait and 10 on Graham Island. In 1987, an intensive basin analysis program was conducted by the Geological Survey of
Canada (GSC), which included in 1988 the acquisition of 7 seismic reflection lines across the basin. Based on the integration of these new data with borehole log, and onshore geological mapping, the QCB is expected to hold significant oil and gas reserves (Whiticar et al., 2003, 2004). The seismic images reveal several subbasins and a complex pattern of faulting (Rohr and Dietrich, 1992).

In this paper we present waveform tomography of line 88-06 (figure 2.1), which crosses the Tertiary sedimentary stratigraphy in Hecate Strait. The geological structure along this line is characterised by Miocene sedimentary subbasins, which are typically half-grabens limited on their west side by a master fault, overlain by a more continuous Pliocene stratigraphy. The sedimentary rocks thin to the north-east, where 20 km of shallow basement rocks are crossed by the seismic line (Rohr and Dietrich, 1992). Line 88-06 was shot using a 6358 in$^3$ airgun array and recorded by a 240-channel streamer hydrophone with a maximum offset of 3770 m and near offset of 186 m. We did not see any evidence of streamer feathering during the conventional processing of these data, for example anomalous stacking velocities for water layer multiples and near-surface scattered arrivals after application of dipmoveout (DMO). We therefore assumed a 2D geometry with no streamer feathering for the waveform tomography of this line. The shot and receiver depths were 12 m, and the shot and receiver intervals were 45 m and 15 m respectively. An analogue bandpass filter was applied to the data prior to digitalisation: 8 Hz with a slope of 6 dB/Octave to 90 Hz with a slope of 72 dB/Octave (Figure 2.2b). A 45-km section of line 88-06 which intercepts the Tyee N-39 exploration well, was used in this study (highlighted in red in Figure 2.1). In this section of the line, the water depth was between 25-50 m.
Figure 2.1: Location map of the Queen Charlotte Basin with seismic reflection lines acquired in 1988. The line in red represents the section of line 88-06 used for this waveform tomography study. The black dots indicate the location of wells. Well Tyee N-39 is located near the western end of line 88-06. QCF indicates the Queen Charlotte Fault.
2.4.1 Geometry and Aliasing

The general similarity of waveform tomography to migration (Mora, 1989; Tarantola, 1984) suggests that the success of the inversion will depend on the spatial distribution of the input data. Aliasing in both shot and the receiver domains can potentially result in artefacts in the constructed models.

The condition for an unaliased wavefield at a given frequency can be expressed as:

\[ \Delta_{samp} \leq \frac{\lambda_{\text{min}}}{2}, \]  

(2.7)

where \( \lambda_{\text{min}} \) is the minimum wavelength of the propagating monochromatic wave, and \( \Delta_{samp} \) is the spatial sampling interval. For a wave propagating in a medium with a minimum velocity \( c_{\text{min}} \), at a frequency \( f \), and emerging with an angle \( \theta \), the wavelength \( \lambda \) of the wave is written as:

\[ \lambda = \frac{c_{\text{min}}}{f \sin \theta}. \]  

(2.8)

When \( \theta \) tends to \( \frac{\pi}{2} \), equation 4.2 becomes:

\[ \lambda = \lambda_{\text{min}} = \frac{c_{\text{min}}}{f}, \]  

(2.9)

and we have the condition:

\[ \Delta_{samp} \leq \frac{c_{\text{min}}}{2f}. \]  

(2.10)

This condition assumes an emerging angle \( \theta = \frac{\pi}{2} \), but since most of the wavefield does not propagate at this limiting angle some amount of aliasing may be tolerable. However, the relationship between data aliasing and the presence of aliasing artefacts in the constructed model is still somewhat unclear, and further theoretical and numerical studies are needed.

After performing experiments with synthetic data at a crustal scale, Brenders and Pratt
(2007b) suggested that the following condition on the data geometry would result in an acceptable velocity model: a receiver interval $\Delta r \leq \Delta_{samp}$ and a shot interval $\Delta s \leq 3\Delta_{samp}$ at the starting frequency.

The aliasing condition can also be expressed using the sparseness of the data. The sparseness of the data is expressed as (Bleibinhaus et al., 2008):

$$N_A = \frac{\Delta}{\Delta_{samp}}, \quad (2.11)$$

where $\Delta$ is either the source interval $\Delta s$ or the receiver interval $\Delta r$, whichever is sparser. The wavefield is completely unaliased when $N_A$ is less than 1. The best result from Brenders and Pratt (2007b) was obtained using at the lower frequency $\Delta s = 2\Delta_{samp}$ and $\Delta r = \Delta_{samp}$, which gives a sparseness $N_A = 2$ and $N_A = 18$ for the lowest and the highest frequencies respectively (0.8 Hz and 7 Hz). In our study, at the minimum usable frequency of 7 Hz (Figure 2.2b) an unaliased wavefield is obtained when $\Delta s$ and $\Delta r$ are less than $\Delta_{samp} = 105.7$ m (using $c_{min} = 1480$ m/s). At the highest frequency used in the inversion (13.66 Hz), an unaliased wavefield is obtained when $\Delta s$ and $\Delta r$ are less than $\Delta_{samp} = 54.1$ m. These requirements are satisfied by the shot interval (45 m) and the receiver interval (15 m) of line 88-06. However to reduce computational cost and to avoid oversampling (since the maximum frequency used is 13.66 Hz) the data were subsampled to use every second shot (that is $\Delta s = 90$ m) and every third receivers ($\Delta r = 45$ m). This new geometry gives $N_A = 0.8$ for the lower frequency and $N_A = 1.6$ for the higher frequency, and appears to be sufficient when compared with the parameters employed by Brenders and Pratt (2007a) and other similar studies (Table 2.1).
Figure 2.2: (a) Field shot gather located at x=8.27 km and (b) the associated amplitude spectrum. Only the initial 3 s of data with every third receiver is shown. Direct (D), refracted (Rf) and a reflected (Re) arrivals are visible.
Table 2.1: Aliasing number $N_A$ for some survey configurations. The first value of $N_A$ is for the lower frequency and the second value for the higher frequency.

<table>
<thead>
<tr>
<th>Authors</th>
<th>Data</th>
<th>Geometry</th>
<th>Frequencies (Hz)</th>
<th>$N_A$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Brenders and Pratt (2007c)</td>
<td>Synthetic</td>
<td>Refraction</td>
<td>0.5-7.5</td>
<td>0.5-4</td>
</tr>
<tr>
<td>Brenders and Pratt (2007a)</td>
<td>Synthetic</td>
<td>Refraction</td>
<td>0.8-7</td>
<td>2-18</td>
</tr>
<tr>
<td>Ravaut et al. (2004)</td>
<td>Real</td>
<td>Refraction</td>
<td>5.4-20</td>
<td>0.5-1.8</td>
</tr>
<tr>
<td>Operto et al. (2006)</td>
<td>Real</td>
<td>Refraction</td>
<td>3-15</td>
<td>3-15</td>
</tr>
<tr>
<td>Pratt (1999)</td>
<td>Synthetic</td>
<td>Crosshole</td>
<td>$125.10^3$-750.10$^3$</td>
<td>0.3-1.6</td>
</tr>
<tr>
<td>Pratt et al. (2004)</td>
<td>Real</td>
<td>Crosshole</td>
<td>100-1000</td>
<td>0.3-3</td>
</tr>
<tr>
<td>Wang and Rao (2006)</td>
<td>Real</td>
<td>Crosshole</td>
<td>190-485</td>
<td>0.1-0.5</td>
</tr>
<tr>
<td>Hicks and Pratt (2001)</td>
<td>Real</td>
<td>Reflection</td>
<td>10-60</td>
<td>0.16-1</td>
</tr>
<tr>
<td>This study</td>
<td>Real</td>
<td>Reflection</td>
<td>7-13.66</td>
<td>0.8-1.66</td>
</tr>
</tbody>
</table>

2.4.2 Starting Model and Convergence

A suitable starting model for waveform tomography has to predict the first arrival to within half a cycle (Sirgue, 2003). From this condition, the relationship between the traveltime error $\delta t$ relative to the total arrival time $T$ of the event can be expressed as (Pratt, 2008):

$$\frac{\delta t}{T} \leq \frac{1}{2N_\lambda},$$

(2.12)

where $N_\lambda = l/\lambda$ is a dimensionless term representing the propagation distance in terms of wavelengths and $l$ is the propagation distance from the source to the receiver (Pratt and Shipp, 1999; Pratt, 2008). This relation shows that small values of $N_\lambda$ are essential; they correspond to lower frequencies which give a less rigid condition on the starting model; they also correspond to wider Fresnel zones and good resolving power at long wavelengths. For a given frequency, small values of $N_\lambda$ correspond to a limited offset; however, longer offset data are also important because they are required to constrain the low wavenumbers in the velocity model (Sirgue and Pratt, 2004). In order to meet the convergence requirement (equation 4.5), the starting model was derived from traveltime tomography using an algorithm based on first arrival traveltime picking (Aldridge and
Oldenburg, 1993). After 30 iterations, a model was obtained with a root mean square misfit of 3.3 ms and maximum absolute misfit of 44.1 ms (Figure 2.3a). A wave propagating at the maximum offset (3770 m) and at the minimum frequency (7 Hz), has an approximate maximum propagation distance in terms of wavelength of $N\lambda=17.8$. This leads to the requirement that the traveltime error relative to the total arrival time ($\delta t/T$) should be less than 0.028 (using equation 4.5). Considering the maximum arrival time of the first arrival (1.731 s) at the maximum offset (3770 m), and the maximum misfit of the model (44.1 ms), we have $\delta t/T = 0.025 < 0.028$. This means that the current model is adequate for use in waveform tomography. For the inversion of frequencies 7-12 Hz, the starting model was resampled to a 30 m grid in agreement with the finite difference stencils implemented in the wave modelling program of 4 grids points per wavelength. Absorbing boundary conditions with 10 grid-points (Clayton and Engquist, 1977) were used along the sides and bottom, and a free surface boundary condition was used at the top of the model to account for any water-layer reverberations in the data. For the inversion of frequencies 12-13.66 Hz, the model obtained after the inversion of frequencies 7-12 Hz was resampled to a 25 m grid.

2.5 Data Preconditioning

The goal of preconditioning is to organise the seismic data in a form suitable for waveform tomography. This implies that any aspect of the data that is not predicted by the 2D acoustic propagation scheme, e.g., shear waves, coherent noise, shot to shot energy variations, amplitude discrepancy and bad traces, should be removed or corrected. In the case of the seismic data from the Queen Charlotte Basin the main preconditioning steps were:

1. F-K filtering, trace editing and lowpass filtering.
2. Shot-to-shot amplitude balancing

3. Time windowing

4. Amplitude scaling for 2D modelling

2.5.1 F-K Filtering, Trace Editing and Lowpass Filtering

One of the characteristics of marine seismic reflection data is the presence of coherent noise, which can arise from guided waves propagating just below the seafloor and from shallow scattering around the seismic line (Larner et al., 1983). On line 88-06, high amplitude coherent noise with a linear moveout velocity around and less than approximately 1000 ms\(^{-1}\) is present on almost all the shot gathers. In order to suppress

![Figure 2.3: Starting model (a) and associated ray density (b) derived from first arrival traveltime tomography. The ray coverage is limited to approximately 600 m depth. The black tick line represents the location of the well.](image-url)
these arrivals, the data were converted to the frequency-wavenumber (F-K) domain and a fan filter was applied. The filter was designed with a polygon to remove all the energy propagating with a speed less than ±1450 ms⁻¹. By applying this filter, all phases with apparent velocity less than that of the direct water wave were removed (Figure 2.4). The data were then carefully inspected to remove any bad traces and further lowpass filtered at 15 Hz, using an Ormsby filter with 0.1 % white noise and corner frequencies: 0-0-15-16 Hz.

2.5.2 Shot-to-shot Amplitude Balancing

Examination of the data also shows that some shot gathers have significantly more energy than others (e.g., Figure 2.5a). This might be the result of variations in field gains, air pressure variations or changes in the relative position of individual airguns in the array. Such shot-to-shot variations can bias the model update during the inversion, with a larger velocity perturbation introduced where the seismic data amplitudes are greater. Normalizing individual traces to their maximum amplitude value (e.g., Ravaut et al., 2004; Operto et al., 2006; Malinowski and Operto, 2008) is a reasonable way to handle the problem; however, by doing so amplitude variations with offset are lost. We therefore chose to balance the amplitude by normalizing each shot gather to the same maximum amplitude value, thus ensuring that amplitude variations with offset are preserved as well as the amplitude variation within individual traces (see Figure 2.5b).

2.5.3 Time Windowing

The purpose of time windowing is to provide the algorithm with data that favor the contribution of first arrivals, direct and refracted energy, at the early stage of the inversion and exclude late arrivals, multiples and scattered energy, that originate from outside the
Figure 2.4: (a) Field shot gather at location x = 8.27 km after low pass filtering to 15 Hz. The arrows indicate coherent noise. (b) The same shot gather after removal of the coherent noise using F-K filtering.

imaging plane. Early arrivals have a greater contribution to the reconstruction of low wavenumbers, and thus help to stabilize the inversion. The window can be increased in the later stages of the inversion to include late arrivals (e.g., Breders and Pratt, 2007a,c). In
Figure 2.5: Illustration of the shot-to-shot amplitude variation in the field data: (a) Concatenated field shot gathers at location $x = 3.77$ km and location $x = 44.9$ km. The energy disparity between the gathers is evident. (b) The same data after correction with shot-to-shot amplitude balancing.

In this study, the data were muted 1.5 s below the first arrivals, using a window of 2 s for the inversion of frequencies ranging from 7-12 Hz. This window essentially contains the most linear components of the wavefield, which include all the direct arrival and refracted energy as well as most of the early wide-angle reflections and diffractions. The window was later increased to 3 s to capture later arrivals for the inversion of frequencies 12-13.66 Hz.
Figure 2.6: RMS Amplitude variation with offset bins (offsets are binned every 45 m) of (a) the field data after F-K and low pass filtering at 15 Hz, (b) the field data after F-K, low pass filtering and amplitude balancing. (c) The amplitude variation of the forward modelled data using the starting traveltime model and (d) the scaled, 2D converted, F-K filtered, low pass filtered and shot-to-shot amplitude balanced field data.

2.5.4 Amplitude Scaling for 2D Modelling

In order to correct the field data from 3D propagation to the 2D propagation assumed in the forward modelling step of the inversion, the data were multiplied by $\sqrt{t}$ to approximately correct the geometrical spreading from 3D to 2D. A single comparison was then made between the RMS amplitude variations with binned offset of the processed shot gathers (2D corrected, time windowed, shot-to-shot amplitude balanced, F-K filtered and lowpass filtered) with the modelled data, derived after forward modelling using the velocity model.
from traveltime tomography (Figures 2.6b and 2.6c). The amplitude disparity between the two datasets is evident. This disparity might be related to the differences in radiation pattern between the field data and the modelled data, to complex geometrical spreading, and to dispersion phenomena not yet considered. In order to correct for these differences, the prepared field data were further multiplied by a scaling factor obtained using linear regression of the logarithm of the RMS amplitude variation with offset bins of the two datasets (Brenders and Pratt, 2007a). The logarithm of the RMS amplitudes with respect to offset bins of the field and the modelled data give two straight lines, $ax + b$ for the field data and $a'x + b'$ for the modelled data, allowing the scaling factor to be determined from:

$$f(x) = \frac{e^{ax+b}}{e^{a'x+b'}}. \quad (2.13)$$

Multiplying the pre-processed field data by this additional factor results in the amplitude behavior of both the forward and the field data being generally similar along the seismic line (Figure 2.6). The preconditioned data, ready for waveform tomography, were finally arranged in reduced time (with a reduction velocity of 3000 ms$^{-1}$) to ensure the efficient inclusion of all arrivals during the inversion (Figure 2.7) and to avoid time wraparound effects during forward modelling.

### 2.6 Waveform Inversion Strategy

Waveform tomography of field seismic data based on the gradient descent method often suffers from convergence problems due to non-linearity. Low frequencies and an accurate starting model, which provides the long wavelength variations in the velocity model, are critical to the success of the inversion. When the low frequencies are missing, in addition to the need for a more accurate starting model, which can sometimes be difficult to obtain from traveltime tomography due to the resolution limit, an inversion strategy that avoids
Figure 2.7: (a) Preconditioned data for waveform inversion. The data contains frequencies 7-15 Hz displayed using a reduction velocity of $v = 3000 \text{ ms}^{-1}$. (b) Associated amplitude spectrum.
convergence to the numerous local minima present at high frequencies is necessary. As the starting frequency of the data used in this study is relatively large (7 Hz), it is necessary to design a specific strategy. Another problem is that the recovery of deeper structures in the velocity model is affected by the amplitude distribution within the seismic data. The early arrivals typically have larger amplitudes than late arrivals, and thus dominate the gradient function used to update the velocity model. Consequently, the shallow parts of the velocity model contribute more to the misfit function than deeper parts and are thus more accurately reconstructed.

The ray coverage from traveltime tomography, which is based on first arrival traveltimes is limited to approximately the top 600 m of the model (Figure 2.3b). A strategy was designed to recover structures down to 1200 m and to ensure sufficient convergence of the inversion to a likely global minimum. This strategy consisted of successively recovering shallow to deep structures in the subsurface velocity model. The shallower structures were recovered using near offset and early arrivals and the deeper structures were recovered by additionally including far offsets and late arrivals. The inversion of higher frequencies ($\geq 10.5$ Hz) was implemented with the additional introduction of attenuation, which improved the convergence and helped to reduce artefacts. Tables 2.2 and 2.3 show a summary of this inversion strategy, and a detailed description is provided below.

2.6.1 Frequency Selection

There are currently no well-defined criteria for choosing frequencies in waveform tomography. The general methodology is based on the Nyquist sampling theorem (e.g., Pratt et al., 2004; Bleibinhaus et al., 2008; Breinders and Pratt, 2007a,c). According to this theorem, the interval $\Delta f$ of frequencies used in the inversion should not exceed the inverse of the cycle duration of the wavefield. However it is also possible (for example Wu and
Toksöz, 1987; Sirgue and Pratt, 2004; Brenders and Pratt, 2007b; Wang and Rao, 2009) to obtain an unaliased image with fewer frequencies.

In this study, two frequency discretisation schemes were evaluated:

1. The first method is based on the Nyquist sampling theorem and is called full waveform tomography (Pratt and Worthington, 1990). The frequencies used are selected every $\Delta f$, with

$$\Delta f = \frac{1}{T_w}. \quad (2.14)$$

$T_w$ is the length of the input window of data, which would imply $\Delta f = 0.5$ Hz for the 2 s window and $\Delta f = 0.33$ Hz for the 3 s window of our dataset. The longer window was used near the end of the inversion procedure to increase the depth of investigation. For every frequency range (Table 2.2), the frequencies were inverted in pairs. Between 7 and 10.5 Hz, 4 pairs of frequencies were successively inverted: (7,7.5), (8,8.5), (9,9.5), and (10,10.5); between 10.5 and 12 Hz, the pairs of frequencies were (10.5,11) and (11.5,12); between 12 and 13.66 Hz the pairs of

<table>
<thead>
<tr>
<th>Freq range (Hz)</th>
<th>Models</th>
<th>$T_{win}$ (s)</th>
<th>$\tau$ (s)</th>
<th>Offset</th>
<th>Depth weighting</th>
<th>Reconstruction</th>
</tr>
</thead>
<tbody>
<tr>
<td>7 to 10.5</td>
<td>$v_p$</td>
<td>2</td>
<td>0.8</td>
<td>near</td>
<td>shallow</td>
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<td></td>
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<td>0.8</td>
<td>far</td>
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<td></td>
<td></td>
<td></td>
<td>1.76</td>
<td></td>
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<tr>
<td>10.5 to 12</td>
<td>$v_p$ &amp; $Q_p$</td>
<td>0.8</td>
<td>near</td>
<td>shallow</td>
<td>shallow</td>
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<tr>
<td>12 to 13.66</td>
<td>$v_p$ &amp; $Q_p$</td>
<td>3</td>
<td>1.2</td>
<td>near</td>
<td>shallow</td>
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<td>2.4</td>
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frequencies were (12,12.33), (12.66,13) and (13.33,13.66). Each pair of frequencies was inverted following the sequence of Table 2.2; i.e. each pair was inverted using four reconstructions from the shallow to deep as described in the next section on layer stripping.

2. The second discretisation scheme, named efficient waveform tomography (after Sirgue and Pratt, 2004) is based on the continuity of the spectral coverage in the wavenumber domain. A given frequency spans an area \((k_{z\text{min}}, k_{z\text{max}})\) in the wavenumber space. The minimum vertical wavenumber \((k_{z\text{min}})\) and the maximum vertical wavenumber \((k_{z\text{max}})\) are respectively (Wu and Toksöz, 1987; Sirgue and Pratt, 2004):

\[
k_{z\text{min}} = \frac{2f\alpha}{c_o} \tag{2.15}
\]

and

\[
k_{z\text{max}} = \frac{2f}{c_o}, \tag{2.16}
\]

where \(c_o\) is the velocity of the model and \(\alpha\) is the cosine of the reflection angle of a wavefield propagating at maximum incidence, and is equivalent to the inverse of the NMO (normal moveout) stretch factor (Sirgue and Pratt, 2004). \(\alpha\) can be expressed as:

\[
\alpha = \frac{1}{\sqrt{1 + (h_{\text{max}}/z_{\text{max}})^2}}, \tag{2.17}
\]

with \(h_{\text{max}}\) the source-receiver half offset and \(z_{\text{max}}\) the maximum depth to be imaged.

