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UNIVERSITY OF CALIFORNIA, SAN DIEGO

Lithosphere structure of the Earth from surface wave tomography

A dissertation submitted in partial satisfaction of the
requirements for the degree
Doctor of Philosophy

in

Earth Sciences

by

Zhitu Ma

Committee in charge:

Professor T. Guy Masters, Chair
Professor Paterno R. Castillo
Professor M. Gabriele Laske
Professor Peter M. Shearer
Professor Daniel M. Tartakovsky

2015

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The dissertation of Zhitu Ma is approved, and it is acceptable in quality and form for publication on microfilm and electronically:

Chair

University of California, San Diego

2015

DEDICATION

To my parents

EPIGRAPH

It is not the critic who counts ... The credit belongs to the man who is actually
in the arena, whose face is marred by dust and sweat and blood; who strives
valiantly; who errs, who comes short again and again, because there is no effort
without error and shortcoming.

– Theodore Roosevelt

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Chapter 3, in full, is a reformatted version of a publication in Geophysical Journal International: Ma, Z., G. Masters, G. Laske, and M.E. Pasyanos (2014), A comprehensive dispersion model of surface wave phase and group velocity for the globe, Geophys. J. Int. 199 (1), 113-135. DOI: 10.1093/gji/ggu246. I was the primary investigator and author of the paper.

Chapter 4, in full, has been submitted for publication of the material as it may appear in Geophysical Journal International: Ma, Z. and G. Masters, Effect of earthquake locations on Rayleigh wave azimuthal anisotropy. I was the primary investigator and author of the paper.

Chapter 5, in full, has been submitted for publication of the material as it may appear in Geophysical Journal International: Ma, Z. and G. Masters, 2D global Rayleigh wave attenuation model by accounting for finite-frequency focusing and defocusing effect. I was the primary investigator and author of the paper.

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Ma, Z. and G. Masters, A new global Rayleigh and Love wave group velocity dataset for constraining lithospheric properties, *Bull. Seism. Soc. Am.*, 104 (4), 2007-2026, DOI:10.1785/0120130320, 2014.

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Laske, G., G. Masters., Z. Ma and M. Pasyanos, Update on CRUST1.0 - A 1-degree Global Model of Earth's Crust, *Geophys. Res. Abstracts*, 15, Abstract EGU2013-2658, 2013.

ABSTRACT OF THE DISSERTATION

Lithosphere structure of the Earth from surface wave tomography

by

Zhitu Ma

Doctor of Philosophy in Earth Sciences

University of California, San Diego, 2015

Professor T. Guy Masters, Chair

The lithosphere is commonly defined as the outermost rigid layer of the Earth, including the crust and the uppermost part of the mantle. It forms the rigid ‘plate’ in plate tectonic theory. To study its large-scale structure globally, I use fundamental mode surface waves with a period range between 25 s and 200 s. A large dataset including all long period seismograms publicly available for earthquakes with magnitudes larger than $Mw = 5.5$ between 1977 and 2007 has been assembled. To analyze such a large dataset, an efficient measurement method has been developed based on a cluster analysis technique to measure the velocity and amplitude variations of the surface wave packages.

These measurements are then inverted for a self-consistent global dispersion model for both Rayleigh and Love waves, which provides the basic input data for

constraining the lithospheric structure at high resolution. These maps of surface wave phase/group velocities match surface tectonics very well. Slow anomalies are found beneath orogenic zones and other regions with thick crust, e.g. the Himalayas and Andes. The Basin and Range province in North America, mid-ocean ridges, and back-arc basins also show up as slow anomalies. Cratons can be seen in the low frequency maps as regions of anomalously high velocities.

I also find that azimuthal anisotropy needs to be included in order to obtain reliable isotropic velocity variations for Rayleigh waves, and the uncertainties in earthquake locations can affect the resulting azimuthal anisotropy. In addition, I apply finite-frequency theory to account for the focusing-defocusing effect when modeling the amplitude data and present a set of 2D global attenuation maps for Rayleigh waves.

The products from this thesis are expected to find wide use by the seismological and the broader geophysical community. One particular application would be to use the maps of phase/group velocity and attenuation to investigate the evolution of the oceanic lithosphere. All measurement and model files produced in this thesis have been made available on the Internet.

Chapter 1

Introduction

1.1 Observation

The practice of observational seismology always starts with analyzing the wiggles of a *seismogram*, which is the recording of the Earth's vibrations after an earthquake. To understand the large scale structure of the lithosphere globally, I focus on studying long period seismograms and, in particular, the surface wave signals with periods between 25 s and 200 s.

Figures 1.1 and 1.2 give examples of these seismograms. This example is for an earthquake which happened in the Indian Ocean and recorded by two stations: WUS in China and MBAR in Uganda. What is immediately clear is the *dispersion* property of the surface waves. Unlike body waves, the propagation speed of surface waves varies substantially as a function of frequency. In this example, we can see that the part of the surface wave package with a period around 50 s arrives the earliest and hence has the highest velocity.