The selection of frequencies is designed in such a way that the minimum wavenumber of the frequency used in any iteration corresponds to the maximum wavenumber of the frequency previously used. In this fashion, there is no overlap of the wavenumbers. This approach produces a significant reduction in the number of frequencies used, and a significant reduction in computational cost. To select the
frequencies, the following formula is used:

\[ f_{n+1} = \frac{f_n}{\alpha}, \]  

(2.18)

where \( f_n \) represent the \( n^{th} \) and \( f_{n+1} \) the \( n^{th}+1 \) frequency used. \( \alpha \) is obtained based on equation 2.17 and was calculated using a maximum offset of 1600 m after observation of the disproportionately small weight of the far offsets in the gradient during the inversion, and the target depth was set to 1200 m. This gives \( \alpha = 0.83 \) allowing the following selection of frequencies: 7, 7.5, 9.03, 10.88 and 13.11. The first 2 frequencies do not respect the selection criteria and were grouped to enhance the signal to noise ratio, which is relatively low at these frequencies; the other frequencies were chosen based on the selection criteria, starting from 7.5 Hz and were inverted one at a time. The inversion proceeded successively to 13.11 Hz, following the sequence of Table 2.3.

The condition for avoiding aliasing in the wavenumber domain (Wang and Rao, 2009) is:

\[ \Delta k_z \leq \frac{1}{z_{max}}; \]  

(2.19)

for two consecutive frequencies \( f \) and \( f + \Delta f \), the sampling interval using the corresponding neighboring wavenumbers can be expressed (using equation 2.15) as:

\[ \Delta k_z = k_{z_{min}}(f + \Delta f) - k_{z_{min}}(f) \]
\[ = \frac{2\alpha \Delta f}{c_0} \]  

(2.20)

\[ \leq \frac{1}{z_{max}}. \]  

(2.21)
From where the condition for avoiding aliasing becomes:

\[
\Delta f \leq \frac{c_o}{2\alpha z_{\text{max}}}; \quad (2.22)
\]

which gives, using the minimum velocity of the model (1480 ms\(^{-1}\)), the condition \(\Delta f \leq 0.74\), which is not respected by the above frequency selection. This discretisation method will produce aliasing of the corresponding seismic trace in time, but will theoretically yield an unaliased image of the target depth as shown by Sirgue and Pratt (2004).

In both frequency selection methods, a low pass filter in the wavenumber domain was applied to the gradient to eliminate wavenumbers greater than \(k_{z_{\text{max}}}\) and to encourage the recovery of low wavenumbers (Sirgue, 2003). For every frequency or group of frequencies used in the inversion, the vertical component of the wavenumber was chosen as \(k_z = k_{z_{\text{max}}}\) and the horizontal component as \(k_x = k_z/5\).

**Table 2.3:** Inversion strategy for efficient waveform tomography. \(T_{\text{win}}\) is the time window of data and Freq range is the frequency range.

<table>
<thead>
<tr>
<th>Freq range (Hz)</th>
<th>Models</th>
<th>(T_{\text{win}}) (s)</th>
<th>(\tau) (s)</th>
<th>Offset</th>
<th>Depth weighting</th>
<th>Reconstruction</th>
</tr>
</thead>
<tbody>
<tr>
<td>7 to 10.88</td>
<td>(v_p)</td>
<td>2</td>
<td>0.8</td>
<td>near</td>
<td>shallow</td>
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<tr>
<td>13.11</td>
<td>(v_p)</td>
<td>3</td>
<td>1.2</td>
<td>near</td>
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2.6.2 Time Damping of the Data Residual and Layer Stripping

Time windowing can be applied in the frequency domain using complex frequencies (e.g., Mallick and Fraser, 1987; Pratt et al., 2004; Brenders and Pratt, 2007a; Sirgue and Pratt, 2003). A complex angular frequency can be defined as:

\[ \omega = 2\pi f + \frac{i}{\tau}, \]

(2.23)

where \( f \) is the real component of the frequency and \( \tau \) a time damping term. The addition of this imaginary component is similar to multiplying the time domain data by a function \( e^{-t/\tau} \). This function serves to damp the contribution of late arrivals in the data residual, and thus enables the selection of time aperture in the frequency domain data. By changing the value of \( \tau \), different time windows can effectively be included. The function assumes data recorded at zero offset. For data recorded at non zero offset, a time shift is applied corresponding to the first arrival time at the given location. This is obtained by further multiplying the data by \( e^{t_o/\tau} \), where \( t_o \) is the first arrival time (Brenders and Pratt, 2007a).

For this study, two values of \( \tau \) were used in sequence for every frequency or frequency group. These values correspond to \( \tau = 0.8 \) s and \( \tau = 1.76 \) s which respectively represent 40 % and 88 % of the 2 s input data window for frequencies 7-12 Hz. The values were changed to \( \tau = 1.2 \) s and \( \tau = 2.4 \) s for frequencies 12-13.66 Hz, which correspond to 40 % and 80 % of the 3 s input data window (see Tables 2.2 and 2.3).

In addition to selecting a time aperture in the data residual, a layer stripping approach was implemented to successively recover shallowmost to deeper layers in the model under reconstruction. The approach was implemented by inverting from near to far offsets using a weighting function that was applied to the data residual with respect to offset, and weighting up the gradient with depth using a depth-dependent weighting function. For every pair of frequencies in the full waveform tomography approach or single frequency
in the efficient waveform tomography approach, the inversion was run in two stages (Tables 2.2 and 2.3):

1. Recovery of the upper portion of the velocity model by focussing on the near offsets (offsets range 300-1600 m) and weighting up the gradient representing the upper part of the model (100-800 m).

2. Recovery of the deeper portion of the model by focussing on the far offsets (offset range 1600-3770 m) and weighting up the gradient with depth (800-1500 m).

The weighting functions consist of vertically and offset dependent cosine tapers. For the shallow reconstruction, the gradient was tapered from 100 m toward the surface and from 800 m toward the bottom of the model and the residual was tapered with offset from 300 m toward the minimum offset (215 m) and from 1600 m toward the maximum offset (3770 m). For the deeper reconstruction, the gradient was tapered from 800 m toward the surface and from 1400 m toward the bottom of the model and the residual was tapered with offset from 1600 m toward the minimum offset and from 3770 m toward the maximum offset.

2.6.3 Introduction of Attenuation

Intrinsic seismic attenuation can be introduced in frequency domain acoustic waveform inversion using complex velocity (Song and Williamson, 1995; Hicks and Pratt, 2001). The complex valued velocity has the form:

\[ v = v_r + iv_i, \]  

with \( v_r \) and \( v_i \) the real and imaginary part of the velocity, from which the seismic \( Q_p \) factor has the form

\[ Q_p = -\frac{v_r}{2v_i}, \]
The attenuation is the inverse of the $Q_p$ factor, $Q_p^{-1}$. Rock properties such as fractures, porosity and rheology all influence attenuation. However, recovery of attenuation and velocity is more challenging than the recovery of velocity alone, because of the difficulty in discriminating scattering attenuation due to small-scale velocity variations, from intrinsic attenuation due to rock properties (Hicks and Pratt, 2001; Kamei and Pratt, 2008). An accurate short wavelength velocity image can help to remove scattering attenuation effects. Kamei and Pratt (2008) proposed a strategy of inverting for the velocity in an initial step, and then inverting for both the velocity and attenuation in a second stage once the velocity image had been sufficiently well resolved. This same strategy was used in our study, and the attenuation was introduced for the inversion of frequencies from 10.5 to 13.66 Hz. Both velocity and attenuation were inverted at these frequencies. The starting model for attenuation was a homogeneous model with $Q_p = 100$.

### 2.6.4 Source Extraction and Model Update

The layer stripping approach comprised four separate inversions, at which new velocity models were computed, as shown by Tables 2.2 and 2.3, and the amplitude and phase of the source wavelet were estimated at each of these steps using Equation 2.6. The source estimate needed for the first inversion was derived from the velocity model from ray-based tomography and a spike wavelet. In each of these four inversion steps, 5 iterations of the linear estimate (2.5) were used to minimize the data residual, resulting in intermediate velocity models, at which the amplitude spectrum, but not the phase spectrum, of the source function was also updated. As the resolution of the velocity model increases with the inclusion of higher frequencies in the inversion, the estimate of the source wavelet becomes more realistic (e.g., Figure 2.8), but the bandwidth of the wavelet is of course limited by the frequencies included in the waveform tomography.
Figure 2.8: (a) Source signature estimated using the starting traveltime tomography model and (b) using full waveform tomography models (velocity and attenuation) for frequencies 7-12 Hz. A more realistic source signature is obtained with the waveform tomography models.

2.7 Results and Discussion

This section presents and discusses the results from the full waveform and efficient waveform tomography approaches. Model assessment was carried out by comparing the P-wave velocity model with the sonic log, the migrated seismic section, and by comparing the modelled seismograms with the field seismograms.

2.7.1 Velocity Model from Low Frequencies

The P-wave velocity models obtained after inverting for two frequency ranges 7-8.5 Hz and 7-10.5 Hz using the full waveform tomography approach are displayed in Figures 2.9a and 2.9b. There is a noticeable increase of the resolution in the velocity models when higher frequencies are included. The most noticeable features in the 7-10.5 Hz model are layering and number of sharp lateral changes in velocity, localized low velocity anomalies (e.g., at x = 13 km, 15 km, 17 km, 20-21 km) that may represent steeply dipping faults.
The results from efficient waveform tomography for frequencies 7-10.88 Hz are displayed in Figure 2.10. Overall, the efficient waveform tomography results appear similar to the full waveform tomography results. All the major features present on the full waveform tomography velocity model are also present on the efficient waveform velocity model. Both models clearly show the same distribution of lateral velocity variation along the line. The most visible differences between the two models (Figures 2.9b and 2.10b) are localized low velocity anomalies and layering which appear clearer on the full waveform tomography result than on the efficient waveform tomography result. Layer contrasts also appear sharper in the full waveform tomography result than in the efficient waveform tomography result. 1D velocity profiles extracted from full waveform tomography and efficient waveform tomography results at several locations along the line confirm the good agreement between the two (Figure 2.11). This shows the ability of efficient waveform tomography to provide acceptable results with fewer frequencies. A similar conclusion was obtained by Breiners and Pratt (2007b) with synthetic refraction data.

To evaluate the effect of data amplitudes on the recovered velocity model, a phase-only inversion was also performed using the full waveform tomography approach for frequencies 7-10.5 Hz. The phase-only inversion was obtained by additionally normalizing the amplitudes of the frequency domain field and predicted data to 1. The result (Figure 2.9c) is similar to the amplitude-plus-phase inversion (Figure 2.9b), but the amplitude-plus-phase velocity model appears to have greater resolution and contains more information on the structure and stratigraphy. For example, the lateral change in velocity between 35-37 km is sharper in the amplitude-plus-phase model than in the phase-only model, demonstrating the value of including amplitude information.
Figure 2.9: Inversion results: (a) full waveform tomography result after 7-8.5 Hz inversion; (b) full waveform tomography result after 7-10.5 Hz inversion and (c) phase-only tomography result after 7-10.5 Hz inversion. Note the greater resolution of the amplitude-plus-phase inversion in (b) compared with the phase-only inversion in (c). The black tick line represents the location of the well.

### 2.7.2 Velocity and Attenuation Models from High Frequencies

For frequencies 10.5-13.66 Hz, both attenuation and velocity models were recovered. The jump from 10.88 Hz to 13.11 Hz in the efficient waveform tomography provided a very noisy attenuation result. Consequently inversion for attenuation was carried out only using full waveform tomography. The failure of efficient waveform tomography to recover an acceptable attenuation model could be related to increased aliasing in the wavenumber domain. Figure 2.12 shows the result of the full waveform tomography inversion with
attenuation for frequencies 10.5-12 Hz. At these frequencies, the simultaneous velocity and attenuation inversion resulted primarily in updates to the attenuation model; no major changes were observed in the velocity model, although some layer contrasts increased. The inversion of frequencies greater than 12 Hz resulted in excessive noise in the attenuation model. Consequently, for frequencies greater than 12 Hz, updates were only applied to the velocity model and the attenuation model was kept unchanged. For comparison, the inversion at these higher frequencies was also performed without attenuation. The result of the inversion without attenuation (Figure 2.13) includes some x-shaped artefacts. In general, the introduction of attenuation improved the speed of convergence in the inversion and significantly reduced artefacts. The x-shape artefacts are due to the increase of the data residual at far offset with increasing frequency. The inversion was halted at 13.66 Hz because no significant improvement was observed and the velocity model was becoming increasingly blurred due to short wavelength artefacts; the 13.66 Hz model already contains some artefacts.
Figure 2.11: 1D profiles of full waveform tomography, efficient waveform tomography and the starting model at 4 locations along the line. (a) At 11.6 km, (b) 23.6 km, (c) 37.1 km and (d) 41.6 km. The black curve represents the starting model, red the full waveform model and blue the efficient waveform model. The figures show a generally good agreement between both the efficient and the full waveform tomography models.
A velocity perturbation model (Figure 2.12c) was derived after subtracting the starting traveltime tomography model from the 7-12 Hz full waveform tomography model. The sedimentary stratigraphy can be more clearly identified in the perturbation than in the velocity model, demonstrating the migration-like solution of waveform tomography. A number of 0.5-0.75 km wide subvertical bands with magnitude -50 to -100 ms$^{-1}$, e.g., at $x = 12$ km, 18 km and 20.5 km are also visible in the perturbation.

Although it is more difficult to assess the quality of the attenuation model, the increase of the convergence rate and the reduction in artefacts in the velocity model suggest that the inclusion of attenuation is in general beneficial. The attenuation values (Figure 2.12a) are generally consistent with laboratory measurements of saturated sandstone (Winker and Nur, 1979), and consistent with the well logs which show that these strata comprise clastic rock with porosity that varies from 0 to 25%. Studies on the characterization of fractures with waveform tomography, suggest that fractures with size smaller than or equal to a half-wavelength of the seismic wavefield i.e. below the resolution limit of waveform tomography act as single scatterers and produce images with strong attenuation values (Rao and Wang, 2009). The distribution of high attenuation values ($Q_p^{-1} > 0.03$) might correspond to the locations of fractures in the vicinity of faults, or alternatively to regions of increased fluid saturation.

2.7.3 Synthetic and Field Shot Gathers

Figure 2.14 shows a comparison between the field seismograms (Figure 2.14a), the modelled seismograms calculated from the velocity and attenuation models obtained after the 7-12 Hz full waveform tomography approach (Figure 2.14b), and the modelled seismograms calculated using the starting model (Figure 2.14c). The seismograms from 3 shot gathers at locations $x = 15.38$ km, $x = 35$ km and $x = 36.7$ km are displayed. Figures 2.14a and 2.14c show the ability of the starting model to predict the first arrival
Figure 2.12: Waveform inversion results for frequencies 10.5-12 Hz. Both the velocity model (a) and the attenuation model (b) were recovered; (c) is the velocity perturbation obtained after subtracting the starting model obtained by ray-based tomography from the 7-12 Hz model. The ellipses in the attenuation model are indicative of some of the areas with higher attenuation values that may represent increased fracturing or increased in fluid saturation. The short black tick line indicates the location of the well log.

traveltimes, but not the complexity of the waveforms, which include the amplitude and phase behavior of the early and late arrivals. Figures 2.14a and 2.14b demonstrate that a significant improvement in the fit to the field data was obtained with waveform tomography. There is a high degree of similarity between the seismograms calculated
Figure 2.13: Waveform inversion results for frequency 13.66 Hz with (a) attenuation and (b) without attenuation. Arrows indicate of x-shaped artefacts. The introduction of attenuation has helped to significantly reduce the amplitude of x-shape artefacts in the velocity model. The short black tick line represents the location of the well log.

from the waveform tomography velocity model and the field seismograms. The shape of the waveforms, the arrival times, as well as the amplitude and phase variations, are very similar within the 2 s data window for all the offsets. The good fit between the two datasets shows the convergence of waveform inversion toward a likely global minimum and the recovery of the background velocities as well as short wavelength layering.

2.7.4 Velocity Model and Sonic Log

A comparison was made between the sonic log and the corresponding 1D velocity profiles at location $x = 19.7$ km, from full waveform and efficient waveform tomography approaches (Figure 2.15). Overall, a good agreement is observed between the sonic log and the waveform tomography results. Our inversion strategy enables a good recovery of structures down to 1200 m, which is deeper than the coverage with traveltime tomography.
Figure 2.14: Three concatenated shot gathers at locations 15.38 km (left), 35 km (center) and 36.7 km (right) of (a) the field data low pass filtered at 12 Hz, (b) the modelled data calculated with the velocity and attenuation models derived from the 7-12 Hz full waveform tomography and (c) the model data calculated with the starting velocity model from traveltime tomography. There is a high degree of similarity between the modelled data using the waveform tomography models and the field data, within the 2 s window. Reduction velocity $v = 3000 \text{ m s}^{-1}$.

(Figure 2.3b). The decrease of velocity at 660 m depth in the sonic log, corresponds to the transition zone between the Pliocene and the upper Miocene sections (Figure 2.16) and is

66
consistent with other related studies in the area (e.g., Rohr and Dietrich, 1992). This transition, the Pliocene - upper Miocene unconformity, starts 50 m deeper in the velocity models than in the sonic log. This is due to the starting traveltome model, which probably did not provide a sufficiently accurate long wavelength structure in this part of the model, because this depth is greater than the first arrival ray coverage. In waveform tomography, the low wavenumbers contribute to the tomographic-like reconstruction. If interval velocities are too slow in the starting model, the layer interface will be shallower in the solution model and if interval velocities are too fast, the interface will be deeper. This might explain the 50 m discrepancy in the Pliocene-upper Miocene unconformity in the 1D velocity profile, as the starting velocity in that part of the model is faster.

Figure 2.15: Sonic log (in red), starting model (dotted black) and (a) full waveform result (in blue); (b) efficient waveform result (in blue) at the well log location. The well log is located at x = 19.7 km, and start from 214.3 m depth. The waveform results correlate with the sonic log better than the starting model derived from traveltome tomography. The major difference between the sonic log and the waveform tomography result is the transition from Pliocene to Upper Miocene at 660 m depth which is 50 m deeper in the waveform tomography results.
2.7.5 Velocity Model and Migrated Seismic Section

In order to check the consistency between the velocity model from waveform tomography and the conventional migrated seismic section, the time-converted 7-12 Hz velocity model from full waveform tomography was superimposed (Figure 2.17a). The correspondence of features in the migration with the results from waveform tomography constitutes an independent validation of the waveform tomography results. Waveform tomography can also complement the interpretation of the migrated section by providing physical properties of the subsurface such as velocity and attenuation. In the following, the migrated seismic section and its interpretation are derived from Rohr and Dietrich (1992) and Whiticar et al. (2003).

The sedimentary section comprises Pliocene and Miocene rocks separated by an angular unconformity (Figures 2.16 and 2.17a). A vertical strike-slip fault at x = 17.5 km marks the south-west margin of an antiform extending from x = 17.5 km to x = 30 km. North-east of the antiform, a deep half-graben is centred near x = 38 km and contains up to 4 s of coherent reflections. The south-west part of the half-graben fill has been inverted into an anticline cut by faults. South-west of the vertical fault at x = 17.5 km, the Miocene section shallows towards Moresby Island on the west (Figure 2.16). Most of the faults interpreted from the migrated section are truncated at or close to the upper Miocene unconformity, but at least one major fault extends into the Pliocene section at x = 17.5 km (Rohr and Dietrich, 1992).

The velocity model (Figure 2.17a) shows a general increase in velocity with depth, which is related to the compaction and lithification of sedimentary rocks with age. The lateral variation in velocity does not follow the seismic reflectors at the apex of the anticline between x = 17.5 km and x = 23 km; the apex of the anticline has velocities approximately 150 m s\(^{-1}\) lower than the rocks at a similar depth. Reduced velocities at the apex of anticlinal folds have also been observed above the Cascadia accretionary wedge.
(Hayward and Calvert, 2007), and the lower velocities there have been attributed to tensile fractures that provide pathways for fluid expulsion.

The velocity perturbation from the 7-12 Hz full waveform tomography result was also superimposed on the migrated seismic section (Figure 2.17b). The orientation of the layering in the perturbation model is consistent with the seismic stratigraphy in the migrated seismic section, and therefore indicates the presence of the anticline. Some sub-vertical bands in the perturbation correlate with faulting, e.g., at x = 6 km and 17.5 km, while others, e.g., at x = 9.5 km and 20.5 km, are located above subvertical faults and are interpreted to be their upward extension. However, it is unclear if all subvertical, lower velocity anomalies arise from faulting, because no fault has been identified close to the vertical zone at x = 12 km, and some of these features may be artefacts similar to those observed in conventional migrations.

Localised low velocity anomalies, which also correlate with higher seismic attenuation values (Figure 2.12b), can be identified deeper in the section, and close to the upper Miocene-Pliocene unconformity, at x = 13 km, 15 km, 17 km, 20-21 km (Figure 2.17a). Broader, negative anomalies can also be identified in the deeper part of the perturbation model, e.g., at x = 11-24 km, 26.5-30 km and 35-40 km (Figure 2.17b). These anomalies either correlate with or closely overlie previously interpreted faults that extend up to the Miocene-Pliocene unconformity. We suggest that these low velocity anomalies represent fracture-related porosity attributable to faulting that extends further upward into the Pliocene section (Figure 2.17a). The sharp lateral changes in velocity between x = 36 km and x = 39 km overlie the tectonic inversion structure in the deep half-graben that has produced a small anticline in the unconformity with apex at x = 37 km. We interpret the lateral velocity variation in the overlying Pliocene section to be also due to deeper faulting that extends upward across the unconformity, for example at x = 36 km. At x = 39-40 km, the velocity model reveals a small flower structure, which is also the upward continuation.
of a deeper fault and indicates strike-slip faulting. The larger QCB has recently been subject to regional transpression, with strike-slip motion distributed across the basin (Rohr and Dietrich, 1992) and we therefore suggest that the faults cutting the Pliocene section are primarily strike-slip in nature.