In many studies, these seismograms are band-passed using a Gaussian filter to isolate a specific frequency band. In this thesis, I also correct for the dispersion properties and the resulting seismograms are shown in Figure 1.2. These seismograms are also normalized to their predicted amplitudes and shifted according to the predicted arrival times based on a 1D Earth model (i.e. assuming Earth properties vary only as a function of depth). We can see that the surface wave package at WUS arrives a lot later than the one at MBAR. I will show later that it is due

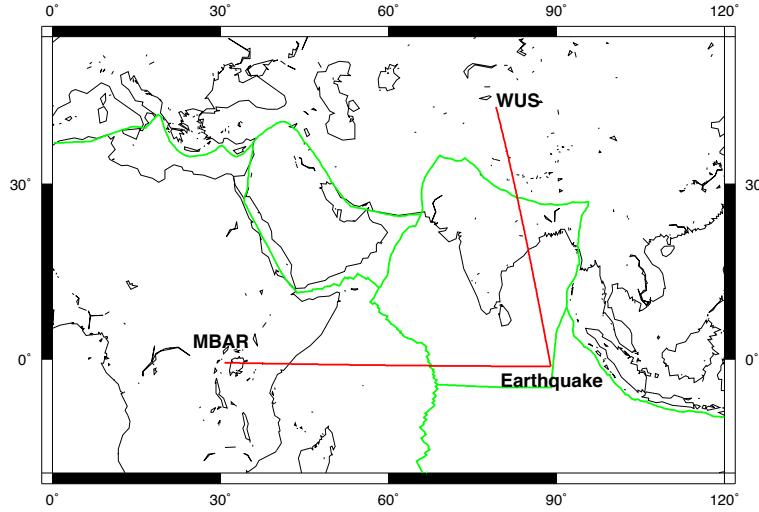


Figure 1.1: The locations of the earthquake and the two stations WUS and MBAR used in Figure 1.2. Also plotted are the plate boundaries from DeMets et al. (1994).

to the thick crust along the WUS path underneath the continent and especially underneath Tibet. The surface wave recorded at MBAR has a smaller amplitude and this is due to the attenuation along the mid-ocean ridge system of the Indian Ocean.

Despite the difference in amplitudes and arrival times, the shapes of the processed seismograms are remarkably similar. This enables us to develop a very efficient method to measure their relative amplitudes and arrival times. Chapters 2 and 3 of this thesis describe the details of this method.

It is also worth pointing out that by using seismograms like this, we can gain a tremendous amount of information along the ray paths that cover part of the Indian Ocean, India, Tibet and Africa, regions that are difficult to get to physically.

Motivated by this, a large dataset including all long period seismograms from 1977 and 2007 that are publicly available from the Incorporated Research Institutions for Seismology (IRIS) has been assembled. While destructive earthquakes are relatively rare, roughly 1-2 earthquakes with magnitudes large enough (I use $M_W \geq 5.5$ in this thesis) to be recorded globally happen everyday (Figure

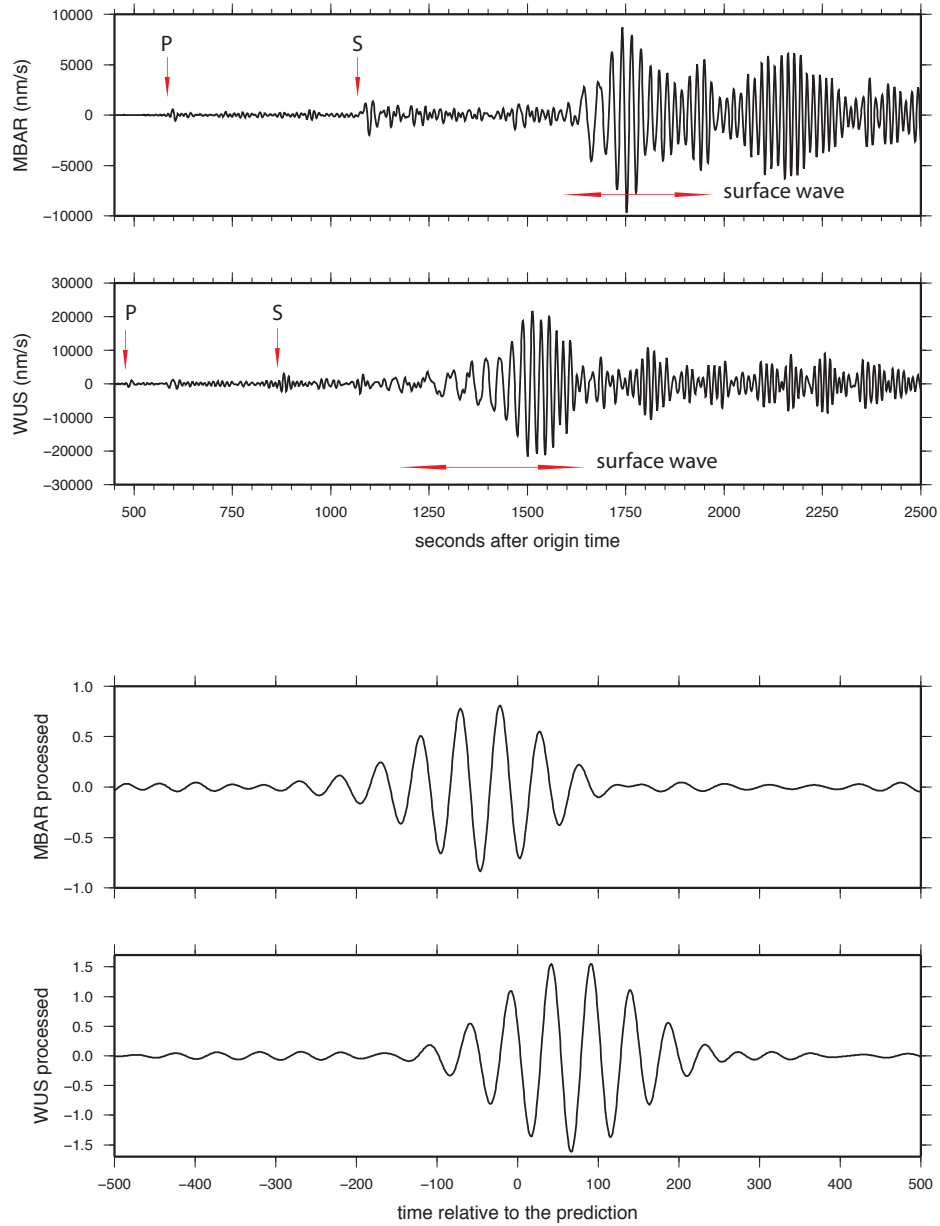


Figure 1.2: Examples of seismograms. The top two panels show the seismograms from stations MBAR and WUS after correcting for instrument response to show the true ground motion. The body wave P and S arrivals are identified. The portion of the surface wave packages this thesis focuses on is also marked. The bottom two panels show the processed seismograms which are band-passed with a center frequency of 20 mHz. Note the time shifts and amplitude variations in the processed seismograms.

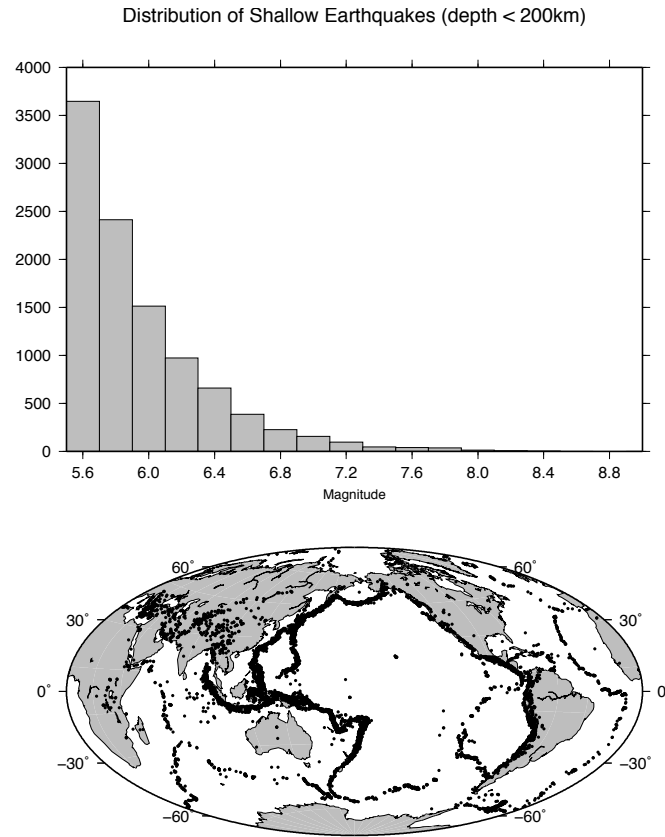


Figure 1.3: Distribution of shallow earthquakes (depth < 200 km) from 1977 and 2007 analyzed in this thesis.