The high lateral resolution of the velocity model is due to the inversion of refracted waveforms, which can include backscattered arrivals generated by steeply dipping velocity anomalies. Therefore the presence of subvertical anomalies in the model is not surprising. Many of the small-scale features in the velocity model correlate with structures previously identified in the migration. However, waveform tomography has also permitted the identification of other structures in the shallower Pliocene section that are not readily interpretable in the conventional migration. Given the tectonic environment, some of these features can be reasonably interpreted as faulting, but other structures are somewhat more enigmatic. 2D waveform tomography cannot account for elastic or 3D effects, such as out-of-plane propagation or scattering, which cannot be readily identified in our velocity model. However, our results demonstrate the potential of waveform tomography to resolve faults and their associated velocity anomalies where they are difficult to discern in conventionally processed seismic sections.
Figure 2.16: Migrated seismic section from line 88-06. The black line represents the Pliocene (P) to Upper Miocene (uM) unconformity and IM represents the Lower Miocene. The white vertical line is the total extent of the recorded sonic log; the black dashed line shows the fault locations and the red line is the basement of the basin. The section was derived and interpreted from Rohr and Dietrich (1992).
Figure 2.17: Migrated seismic section superimposed on (a), the time-converted velocity model from the 7-12 Hz full waveform tomography results and (b), the time-converted velocity perturbation model derived from the 7-12 Hz full waveform tomography results. The white vertical line shows the sonic log truncated to the maximum depth coverage of the velocity model. The black line represents the Pliocene (P) to the Upper Miocene (uM) unconformity. The black dashed line shows the location of previously interpreted faults and the white dashed line indicates proposed upward continuation into the Pliocene stratigraphy.
2.8 Conclusion

With an appropriate preconditioning of the data and a carefully selected inversion strategy, waveform tomography can be successfully applied to offset limited seismic reflection data with a starting frequency as high as 7 Hz. The preconditioning of the data is designed to convert the field data to a form similar to that predicted by the acoustic waveform modelling algorithm that is used. The preconditioning of the seismic data from the Queen Charlotte Basin comprised trace editing, coherent noise removal, shot-to-shot amplitude balancing and 2D amplitude correction. The starting P-wave velocity model was derived from ray-based travel time tomography, and care was taken to ensure that it predicted the first arrival waveforms of the field data to within half a cycle. The inversion strategy is intended to mitigate the non-linearity inherent in the inversion. This strategy utilised sequential time damping (from lower to higher values) for every frequency group, in combination with a layer stripping approach implemented by inverting from the near to far offsets, and weighting the gradient with depth to boost in the misfit function the contribution of the layer under reconstruction. The sequential time damping enables the successive inclusion in the inversion of more data that illuminate the deeper parts of the model. Two frequency discretisation schemes for a frequency range of 7-13.66 Hz were tested: the first, named full waveform tomography used 18 frequencies and the second, named efficient waveform tomography used only 5 frequencies with a significant reduction in computational cost. Both discretisation schemes converged to similar results, showing the robustness of the inversion strategy but the result of full waveform tomography appeared to have slightly greater resolution. Attenuation was introduced into the full waveform tomography approach for higher frequencies (≥ 10.5 Hz) and its introduction appears to significantly reduce x-shape artefacts at higher frequencies in the velocity model. The modelled data from waveform tomography fit the field data with a
high degree of accuracy. Higher seismic attenuation values identified in the attenuation model correlate with localized low velocity anomalies identified in the velocity model that are associated with faulting. A good agreement was generally observed between the velocity models, the sonic log and the migrated seismic section, indicating the reliability of the results. The velocity model derived by waveform tomography also permitted the identification of structures in the shallower Pliocene section that are not readily interpretable in the conventional migration.

2.9 Acknowledgments

We are grateful to Gerhard Pratt, who provided us with his waveform tomography code. Well logs from the Tyee N-39 were provided by Divestco. The seismic data were plotted with SeismicUnix and Gnuplot. SeismicUnix was also used partially to process the data. This project was funded by the Natural Sciences and Engineering Research Council of Canada.
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Chapter 3

Seismic velocity and attenuation structures of the Queen Charlotte Basin from full waveform tomography of seismic reflection data

A version of this manuscript was accepted for publication with minor revisions as:

3.1 Abstract

We apply visco-acoustic waveform tomography to four seismic reflection lines from the central and northern part of the Queen Charlotte Basin, and using frequencies of 7-12 Hz, we estimate the compressional velocity and attenuation above a depth of approximately 1.2 km. We refine our previously published inversion strategy by alternating between phase-only and amplitude-plus-phase velocity inversion for the first two pairs of frequencies used, and add a second step, in which we invert for attenuation from the lowest frequency using the final recovered velocity model and an initial homogeneous $Q_p$-model. Our recovered velocity and attenuation models show an overall good
correlation with the available sonic and gamma ray logs. Modeled seismic data matches
the field data well and 1D velocity and attenuation profiles extracted at line intersections
show a good correlation, thus demonstrating the robust nature of the results. Recovered
velocities aid in interpreting shallow structures not readily identifiable on the conventional
migration such as Quaternary strata and Pliocene faulting. Recovered attenuation values
in the sedimentary rocks are generally consistent with saturated sandstones and consistent
with the geology interpreted from well logs. Localized regions of elevated attenuation and
associated low velocities correlate with siltstones and shales, the presence of
hydrocarbons, or inferred increases in porosity due to fracturing. Seafloor pockmarks,
where venting of gas occurs, are underlain by low velocities and an anomalous attenuation
variation, and pipe-like gas chimneys are interpreted in two other areas of Hecate Strait.
Igneous basement is associated with high velocity and high attenuation in its uppermost
part, suggesting the presence of volcanic rocks, but the elevated attenuation may also be
due to scattering and elastic mode conversions not included in the visco-acoustic
inversion.

3.2 Introduction

The Queen Charlotte sedimentary Basin (QCB) is the largest Tertiary Basin on the west
coast of Canada with an area of approximately 80,000 km² (500 km long and 150-200
km wide) (Whiticar et al., 2003). This large basin is bounded to the south and to the
north by Vancouver Island and Alaska respectively, and is terminated to the east by the
Coast Plutonic Complex and to the west by the Queen Charlotte Fault, which separates
the North American Plate from the Pacific Plate (Woodsworth, 1991)(Figure 3.1). The
QCB comprises up to 6 km of Tertiary sedimentary and volcanic rocks above the igneous
basement. The basin underlies Dixon Entrance, Hecate Strait and Queen Charlotte Sound
along the western margin of British Columbia (Rohr and Dietrich, 1992).

The geology of the QCB, its evolution and the hydrocarbon potential of the basin have been thoroughly documented (e.g., Rohr and Dietrich, 1992; Woodsworth, 1991; Whiticar et al., 2003; Lyatsky, 1993). Information about the offshore geology has been obtained from 8 petroleum-exploration wells drilled in the basin in the 1960s, and from seismic reflection surveys, including one acquired in 1988 by the Geological Survey of Canada (GSC). In addition, many other geological and geophysical studies, including high-resolution bathymetry and seafloor sampling, have been carried out in the basin (e.g., Irving et al., 2000; Halliday et al., 2008; Barrie et al., 2011).

The purpose of this work is to further the study of the basin by quantitatively imaging its upper structure using 2D visco-acoustic frequency domain full waveform tomography applied to the limited offset seismic reflection data collected in 1988 by the GSC. P-wave velocities and inelastic attenuation models are derived from the inversion of 4 lines: 88-07, 88-06, 88-05 and 88-04 (Figure 3.1). These lines are located in Hecate Strait and partially in Dixon Entrance. The expected depth of coverage is not more than 1200 m due to the limited maximum offset of the data (3770 m). The top 1200 m of the QCB contains Tertiary rocks of the Skonun and Masset formations which are overlain by Quaternary rocks. The Skonun Formation comprises interbeded sandstones, shales and siltstones while the Masset Formation consists mostly of basalts and volcaniclastic rocks (Dietrich, 1995).

Waveform tomography has the potential to effectively image subsurface structure in highly complex areas. It can therefore have a significant impact in the hydrocarbon industry, where image distortions due to shallow gas and complex geology have hampered the effectiveness of conventional imaging techniques. The high-resolution of waveform tomography comes from the fact that all wave modes, which include diffractions, guided waves, reflection and other scattering effects are taken into account during the inversion procedure. Unfortunately, the inclusion of all waveform modes is both computationally
expensive and it increases the non-linearity of the inversion. The non-linearity can be mitigated by a good starting model, careful preconditioning of the data and the use of very low starting frequencies. When low starting frequencies are not available, an efficient inversion strategy must be selected to ensure sufficient convergence to the global minimum (e.g., Sirgue and Pratt, 2003; Takam Takougang and Calvert, 2010, 2011). Previous synthetic tests (e.g., Brossier et al., 2009; Barnes and Charara, 2008) have shown that acoustic waveform tomography can be effectively used with elastic data when the data are dominated by P-waves, and S-waves are of low amplitudes. This is the case in shallow marine environments, such as the QCB, where S waves and P-to-S conversions are weak due to the presence of soft sediments which imply a gradual change in S-wave velocity with depth, and small velocity contrasts. However, the inversion might be more challenging on the eastern side of the QCB where P-to-S wave conversion is expected at the top of the shallow igneous basement and at the seafloor.

We begin by reviewing the geological setting of the basin, focusing on its history, structure and stratigraphy. We then present the characteristics of the seismic survey, followed by the data preconditioning and inversion strategy. The preconditioning of the data and inversion strategy (full waveform tomography approach) was described in detail in Takam Takougang and Calvert (2011) refered to as paper 1. Therefore only a general review and some subsequent improvements to the methodology are provided here. The last part of the paper focuses on interpretation of the derived velocity and attenuation models. Our velocity perturbation models are also compared to conventional migrated sections from all the lines, and correlations between the velocity models, attenuation models and sonic logs are presented to check the consistency of the results. A comparison is also made between the field data and the synthetic data derived from the recovered velocity and attenuation models.
Figure 3.1: Location map of the Queen Charlotte Basin showing the seismic reflection lines acquired in 1988. The 4 lines used in this study are highlighted in red. The black dots indicate the location of wells. Wells Tyee N-39, Sockeye B-10 and Murrelet L-15 tie or are very close to the lines and were used for model assessment. The gray rectangles represent the study area of Halliday et al. (2008) (the big rectangle) and of Barrie et al. (2011) (the small rectangle) where pockmark structures were identified. QCF indicates the Queen Charlotte Fault and PLF the Principe-Ladero Fault.
3.3 Geological Setting

The Queen Charlotte Basin (QCB) lies east of the Queen Charlotte Fault, which forms part of the Pacific-North America Plate Boundary. The basin’s hydrocarbon potential is considered to be the greatest amongst the basins along the northwest margin of North America (Dietrich, 1995; Higgs, 1991). The geological history of the QCB is linked to the evolution of the Pacific continental margin and associated convergent and transcurrent plate interactions (Lewis et al., 1991; Rohr and Dietrich, 1992). The QCB first developed during the Eocene over the Middle to Late Triassic Wrangellia Terrane which, with other parts of the Insular Belt, accreted to North America during the Middle Jurassic (Van der Heyden, 1992). The Insular Belt is the westernmost tectonic belt of the Canadian Cordillera. The northern section of the Insular Belt is separated from the Pacific Plate by the dextral Queen Charlotte strike-slip fault.

There were three successive phases of basin development: transtension, transform-margin and transpression (Rohr and Dietrich, 1992; Irving et al., 2000). The transtensional phase started during the Eocene and lasted more than 25 million years. During this initial stage, most of the Miocene volcanics and associated sediments were deposited into a basin that developed due to east-west crustal extension (Irving et al., 2000). During the pure transform phase of basin development, east-west extension ceased, and igneous activity decreased and finally ceased. Subsidence continued during the Middle and Late Miocene, with continuous accumulation of sediments. The youngest transpressional phase, which commenced in the late Miocene or earliest Pliocene, was characterized by subsidence in the basin with relatively continuous sediment deposition. The relative plate motion became highly oblique and convergent, resulting in the uplift and erosion of the Queen Charlotte Islands (Irving et al., 2000).

Our area of study, the northern part of the basin, encompasses Hecate Strait and part of
Dixon Entrance. Miocene structures in Hecate Strait are mainly oriented northwest, and are sub-parallel to the plate boundary. Converging and diverging fault segments are present, which result in a complex network of faults and subbasins, typically half-grabens bounded on their west side by a master fault. Dixon Entrance is underlain by a set of fault-bounded subbasins separated by basement platforms, which outcrop or lie close to the seafloor (Rohr and Dietrich, 1992).

3.3.1 Structure and Stratigraphy

The QCB is underlain by Mesozoic and Tertiary volcanic, plutonic and sedimentary rocks (Dietrich, 1995). The structure of the basin can be subdivided into 3 major sections from bottom to top (Figure 3.2):

1. Lower Triassic-Lower Jurassic
   
   The igneous basement of the basin comprises volcanic rocks of the Triassic Karmutsen Formation, which are up to 4600 m thick (Dietrich, 1995). The Karmutsen strata are conformably overlain by up to 600 m of Upper Triassic and Lower Jurassic limestones, sandstones and shales of the Kunga Group, and approximately 300 m of deep marine shale, siltstones, sandstones and mudstones of the Lower Jurassic Maude Group. The Kunga and Maude groups are considered to be the principal oil source rocks in the basin (Dietrich, 1995).

2. Middle Jurassic-Cretaceous
   
   The Maude Group unconformably underlies 800 m of volcanics and volcaniclastic rocks of the Middle Jurassic Yakoun and Moresby groups, which lie unconformably below up to 2500 m of Upper Jurassic-Cretaceous sandstone, shale and conglomerate of the Longarm Formation and the Queen Charlotte Group. Middle Jurassic and older rocks are locally intruded by late Middle to Late Jurassic plutons. Upper parts of the
Queen Charlotte Group locally contain volcanic rocks (Dietrich, 1995).

3. Tertiary

The Tertiary formations are subdivided into Paleogene and Neogene strata. The Paleogene strata (Lower Tertiary) are 1000 m thick and comprise mostly volcanic flows with some conglomerates, black shale and sandstones. The Neogene strata comprise volcanic and sedimentary rocks of the Masset and the Skonun formations respectively. The Masset Formation comprises up to 4000 m of basalts, rhyolite flows, and pyroclastics (Dietrich, 1995). The Skonun Formation consists of interbedded sandstones, shales, conglomerates and lignites (coals), and reaches a thickness of 6 km in the offshore part of the basin. The Paleogene strata have limited reservoir quality, whereas the Masset Formation has good reservoir properties (Dietrich, 1995).

4. Quaternary

The Skonun and Masset formations are unconformably overlain by the Quaternary Cape Ball Formation, which contains glacial and post-glacial deposits. Glacial sediments are divided into three units: ice-contact tills, glaciomarine mud, and ice-distal mud to sandy-mud. This interpretation is based on surficial geology of the northwestern Canadian continental shelf, which includes the QCB (e.g., Luternauer et al., 1989; Barrie et al., 1991, 2006; Halliday et al., 2008).
<table>
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<th>Age</th>
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<tr>
<td>Lower</td>
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<td>Karmutsen</td>
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<td>Jurassic</td>
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<td>Lower</td>
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<td>Middle</td>
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<td>Cretaceous</td>
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<td>Lower</td>
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<tr>
<td>Paleogene</td>
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<tr>
<td>Neogene</td>
<td>4000</td>
<td>Skonun</td>
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<td>Tertiary</td>
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Figure 3.2: Generalized stratigraphy of the QCB (after Dietrich, 1995). (*) Section covered by waveform tomography.
3.4 **Seismic Reflection Surveys**

Eight regional 2D seismic reflection lines were acquired in 1988 by the GSC (Figure 3.1). These lines were shot offshore, using a 6358 in$^3$ airgun array and recorded using a 240-channel hydrophone streamer with a minimum offset of 185 m and a maximum offset of 3770 m. The shot and receiver intervals were 45 m and 15 m respectively, and the shot and receiver depths were 12 m. A bandpass filter was applied before digitalization: 8 Hz with a slope of 6 dB/Octave to 90 Hz with a slope of 72 dB/Octave. The four lines used for this study of Hecate Strait and partially Dixon Entrance are 88-04, 88-05, 88-06 and 88-07. Three exploration wells (Tyee N-39, Sockeye B-10 and Murrelet L-15) tie, or are very close to the intersection points of these seismic lines (Figure 3.1).

3.5 **Full Waveform Tomography**

We used the 2D visco-acoustic frequency domain waveform tomography approach as described in detail by Pratt et al. (1998), Pratt (1999) and Pratt and Shipp (1999). In essence, the objective of the inversion is to generate an update to a starting velocity, and when used, attenuation model, by iteratively minimizing the misfit between the forward modeled data and the field data, using the negative gradient. The gradient is computed by multiplying in the frequency domain the forward modeled wavefield by the backward propagated data residual.

For a successful waveform inversion, the convergence criteria implies that the first arrival traveltimes calculated using the starting velocity model should match first arrivals from the field data to within half a cycle (Sirgue, 2003). Also, since we are using a 2D acoustic wave propagation scheme to model the 3D propagation of elastic data, amplitude correction is necessary, as well as the removal of elastic modes such as ground roll or interface waves, not included in the acoustic wave propagation. The convergence criteria
can be formulated mathematically using the travel time error $\delta t$ relative to the total arrival time $T$ of the event as (Pratt, 2008):

$$\frac{\delta t}{T} \leq \frac{1}{2N_\lambda},$$

(3.1)

where $N_\lambda = l/\lambda$ is a dimensionless term representing the propagation distance in terms of wavelengths and $l$ is the propagation distance from the source to the receiver. Equation 4.1 shows that small values of $N_\lambda$ are essential; they correspond to lower frequencies or, for a given frequency, to short offset data. However, longer offset data are also essential, because they are required to constrain the low wavenumbers in the velocity model (Sirgue and Pratt, 2004). When low frequencies are missing from the data, a specific inversion strategy is necessary in addition of a very good starting model.

The QCB survey is characterized by a relatively high starting frequency (7 Hz) and a limited maximum offset (3770 m), and we designed a specific strategy to improve convergence to a likely global minimum. This strategy also recovers structures to a depth of approximately 1200 m, which would not otherwise have been possible due to the amplitude distribution in the data. Early arrivals typically have larger amplitudes than late arrivals and thus dominate the gradient function, from which the velocity model update is calculated. Consequently, the shallow parts of the velocity model contribute more to the misfit function than deeper parts and are thus more accurately reconstructed. As mentioned in the Introduction, a detailed description of this inversion strategy (full waveform tomography approach), as well as the preconditioning of the data are provided in paper 1. It should be noted that, unlike in paper 1, we applied waveform tomography to a maximum frequency of 12 Hz to reduce the computational cost, because the four lines used in this study are longer, 93-154 km, whereas the line in paper 1, which had a length of 45 km, was inverted to a maximum frequency of 13.66 Hz. Below is a summary of the
main steps involved:

### 3.5.1 Data Preconditioning

The main preconditioning steps of the 4 seismic lines are listed below:

1. **F-K filtering, trace editing and lowpass filtering**

   An F-K filter was used to remove steeply dipping coherent noise present in almost all the shot gathers. The data were then carefully inspected to remove bad traces and were finally lowpass filtered at 15 Hz to include only the frequencies necessary for the inversion.

2. **Shot-to-shot amplitude balancing and 2D amplitude scaling**

   In order to balance the energy of each shot gather, and consequently, ensure a uniform model update along a seismic line during the inversion, amplitudes in each shot gather were normalized to the same maximum value. This normalization does not change the relative variation in amplitudes within individual traces, or the amplitude variation with offset. The data were then scaled to simulate 2D modeling by first multiplying every trace by $\sqrt{t}$, and second by a scale factor obtained from a comparison of the logarithm of the Root Mean Square (RMS) amplitude variation with offset of the field and the modeled data. Finally, the data were muted 1.5 s after the first arrivals and a window of $T_w = 2$ s was used for the inversion.

3. **Data subsampling**

   For line 88-06, 88-05 and 88-07, the data were resampled to every second shot and every third receiver; this increases the shot interval to 90 m and the receiver interval to 45 m. For line 88-04, the shot interval was increased to every 4 shots giving a shot interval of 180 m, but the receiver interval remained equal to 45 m. This subsampling was carried out to reduce the computation time, and to avoid oversampling since
the maximum frequency used during the inversion is 12 Hz. These new geometries appear to be sufficient to avoid aliasing during waveform inversion, as discussed in paper 1.

### 3.5.2 Inversion Strategy

The starting model for each of the 4 seismic lines was derived using traveltime inversion of first arrival traveltimes (Aldridge and Oldenburg, 1993). The initial model for travel time inversion was calculated using a velocity gradient of 1.2 s$^{-1}$ in the sediment and a velocity gradient of 2.5 s$^{-1}$ to account for the transition from sedimentary to igneous rocks. The velocity of the water layer was kept constant at 1480 m$^{-1}$. 20 iterations were computed for each line and a root mean square misfit (rms) of approximately 5.8 ms was obtained for lines 88-06, 88-04 and 88-07 and 3.6 ms for line 88-05. These results predicted the first arrival traveltimes to within half a cycle and respected the condition of Equation 4.1 (See paper 1 for detailed explanations).