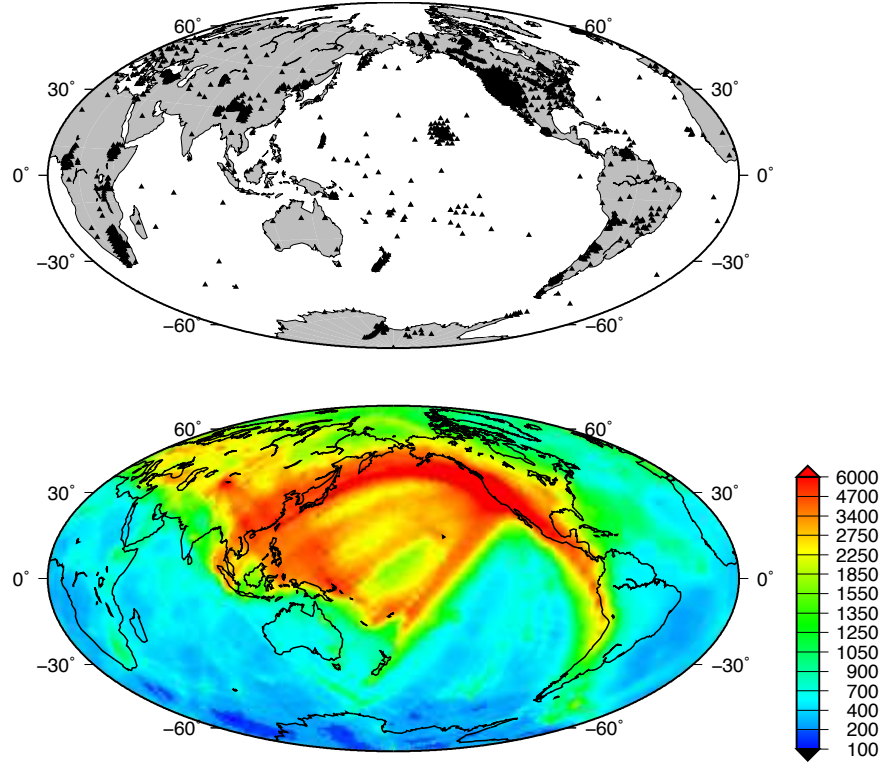


Figure 1.4: top: Distribution of seismic stations used in this thesis. bottom: the number of ray paths passing each of the $1^\circ \times 1^\circ$ equal area block.

1.3).

Thanks to the world-wide efforts on improving seismic recordings since the early 1960s (see a thorough review by Agnew et al., 2002), the Earth vibrations from these earthquakes can now be routinely and robustly recorded by a large array of seismic stations (Figure 1.4). A total of about 2.5 million seismograms from these stations have been analyzed during my Ph.D. and the products of this analysis will be presented in later chapters of this thesis.

1.2 Theory

1.2.1 Describing a seismogram

The standard theory to describe a long period seismogram is based on the normal mode theory which was systematically formulated in Gilbert (1971). A

thorough treatment of this formulation can be found in Dahlen & Tromp (1998). In this section, I will give a brief summary of the part of the theory that is relevant to this thesis. Definitions of key scientific terms will be in *italics*.

The fundamental physics is no more than Newton's second law and we aim to solve the following equation in spherical coordinates (r, θ, ϕ) (eqn 9.1 in Nolet, 2008)

$$\rho \frac{\partial^2 \mathbf{s}}{\partial t^2} + \mathcal{L} \mathbf{s} = \mathbf{f} \quad (1.1)$$

where ρ is the density, $\mathbf{s}$ is the displacement field and $\mathbf{f}$ represents the source. The symbolic operator $\mathcal{L}$ is linear and includes the derivatives operating on elastic constants and displacements.

On a spherical Earth, a convenient way to solve this equation is to decompose the displacement (which is a vector field) using (details in section 8.6.1 in Dahlen & Tromp, 1998)

$$\mathbf{s} = \hat{\mathbf{r}}U + \nabla_1 V - \hat{\mathbf{r}} \times (\nabla_1 W) \quad (1.2)$$

where U, V, W are scalars and are often expanded in spherical harmonics:

$$U(r, \theta, \phi) = \sum U_l^m(r) Y_l^m(\theta, \phi) \quad (1.3)$$

$$V(r, \theta, \phi) = \sum V_l^m(r) Y_l^m(\theta, \phi) \quad (1.4)$$

$$W(r, \theta, \phi) = \sum W_l^m(r) Y_l^m(\theta, \phi) \quad (1.5)$$

and $U_l^m(r)$, $V_l^m(r)$, $W_l^m(r)$ are now functions only of the radius. The surface gradient in eqn 1.2 is simply:

$$\nabla_1 = \hat{\boldsymbol{\theta}} \frac{\partial}{\partial \theta} + \frac{1}{\sin \theta} \hat{\boldsymbol{\phi}} \frac{\partial}{\partial \phi} \quad (1.6)$$

It can be shown that after inserting eqn 1.2 into eqn 1.1, the equations determining U and V are decoupled from the ones determining W (eqn 8.43-8.45 in Dahlen & Tromp, 1998). In normal mode literature, one usually calls the displacements related to U and V *spheroidal* modes and the ones related to W *toroidal* modes.