The waveform tomography inversion strategy consisted of successively recovering shallow to deep structures in the subsurface velocity and attenuation models (a layer stripping approach). Shallow structures were recovered using near offset and early arrivals and deeper structures were recovered by additionally including far offsets and late arrivals. This was accomplished by weighting the gradient with depth and weighting the data residual with offset to focus on the region under reconstruction. Complex frequencies were used for the selection of time aperture in the frequency domain data. A complex angular frequency can be defined as (Mallick and Fraser, 1987; Pratt et al., 2004):

$$\omega = 2\pi f + i/\tau.$$  \hspace{1cm} (3.2)

The imaginary component $1/\tau$ serves to damp the contribution of late arrivals in the data.
residual, and thus enables the selection of time aperture in the frequency domain data. For the shallow and the deep reconstructions, 2 consecutive values of $\tau$ were used. These values correspond to $\tau = 0.8$ s and $\tau = 1.6$ s, which respectively represent 40 % and 80 % of the input data. Pairs of frequencies to include in the inversion were selected every $1/T_w = 0.5$ Hz from 7 to 12 Hz. For every pair of frequencies, 5 iterations were performed for each selected $\tau$, making a total of 10 iterations for the shallow reconstruction and 10 iterations for the deep reconstruction (see Figure 3.3). The source signature was estimated as described in paper 1.

This inversion strategy was successfully applied in paper 1 to image a section of line 88-06. However, on line 88-05 we observed a lack of convergence in the inversion due to shadow zones in the input data, i.e. regions with very weak amplitudes. In those areas, the inversion has difficulty in recovering the correct phase of the data. For this reason, we refined our inversion strategy by alternating between phase-only and amplitude-plus-phase inversion for the early frequencies, i.e. frequencies ranging between 7 Hz and 8.5 Hz. The phase-only inversion was performed with $\tau = 0.8$ and the amplitude-plus-phase inversion with $\tau = 1.6$. No change in the inversion strategy was made for the other frequencies. The phase-only inversion improves the recovery of the correct phase in shadow zones at early stages of the inversion.

We also refined the attenuation inversion strategy. Instead of introducing attenuation only at higher frequencies ($f \geq 10.5$ Hz), which produced a heterogeneous attenuation model due to the lack of low frequencies, we introduced attenuation from the starting frequency (7 Hz). We therefore performed a 2 step inversion procedure: The first step consisted of inverting only for velocity from the lowest to the highest frequency, and the second step consisted of inverting for both velocity and attenuation. During the second step, mostly the attenuation model was updated, since a high resolution velocity model had already been obtained from the first step. This strategy enables us to obtain a more
laterally continuous attenuation model. The starting model for attenuation in all the lines was a homogeneous $Q_p$-model with $Q_p = 100$.

Figure 3.3: Schematic view of the inversion strategy used. The inversion is performed in 2 stages and the data are inverted in pairs from 7 Hz to 12 Hz with a step length of 0.5 Hz. During stage 1, only velocities are inverted and during stage 2, both velocities and attenuation are inverted. For frequencies 7-8.5 Hz, phase-only, and amplitude-plus-phase inversion are performed successively for every pair of frequencies to improve the fitting (stage 1a).

3.6 Waveform Tomography Results

The starting models and ray densities derived from first arrival traveltime tomography are displayed in Figures 3.4 and 3.5. Velocity, attenuation and perturbation velocity models derived using waveform tomography of lines 88-04, 88-05, 88-06 and 88-07 are displayed in Figures 3.6, 3.7, 3.8 and 3.9 respectively. The perturbation models (Figures 3.6c, 3.7c, 3.8c and 3.9c) were obtained for every line by subtracting the starting model from the final model and dividing the result by the starting model.

Generally the results show two main geological features: the shallow sedimentary stratigraphy and basement of mostly Triassic igneous rocks (Rohr and Dietrich, 1992). The sedimentary rocks are characterized by velocities of approximately 1750-3000 ms$^{-1}$ and by attenuation of $Q_p^{-1} \leq 0.04$ with areas of attenuation as high as 0.15 (Figure 3.10a),
whereas the basement rocks have velocities of 3500-5500 ms$^{-1}$ and inversion indicates attenuation generally between 0.01-0.07 (Figure 3.10b). Elevated attenuation ($Q_p^{-1} \geq 0.07$) is present mostly at the top of the igneous basement.
**Figure 3.4:** (a) 88-04 Starting velocity model derived by first arrival tomography, and associated ray density (b). (c) 88-07 Starting velocity model derived by first arrival tomography and associated ray density (d). The ray coverage is limited to approximately 600 m depth but becomes deeper in igneous basement due to a greater velocity gradient in the initial velocity model. The white line represents the seafloor topography.
Figure 3.5: (a) 88-06 Starting velocity model derived by first arrival tomography and associated ray density (b). (c) 88-05 Starting velocity model derived by first arrival tomography and associated ray density (d). The ray coverage is limited to approximately 600 m depth but becomes deeper in igneous basement due to a greater velocity gradient in the initial velocity model. The white line represents the sea floor topography.
Figure 3.6: 88-04 waveform tomography results. (a) velocity model; (b) attenuation model; (c) fractional velocity perturbation. Localized low velocity zones (LVZ) and regions with interpreted shallow faulting (F) are indicated. Arrows in (b) indicate regions of high attenuation in sediment. The white line represents the seafloor topography.
Figure 3.7: 88-05 waveform tomography results. (a) velocity model; (b) attenuation model; (c) fractional velocity perturbation and (d) zoomed velocity model between x = 31-69 km to highlight the pockmark structures and interpreted pipe-like gas chimneys. Localized low velocity zones (LVZ) and regions with interpreted shallow faulting (F) are indicated. Arrows in (b) indicate regions of high attenuation in sediment. Pipe-like gas chimneys are associated with low velocities and high attenuation. The white line represents the seafloor topography.
Figure 3.8: 88-06 waveform tomography results. (a) velocity model; (b) attenuation model; (c) fractional velocity perturbation and (d) zoomed velocity model between x = 0-32 km. Localized low velocity zones (LVZ), regions with interpreted shallow faulting (F) and antiform A1 are indicated. Arrows in (b) show regions with high attenuation in sediments and the white line represents the seafloor topography.
Figure 3.9: 88-07 waveform tomography results. (a) velocity model; (b) attenuation model; (c) fractional velocity perturbation. Localized low velocity zones (LVZ) and regions with interpreted shallow faulting (F) are indicated.
Figure 3.10: (a): Cross-plotting of attenuation ($Q_p^{-1}$) versus velocity ($V_p$) in sedimentary rocks for all four seismic lines; velocities in sedimentary rocks (1750-3000 ms$^{-1}$) are generally associated with attenuation values $Q_p^{-1} \leq 0.04$. Elevated attenuation ($\geq 0.05$) is due to localized low velocity zones. (b): Cross-plotting of attenuation ($Q_p^{-1}$) versus velocity ($V_p$) at igneous basement; velocities at igneous basement (3500-5000 ms$^{-1}$) are associated with attenuation values $Q_p^{-1} = 0.01$-0.07, and attenuation as high as 0.1 is present mostly at top of basement.
3.6.1 Model Consistency

1D Profiles at Line Intersections

The extracted 1D velocity profiles are generally similar at line intersections (Figure 3.11), thus showing the robustness of the inversion strategy selected. Small differences between various lines at the same well may be related to slight differences in the starting models used for the different lines. The extracted 1D attenuation profiles at the intersection of the lines are also generally similar (Figure 3.12), with the major horizons being consistently recovered in all the lines. However, it is clear that the attenuation results are not as robust as those for velocity. Attenuation models mostly depend on the amplitudes of waveforms in contrast to the velocity models which mostly depend on arrival time and phases of the waveforms. Since a number of variable factors not related to attenuation such as source and receiver coupling, coherent noise and radiation patterns affect the amplitude of waveforms, the inverted attenuation models are likely to exhibit greater variation than velocity models. Also, although we corrected the amplitudes of the 3D elastic input data to behave similar to 2D acoustic data, some differences may still persist and affect our estimate of attenuation. However, our recovered attenuation models are more consistent than the attenuation model obtained in paper 1, using a section of line 88-06. This difference is due to the fact that in paper 1, we recovered the attenuation model starting from 10 Hz; whereas in this study we recovered the attenuation models from the minimum frequency, 7 Hz. These results therefore show that a better estimate of attenuation can be obtained when low frequencies are available in the field data.
Figure 3.11: 1D profiles of waveform tomography velocity models at the intersection between line 88-04 and line 88-05 (a), line 88-04 and line 88-06 (b), line 88-04 and line 88-07 (c), line 88-06 and line 88-07 (d). The good agreement between various profiles at the intersection of lines shows the robustness of the inversion strategy.
Figure 3.12: 1D profiles of waveform tomography attenuation models at the intersection between line 88-04 and line 88-05 (a), line 88-04 and line 88-06 (b), line 88-04 and line 88-07 (c), line 88-06 and line 88-07 (d).
Synthetic Modeling

Synthetic and field data are also used to check the reliability of the results. Synthetic data were obtained after forward modeling in the frequency domain using the velocity and attenuation models from waveform tomography at 12 Hz, and the results are compared using a true-amplitude display with the field data in a common offset gather corresponding to an offset of 2.915 km. Comparison in the common offset domain has the advantage of clearly showing the match along the entire line. Only the comparisons for lines 88-05 and 88-06 are shown (Figures 3.13 and 3.14), but similar results were obtained for the remaining lines. In the sedimentary section, the arrival times, the waveform shapes of both the first and late arrivals within the 2 s window, and the amplitude variations are well reproduced in the synthetic data, between receiver locations 0-57 km on line 88-06 and between 0-67 km and 86-92 km on line 88-05. Diffraction patterns are well reproduced at receiver locations 50-56 km on line 88-05 and at 40-46 km on line 88-06. Shadow zones, i.e. areas with weak amplitudes in the sedimentary section are also well reproduced at 0-5 km and 40-46 km on line 88-05 (Figures 3.13a and 3.13b). Just below the seafloor, the delayed arrivals, centered at 48 km and the overlaying shadow zone at 40-46 km on line 88-05 are also well reproduced in the synthetic data using a nearer offset of 2.015 km (Figures 3.13c and 3.13d). Where the igneous basement is shallow, i.e. at 58-88 km on line 88-06 and 70-86 km on line 88-05, the match between synthetic and field data is poor. For example, late arrivals at 64-66 km and 74-78 km on line 88-06 are not well reproduced in the synthetic data. This is not a surprising result as P-waves propagating in this area are strongly attenuated, partially due to heterogeneity, but amplitudes are also reduced by S-wave conversion, which is not included in the inversion. Converted S-waves are expected to be stronger in this area due to the greater velocity contrast between low velocity sedimentary and high velocity igneous rocks.
Figure 3.13: 88-05 common offset gather (offset = 2.915 km) of (a) field data low pass filtered to 12 Hz and (b) modeled data calculated with the velocity and attenuation models derived from the 12 Hz inversion. There is a high degree of similarity between the modeled data and the field data, within the 2 s window, mostly in the sediment, i.e. between 0-70 km and 90-95 km. At offset = 2.015, the low velocity region beneath first arrivals at 47-50 km in the field data (c). The feature is reproduced in the modeled data (d). Reduction velocity $v = 5000 \text{ ms}^{-1}$.
**Figure 3.14:** 88-06 common offset gather (offset = 2.915 km) of (a) field data low pass filtered to 12 Hz and (b) modeled data calculated with the velocity and attenuation models derived from the 12 Hz inversion. There is a high degree of similarity between the modeled data and the field data, within the 2 s window in the sediment, i.e. between 0-57 km. At the basement, i.e. between 58-90 km, the match is still good for the early arrivals, but poor at late times. The reduction velocity used is $v = 4000 \text{ ms}^{-1}$.

### 3.6.2 Well Ties

**Sonic Logs**

In order to assess the quality of the waveform tomography results, a comparison was made between the available sonic logs and 1D velocity profiles extracted from the seismic lines. Three seismic lines, 88-04, 88-06 and 88-07 intersect near the Tyee N-39 well. 1D velocity profiles were extracted from the starting and final velocity models and superimposed on the sonic log from the well (Figure 3.15). A generally good match exists between the sonic log and the 12 Hz final models, showing the reliability of the results. The most impressive result is obtained with line 88-06 (Figure 3.15a) where the Pliocene/upper-Miocene unconformity
is characterized by a decrease in velocity, and the presence of coal at 660 m is well resolved. This match is actually better than the one obtained in paper 1, where a 50-m offset was observed between the sonic log and the final velocity profile near the unconformity; this improvement is explained by the changes made to the inversion strategy. A mismatch is however evident at 1100 m where an increase in velocity of approximately 200 ms$^{-1}$ is observed. On lines 88-04 and 88-07 (Figures 3.15b and 3.15c), the low velocity layer at 550-600 m is well resolved, but the underlying higher velocity layer at 600-660 m is not fully recovered. These mismatches may be related to anisotropy or to the increase in non-linearity in the inversion: small changes in layer thickness and velocity can have a disproportionate effect on the data, leading to convergence problems (Pratt et al., 2004).

The Sockeye B-10 and the Murrelet L-15 wells were superimposed on the velocity models from lines 88-05 and 88-04 respectively (Figures 3.15d, 3.15e and 3.15f). A generally good match exists between the Murrelet L-15 sonic and line 88-04 where the starting 1D velocity profile matched the sonic velocities well. The tie of the Sockeye B-10 sonic log with line 88-05 (Figure 3.15d) and line 88-04 (Figure 3.15e) shows a good recovery of the decrease in velocity at 950 m, associated with oil staining, but there are mismatches at 200-750 m on line 88-05, with velocities from the 1D profile faster than the sonic velocities and an unexplained increase in velocity at 400-550 m in the 1D profile from line 88-04. The Sockeye B-10 well, however, does not exactly tie to lines 88-05, and 88-04 and the location of the well was projected to the nearest points on the respective lines where the 1D profiles were extracted. The Sockeye B-10 well is approximately 2 km from line 88-04 and 0.5 km from line 88-05. In contrast, the Tyee N-39 well is only 0.2 km from line 88-06. Consequently, the mismatches observed between the sonic velocity of the Sockeye B-10 well and the 1D velocity profiles from lines 88-04 and 88-05 may be related to differences in geological structure between the well location and the projected location on the lines. In all cases, the final 7-12 Hz waveform tomography model fit the sonic log
better than the starting model from traveltime tomography.

Figure 3.15: (a) to (f): Comparison of sonic logs with velocities estimated at well locations by waveform tomography. The waveform results correlate with the sonic log better than the starting model derived from traveltime tomography, and there is a general good agreement between the waveform tomography results and the sonic logs. The lithologies (S: sandstone, Sl-Si: Silstone and shale) are derived from Shouldice (1971); Q is Quaternary.

Estimated Clay Content

Intrinsic seismic attenuation can potentially predict rock properties such as porosity, permeability and clay content. A relationship linking attenuation ($Q_p^{-1}$) and clay content
in sandstones was shown by Klimentos and McCann (1990), and can be written as:

\[ Q_p = 179C^{-0.843}; \]  

(3.3)

where \( C \) is the percent clay content by volume. We used this relation (3.3) to estimate the percentage of clay content in the sedimentary rocks at two well locations, where siltstones and shales were found in the well (Murrelet L-15 and Tyee N-39 wells). The results are compared with the estimated percentage of clay from gamma ray logs at the respective wells using the relation of Steiber (1973) (Figure 3.16). It is obvious that, although the percentage of clay estimated from the gamma ray logs is greater in places than that from the seismic attenuation values, a clear correlation exists. The peak in clay volume at 700-900 m depth at the Tyee N-39 well (Figures 3.16b and 3.16c) is relatively well resolved. Also, layers with alternating higher and lower clay content estimated from attenuation values at the Murellet L-15 well correlate consistently with the estimated clay content from the gamma ray log (Figure 3.16a). It should be noted that these estimates are theoretical and that the percentage of clay estimated from the attenuation values and from the gamma ray logs requires geological calibration. However, the correlation between the two estimates demonstrates the potential use of inverted attenuation models to estimate physical properties.
3.7 Structure and Lithology of the Queen Charlotte Basin

In this section, we will start by reviewing the structures and faulting previously identified in the basin from migrated sections (Rohr and Dietrich, 1992; Whiticar et al., 2003). Then, we will describe the general stratigraphy and faulting of the basin, in the context of the
waveform tomography models. We will use the interpretation of Rohr and Dietrich (1992) and the superposition of time converted waveform tomography velocity models on migrated sections to interpret the shallow basin stratigraphy, which mainly includes the Pliocene and the Quaternary sections. The derived seismic attenuation will be characterized, and interpreted in terms of previous estimates for sedimentary and igneous rocks.

### 3.7.1 Seismic Reflection Profiles

#### Line 88-04

In southern Hecate Strait, line 88-04 intersects the Murrelet L-15 well, which was drilled at x = 32 km (near CDP = 2008) into a deep half-graben containing up to 5.5 km of Tertiary strata (Rohr and Dietrich, 1992) (Figure 3.17). The half-graben bounding fault, which is located at x = 39 km (CDP = 2475), defines the southeast side of Moresby Ridge, a basement high comprising Mesozoic volcanic rocks and overlain by less than 500 m of Pliocene sedimentary rocks (Rohr and Dietrich, 1992). At 28-36 km (CDP = 1742-2275), the half-graben fill has been folded and inverted into an anticline during the Late Miocene and Pliocene. At 70-105 km (CDP = 4542-6875), the extensional basin has also been inverted, resulting in a succession of faults, which have uplifted Miocene strata, e.g., at x = 82, 84, 88 and 96 km. Tertiary strata to the north-west are extensively disrupted by upward branching faults with various dips. Antiform A1, which is also identified on lines 88-06 and 88-07 occurs between 135 and 154 km.

#### Line 88-05

At x = 23 km, CDP = 1408, a fault is imaged on the migrated section of line 88-05 (Figure 3.18). The hanging wall of this fault has been inverted into a fold, which was penetrated by the Sockeye B-10 well at location x = 18.3 km. An antiform, between x =
54-60 km, was created by inversion of the southwest part of a half-graben (CDP = 3600-4800), and is cut at x = 60 km (CDP = 3875) by an east dipping fault. Between x = 46-78 km (CDP = 2942-5055), upper Miocene stratigraphy thins to the north-east and onlaps the Principe-Ladero Basement High (PLBH) at x = 79 km. Basement rocks of the PLBH occur just beneath the seafloor from x = 79 km to x = 87 km. A steeply-dipping extensional fault on the north-east margin of the PLBH marks the southwest edge of another graben. Two faults at x = 90 km and x = 93 km locally offset the graben fill.

**Line 88-06**

The Tyee N-39 well, which is located at x = 19.7 km (CDP = 1188), penetrated a faulted NW-trending anticline, known as the Tyee structure (Rohr and Dietrich, 1992) which occurs in Miocene strata, between x = 18 km to x = 35.2 km (CDP = 1075-2222) (Figure 3.19). The Tyee structure is also imaged on lines 88-07 and 88-04. In the overlying Pliocene strata, the sedimentary rocks have been folded into an antiform (A1) between x = 14.3 km and x = 35.2 km (CDP = 828-2222), and also folded at x = 11.5 km; these strata are cut by upward branching faults as a consequence of Pliocene and Late Miocene deformation. At x = 18 km (CDP = 1075), a sub-vertical fault cuts the Pliocene rocks and delimits the Tyee structure to the west. South-west of the Tyee structure, sedimentary rocks shoal toward Moresby Island, and are cut by vertical faults at x = 4.5, 5 and 9 km. Between x = 36 and 38.5 km (CDP = 2275-2442), a small anticline in the upper-Miocene strata with apex at x = 37 km (A2 in Figure 3.19) originated in the Miocene by shortening of a deep half graben centered near x = 40 km. A small faulted anticline is present at x = 50-56 km. From x = 58 km to the end of the line, Tertiary strata thin and onlap the southwestern flank of the PLBH, where igneous basement rocks occur at depths as shallow as 250 m.
Line 88-07

North-east of the Tyee structure imaged between x = 10-26 km (CDP = 542-1608), line 88-07 crosses a basement high at CDP = 3000-3501, which separates two subbasins with different structural characteristics (Figure 3.20). South-west of the high, Miocene strata are unconformably overlain by flat-lying Pliocene strata. North-east of the basement high, both Miocene and Pliocene sections are folded in a broad anticline centered at x = 66 km (CDP = 4274) and cut by a vertical fault at x = 63.5 km (CDP = 4108) that offsets underlying basement rocks. North-east of this structure, the PLBH and adjacent graben (Feeney basin) occur at CDP = 5000-7000. The Feeney basin was interpreted by Rohr and Dietrich (1992) as a basin that may have initially formed as a graben, and was then asymmetrically uplifted and eroded. Several Miocene faults are present along the line. Pliocene deposits pinch out on basement highs between subbasins, which become shallower to the north-east, greatly reducing the thickness of the Pliocene sediments to only few tens of meters in Dixon Entrance. Line 88-07 crosses the Principe Ladero Fault (PLF) at CDP = 7001.