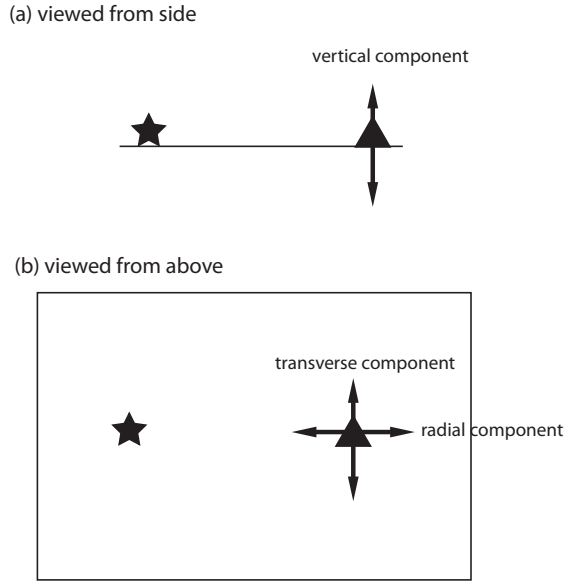


Figure 1.5: Definition of the different components of the recorded displacement field at a seismic station. The source is presented by a star symbol and the receiver by a triangle symbol. Note that not to confuse the radial component used here with the radial direction of the Earth, which corresponds to the vertical component.

A seismogram on either the vertical, radial or transverse component (definition shown in Figure 1.5) can then be written asymptotically as (eqn 11.33-34, 11.42-44 in Dahlen & Tromp, 1998)

$$s(\theta, \phi, t) = \int A(\omega) e^{i[\omega t - k(\omega)\Delta]} d\omega \quad (1.7)$$

where the dependence of the wave number k on frequency ω is explicitly written out. All the terms in eqn 1.7 can be computed based on the scalars U , V and W for a particular Earth model and a given earthquake source.

Because W is decoupled from U and V , in the surface wave literature, one usually calls the seismograms composed of fundamental toroidal modes *Love* waves (named after A. E. H. Love) and the ones composed of fundamental spheroidal modes *Rayleigh* waves (named after Lord Rayleigh). Typically, Love waves are measured on the transverse component and Rayleigh waves are measured on the radial and vertical components.

In surface wave analysis, one usually chooses to fix the frequency ω and think of wave propagation using *phase velocity* c (or slowness $s = 1/c$):

$$c(\omega) = \omega/k(\omega) \quad (1.8)$$

Another important observable is the *group velocity* U , which is defined as

$$U(\omega) = d\omega/dk(\omega) \quad (1.9)$$

It is related to the point where the phase $\omega t - k(\omega)x$ is stable and contributes the most to the integral in eqn 1.7.

Figure 1.6 shows the difference between the phase and group velocity. The velocity of the peaks and troughs correspond to the phase velocity while the peaks of the envelopes (hence the energy) correspond to the group velocity.

In a realistic Earth with 3D structure, one can think of replacing the wave number k with $\bar{k}$ which is defined as

$$\overline{k(\omega)} = \frac{1}{\Delta} \int_0^\Delta k(\omega, \theta, \phi) d\gamma \quad (1.10)$$

where Δ is the epicenter distance from the source to the receiver. In this thesis, the integral is computed along the minor arc connecting the source and receiver and k is varying as a function of colatitude θ and longitude ϕ on the surface of the Earth. Eqn 1.10 can be proven to be a good approximation even when the actual path deviates from the minor arc path as a result of Fermat's principle (eqn 16.109 in Dahlen & Tromp, 1998). It can also be validated based on the path integral method (eqn 42-43 in Woodhouse & Wong, 1986).

In a realistic Earth model, one is also interested in studying the *azimuthal anisotropy*, which means the wave propagation speed can be a function of wave propagation direction. In weakly anisotropic materials, most studies adopt the following form (Smith & Dahlen, 1973):

$$c = c_0 + A_1 \cos 2\phi + A_2 \sin 2\phi + A_3 \cos 4\phi + A_4 \sin 4\phi \quad (1.11)$$

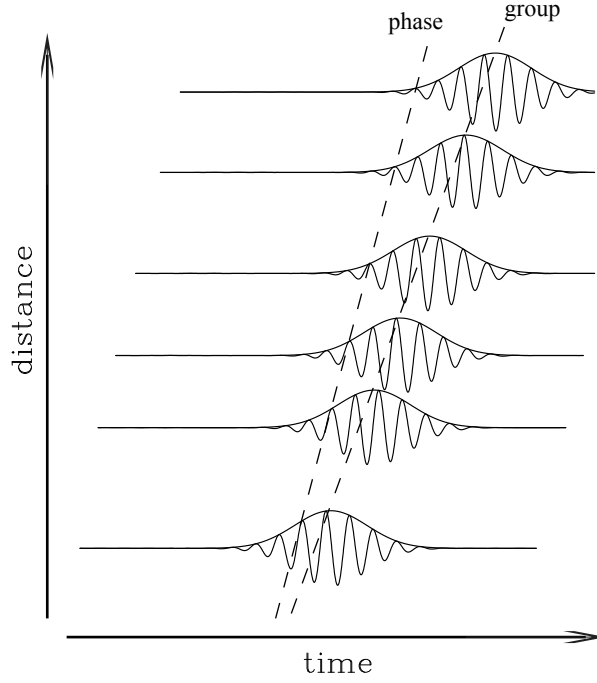


Figure 1.6: A schematic plot showing the difference between phase and group velocity. Note that the phase velocity is higher than group velocity for the Earth.

where phase velocity c is the sum of an *isotropic* term c_0 and four azimuthal anisotropic terms. ϕ is the ray propagation direction, measured clockwise from the North. I will focus on the 2ϕ terms for Rayleigh waves in this thesis.

The perturbation of the amplitude $A(\omega)$ in eqn 1.7 for a 3D Earth is more complicated. It can be expressed as (e.g. Ekström et al., 1997; Dalton & Ekstrom, 2006)

$$A = A_S A_R A_\Delta A_F A_Q \quad (1.12)$$

where A_S is the source excitation, A_R is the receiver response, A_Δ is the geometrical spreading, A_F is the *focusing-defocusing* term due to elastic velocity perturbation and A_Q is the decay factor due to *attenuation*.

Attenuation, which is also called the quality factor, is given by the dimen-

sionless quantity Q in terms of the fractional energy loss per cycle (Shearer, 2009, eqn 6.82)

$$\frac{1}{Q(\omega)} = -\frac{\Delta E}{2\pi E} \quad (1.13)$$

where E is the peak strain energy and $-\Delta E$ is the energy loss per cycle. Strictly speaking, the definition above is for ‘temporal’ Q_t but we normally measure ‘spatial’ Q_s for surface waves. The two are related to each other by

$$\frac{1}{Q_t} = \frac{U}{c} \frac{1}{Q_s} \quad (1.14)$$

I will discuss spatial Q_s exclusively and, for simplicity, I will drop the subscript s for the rest of this thesis.

Among all the terms in eqn 1.12, A_Q is of primary interest. But it is clear that to derive A_Q from the raw measurement A , one has to correct for the other terms first. $A_\Delta = 1/\sqrt{\sin \Delta}$ is well characterized. The source excitation can be computed (eqn 11.34 in Dahlen & Tromp, 1998) given the Centroid Moment Tensor catalog (Ekström et al., 2012). However, neither A_S nor A_R is precisely known due to incomplete knowledge of the sources and receivers. Fortunately, methods have been developed to handle these uncertainties. Two different theories exist to model the focusing-defocusing term (Woodhouse & Wong, 1986; Zhou et al., 2004) and chapter 5 of this thesis will compare these two theories.

1.2.2 Tomography

The concept of *tomography* will be used extensively throughout this thesis. Introductory material can be found in section 5.6 in Shearer (2009). Interested readers can also find the state-of-art of the practice of seismic tomography in Nolet (2008). Here, I will present a simple example to explain the basic concept for general readers.

Consider the problem illustrated in Figure 1.7. One is interested in knowing the velocity (or equivalently the slowness $s = 1/c$) of each of the nine cells. Suppose the researcher can measure the time it takes for the seismic ray to travel along one

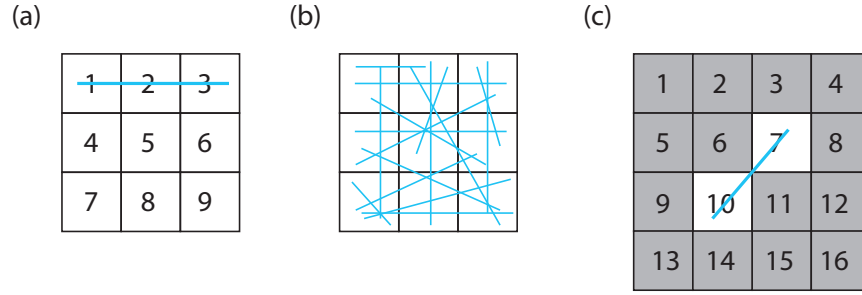


Figure 1.7: An example explaining the basic concept of tomography. (a) An example of one ray path and the cell numbering. (b) After measuring many paths, the tomography problem can now be solved. (c) Occasionally, trade-offs exist among a few cells.