Summary

In much of the basin, sedimentary strata are folded, with several anticlines present, and cut by a complex network of faults, which have created a number of Miocene sub-basins that are partially inverted and unconformably overlain by Pliocene strata. To the east and north, the thickness of Pliocene sediments decreases to almost zero.
Figure 3.17: (a) 88-04 interpreted migrated seismic section (Rohr and Dietrich, 1992). The black lines represent the Pliocene (P) to upper Miocene (uM) unconformity, and the black dashed lines show the location of previously interpreted faults; (IM) represents the lower Miocene. The white vertical line is the total extent of the well; the brown and purple lines are the top of basement and volcanic rocks respectively. (b) The same migrated seismic section truncated at 1.5 s and superimposed on the time-converted velocity model from the 12 Hz full waveform tomography result. The white lines show the interpreted Quaternary (Q) to Pliocene (P) unconformity and the white dashed lines indicate interpreted faults, based on waveform tomography results.
Figure 3.18: (a) 88-05 interpreted migrated seismic section (Rohr and Dietrich, 1992). The black lines represent the Pliocene (P) to upper Miocene (uM) unconformity, and the black dashed lines show the location of previously interpreted faults; (lM) represents the lower Miocene. The white vertical line is the total extent of the well; the brown and purple lines are the top of basement and volcanic rocks respectively. (b) The same migrated seismic section truncated at 1.5 s and superimposed on the time-converted velocity model from the 12 Hz full waveform tomography result. The white lines show the interpreted Quaternary (Q) to Pliocene (P) unconformity and the white dashed lines indicate interpreted faults, based on waveform tomography results.
Figure 3.19: (a) 88-06 interpreted migrated seismic section (Rohr and Dietrich, 1992). The black lines represent the Pliocene (P) to upper Miocene (uM) unconformity, and the black dashed lines show the location of previously interpreted faults; (lM) represents the lower Miocene. The white vertical line is the total extent of the well; the brown and purple lines are the top of basement and volcanic rocks respectively. (b) The same migrated seismic section truncated at 1.5 s and superimposed on the time-converted velocity model from the 12 Hz full waveform tomography result. The white lines show the interpreted Quaternary (Q) to Pliocene (P) unconformity and the white dashed lines indicate interpreted faults, based on waveform tomography results. Antiforms A1 and A2 are also indicated.
Figure 3.20: (a) 88-07 interpreted migrated seismic section (Rohr and Dietrich, 1992). The black lines represent the Pliocene (P) to upper Miocene (uM) unconformity, and the black dashed lines show the location of previously interpreted faults; (lM) represents the lower Miocene. The white vertical line is the total extent of the well; the brown and purple lines are the top of basement and volcanic rocks respectively. (b) The same migrated seismic section truncated at 1.5 s and superimposed on the time-converted velocity model from the 12 Hz full waveform tomography result. The white lines show the interpreted Quaternary (Q) to Pliocene (P) unconformity and the white dashed lines indicate interpreted faults, based on waveform tomography results. LVZ are localized low velocity zones.
3.7.2 Stratigraphy and Lithology from Velocity Models

The seafloor, whose depth was measured during the seismic survey with an acoustic echo-sounder, is represented in the velocity models by a fractional velocity perturbation of approximately -0.1, at x = 0-65 km, 40-95.76 km and 50-92.79 km on lines 88-04 (Figure 3.6c), 88-05 (Figure 3.7c) and 88-06 (Figure 3.8c) respectively. This negative velocity perturbation may be a consequence of the acoustic waveform inversion of elastic seismic data generated by the seafloor. The inversion may accommodate the change in amplitude and phase due to mode conversion at the seafloor by slightly underestimating the velocity.

Deeper and broader, at approximately 500-600 m depth between x = 70-144 km, 10-40 km and 4-34 km on lines 88-04 (Figure 3.6c), 88-05 (Figure 3.7c) and 88-06 (Figure 3.8c) respectively, the top of a fractional velocity perturbation of approximately -0.1 lies close to the previously interpreted Pliocene/upper Miocene unconformity. However, the correlation is not exact because the perturbation is also controlled by the velocity gradient of the starting model. On line 88-07 for example, there is no significant negative velocity perturbation at the Pliocene/upper Miocene unconformity at x=0-24 km (Figures 3.9c and 3.20b), because the starting velocity gradient is close to the final model (Figure 3.15c). This result demonstrates the potential ambiguity that can arise from using the perturbation models for interpretation of basin stratigraphy, although horizons often appear more continuous.

At approximately 0.4-0.5 km depth, a vertical increase in velocity from 2300 ms\(^{-1}\) to 2600 ms\(^{-1}\), which does not correlate with the deeper Pliocene/upper Miocene unconformity is observed in sedimentary rocks (x = 0-50 km) on line 88-06 (Figures 3.8a and 3.19). We interpret this increase in velocity as the Quaternary/Pliocene unconformity (Figure 3.19), because 450 m of Quaternary rocks were identified in the Tyee N-39 well located at x = 19.7 km (Shouldice, 1971). Although not originally interpreted on the
migrated sections, more recent work on the surficial geology of the offshore Canadian shelf, including Dixon Entrance, Hecate Strait and Queen Charlotte Sound has identified Quaternary sediments a few hundred meters thick, underlain by Pliocene sediments (Luternauer et al., 1989; Barrie et al., 1991, 2006; Halliday et al., 2008). We also interpret the shallow increase in velocity on line 88-05 between 0 and 70 km (Figure 3.18a) and 88-07 between 0-70 km (Figure 3.20a) as the Quaternary/Pliocene unconformity.

Using the Pliocene/upper Miocene unconformity identified by Rohr and Dietrich (1992) and our interpretation of the Quaternary/Pliocene unconformity, it appears that Quaternary sedimentary strata are commonly characterized by velocities in the range 1750-2300 ms\(^{-1}\). The underlying Pliocene strata have velocities in the range 2300-2700 ms\(^{-1}\), and the upper Miocene strata 2700-3000 ms\(^{-1}\). Velocities generally increase with depth due to compaction and lithification. On line 88-04, at 0.3-0.7 km depth and between x = 70-105 km (Figure 3.17), the upper Miocene section exhibits lower velocities (2100-2800 ms\(^{-1}\)) than expected. These low velocities may be due to the abundance of faults in the region which may have increase fracturing and porosity in the sedimentary rocks and consequently reduced the velocity. The igneous basement, which is mostly Triassic, and can be identified on the migrated sections, exhibits velocities in the range 3500-5000 ms\(^{-1}\). The high velocity at the basement may be due to volcanic or granitic rocks. However, it is not possible using seismic velocity alone to clearly identify the nature of these rocks, because both volcanic and granitic rocks have a similar velocity range. We will constrain the possible nature of these rocks using inverted attenuation models in a later section.

Subparallel sedimentary layering is evident in the velocity models; for example on line 88-05 at x = 47-78 km and 87-95 km (Figures 3.7a and 3.7c), on line 88-06 at x = 4-36 km (Figure 3.8a), line 88-04 at x = 0-36 km and 120-134 km (Figures 3.6a and 3.6c) and line 88-07 at x = 2-10 km (Figure 3.9c) with velocities ranging between 1800 and
2300 ms\(^{-1}\) and fractional perturbation between 0.1 and 0.2. These shallow sediments are nonmarine as shown by the exploration wells Tyee N-39, Sockeye B-10 and Murrelet L-15 (Shouldice, 1971; Rohr and Dietrich, 1992). The presence of layering suggests fairly continuous, widespread sediment deposition during the Quaternary and the Pliocene.

### 3.7.3 Interpretation of Faults

Structural features identified in the velocity and velocity perturbation models may be due to faults, folds, anticlines or synclines. Correlation of the velocity models with interpreted migrated sections suggests that subvertical sharp lateral changes in velocity are commonly related to faults (e.g., Takam Takougang and Calvert, 2011). On all the lines, most of the upper Miocene faulting identified on the migrated sections correlates with subvertical sharp lateral decreases in velocity. We therefore interpret such changes in velocity to be due to faults. Lateral changes in the velocity perturbation of line 88-04 between 70-152 km, due to changes in velocity of approximately 3000-2500 ms\(^{-1}\) at 0.6 km depth, correlate with deformation observed on the migrated section (Figures 3.6 and 3.17); perturbation contrasts between -0.1 to 0.1 can be identified at x = 74, 78, 82, 84, 86, 96, 100, 106, 118, 123, 130 and 139 km (Figure 3.6c). We interpret these changes in velocity to be due to faults which correlate with previously identified faults on the migrated section. These faults result from the inversion of the extensional basin at 70-105 km, and from the formation of anticline A1 at 130-140 km (Rohr and Dietrich, 1992). Sudden northwest dipping decreases in velocity at 4 km and 143 km, which do not correlate with Miocene faulting are also interpreted as faults (Figures 3.6c and 3.17).

On line 88-06, below 0.6 km depth, lateral velocity variations in the perturbation model with values of -0.1 to -0.05 at x = 13.5, 18, 21, 23 km correlate with previously identified faults in the Miocene stratigraphy (Figures 3.8c and 3.19). We interpret some lateral velocity changes in the overlaying Pliocene section as the upward continuation of
these faults across the Pliocene/upper Miocene unconformity, associated with the formation of the Tyee structure. These faults cross antiform A1, and some appear to correlate with a shallow subvertical positive velocity perturbation of approximately 0.1 (Figures 3.8c). Usually, faults are associated with a decrease in velocity. So these sub-vertical features may be artifacts similar to those observed in conventional migration. Two faults, at x = 36 and 38.5 km are also interpreted as the upward continuation of Miocene faulting across the unconformity, and associated with the shortening of a deep half-graben centered near 39.7 km.

On line 88-07, (Figures 3.9a and 3.20) lateral variations in velocity, at x = 9 and 16 km cut antiform A1 and are interpreted, as on line 88-06, to extend upward across the Pliocene/Miocene unconformity because they correlate with Miocene faulting. Sharp lateral changes in velocity, across and close to antiform A1 at x = 12, 14, 22 and 26 km are also interpreted as faults. A localized steeply dipping zone of lower velocity at 77 km, which is also associated with negative fractional velocity of approximately -0.1 (Figure 3.9c) is interpreted as a fault.

On line 88-05, a fault, imaged on the migrated section at x = 23 km (CDP = 1501) (Figure 3.18) can also be recognized on the perturbation velocity model at x = 23-24 km where east-dipping sedimentary stratigraphy are disrupted by a change in the velocity perturbation image of -0.1 to 0.15 (dashed line in Figure 3.7c). Both the hanging wall and the footwall of the fault were subject to deformation, and in particular, from 0.6 km depth downward, between x = 24 to 30 km, east deeping Miocene sedimentary deformation is evident on the velocity perturbation model, shown by alternating horizons with a magnitude of -0.1 to 0.1. Subvertical velocity anomalies of approximately 2500 ms$^{-1}$, corresponding to perturbation velocities of approximately -0.1 at x = 4, 6, 8, 47 and 51 km, are interpreted to be due to Pliocene faulting, and some of these faults (x = 6, 51 km), which correlate with previously identified faults in the Miocene, are interpreted as the
extension of Miocene faults (Figure 3.18). Deeper, at x = 32 km, a sharp lateral change in velocity from 2500 ms\(^{-1}\) to 3000 ms\(^{-1}\) (Figure 3.7a) corresponds on the migrated section to the location of a west-dipping fault (approximately CDP = 2100). The fault at x = 59 km, which marks the northeast margin of a small anticline, is also interpreted to continue upward into the Pliocene section.

3.7.4 Seismic Attenuation in Sediment

Well log reports indicate that the sedimentary strata comprise clastic rocks composed mostly of sandstones with some siltstones, mudstones, clay and coal. The attenuation values recovered in these sedimentary rocks, \(Q_{p}^{-1} \leq 0.04\) (Figures 3.6b, 3.7b, 3.8b and 3.9b), are generally consistent with laboratory measurements of saturated sandstone (Winker and Nur, 1979) and therefore consistent with the well logs (Figure 3.15). However, regions of locally elevated attenuation (\(Q_{p}^{-1} \geq 0.06\)) are identified in the sedimentary rocks and these regions appear to correlate with localized low velocity anomalies with values of 2200-2600 ms\(^{-1}\) (Figure 3.10a), for example at approximately 800 m depth, between x = 2 and 38 km on line 88-06 and 88-05, and at x = 2, 24-30, 74-102, 112-154 km on line 88-04. Layering, with attenuation values of \(Q_{p}^{-1} = 0.03-0.04\) is present on lines 88-06, 88-05 and mostly 88-04. However, on line 88-07, between 38 and 133 km, the sedimentary rocks with attenuation values \(Q_{p}^{-1} = 0.03 – 0.04\) are less continuous along the line and this observation is consistent with the velocity stratigraphy (Figure 3.9) within the same interval. These differences in attenuation stratigraphy, may be related to a variation in the nature of sedimentary rocks, from Hecate Strait (lines 88-04, 88-05, 88-06) to southern Dixon Entrance (line 88-07). Localised zones with high attenuation values (\(Q_{p}^{-1} = 0.06\)) along the line, for example in the Feeney basin may be due to the presence of volcaniclastic sediments.
3.7.5 Seismic Attenuation in Igneous Rocks

Figures 3.6b, 3.7b, 3.8b and 3.9b show high inverted attenuation values in the igneous basement on all lines (88-06 at x = 62-92.79 km, 88-05 at x = 70-95.76 km, 88-04 at x = 0-40 km, 50-155 km and 88-07 at x= 48-54 km, 76-88 km and 108-133 km). In these areas, there appears to be a correlation between high attenuation values ($Q_p^{-1} = 0.05-0.15$) and high velocity ($\geq 3500$ ms$^{-1}$). The recovered attenuation values may be too high in places due to elastic amplitude losses in the field data that are not included in the viscoacoustic forward modelling. Mode conversions where large physical property contrasts are present, such as at the top of igneous basement, may thus increase our estimates of attenuation because this amplitude loss can be incorporated into the attenuation. Most of the attenuation estimated for the igneous basement is in the range $Q_p^{-1} = 0.01-0.07$ (Figure 3.10b). These attenuation values, which correspond to low $Q_p$ values ($Q_p = 14-100$), are consistent with laboratory measurements of intrinsic attenuation in oceanic basalts under saturated conditions and elevated pore pressure (100 MPa). $Q_p$ values range from 8 for high porosity Lau Basin basalt to 85 for Atlantic oceanic crust (Wepfer and Christensen, 1990). Our estimated attenuation and velocity values are therefore generally consistent with volcanic rocks in the upper part of the igneous crust. This interpretation is consistent with previous geophysical studies of the Moresby Ridge, a shallow basement block in the western part of the QCB basin, where high gravity and magnetic anomalies have been interpreted to indicate the presence of Mesozoic volcanic rocks (Young, 1981), and to other related studies (Table 3.1). Generally, attenuation in basalt is expected to be low ($Q_p=400-600$), but attenuation can be significantly higher where porosity is great. The high attenuation values (low $Q_p$) shown in Table 3.1 are interpreted to be due to a high porosity and alteration (Wepfer and Christensen, 1990, 1991).
Table 3.1: Intrinsic \(Q_{int}\) and effective \(Q_{eff}\) attenuation values for respectively laboratory experiments (Lab) using ultrasonic pulse-echo and in-situ measurement of basalts using Vertical Seismic Profile (VSP).

<table>
<thead>
<tr>
<th>Locations</th>
<th>Data</th>
<th>(Q_{int})</th>
<th>Authors</th>
</tr>
</thead>
<tbody>
<tr>
<td>Juan de Fuca basalt</td>
<td>Lab</td>
<td>31</td>
<td>Wepfer and Christensen (1990, 1991)</td>
</tr>
<tr>
<td>Juan de Fuca basalt</td>
<td>Lab</td>
<td>11-17</td>
<td>Tompkins and Christensen (1999)</td>
</tr>
<tr>
<td>Oman Ophiolite basalt</td>
<td>Lab</td>
<td>124</td>
<td>Wepfer and Christensen (1990, 1991)</td>
</tr>
<tr>
<td>Lau Basin basalt</td>
<td>Lab</td>
<td>8</td>
<td>Wepfer and Christensen (1990, 1991)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Locations</th>
<th>Data</th>
<th>(Q_{eff})</th>
<th>Authors</th>
</tr>
</thead>
<tbody>
<tr>
<td>Faroe lower series (Lopra)</td>
<td>VSP</td>
<td>13-24</td>
<td>Ker and Rodriguez (2002)</td>
</tr>
<tr>
<td>Norwegian margin (hole 642E)</td>
<td>VSP</td>
<td>25</td>
<td>Rutledge and Winkler (1989)</td>
</tr>
<tr>
<td>Rockall Trough 164/07-1</td>
<td>VSP</td>
<td>15-35</td>
<td>Maresh et al. (2006)</td>
</tr>
</tbody>
</table>

3.8 Discussion

Although there is generally a good correlation between velocity models derived by waveform tomography and the migrated seismic sections, we are able to identify a number of features in the velocity models that cannot be readily seen on the migrated sections. These relatively shallow features, which include shallow Pliocene faulting, affect refracted arrivals that are typically muted during conventional seismic data processing, but are included in the waveform tomography. Thus waveform tomography complements interpretations derived from conventional migrated sections. In addition, the seismic velocity of certain geological features such as the apex of antiforms can be quantified. Since visco-acoustic inversion produces estimates of both velocity and attenuation in the subsurface, it is also possible to identify anomalous regions by cross-plotting these two parameters (Figure 3.10). These anomalous regions (high attenuation associated with low velocity) correspond to geological features such as increase in clay content, seafloor pockmarks, gas chimneys and oil staining. Below, we present these features and discuss their possible nature and origin.
3.8.1 Faults Extending up into Pliocene

In general, many of the faults identified on the migrated sections by Rohr and Dietrich (1992) and Whiticar et al. (2003) are restricted to below the Pliocene/upper Miocene unconformity. These faults developed during the transtensional phase of basin formation, which ended in the late Miocene (Rohr and Dietrich, 1992). Using the velocity models derived by waveform tomography, some of these Miocene faults are interpreted to extend shallower into the Pliocene section. These faults were identified at x = 36 and 38.5 km on line 88-06, across antiform A1 on lines 88-06 and 88-07 (Figures 3.19 and 3.20), at 6 km and between 50-70 km on line 88-05 (Figure 3.18). The upward extension of these faults, into the Pliocene section, suggests that the transtensional phase of the basin formation may have ended more recently, during the Pliocene, than previously supposed.

3.8.2 Velocity at Apex of Antiforms

Typically, isovelocity contours in the derived velocity models do not follow the stratigraphy in anticlines, and velocities estimated at the apex appear to be anomalously low. A decrease in velocity of approximately 150 ms$^{-1}$ is observed at the apex of antiform A1 at x = 18-23 km on line 88-06 (Figure 3.8a), x = 16-22 km on line 88-07 (Figure 3.9a) and x = 136-148 km on line 88-04 (Figure 3.6a). Other decreases in velocity are observed at the apex of anticlinal folds on line 88-07 at x = 63-72 km and on line 88-06 between x = 36 and 38.5 km, where a sharp lateral decrease in velocity from approximately 2500 to 2300 ms$^{-1}$ overlies a small faulted anticline in the upper-Miocene strata with apex at x = 37 km (A2 in Figure 3.19). We interpret these decreases in velocity to be the result of tensile fractures that increase the porosity and consequently lower the velocity. Tensile fractures are probably due to folding during basin inversion; e.g., anticline A2 originated in the Miocene by shortening of a deep half graben centered near x = 39.7 km (Rohr and Dietrich,
Decreases in velocities at the apex of anticlinal folds have also been observed at the Cascadia accretionary wedge (Hayward and Calvert, 2007).

### 3.8.3 Porosity and Oil Staining in Sediments

Localized low velocity zones with velocities of 2200-2300 ms\(^{-1}\), associated with elevated attenuation are present at 700-1000 m depth, between x = 10-32 km on line 88-05 (Figures 3.7) and at x = 3 km, 28-30, 74-100, 118-123 km, and 138-152 km on line 88-04 (Figure 3.6). Many of these zones are close to faults or folds and the low velocity and high attenuation may be due to elevated fracture porosity. However, the decrease in velocity and associated increase of attenuation on line 88-05 between x = 16-20 km and on line 88-04 between x = 80-82 km may be linked to the oil staining identified in the Sockeye B-10 well between 950-1050 m. A comparison of the 1D attenuation extracted from line 88-04 and 88-05 with the lithology log from the well (Figures 3.16d and 3.16e) shows that the strong increase in attenuation correlates with the oil staining. Elevated attenuation \( (Q_p^{-1} = 0.06-0.015) \) is consistent with attenuation measured in oil sandstones and gas sandstones (Klimentos, 1995). We can therefore suggest that the elevated attenuation between 8 and 20 km and at 29-34 km on line 88-05 and between 74 and 100 km on line 88-04 is due to the presence of hydrocarbons, perhaps in relatively small quantities.

### 3.8.4 Increase in Clay Content

An almost continuous layer of high attenuation \( (Q_p^{-1} = 0.05-0.1) \), associated with localized low velocity zones (2000-2300 ms\(^{-1}\)) at 700-900 m depth, is present between x = 2-35 km on line 88-06 (Figure 3.8). Line 88-06 intersects line 88-04 near the Tyee N-39 well and elevated attenuation associated with low velocity zones is also present between x = 136-140 km on line 88-04 (Figure 3.6) at the same depth. This layer of elevated attenuation on
3.8.5 Seafloor Pockmarks and Gas Chimneys

A V-shaped region of anomalously low velocity is present just below the seafloor on line 88-05 at x = 47-51 km. Velocities in this anomalous zone are approximately 1750-1850 ms$^{-1}$ (Figures 3.7a and 3.7d), compared with surrounding values of 2000-2200 ms$^{-1}$ at a depth of approximately 300 m. This anomalous zone corresponds to a chain of pockmarks in the seafloor where authigenic carbonate chimneys occur (Figure 3.1) (Halliday et al., 2008; Barrie et al., 2011). In the vicinity of the pockmarks, seismic arrivals are clearly delayed by the lower velocities (Figure 3.13). We suggest that the V-shaped low velocity zone is the subsurface expression of hydrocarbon seepage, which has given rise to the pockmarks and carbonate chimneys with the low velocity due to the presence of gas. Authigenic carbonates are considered to be derived by microbial oxidation of hydrocarbon bearing fluids (Hovland et al., 2005; Roberts and Aharon, 1994; Wallace et al., 2006), and active venting of gas has been observed from at least 27 individual carbonate chimneys over the area (Barrie et al., 2011). Fluids may have migrated upward to the seafloor along underlying faults, which we interpret within the Pliocene section, extending to the Quaternary unconformity. Below the pockmark structure, elevated attenuation of $Q_p^{-1} = 0.05-0.1$ is present at 900-1100 m below the seafloor (Figure 3.7b). The high attenuation at this depth might be related to the
presence of gas (Klimentos, 1995; Parra et al., 2006), but low attenuation in the overlaying sediments (200-800 m) may be due to carbonate precipitation within the clastic rocks.