specific path (shown in blue) and the geometry of the path is known (Figure 1.7a), an equation can then be set up based on this single measurement:

$$l_1^1 s_1 + l_1^2 s_2 + l_1^3 s_3 = t_1 \quad (1.15)$$

where l_i^j means the path length in cell j from measurement i , s_j is the slowness of the j th cell and t_i is the measured time. One usually manages to measure many more paths before attempting to solve for each s_j (Figure 1.7b). After the researcher has measured many paths, the tomography problem can now be set up as:

$$Ax = b \quad (1.16)$$

where the matrix A represents all the individual path lengths in each cell from each measurement and b represents the measured travel times. To be explicit, the first row of the system is simply:

$$\begin{pmatrix} l_1^1 & l_1^2 & l_1^3 & 0 & 0 & 0 & 0 & 0 & 0 \\ \vdots & & & & & & & & \\ \vdots & & & & & & & & \\ \vdots & & & & & & & & \\ \vdots & & & & & & & & \end{pmatrix} \begin{pmatrix} s_1 \\ s_2 \\ \vdots \\ s_9 \end{pmatrix} = \begin{pmatrix} t_1 \\ \vdots \\ \vdots \\ \vdots \\ \vdots \end{pmatrix} \quad (1.17)$$

A slightly more complicated situation is shown in Figure 1.7c. Most of the cells are well constrained (colored grey) except the cells No.7 and 10 where only one path is recorded. In this case, a trade-off exists between these two cells as one of them can have arbitrarily high velocity while the other one can have very low velocity, and vice versa. A reasonable solution is to ask for a smooth model, which means the velocity of these two cells cannot be far away from their adjacent ones. To accomplish this, a discrete Laplacian operator $D = \nabla^2$ in the form of

$$\nabla_j^2 = \frac{1}{4}(x_{\text{left}} + x_{\text{right}} + x_{\text{up}} + x_{\text{down}}) - x_j \quad (1.18)$$

is commonly used and added to eqn 1.16:

$$\begin{bmatrix} A \\ \lambda D \end{bmatrix} x = \begin{bmatrix} b \\ 0 \end{bmatrix} \quad (1.19)$$

For cells No.7 and 10, it means adding these two constraints:

$$\begin{aligned} s_7 &= 1/4(s_3 + s_6 + s_{11} + s_8) \\ s_{10} &= 1/4(s_6 + s_9 + s_{11} + s_{14}) \end{aligned} \quad (1.20)$$

Parameter λ controls the level of smoothing. Its choice can vary from problem to problem and methods of choosing this parameter will be described later in this thesis.

Solving eqn 1.19 for the unknown x is the essential goal of any tomography study. In many realistic problems, this system can be quite large (500,000 rows and 41,000 unknowns in my case). Fortunately, the matrix A is quite sparse and efficient algorithms have been developed (e.g. Paige & Saunders, 1982) to solve the problem.

When the width of the seismic rays is considered as in Chapter 5 in this thesis, one replaces the ray path in eqn 1.15 with the corresponding kernel