At x = 38-40 km and deeper at x = 67 km, on line 88-05, there are narrow, steeply dipping zones of low velocity in the range of 2100-2250 ms$^{-1}$ that are also associated with high attenuation values: $Q_p^{-1} = 0.08-0.1$ at x= 66 km and $Q_p^{-1} = 0.05-0.09$ (0.09 just below the seafloor) between x = 38-40 km (Figures 3.7a, 3.7d and 3.7b). These elevated attenuation values are consistent with gas sandstones or oil sandstones (Klimentos, 1995). We speculate that these features are pipe-like chimneys resulting from gas ascension. The upward migration of gas through chimneys located between x = 38-40 km and at x= 67 km might be responsible for the presence of a zone with relatively weak amplitude in the input data at 40-46 km (Figure 3.13), and a layer with elevated attenuation between x = 62-65 km at the seafloor (Figure 3.7b) respectively. Gas chimneys were also identified by Hicks and Pratt (2001) over a gas-sand deposit, using attenuation models from waveform tomography.

3.9 Conclusion

The Queen Charlotte basin is a complex network of half-grabens and other sub-basins containing variably deformed stratigraphy. We applied visco-acoustic waveform tomography to four seismic reflection profiles, which extend across the basin, from Hecate Strait to Dixon Entrance. The maximum depth of penetration was restricted to 1200 m due to the 3770 m maximum offset of the input data. A carefully designed preconditioning of the data, comprising F-K filtering and amplitude corrections, and a specific inversion strategy were applied to ensure convergence to a likely global minimum. The P-wave velocity and attenuation models show a good correlation with the available sonic, gamma ray logs, and with interpreted migrated seismic sections from previous studies. 1D
velocity and attenuation profiles extracted at the line intersections also show good correlation, indicating the robustness of the inversion strategy used, and synthetic data calculated from the final 12 Hz waveform tomography velocity models match the field data well.

Using velocity models from waveform tomography, it was possible to identify Quaternary strata and shallow faulting in the Pliocene section that were not interpretable on the original migrated sections. Also, some faults previously identified on the migrated section, below the Pliocene/upper Miocene unconformity, can now be extended upward into the Pliocene section. The presence of faulting in the Pliocene suggests that the transtensional phase of basin evolution continued longer than previously supposed.

P-wave velocities in sediments range between 1750-3000 ms$^{-1}$ and attenuation values ($Q_p^{-1} \leq 0.04$) are generally consistent with values for saturated sandstones, but in some areas elevated attenuation ($Q_p^{-1} = 0.05-0.15$) is associated with localized low velocities (2200-2600 ms$^{-1}$). In basement areas, the estimated attenuation values ($Q_p^{-1}$) are generally between 0.01 and 0.1 and there is a correlation between high attenuation values ($Q_p=0.05-0.1$) and high velocity ($V_p \geq 3500$ ms$^{-1}$). These rocks are interpreted to be volcanic, but the elevated effective attenuation could also be an artifact due to scattering or mode conversion effects, which are not included in the inversion.

Decreases in velocity of approximately 150 ms$^{-1}$ at the apex of anticlinal folds were observed on lines 88-06, 88-04 and 88-07 and are interpreted to be due to tensile fracturing that increases the porosity and consequently lowers the seismic velocity. Elevated attenuation, associated with localized low velocity zones are present deeper on lines 88-04, 88-05 and 88-06 (at 700-1000 m depth), and are close to the location of faults and folds. These regions with elevated attenuation and localized low velocities may be due to fracture related porosity. However, the correlation of high attenuation with greater clay content and the presence of siltstones and shale in the Tyee N-39 well at 700 m depth,
suggest that elevated attenuation at 2-36 km on line 88-06 and 136-140 km on line 88-04 is related to the presence of siltstones and shale. High attenuation, associated with anomalously low velocity zones was also found at the Sockeye B-10 well location on lines 88-05 and 88-04, and in this case, correlates with the occurrence of oil staining in the well at 950-1050 m depth.

A shallow V-shape low velocity anomaly was identified on line 88-05 immediately beneath a chain of seafloor pockmarks, where authigenic carbonate chimneys, which are interpreted to be related to hydrocarbon seepage, occur. West of the V-shape anomaly at x = 38-40 km and deeper at x = 67 km on line 88-05, sub-vertical zones of low velocity associated with high attenuation values are interpreted as pipe-like chimneys resulting from gas ascension.

### 3.10 Acknowledgments

We are grateful to Gerhard Pratt, who provided us with his waveform tomography code. Well logs from the Tyee N-39, Sockeye B-10 and Murrelet L-15 were provided by Divestco. The seismic data were plotted with SeismicUnix and Gnuplot. SeismicUnix was also used partially to process the data. This project was funded by the Natural Sciences and Engineering Research Council of Canada. The final manuscript was improved by comments from Ian Jones and by three anonymous reviewers.
Reference List


Seismic waveform tomography across the Seattle Fault Zone in Puget Sound: Resolution analysis and effectiveness of visco-acoustic inversion of visco-elastic data

A version of this manuscript was submitted for publication as:

4.1 Abstract

Visco-acoustic waveform tomography was applied to marine seismic reflection data across the Seattle fault zone in Puget Sound. Using the recovered velocity and attenuation models, we performed a set of synthetic visco-acoustic and visco-elastic checkerboard tests, and compared the results to verify the effectiveness of applying acoustic waveform tomography to visco-elastic field data. Visco-acoustic waveform tomography provides higher resolution velocity and attenuation models than ray-based tomography, but artifacts
due to mode conversions are present at layer interfaces where the velocity contrast is high. Mode conversions also affect attenuation values, which are too high in places because visco-acoustic inversion compensates the loss of amplitude due to mode conversion by over-estimating the attenuation. A comparison of the inverted attenuation model from visco-elastic modeling and the inverted attenuation model from visco-acoustic modeling suggests that inverted attenuation values can be reliable when the velocity gradient is low. Joint interpretation of the derived velocity and attenuation models enables us to identify Quaternary (glacial and post-glacial Pleistocene), Tertiary sedimentary and Eocene volcanic rocks. Several shallow north dipping faults, anticlines and a syncline are identified across the Seattle uplift and the Seattle fault zone. Our interpretation of faults using the velocity model, attenuation model and migrated section is consistent with two possible published models of the Seattle Fault Zone: either a thrust fault that accommodates north-south shortening by forming a fault-propagation fold with a forelimb breakthrough, or part of a passive roof duplex in which the Seattle Fault Zone is located at the leading edge of a triangle zone that is propagating into the Seattle Basin.

4.2 Introduction

The Seattle Fault Zone (SFZ) is located in Puget Sound near the densely populated greater Seattle metropolitan area (Figure 4.1). A M7 earthquake occurred on the SFZ about 1,100 years ago (Bucknam et al., 1992, 1999), and the SFZ is considered a major seismic hazard. A number of geophysical surveys, including gravity, magnetics and seismic, have been acquired to understand and map subsurface structures related to faults in this area. Structural interpretations of the seismogenic faults based on seismic surveys have been presented (Johnson et al., 1994; Pratt et al., 1997; Liberty and Pratt, 2008; Brocher et al., 2004), but these interpretations are ambiguous due to the absence of clear reflections from
fault planes, or from stratigraphy within the fault zone (Brocher et al., 2004; Calvert et al., 2003). Ray-based tomography velocity models were derived across the SFZ by Calvert et al. (2001, 2003), and the results show along-strike variations in both the number of fault splays and the degree of uplift of the Tertiary lithologies over a distance of less than 3 km. In this paper, we present high resolution velocity and attenuation models from full waveform tomography across the SFZ. Our aim is to utilize the higher resolution velocity and attenuation models available from waveform tomography to identify structures not visible in conventional seismic images and to refine the interpretation of this complex area.

Recent studies, including cross-hole, seismic reflection and refraction surveys have demonstrated the ability of waveform tomography to provide velocity models of the subsurface with a resolution far greater than is possible using traveltime tomography (Ravaut et al., 2004; Pratt et al., 2004; Operto et al., 2006; Brenders and Pratt, 2007; Malinowski and Operto, 2008; Bleibinhaus et al., 2008; Takam Takougang and Calvert, 2011). Due to the greater non-linearity and computational cost of elastic inversion, most of the applications of waveform tomography are based on acoustic or visco-acoustic modeling of the visco-elastic field data. This approximation is based on the assumption that the data are dominated by unconverted P-waves, and therefore the subsurface can be modeled as an acoustic medium. However, when S-waves and converted P-waves are significant in the data, the acoustic approximation breaks down, leading to errors in modeling the amplitudes of the field data, especially the variation of amplitudes with offset. Mulder and Plessix (2008) and Barnes and Charara (2009) have shown with a suite of synthetic marine data examples that in cases where S-waves are significantly present in the data, acoustic waveform tomography provides reliable results only when the S-wave velocity varies slowly relative to the wavelength of the seismic data. When S-waves are removed from the data, for example by muting, if the transmitted amplitude Versus Offset
(AVO) effect due to S-conversion is significant, acoustic waveform tomography leads to reliable results only for the near offset data (Barnes and Charara, 2009). In order to correct for the inability of the acoustic approximation to model fully the AVO effects of the visco-elastic field data, it is common practice to normalize the amplitudes of the field data. One approach to normalization is to scale individual traces to a common maximum amplitude (Ravaut et al., 2004; Operto et al., 2006; Malinowski and Operto, 2008), thereby ignoring the AVO effects and considering only the amplitude variation within individual traces. Another approach to correcting the field data consists of scaling the amplitude of the field data using a comparison with data modeled using the acoustic wave equation so that the AVO of both data becomes similar (Brenders and Pratt, 2007; Takam Takougang and Calvert, 2011; Delescluse et al., 2011). It is also possible to match the AVO of the visco-elastic field data by introducing attenuation into the acoustic modeling (Pratt, 1999). In all cases, the physics of the problem is imperfectly represented, and a significant question remains as to how well the acoustic or visco-acoustic approximation can model visco-elastic field data. In other words, how reliable are the results of visco-acoustic waveform tomography of visco-elastic data?

We study the practical validity of visco-acoustic waveform tomography of marine seismic reflection data from the SFZ, to a maximum depth of 1.3 km. These results could also have implications for shallow marine seismic reflection data from other complex areas. We apply waveform tomography to a section of line PS-2 from the Seismic Hazards Investigation in Puget Sound (SHIPS) survey (Figure 4.1), and following the preconditioning and inversion strategy presented in Takam Takougang and Calvert (2011, 2012), we derive velocity and attenuation models to a maximum frequency of 17.6 Hz. We compute a set of checkerboard tests to evaluate the quality of velocity models derived from data modeled using visco-acoustic and visco-elastic algorithms. We show from these results that the visco-acoustic approximation of the seismic wavefield is valid for much of
the velocity and attenuation models, but some artifacts can be present near sharp medium parameter contrasts. Using the recovered velocity and attenuation models from waveform tomography, we provide a geological interpretation of the lithology and faults to a depth of 1.2-1.3 km.

4.3 Geological Setting

The SFZ is a complex zone of deformation with multiple thrust faults that extends across the Puget Lowland of western Washington, USA. The Puget Lowland is a forearc basin located between the Cascade volcanic arc in the east and the Olympic Mountains to the west, which form part of the accretionary complex of the Cascadia subduction zone (Johnson et al., 1994). The Puget Lowland lies above the Eocene suture zone, between Crescent Terrane volcanic rocks and the older Tertiary North American crust (Johnson et al., 1996). Subduction along the continental margin is accommodated in part by north-south shortening within the Puget Lowland (Pratt et al., 1997; Van Wagoner et al., 2002) resulting in the formation of a series of basins and uplifts separated by faults. Among the faults beneath the Puget Lowland, the SFZ has produced the greatest vertical offset of the igneous basement (Liberty and Pratt, 2008).

The east-trending SFZ offsets the Seattle uplift to the south, where Early Tertiary basalts of the Crescent Formation were uplifted, and the 9-km deep Seattle Basin to the north, where Tertiary bedrock is buried under at least 7 km of sedimentary strata of the younger Blakeley and Blakely Harbor formations (Pratt et al., 1997; Johnson et al., 1994). The Crescent Formation, the upper part of which consists of interbedded marine basalts and sedimentary rocks, forms the basement to the western part of the Seattle Basin (Johnson et al., 1994). The age of the Blakely Harbor Formation, which consists of nonmarine sandstone, conglomerate and silstone, is between 24 Ma and 2 Ma (Johnson
et al., 1994). The Blakely Harbor Formation is 3010 m thick and is underlain by the Blakeley Formation, which immediately overlies the Crescent Formation (Figure 4.2).

![Map of the Puget Lowland with line PS-2 from the SHIPS seismic survey. The section highlighted in red was used for this study. The locations of the Kingston 1 and Socal-Shroder 1 wells are indicated. A indicates Alki point; A1-Blake Island anticline and A2-Winghaven Park anticline.](image-url)

**Figure 4.1:** Map of the Puget Lowland with line PS-2 from the SHIPS seismic survey. The section highlighted in red was used for this study. The locations of the Kingston 1 and Socal-Shroder 1 wells are indicated. A indicates Alki point; A1-Blake Island anticline and A2-Winghaven Park anticline.
The Blakely Harbor Formation comprises sandstone, silstone, shale, and conglomerate deposited in a deep-marine, submarine-fan environment (Johnson et al., 1994). Quaternary sediments (900 m) overlay the Blakely Harbor Formation and comprise Pleistocene glacial and post-glacial to Holocene deposits (Johnson et al., 1994; Calvert et al., 2003). The Pleistocene glacial deposits consist of clay, silts, sand, and interbedded sand and clay (Tubbs, 1974).

4.4 Structure of the SFZ

The SFZ was first inferred from gravity data in 1965, and the fault zone was interpreted to comprise two steeply north-dipping normal faults with 11 km of vertical displacement (Daneš et al., 1965). More recent information about the structure of the Puget Lowland has been derived from gravity, magnetics and mostly seismic data (Johnson et al., 1999; Blakely et al., 2002; ten Brink et al., 2002; Calvert et al., 2003). The structure of the SFZ is constrained by seismic reflection profiles acquired in Puget Sound and Lake Washington. However, the location of the individual faults is uncertain, because they are steeply dipping and are interpreted to lie within zones of little or no reflectivity (Calvert et al., 2003; Liberty and Pratt, 2008). Consequently, there are a number of different interpretations. Johnson et al. (1994, 1999) proposed a model of the SFZ consisting of four sub-parallel, south-dipping faults with dominant reverse or thrust displacements. This model was later adopted by Calvert et al. (2003) in their combined interpretation of seismic reflection data and P-wave velocity models derived from first arrival tomography. Johnson et al. (1994) proposed a mean dip of 45-65° for the SFZ using seismic reflection profiles, while Calvert et al. (2001) proposed a dip of 60° for the northernmost fault of the SFZ. Pratt et al. (1997) proposed a dip of 45° for the upper 6 km of the fault, decreasing to 20-25° at 6-16 km depth. Based on observed seismicity and the focal mechanisms of
small earthquakes along the SFZ, Van Wagoner et al. (2002) proposed that the SFZ is south-dipping with a subvertical dip, and argued that seismicity in the SFZ does not show evidence of alignment along the proposed south dipping thrust model of Johnson et al.
(1999). ten Brink et al. (2002) used a combination of seismic, gravity and elastic dislocation modeling to propose a model for the SFZ geometry that would have generated the observed shoreline uplift in the large earthquake approximately 1100 years ago. This model also shows that the interpretation of the SFZ, with 3 to 5 south-dipping thrust faults, is inconsistent with density models derived from gravity data. ten Brink et al. (2002) interpret the SFZ as a thrust fault dipping to the south with an estimated dip of 41-49°. On the basis of onshore seismic reflection surveys, Liberty and Pratt (2008) modeled the SFZ as a thrust fault that accommodates north-south shortening by forming a fault-propagation fold with a forelimb breakthrough and backthrusts. Contrary to the interpretation of the Seattle faults as south-dipping (Johnson et al., 1994; Calvert et al., 2003), airborne laser terrane mapping has revealed the presence of postglacial Holocene fault scarps, which indicate north-dipping thrusting at the surface along strands of the SFZ on Bainbridge Island, west of Puget Sound (Nelson et al., 2003). Brocher et al. (2004), using gravity, magnetics and seismic data, interpreted the Seattle uplift as a passive roof duplex bounded by a floor thrust and a roof thrust that merge in the subsurface, forming a triangle zone within the Seattle uplift. The SFZ was interpreted by Brocher et al. (2004) as a south-dipping reverse fault forming the leading edge of the triangle zone, and shallow north-dipping splay thrust accommodate the northward propagation of the triangle zone into the Seattle Basin. This interpretation reconciles the south-dipping thrust of the Seattle Fault with the north-dipping Holocene thrusting identified at the surface.

4.5 Waveform Tomography

We applied visco-acoustic waveform tomography (Pratt et al., 1998; Pratt, 1999) to a section of line PS-2 from the SHIPS seismic reflection data between shot points 500 and 925. The SHIPS survey was acquired in 1998, as part of a larger wide-angle and
multichannel reflection survey aimed at mapping the subsurface architecture and relating it to mapped surface structures (Fisher et al., 1999). Line PS-2 was shot north-south in Puget Sound. The source was a 13-gun tuned airgun array with a total volume of 4838 in$^3$. The shot interval was 50 m, and the data were recorded by a 96-channel hydrophone streamer with 25 m group interval, a minimum offset of 200 m and a maximum offset of 2575 m. The 50 m shot interval yielded a nominal 24-fold stacked reflection section. This section of line PS-2 crosses the SFZ, the Seattle uplift, and part of the Seattle basin, and has a total length of 23.8 km.

The preprocessing of the seismic data comprised data editing to remove noisy traces, shot-to-shot energy balancing and 2D amplitude scaling of the 3D field data, following the procedure outlined in Takam Takougang and Calvert (2011). The data were muted 1.5 s below the first arrivals and a time window of $T_w=2.5$ s was used for the inversion. Our inversion strategy, which recovers velocity and attenuation models sequentially, using a layer stripping approach for the reconstruction of each model, is outlined in detail in Takam Takougang and Calvert (2011, 2012). Generally, the layer stripping approach recovers shallow to deep structures by varying data aperture in the frequency domain, and using depth and offset weighting to boost the contribution of the layer under reconstruction in the misfit function. The velocity model is reconstructed initially using all frequencies, from the lowest to the highest, and later the attenuation model is recovered in a second step. The minimum frequency used for this data was 4.4 Hz and we performed the inversion up to 17.6 Hz, using the frequencies in triplets every 0.4 Hz ($1/T_w$).

The sparseness of the data ($N_A$), the ratio between the source interval or the receiver interval, whichever is sparser ($\Delta$) to the spatial sampling interval ($\Delta_{samp}$) at a given frequency $f$, was calculated to select the maximum usable frequency for the inversion necessary to avoid aliasing. Ideally, this ratio should be less or equal to 1 and can be
written as (see Takam Takougang and Calvert (2011)):

\[ N_A = \frac{\Delta}{\Delta_{samp}} = \frac{2 \Delta f \sin \theta}{c_{min}}, \]  

(4.1)

where \( \theta \) is the emerging angle of the propagating wavefield and \( c_{min} \) is the minimum velocity in the velocity model. For line PS-2, the shot interval of 50 m and the practical minimum velocity of 1480 ms\(^{-1} \) give sparseness values of \( N_A = 0.3 \) for the minimum frequency of 4.4 Hz and \( N_A = 1.17 \) for the maximum frequency of 17.6 Hz when the propagating wavefield emerges at an angle of \( \pi/2 \). Other studies suggest that this value is sufficient for the successful application of waveform tomography (see Takam Takougang and Calvert (2011)). The condition \( N_A = 1 \), which implies an upper frequency of 15 Hz, may be too restrictive as most of the seismic waves emerge with angles less than \( \pi/2 \).

The starting velocity model for waveform tomography was derived from traveltime tomography using an algorithm based on first arrival traveltime picks (Aldridge and Oldenburg, 1993). The travel times of first arrivals were picked with an accuracy estimated to be within \( \pm 8 \) ms. The initial model for traveltime tomography, was designed with a velocity gradient of 1.68 s\(^{-1} \). The gradient was increased to 5.25 s\(^{-1} \) where the basement is shallow, i.e between \( x = 0-9 \) km, and was reduced to 1.0 s\(^{-1} \) where the velocity reached 4000 ms\(^{-1} \) to prevent the inclusion of anomalously large velocities in the model. Velocities above the seafloor, which varied between a depth of 15 m and 230 m, were set to a constant value of 1480 ms\(^{-1} \) during the inversion. 25 iterations were computed, which reduced the root-mean-square (RMS) misfit from 90.458 ms to 7.374 ms. The recovered velocity model was then tested to ensure that it was able to predict first arrival traveltimes to within half a cycle. The starting attenuation model for waveform tomography was a homogeneous attenuation model with \( Q_p = 100 \) below the seafloor and \( Q_p = 10000 \) in the water layer (we assumed no attenuation in the water layer).


4.6 Inversion Results

The recovered velocity model and associated ray density from traveltime tomography are displayed in Figure 4.3. The ray coverage from traveltime tomography is limited to a maximum depth of approximately 800 m, but our waveform tomography inversion strategy allows us to recover structures down to approximately 1300 m, as shown by the resolution tests in the following section. The inversion results from full waveform tomography using frequencies up to 17.6 Hz after the first step without attenuation, and after the second step with attenuation (final model) are displayed in Figure 4.4. In general, velocity models derived by waveform tomography have higher resolution than the starting model from ray-based tomography. A number of features of the waveform tomography velocity models are not present in the starting model, for example the localized low velocity zones (LVZ) at $x = 10$-14 km within the Seattle fault zone and at $x = 15$-23.8 km. Two anticlines centered at $x = 4$ km and $x = 8$ km and a syncline centered at $x = 6$ km are well imaged in the final model (Figure 4.4b). The recovered attenuation model (Figure 4.4c) shows an undulating layer of high attenuation ($Q_p^{-1} = 0.09-0.27$) that appears to be continuous along the line. Artifacts characterized by anomalously low velocity at layer interfaces, are present in the waveform tomography velocity models. For example, at approximately 430 m depth, low velocities of $v \approx 1300 \text{ ms}^{-1}$ are present at $x = 2$-10 km and $x = 13$-22 km. Figure 4.5 shows 1D velocity profiles extracted at $x = 7.5$ km, 9.3 km and 15.7 km from the final velocity model (after stage 2), from the velocity model after stage 1, and from the starting model from traveltime tomography. The abnormal decrease in velocity to $1300 \text{ ms}^{-1}$ below the seafloor, i.e., at $z = 200$ m depth and deeper at $z = 430$ m depth, which corresponds to the interface between two layers in the shallow sedimentary rocks, is absent in the starting model, but is more pronounced in the waveform tomography velocity model without attenuation (stage 1) than the model with attenuation (stage 2), suggesting that some of these artifacts have
been reduced by the introduction of attenuation into the inversion. These artifacts, which are likely related to elastic mode conversion, will be discussed in more detail in the next section.

In the absence of well log data that tie to the seismic line, one way of determining the quality of the result is to compare the field data with synthetic data obtained using the recovered velocity and attenuation models (Figure 4.6). It is interesting to note the high similarity between synthetic and field data at both near offset (Figures 4.6c and 4.6a) and far offset (Figures 4.6d and 4.6b). Some mismatches are however evident, in particular at x = 10-12 km, i.e., within the SFZ. While the fit is good at near offset, it is not clear at far offset, because the strong attenuation of the seismic waveforms within this area reduces amplitudes to a level that cannot be well recovered during the inversion. The strong attenuation of seismic waves in this area led to the absence of clear reflections in the migrated seismic section making interpretation difficult (Johnson et al., 1999; Calvert et al., 2003).

![Figure 4.3](image)

**Figure 4.3:** (a) Starting velocity model derived from first arrival traveltime tomography. (b) ray density. The white line represents the seafloor topography.
Figure 4.4: Waveform tomography results using frequencies from 4.4 Hz to 17.6 Hz: (a) Velocity model after the first step in which attenuation was not included; (b) velocity model after the second step (final model) and (c) associated attenuation model. Arrows indicate anomalously low velocity at layer interfaces due to mode conversion. These artifacts are stronger in the model without attenuation (a).

4.7 Model Assessment: Visco-Acoustic and Visco-Elastic Modeling

In order to thoroughly investigate the nature of possible artifacts present in the velocity model, and to understand the reliability of the recovered velocity and attenuation models from visco-acoustic waveform tomography, we performed a set of visco-acoustic and visco-elastic checkerboard tests. The synthetic visco-acoustic and visco-elastic data are derived from forward modeling using the velocity and attenuation models perturbed by
Figure 4.5: 1D profiles from the starting model and waveform tomography inversion results, after stage 1 and stage 2 (final model). The arrows indicate decrease in velocity below the seafloor (at 200 m) and at layer interfaces due to mode conversion. These effects have been partially reduced with the introduction of attenuation during stage 2 of inversion.

checkerboard patterns. By comparing the checkerboard test results from visco-acoustic and visco-elastic data, we estimate the resolution limit, and provide a possible explanation for the nature of artifacts.

4.7.1 Methodology

We designed a set of checkerboard velocity and attenuation models consisting of alternating rows and columns of positive and negative velocity and attenuation perturbations, that we superimposed on the final velocity and attenuation models from waveform tomography, with the exception of the water layer. The perturbations are a constant percentage of the actual velocity and attenuation values, and are therefore spatially variable. The percentage of perturbation of the velocity model was set to 5 % which represents a velocity perturbation of $\pm 100 \, \text{ms}^{-1}$ at 2000 ms$^{-1}$ and a perturbation of $\pm 175 \, \text{ms}^{-1}$ at 3500 ms$^{-1}$. The perturbation for attenuation was set to 50 %, to obtain
Figure 4.6: Comparison of synthetic and field data low pass filtered to 17.6 Hz in common offset gathers. (a) and (b) synthetic data, (c) and (d) field data. The white line shows first arrival picks where waveform amplitudes are weak.

detectable attenuation variation. 50% perturbation gives a variation of ±0.05 attenuation at an attenuation value of $Q_p^{-1} = 0.1$. We perturbed the velocity and attenuation models independently, i.e. perturbing the velocity assuming no attenuation for the velocity model assessment and perturbing the attenuation while leaving the velocity unchanged for the attenuation model assessment. The checkerboard test used the perturbed models to generate visco-acoustic and visco-elastic data by forward modeling using the same acquisition geometry as the field data, a Ricker wavelet with a dominant frequency of 7 Hz as the source, and a density calculated with the Gardner’s relation. The inverse problem in this experiment consists of running waveform inversion using the newly created dataset, following the same inversion strategy as with the field data, and using the unperturbed velocity and attenuation models as the starting models.
For the frequency domain visco-acoustic modeling, we computed the synthetic data between frequencies 0.25 Hz and 18 Hz using 2D finite difference modeling in the frequency domain (Pratt et al., 1998; Pratt, 1999). Since the data are synthetic visco-acoustic, no preconditioning was necessary prior to using them in the inversion.

For the time domain visco-elastic modeling, we used the 2D time domain visco-elastic finite difference code of Robertsson et al. (1994) to generate the synthetic data. To obtain an S-wave velocity model, we used the relationship between S-wave velocity \( v_s \) and P-wave velocity \( v_p \) from Hamilton (1979) for saturated marine terrigeneous sediments, which is consistent with the shallow Tertiary and Quaternary turbidites, sand, gravel, clay and silts beneath Puget Sound (Tubbs, 1974; Rau and Johnson, 1999). The relationship between \( v_p \) and \( v_s \) is defined in Table 4.1.

Table 4.1: \( v_p \) and \( v_s \) relations used to generate the visco-elastic data

<table>
<thead>
<tr>
<th>Subsurface</th>
<th>( v_p ) (kms(^{-1}))</th>
<th>( v_s ) (kms(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>1.480</td>
<td>0</td>
</tr>
<tr>
<td>Sediments</td>
<td>from 1.512 to 1.555</td>
<td>3.884(v_p)-5.757</td>
</tr>
<tr>
<td>Sediments</td>
<td>from 1.555 to 1.650</td>
<td>1.137(v_p)-1.485</td>
</tr>
<tr>
<td>Sediments</td>
<td>from 1.650 to 2.150</td>
<td>0.991-1.136(v_p)+0.47(v_p^2)</td>
</tr>
<tr>
<td>Sediments</td>
<td>greater than 2.150</td>
<td>0.78(v_p)-0.962</td>
</tr>
</tbody>
</table>

The seismic attenuation values for P-waves and S-waves are comparable in saturated sediment (Winker and Nur, 1979; Murphy, 1982). Therefore, for simplicity, we used the derived attenuation from the field data inversion as the intrinsic attenuation for both P-waves and S-waves in this test.

Our goal is to recover the checkerboard patterns imposed on the models. If the recovered velocity and attenuation models from the field data inversion are reliable, then we should be able to recover the checkerboard patterns, and the size of the checkerboard cells gives an idea of the resolution of the image. Since we are using a visco-acoustic
waveform tomography method, the inversion of the synthetic visco-elastic data will enable us to practically check the effectiveness of applying visco-acoustic waveform tomography to visco-elastic data. Two different sizes for the squares of the checkerboard pattern were used: 400 m and 200 m. The resolution of waveform tomography is of the order of the wavelength of the propagating wavefield (Wu and Toksöz, 1987; Williamson, 1991; Pratt et al., 2002), and since the average velocity of the model is \( c_0 = 2800 \text{ ms}^{-1} \), 400 m corresponds to the wavelength (\( \lambda = c/f \)) at a frequency of 7 Hz, and 200 m corresponds to the wavelength at a frequency of 14 Hz. Consequently, the inversion for the 400 m perturbation size was computed between the frequencies 4.4 Hz and 12 Hz and the inversion for the 200 m perturbation size between 4.4 Hz and 17.6 Hz.

In the following sections, we present the results of the visco-acoustic checkerboard tests, followed by the visco-elastic tests and compare both to assess the validity of our application of acoustic waveform tomography to elastic data from Puget Lowland, and other similar marine datasets.

### 4.7.2 Visco-Acoustic Results

**Velocity Model**

The recovered checkerboard patterns for the velocity model after the inversion of the newly created dataset are displayed in Figures 4.7a and 4.7c. With the exception of the edge of the model between \( x = 0-1.5 \text{ km} \) where the coverage is poor due to the acquisition geometry, the perturbation patterns for both the 400 m size and the 200 m size have been well recovered above 1300 m depth in the northern part of the model and above 900 m depth in the southern part of the model between \( x = 2-9.5 \text{ km} \). The limited depth in the southern part of the model is due to the presence of shallow high velocity rocks in the Seattle uplift which focus seismic waves close to the seafloor.
The relatively good recovery of the 200 m and 400 m checkerboard perturbation size illustrates the greater resolving power of waveform tomography relative to travel time tomography. Corrugation tests which are 1D version of the 2D checkerboard tests were performed on a traveltime tomography velocity model derived on the same line by Calvert et al. (2003), and the best lateral resolution obtained extended to depth of only 500-800 m with a corrugation of half wavelength 800 m. The corrugation with a half wavelength of 400 m was recovered to a depth of 400-600 m south of the SFZ, but was not well recovered to the north. The corrugation with a half wavelength of 200 m was not well recovered at all. With waveform tomography, we should theoretically be able to recover a 160 m checkerboard perturbation, which is the wavelength at 17.6 Hz with an average velocity of 2800 ms\(^{-1}\) but, we did not carry out this test due to computational limitations.

**Attenuation Model**

The apparent seismic attenuation \(Q^{-1}\) can be expressed as the combination of intrinsic attenuation (\(Q_{int}^{-1}\)), describing the energy loss from inelastic absorption related to material properties, and scattering attenuation (\(Q_{sc}^{-1}\)) arising from energy losses due to transmission, scattering from small scale heterogeneities, etc (Spencer et al., 1982; Richards and Menke, 1983). For visco-acoustic modeling, we can write:

\[
Q^{-1} = Q_{acoustic}^{-1} = Q_{int}^{-1} + Q_{sc}^{-1}. \tag{4.2}
\]

Under the assumption that the quality factor \(Q_{int}\) is large and frequency independent, intrinsic attenuation is included in acoustic waveform inversion through the use of complex velocities (Song and Williamson, 1995; Hicks and Pratt, 2001) and has the form:

\[
Q_{int} = \frac{-v_{\text{real}}}{v_{\text{img}}}, \tag{4.3}
\]
where $v_{\text{real}}$ is the real part, and $v_{\text{img}}$ the imaginary part of the complex velocity. In principle, the intrinsic attenuation can be successfully recovered during the inversion when the velocity model is accurate, so the inversion can discriminate between intrinsic and scattering attenuation due to small scale velocity variation (Kamei and Pratt, 2008; Rao and Wang, 2008). For a given frequency however, the inversion cannot recover velocity fluctuations with a wavenumber $K$ higher than $2K = 2/\lambda$ (Wu and Toksöz, 1987). If the resolution of the velocity model is too low, the inversion will compensate the observed change in waveform amplitude due to scattering by including it in the attenuation image.

The checkerboard attenuation model used to model the synthetic data is the intrinsic attenuation, and if the inversion is successful, we should be able to recover the checkerboard pattern, since we used the “true” velocity model during the inversion, i.e., the unperturbed velocity model used during the modeling of the data.

The recovered attenuation checkerboard pattern shows that we were able to successfully recover intrinsic attenuation of the high attenuation layer along the model using both the 400 m and 200 m perturbations (Figures 4.8a and 4.8c). The use of the “true” velocity model during the inversion helps to avoid the introduction of scattering attenuation into the attenuation model.

### 4.7.3 Visco-Elastic Results

**Velocity Model**

Figures 4.7b and 4.7d show the results of the inversion for P-wave velocity in the visco-elastic checkerboard tests. The 400 m checkerboard perturbation has been well recovered, and the quality of the model is very similar to the acoustic case. Artifacts are, however, present in the model at $x = 1.8$-9 km and 13-20 km and correlate with the anomalously low
velocity values at two interfaces in the final velocity model. The 200 m perturbation was well recovered at x = 15-23.8 km to a depth of approximately 1000 m and to shallower depth between 1.8-10 km, but the model contains several artifacts due to elastic effects.

The results show that the recovered velocity model from the application of acoustic waveform inversion to marine elastic data is generally valid, but the resolution of the image is degraded by elastic artifacts. In this situation, the inversion may not be able to recover small-scale velocity perturbations. The absence of these artifacts in the visco-acoustic model indicates that they are probably related to elastic mode conversion at the interfaces. The acoustic inversion of elastic data may compensate for the decrease in amplitude and change in phase of the elastic waveform from mode conversion at layer interfaces by anomalously decreasing the velocity.

Attenuation Model

The test results for the 400 m attenuation perturbation (Figure 4.8b) show that the checkerboard pattern within the high attenuation layer has been recovered, but the quality of the image is significantly lower than that from visco-acoustic checkerboard test. The attenuation model appears to be more sensitive to elastic artifacts than the velocity model and the inversion failed to recover the 200 m attenuation perturbation (not shown here) due to the stronger artifacts from mode conversion. Seismic P-S and S-P mode conversion play an important role in reducing the amplitude of the seismic wavefield (Malin and Phinney, 1985). This loss of amplitude is incorporated in the attenuation model during the waveform tomography process. We can try to quantify this effect by comparing the results of acoustic and elastic modeling. Let $Q_{P-S}^{-1}$ be a quality factor representing the effects of loss of amplitude from mode conversion in the elastic data. The expression for P-wave
attenuation of the visco-elastic modeled data takes the form:

\[ Q_{\text{elastic}}^{-1} = Q_{\text{int}}^{-1} + Q_{\text{sc}}^{-1} + Q_{P-S}^{-1}. \]  \hspace{1cm} (4.4)

Combining equations 4.2 and 4.4 we have:

\[ Q_{\text{elastic}}^{-1} = Q_{\text{acoustic}}^{-1} + Q_{P-S}^{-1}. \]  \hspace{1cm} (4.5)

When applying visco-acoustic inversion to visco-elastic data, attenuation due to mode conversion \((Q_{P-S}^{-1})\) will be present as artifacts on the attenuation model. From equation 4.5, we can estimate the attenuation due to P-S mode conversion artifacts by subtracting the recovered attenuation of visco-acoustic data from the recovered attenuation of visco-elastic data (equation 4.6):

\[ Q_{P-S}^{-1} = Q_{\text{elastic}}^{-1} - Q_{\text{acoustic}}^{-1}; \]  \hspace{1cm} (4.6)

Figure 4.9 shows the estimated attenuation from mode conversion, using equation 4.6. It appears that most of the mode conversion artifacts are present on the southern side of the model, where the high velocity contrasts are shallower. On the northern side of the model, where the velocity contrasts are lower, the artifacts are weaker. This implies that the P-wave attenuation values derived from the application of visco-acoustic waveform tomography to visco-elastic data provide better results when the velocity gradient is small. The artifacts with negative attenuation in Figure 4.9 indicate areas where attenuation has been overestimated whereas the positive values indicate areas where the attenuation was underestimated.
4.7.4 Summary

The application of visco-acoustic waveform tomography to marine visco-elastic data is valid when high velocity and density contrasts are not present in the model. Elastic mode conversion artifacts in the recovered velocity and attenuation models limit model resolution, with the attenuation model being more sensitive to these artifacts than the velocity model. The checkerboard tests show that the derived velocity model from the Puget Lowland data has a resolution of 400-200 m, but the resolution of the attenuation model is limited to approximately 400 m.
Figure 4.7: 400 m and 200 m recovered velocity perturbation pattern from the acoustic ((a) and (c)) and elastic ((b) and (d)) modeled data. The quality of the elastic model is comparable to the acoustic model for the 400 m perturbation size, but for the 200 m perturbation size, the quality of the model is degraded in places by artifacts due to elastic mode conversions (indicated with arrows).
(a) Visco-acoustic

(b) Visco-elastic

(c) Visco-acoustic

Figure 4.8: 400 m and 200 m recovered attenuation perturbation patterns from the visco-acoustic ((a) and (c)) and visco-elastic ((b)) modeling. The undulating high attenuation layer along the model has been well recovered in the visco-acoustic modeling. The checkerboard pattern is less well recovered with visco-elastic modeling and artifacts are present between 1 and 12 km.

Figure 4.9: Estimated attenuation due to mode conversion. The greatest attenuation is present at 1-9 km, which corresponds to areas of high velocity gradients.
4.8 Geological Interpretation

4.8.1 Lithology from Velocity and Attenuation

South of the SFZ and in the Seattle basin (Figures 4.10a and 4.10b), the velocity model from waveform tomography shows just below the seafloor, a unit with velocities in the range 1450-1700 ms$^{-1}$ and a thickness varying between 100 and 300 m. This unit is absent within the SFZ and north of x = 20 km where higher velocities are present close to the seafloor. We interpret this unit to be unconsolidated postglacial Quaternary sedimentary rocks, because these strata fill small basins in the underlying higher velocity strata and the velocity range is consistent with post-glacial Quaternary sediment found in other areas, for example in the Queen Charlotte Basin (Halliday et al., 2008; Takam Takougang and Calvert, 2012). Johnson et al. (1999) have interpreted these rocks to be late Pleistocene to Holocene (post-glacial) in age. Under this unit, which is bounded by an angular unconformity (U1), the velocity increases more rapidly and the underlying strata are characterized by velocities in the range 1700-2300 ms$^{-1}$. The Kingston 1 well, which lies 30 km from the SFZ (Figure 4.1) shows that sonic velocities for the Pleistocene strata are in the range 1600-2200 ms$^{-1}$ (Brocher and Ruebel, 1998); velocities as high as 2090 ms$^{-1}$ and 2395 ms$^{-1}$ were found in Pleistocene strata within 40 m of the surface in an onshore seismic refraction survey (Williams et al., 1999). We therefore interpret this unit to be a glacial Pleistocene deposit. Undulating high attenuation values in the range $1000/Q = 100-270$ or $Q = 4-10$ appear to be globally continuous along the line (Figures 4.10c and 4.10d). We interpret attenuation values in the range $1000/Q = 50-150$ or $Q = 7-20$ associated with the glacial Pleistocene velocities in the range 1700-2300 ms$^{-1}$ to be due to silts and clay from the Lawton Clay Member deposited during the Fraser glaciation (Tubbs, 1974). Poorly to semi consolidated Pleistocene clays were found in the Kingston 1 well above 500 m depth (Rau and Johnson, 1999), and similar inverted
seismic attenuation values for silts and clay were identified in the Queen Charlotte Basin (Takam Takougang and Calvert, 2012). Another unconformity, U2, separates the glacial Pleistocene deposit from Tertiary sedimentary rocks characterized by velocities in the range 2400-3000 ms$^{-1}$. We interpret the Tertiary sediment within the SFZ to be Oligocene Blakeley Formation, which is exhumed at Alki Point, 1 km east of line PS-2 (Calvert et al., 2003), because most of the Blakeley Formation is characterized by velocities between 2400 and 3400 ms$^{-1}$ (Brocher and Ruebel, 1998). The deeper Tertiary sedimentary rocks within the Seattle basin were interpreted by Pratt et al. (1997) and Brocher et al. (2004) as Blakely Harbor Formation.

South of the line, within the Seattle uplift at x = 0-9 km, high velocities in the range 3100-4000 ms$^{-1}$ associated with attenuation as high as 1000/Q = 270 or Q≈ 4 occur at depths of 500-1000 ms. This velocity range is consistent with volcanic rocks from the Puget group in the Socal-Schroeder 1 well (Brocher and Ruebel, 1998), and based on ties to the Kingston 1 well and a P-wave velocity model, ten Brink et al. (2002) have interpreted the 4000 ms$^{-1}$ isovelocity contour at 1-1.5 km depth to be the top of Eocene marine strata. The elevated attenuation values, which are overestimated in places due to elastic mode conversion at x = 2-8 km, are consistent with attenuation of volcanic rocks found in various areas such as, the Queen Charlotte Basin (Takam Takougang and Calvert, 2012) or the Lau basin (Wepfer and Christensen, 1990). Outcrops of Eocene basalts were identified east and west of Puget sound (Haeussler and Clark, 2000) and a number of geological and geophysical data suggest that Eocene volcanic rocks are shallow or crop out along the Seattle uplift (Brocher et al., 2004). We therefore interpret the high velocity values associated with elevated attenuation in the southern part of the line to be due to Eocene volcanic rocks of the Crescent terrane.
Figure 4.10: Interpretation of stratigraphy and faults in the final velocity model (a), velocity model and migrated section (b), attenuation model (c) and the attenuation and migrated section (d). F1-F9 are thrust faults; Eo-Eocene volcanics; Ax-Axial surface within the monocline; P-glacial Pleistocene; U1-Pleistocene-Holocene/glacial Pleistocene unconformity; U2-Tertiary/Quaternary unconformity; A1-Blake Island anticline; A2-Winghaven Park anticline; S-syncline; Bl-Blakeley Formation; m shows seafloor multiples.
4.8.2 Geometry and Faulting

Our interpretation of faults is based on undulations and sharp lateral variations in the velocity model, and their correlation with features in the migrated seismic section and the attenuation model. The migrated seismic section was mostly used for the interpretation of deeper faults, i.e. below the maximum depth of the velocity model (brown dashed lines in Figure 4.11).

In the deeper part of the velocity model, we identified 9 shallow faults denoted F1-F9 (Figure 4.10), which also appear to correlate with changes in the attenuation model as well. Faults F1-F4 are based on sharp velocity variation in the SFZ and are interpreted as high-angle faults, whereas faults F5-F9 are interpreted from lateral velocity variations within the Blakes Island Anticline (A1) and the Winghaven Park Anticlines (A2) (Brocher et al., 2004), and have moderate dips. Anticlines A1 and A2 are well imaged in our velocity model and are separated by a syncline (S). Faults F1 and F2 are characterized by a sharp lateral decrease in velocity at $x = 12-13.5$ km, and intersect in the subsurface at $\approx 900$ m depth. Fault F1 is south dipping with a dip of roughly $80^\circ$ whereas fault F2 is north dipping with a lower dip. Fault F1 and F2 delimit a region characterized by lower velocity and discontinuities in attenuation values, which appear to correlate with the continuation of the Vasa Park trench located east of Puget Sound near Lake Sammamish (Liberty and Pratt, 2008). North-dipping fault F3 is characterized by a discontinuity in the velocity model at $x = 10.5$ km, where velocities appear offset. Fault F4 is located at $x = 9$ km where Pleistocene velocities are concave, and there is a dislocation in the attenuation model. F4 is north dipping with a dip of $\approx 60^\circ$. Faults F5-F9 are sub-parallel and are all north dipping with a moderate dip of $\approx 40^\circ$-$45^\circ$. The north-dipping orientation of these shallow faults is in agreement with physiographic and paleoseismic evidence for dominantly north-dipping Holocene thrust faults at the surface. Airbone laser terrane mapping has revealed the presence of postglacial Holocene fault scarps indicative of
north-dipping thrust faults along strands of the SFZ on Bainbridge Island, west of the Puget Sound (Nelson et al., 2003). A south-dipping low velocity zone with a dip of approximately 30°-40° can be identified on the southern limb of anticline A2 between x = 1-3 km in the velocity model (Figure 4.10a), but no south-dipping features can be easily identified here on the migrated section (Figure 4.10b). The pronounced north-dipping Seattle monocline at x = 12-15 km was interpreted by Brocher et al. (2004) as forming the southern margin of the Seattle basin. This feature, which has an important role in the structural interpretation of the SFZ, was first recognized by Pratt et al. (1997) as an inflection (their axial surface F) in the Quaternary/Tertiary unconformity and they related it to fault-propagating folding. The Seattle monocline is characterized in our attenuation model by a discontinuity at x = 14.5 km, which corresponds to the axial surface F of Pratt et al. (1997), and coincides with the synclinal axial surface identified on land seismic data collected east of Puget Sound by Liberty and Pratt (2008). We refer to this discontinuity as the axial surface (Ax) (Figures 4.10 and 4.11).

Numerous structural models have been proposed for the SFZ (Johnson et al., 1996; Pratt et al., 1997; Calvert et al., 2003; Brocher et al., 2004; Liberty and Pratt, 2008); however, based on the orientation of our interpreted faults, F1-F9, We examine two hypothetical models (Figure 4.11):

1. A large thrust fault that accommodates north-south shortening, by forming a fault-propagation fold with a forelimb breakthrough. This interpretation was invoked by Liberty and Pratt (2008) for onshore seismic data collected east of Puget Sound. They interpreted the synclinal axial surface as the deformation front and the exposed fault at the Vasa Park trench as the forelimb breakthrough. The backthrust was interpreted to be adjacent to the Newport Hills anticline to the south, and the presence of Eocene rocks that crop out on the Newport Hill anticline was interpreted to be due to the motion of the main thrust fault. This interpretation was postulated to be valid both
east and west of Puget Sound (Liberty and Pratt, 2008).

2. A passive-roof duplex (Brocher et al., 2004). In this case, the SFZ is located at the leading edge of a triangle zone within the Seattle uplift. The Seattle uplift is underlain by a passive roof duplex bounded top and bottom by roof and thrust faults respectively with opposite senses of vergence.

Discriminating between these two interpretations is not straight forward due to the absence of clear reflections in the seismic reflection image, especially within the SFZ. We extrapolate faults F1-F9 from the final velocity model on the migrated seismic section to discuss the possible interpretation.

If the south-dipping fault F1 extends deeper, i.e., to a depth of approximately 5 km, then the SFZ beneath Puget Sound can be interpreted as a fault-propagating fold with a forelimb breakthrough (Figure 4.11a). In this case, we can draw the deformation front from the axial surface (Ax), and interpret fault F1 as the forelimb breakthrough. This interpretation is, however, problematic because the identification of this fault on the migrated section is ambiguous. Also, the backthrust in this model of the SFZ is not clear in either the velocity model or the seismic reflection image. If we transpose the interpretation from Liberty and Pratt (2008) to line PS-2, then the backthrust should be located adjacent to the Winghaven Park anticline (A2) to the south (F9). The presence of the south-dipping low velocity zone on the southern limb of anticline A2, adjacent to the backthrust may support this interpretation if the zone represents stratigraphy, because south-dipping strata were also identified on the southern limb of the Newcastle Hills anticline, adjacent to the backthrust, east of Puget Sound (Liberty and Pratt, 2008). However, the low velocity zone cuts across reflection in the migrated section, and it may be alternatively interpreted as an autithetic fault zone. The velocity model indicates that Eocene rock in the anticline does not crop out at the seabottom here, suggesting that any backthrust here would be older (at least Eocene) and not part of the modern SFZ. Faults F2-F5 are most likely accommodating
north dipping strata. Fault F6 can also be interpreted as a backthrust, but at that location as well, no Eocene rocks are present at the seabottom, contradicting the hypothesis that the backthrusts fault must have substantially uplifted Eocene rocks Liberty and Pratt (2008). Therefore, the location of the backthrust fault in this interpretation remains problematic. It is however possible that the backthrust is located further to the south, outside our study area.

If the north-dipping fault F2 extends deeper, then our interpretation could support the SFZ as being part of a passive-roof duplex (Figure 4.11b). The floor thrust is south dipping, and merges with the roof thrust in the subsurface, forming a triangle zone beneath the Seattle uplift. The roof thrust is interpreted to lie along the contact between Tertiary sedimentary rocks and the top of the Crescent formation. The north-dipping monocline, referred to by Brocher et al. (2004) as the Seattle monocline, forms the tip of the triangle zone and the southern margin of the Seattle basin along the leading edge of the Seattle fault zone. Faults F2 to F9 are rooted in the roof thrust and accommodate with anticlines the northward propagation of the triangle zone into the Seattle basin. It is interesting to note that the location and orientation of fault F2-F9 closely match the interpretation of Brocher et al. (2004). The interpretation of the SFZ as part of a passive roof duplex was however criticized by Liberty and Pratt (2008), who argued that the presence of Eocene rocks at the surface in the Newcastle Hills demonstrates an inconsistency between the amount and timing of shortening in the roof thrust and the basement thrusts.
Figure 4.11: Superposition of the final velocity model on the migrated seismic reflection section truncated at 6 km depth. (a) shows fault propagation fold interpretation and (b) the passive roof duplex interpretation. The dashed brow lines show the prolongation of faults F2-F9 on the migrated section. The base of the Miocene Blakely Harbor Formation (pink line) is derived from (Brocher et al., 2004). The gray lines show in (a) the interpretation inferred from (Liberty and Pratt, 2008) and in (b) that from (Brocher et al., 2004). Eo-Eocene volcanics; Ax-Axial surface within the monocline; P-glacial Pleistocene; U1-Pleistocene-Holocene/glacial Pleistocene unconformity; U2-Tertiary/Quaternary unconformity; A1-Blake Island anticline; A2-Winghaven Park anticline; Bl-Blakeley Formation; S-syncline; Bh-Blakely Harbor Formation; Cr-Crescent formation; m is seafloor multiples.
4.9 Conclusion

We used SHIPS seismic reflection data from Puget Sound to test the reliability of the application of visco-acoustic waveform tomography to marine visco-elastic data. Using the velocity and attenuation models recovered from the field data, we derived synthetic acoustic and elastic data that we used in a series of checkerboard tests. The results of these tests show that the application of visco-acoustic waveform tomography to marine visco-elastic data is valid when no strong velocity gradient or velocity contrasts are present. When strong velocity gradients exist, the quality of the model is degraded by artifacts arising from elastic mode conversions at layer interfaces, which cannot be accurately modeled using acoustic waveform tomography. When the velocity gradient is low, the mode conversion artifacts become weaker, and the quality of the model improves significantly. With the SHIPS seismic reflection data, the checkerboard tests show that our recovered velocity model has a resolution of 400-200 m whereas the resolution of the attenuation model, which is more sensitive to the mode conversion artifacts, is limited to 400 m.

Our joint interpretation of the velocity model, attenuation model and the seismic reflection profile defines two Quaternary units separated by an unconformity. The top unit is interpreted to be post-glacial and the bottom unit is interpreted as glacial Pleistocene sedimentary rocks. The glacial Pleistocene unit is separated from Tertiary sedimentary rocks by an angular unconformity. The southern part of the line within the Seattle uplift is characterized by the presence of Eocene volcanic rocks at depths as shallow as 500 m. We identified anticlines, a syncline and 9 north-dipping shallow faults across the Seattle uplift and the Seattle fault zone. Based on the geometry of our interpreted faults two hypothetical models may apply to the SFZ. The SFZ can be described as large south-dipping thrust fault that accommodates north-south shortening, by forming a
fault-propagation fold with a forelimb breakthrough, or alternatively the SFZ is located at the leading edge of a passive roof duplex, in which a floor thrust and a roof thrust merge in the subsurface forming a triangle zone that is propagating into the Seattle basin.

4.10 Acknowledgments

We are grateful to Gerhard Pratt, who provided us with his waveform tomography code, and to Johan Robertson for the elastic modeling code. Rie Kamei, provided helpful suggestions for the elastic checkerboard testing. We used SeismicUnix to plot and partially process the data. This project was funded by the Natural Sciences and Engineering Research Council of Canada.
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Conclusions and Discussion

Visco-acoustic waveform tomography is a powerful imaging technique, but its successful application requires a robust inversion strategy and carefully selected preconditioning of the field data. I developed a specific inversion strategy and data preconditioning to successfully apply visco-acoustic waveform tomography to marine seismic reflection data from the QCB of western Canada and the SFZ, Washington. The absence of strong velocity contrasts in the velocity model appears to be a critical condition for the successful application of the method to marine visco-elastic data.

5.1 Data Preconditioning and Inversion Strategy

The preconditioning of the marine seismic reflection data from the QCB and the SFZ consisted of trace editing and coherent noise removal using f-k filtering, time windowing, shot-to-shot amplitude balancing and 2D amplitude correction. The purpose of this preconditioning was to arrange the data in a form similar to that predicted by the acoustic wave equation.

The inversion strategy was designed to mitigate non-linearity, and to ensure successful application of the method by encouraging the convergence toward a global minimum. The
main challenges were the limited maximum offset (3770 m for the QCB data and 2575 m for the SFZ data), which makes it difficult to recover deeper structures, and the relatively high starting frequency (7 Hz for the QCB data and 4.4 Hz for the SFZ data) that increases the non-linearity. My inversion strategy consisted of successively recovering shallow to deep structures in the subsurface velocity and attenuation models by using sequential time damping (from lower to higher values) for every frequency group, in combination with a layer stripping approach implemented by inverting from the near to far offsets, and weighting the gradient with depth to boost in the misfit function the contribution of the layer under reconstruction. The sequential time damping enables the successive inclusion in the inversion of more data that illuminate the deeper parts of the model. The starting P-wave velocity model was derived from ray-based travel time tomography, and care was taken to ensure that it predicted the first arrival waveforms of the field data to within half a cycle. This inversion strategy was successfully applied to image a section of line 88-06 from the QCB, to a depth of 1.2 km, twice the coverage of the traveltime tomography velocity model. Both the full waveform and efficient waveform tomography approaches were evaluated. The full waveform tomography and efficient waveform tomography approaches converged to similar results, showing the robustness of the inversion strategy, but the results from the full waveform tomography approach appeared to have a slightly greater resolution. The full waveform tomography approach was subsequently used in all lines. Initially, on a section of line 88-06 of the QCB, attenuation was introduced in the inversion at higher frequencies ($\geq 10.5$ Hz), but the inverted attenuation model looked unrealistically heterogeneous, and the presence of shadow zones (zones with relatively weak seismic amplitude) made the successful inversion of line 88-05 of the QCB difficult. Therefore, I refined the inversion strategy by alternating between phase-only and amplitude-plus-phase velocity inversion for the first two pairs of frequencies, and added a second step, in which I inverted for attenuation from the lowest frequency using the final
recovered velocity model and an initial homogeneous $Q_p$-model. This inversion strategy gave the best velocity and attenuation images from lines 88-04, 88-05, 88-06 and 88-07 of the QCB data and from line PS-2 of the SFZ data. There was a good correlation of the P-wave velocity and attenuation models with the available sonic logs, gamma ray logs and interpreted migrated seismic sections from previous studies and between field data and modelled data.

5.2 Reliability of Visco-acoustic Waveform Tomography of Marine Visco-elastic Data

The use of acoustic and visco-acoustic approximations to model visco-elastic data is common to reduce the computational cost, but it is also essential to understand the requirements and conditions under which this approximation works best. The quality of the inverted velocity and attenuation models at basement characterized by velocity $\geq 4000$ ms$^{-1}$ looks significantly degraded, especially in the QCB. Also, anomalously low velocities are present in places at the seafloor and at layer interfaces, mostly in the SFZ velocity model.

Visco-acoustic and visco-elastic modelling were performed with the recovered velocity and attenuation models from the SFZ data, and the derived synthetic data were used in a series of checkerboard inversion tests to better understand the reliability of the results. The results of the tests show that the application of visco-acoustic waveform tomography to marine visco-elastic data is valid when strong velocity gradient are not present in the model. When strong velocity gradient are present, the quality of the model is degraded by the presence of elastic mode conversion artifacts, which reduce the expected resolution at higher frequencies. When the velocity gradient is low, the mode conversion artifacts become weaker and the quality of the model significantly improves. This explains the
poor quality of the result at basement were strong velocity gradient are present and the anomalously low velocity at the seafloor. The checkerboard tests show that the resolution of the recovered velocity model from the SFZ data, inverted up to 17.6 Hz, is around 400-200 m whereas the resolution of the attenuation model, which is more sensitive to mode conversion artifacts is not less than 400 m.

5.3 Geological Interpretation of the Results from the QCB

The application of waveform tomography to the QCB, from Hecate Strait to Dixon Entrance to a depth of approximately 1.2 km, enables the identification of a number of shallow geological structures. The velocity models enable the identification of Quaternary strata and shallow faulting in the Pliocene section, not readily interpretable on the original migrated sections and the upward extension into the Pliocene of some faults previously identified on the migrated section below the Pliocene/upper Miocene unconformity, suggesting that the transtensional phase of basin evolution continued longer than previously supposed. Decreases in velocity of approximately 150 ms\(^{-1}\) at the apex of anticlinal folds on lines 88-06, 88-04 and 88-07 are interpreted to be due to tensile fracturing.

Joint interpretation of P-wave velocities and attenuation models in sediments show that they are generally consistent with values for saturated sandstones, with the exception of areas of elevated attenuation associated with localized low velocities. These regions identified on lines 88-04, 88-05 and 88-06 at 700-1000 m depth correlate with greater estimated clay content and with the presence of siltstone and shale in the Tyee N-39 well and are therefore interpreted to be due to siltstone and shale. High attenuation associated with anomalously low velocity zones was also found at the Sockeye B-10 well location on lines 88-05 and 88-04, and in this case, correlates with the occurrence of oil staining in the well at 950-1050 m depth. A shallow V-shape low velocity anomaly was identified on line
88-05 immediately beneath a chain of seafloor pockmarks, where authigenic carbonate chimneys, which are interpreted to be related to hydrocarbon seepage, occur. West of the V-shaped anomaly and deeper on line 88-05, sub-vertical zones of low velocity associated with high attenuation values are interpreted as pipe-like chimneys resulting from gas ascension.

Basement rocks are characterized by high attenuation values and high velocity and are interpreted to be volcanic, but the elevated attenuation is also due to elastic mode conversion artifacts, which are particularly strong where velocity gradients are high.

5.4 Geological Interpretation of the Results from the SFZ

Across the SFZ, the joint interpretation of the velocity model, attenuation model and the seismic profile shows two Quaternary units separated by an unconformity. The top unit is interpreted as Post-glacial and the bottom unit is interpreted as Glacial Pleistocene. The Glacial Pleistocene unit is separated from the Tertiary sediment by an angular unconformity. The southern part of the line within the Seattle uplift is characterized by the presence of shallow Eocene volcanic rocks.

We identified anticlines, a syncline and 9 north-dipping shallow faults across the Seattle uplift and the Seattle fault zone. Based on the geometry of the interpreted shallow faults, two interpretations of the SFZ are possible:

1. A passive roof duplex bounded by a floor thrust and a roof thrust that merge in the subsurface forming a triangle zone within the Seattle uplift. The SFZ is located at the leading edge of a triangle zone and the interpreted shallow faults are rooted into the roof thrust, accommodating the northward propagation of the triangle zone into the Seattle basin.

2. A large thrust fault that accommodate north-south shortening by forming a
fault-propagation fold with a forelimb breakthrough. The main fault strand is identified from an axial surface where the deformation front is located; the forelimb breakthrough is located beneath a trench and the shallow faults identified across the SFZ accommodate north-dipping strata. However, the location of the backthrusts fault in this interpretation remains problematic.

5.5 Suggestions for Future Studies

The study of the reliability of visco-acoustic waveform tomography of marine visco-elastic data showed that visco-acoustic waveform tomography works well when high velocity gradients are absent from the model, as is the case with most of the QCB and SFZ data, but with the exception of areas where shallow basement rocks are present. The failure of visco-acoustic waveform tomography in areas where high velocity gradients are present constitutes a handicap to the utilization of the acoustic method, and it would be interesting to find a way to resolve this problem. A solution could be the direct use of the visco-elastic waveform tomography, but as mentioned earlier, its application remains a challenge because of computational cost and its greater non-linearity, which necessitates a robust inversion approach (e.g., Mulder and Plessix, 2008; Brossier et al., 2009). Another solution could be the use of visco-acoustic waveform tomography implemented with a correction in the acoustic wave equation during the forward modelling step that will take into consideration the difference in waveform properties between the predicted acoustic and elastic data. A correction in acoustic amplitude can be estimated by adding a virtual source in the acoustic wave-equation, obtained by comparison with the elastic wave-equation. This approach was investigated by Chapman et al. (2010) with synthetic modelling experiments, and the results look promising.
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Estimates of Computation Times for Waveform Tomography

The version of the frequency domain waveform tomography code provided by Prof. Gerhard Pratt in 2008, does not support the standard geometry for marine seismic reflection surveys. I redesigned the geometry to conform with the requirement of the code, by adding receivers along the entire line of the survey. In this new configuration, for a given shot, only the receivers corresponding to the position of receivers in the field survey are active, and the remaining are dead.

Computation times per iteration for waveform tomography varied depending on the number of sources and receivers, the size of the finite-difference grid and the number of frequencies or group of frequencies used. The inversion includes forward modelling, back-propagation of data residuals, computation of the gradient and model update. The following computation times were obtained when performing the tomography on a PC-based workstation and on a cluster under unix.
A.1 PC-based: RAM = 8Gb with processor Intel Core 2 Duo @ 2.13 GHz

- Half of line 88-06
  Here I used 458 shots, 994 receivers along the line with 80 receivers active per shot, and a grid of 994x51. The computation time for forward modelling was approximately 30 min. For the full waveform inversion, two frequencies were used per group of frequencies and the inversion was performed from 7.5 Hz to 13.66 Hz. The computation time was approximately 4 days for stage 1 (velocity only) and 4 days for stage 2 (velocity and attenuation) which give a total of 8 days.

A.2 Cluster: RAM = 128 Gb, 4 Xeon X7560 processors with 8 cores each @ 2.26 GHz

For the following lines, the code which runs in serial mode was only able to use 1 processor at a time

- Line 88-05
  I used 1022 shots, 2123 receivers with 80 receivers active per shot, and a grid of 2757x44. The computation time for forward modelling was approximately 1.5 hours. For the inversion, two frequencies were used per group of frequencies and the inversion was performed from 7.5 Hz to 12 Hz. The computation time was 9 days for stage 1 (velocity only) and 9 days for stage 2 (velocity and attenuation) which give a total of 18 days.

- Line 88-06
  I used 985 shots, 2058 receivers with 80 receivers active per shot, and a grid of 2672x44. The computation time for forward modelling was approximately 1.5 hours. For the inversion, two frequencies were used per group of frequencies and
the inversion was performed from 7.5 Hz to 12 Hz. The computation time was approximately 9 days for stage 1 (velocity only) and 9 days for stage 2 (velocity and attenuation) which give a total of 18 days.

- **Line 88-07**
  
  I used 1435 shots, 2956 receivers with 80 receivers active per shot, and a grid of 3828x44. The computation time for forward modelling was approximately 2.5 hours. For the inversion, two frequencies were used per group of frequencies and the inversion was performed from 7.5 Hz to 12 Hz. The computation time was approximately 14 days for stage 1 (velocity only) and 14 days for stage 2 (velocity and attenuation) which give a total of 28 days.

- **Line 88-04**
  
  I used 843 shots, 3452 receivers with 80 receivers active per shot, and a grid of 4466x44. The computation time for forward modelling was approximately 1 hour. For the inversion, two frequencies were used per group of frequencies and the inversion was performed from 7.5 Hz to 12 Hz. The computation time was 7 days for stage 1 (velocity only) and 7 days for stage 2 (velocity and attenuation) which give a total of 14 days. In this case, although line 88-04 is longer than lines 88-06 and 88-07, the smaller number of shots significantly reduced the computation time.

- **Line PS-2**
  
  I used 408 shots, 946 receivers with 96 receivers active per shot, and a grid of 816x51. The computation time for forward modelling was only about 20 minutes. For the inversion, three frequencies were used per group of frequencies and the inversion was performed from 4.4 Hz to 17.6 Hz. The computation time was approximately 2 days for stage 1 (velocity only) and 2 days for stage 2 (velocity and attenuation) which give a total of 4 days.