$$t_i = \int K s \, dA = \sum_j K_i^j s_j \quad (1.21)$$

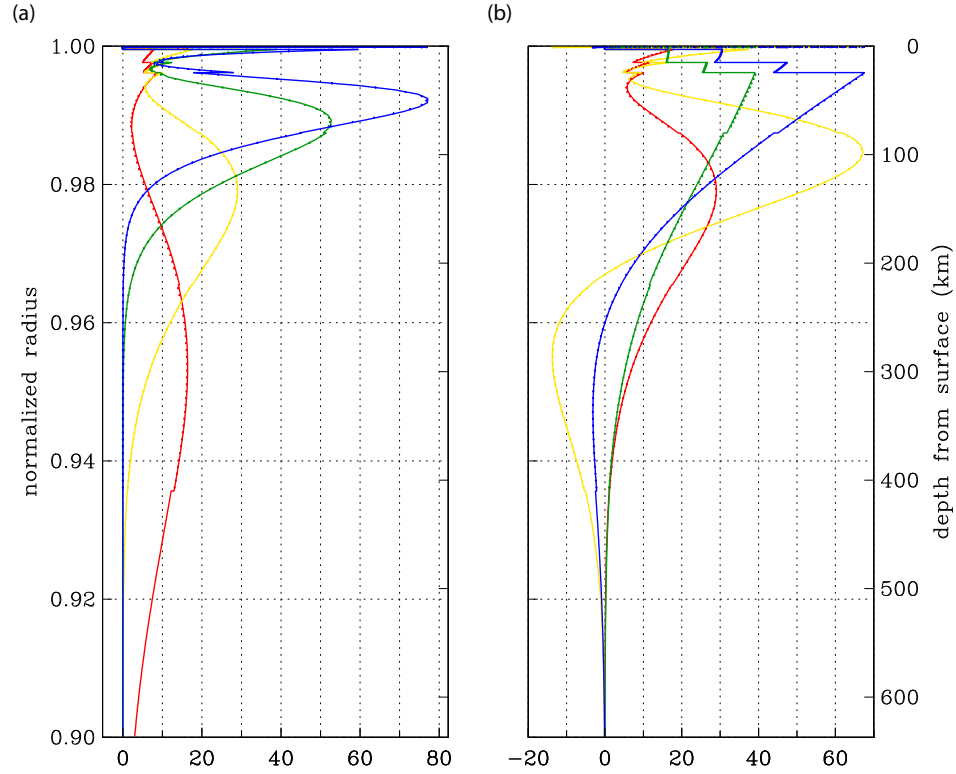


Figure 1.8: Examples of surface wave sensitivity kernels. (a) Sensitivity kernels for Rayleigh waves at different frequencies. red: 5 mHz; yellow: 10 mHz; green: 20 mHz; blue: 30 mHz. (b) Sensitivity kernels for different wave types at 10 mHz. red: Rayleigh wave phase velocity; yellow: Rayleigh wave group velocity; green: Love wave phase velocity; blue: Love wave group velocity. The Earth model used in the computation is the 1D reference model STW105 in Kustowski et al. (2008).

where the integral is done over the area of the kernel. The rest of the theory remains the same.

1.2.3 Sensitivity kernels

This thesis studies the values of surface wave observables at different latitudes, longitudes and frequencies, i.e. to derive $\delta c(\omega, \theta, \phi)$, $\delta U(\omega, \theta, \phi)$ and $\delta Q(\omega, \theta, \phi)$ of our Earth, where the δ means the deviations from a standard reference 1D Earth model.

One can relate these observables that are measured at different frequencies to properties at different depths by considering (use phase velocity as an example)

$$\frac{\delta c}{c}(\omega) = \int_0^1 K_{Vs}(r, \omega) \frac{\delta V_s}{V_s} + K_{Vp}(r, \omega) \frac{\delta V_p}{V_p} + K_\rho(r, \omega) \frac{\delta \rho}{\rho} dr \quad (1.22)$$

where the K are the so called *sensitivity kernels*. The radius r has been normalized by Earth radius $R_a=6371$ km in the equation. For surface waves, the dominant term is K_{Vs} . The details of computing these kernels are in section 9.4 and 11.8 in Dahlen & Tromp (1998) and will not be repeated here.

Strictly speaking, eqn 1.22 is only valid when the perturbation is ‘small’ (which turns out to be a reasonable assumption in many cases), but it gives the readers a direct view of how a certain surface wave observable is sensitive to properties at depth. Figure 1.8a shows examples of these kernels for Rayleigh waves at different frequencies. A general rule is that longer periods can ‘see’ structure at greater depth. Figure 1.8b shows kernels for different wave types at the same frequency. The general rule is, at the same frequency, group velocity is more sensitive to shallow structure than phase velocity and Love waves are more sensitive to shallow structure than Rayleigh waves. It is clear that the depth variation of Earth properties can be recovered by combining surface wave observations from different wave types and different frequencies. This is the main idea of how to use surface wave observables to study Earth structure.

1.3 Motivation

The ‘lithosphere’ is commonly defined as the outermost rigid layer of the Earth, including the crust and the uppermost part of the mantle. On a short time scale, many of the Earth’s dramatic processes humans can experience (e.g. earthquakes and volcanic eruptions) happen in the lithosphere. On a long time scale, the different pieces of the lithosphere can be thought of as the ‘plates’ in plate tectonic theory and the relative motion among these plates through millions of years is one of the major forces that shape our Earth’s surface.

Among various methods available to study lithospheric structure, surface waves provide one of the most valuable constraints (e.g. Trampert & Woodhouse, 1995; Ekström et al., 1997; Shapiro & Ritzwoller, 2002; Nettles & Dziewoński, 2008;

Ritsema et al., 2011; Ekström, 2011). Surface waves are the dominant signals in long period seismograms for shallow earthquakes. In contrast to body waves, surface waves at different frequencies are sensitive to structures at different depths (Figure 1.8), leading to good resolving power in the radial direction (Nettles & Dziewoński, 2008). They can be easily identified and have allowed early investigators to identify large scale lithospheric structures (e.g. Brune, 1969; Dorman, 1969; Knopoff, 1972).

With the rapid development of modern seismometers and the expansion of global seismic networks, surface waves are now providing complete global coverage (Figure 1.4), enabling studies of many remote regions, e.g. the Arctic (Levshin et al., 2001), Antarctica (Ritzwoller et al., 2001), Africa (Li & Burke, 2006), South America (Feng et al., 2004). Recently, these surface wave dispersion measurements and the resulting phase and group velocity maps have also become the essential datasets to constrain shallow upper mantle structure in global mantle models (e.g. Masters et al., 2000; Kustowski et al., 2008; Ritsema et al., 2011).

Despite the swift increase of surface wave datasets, many of the measurement algorithms for group velocity are based on the one first introduced by Dziewoński et al. (1969), or a modified version of it by Levshin et al. (1992). This traditional method measures the dispersion curve for each source-station pair at one time and requires analysts to accurately identify the surface wave signal and decide the appropriate frequency range over which reliable measurements can be made (Ritzwoller & Levshin, 1998).

For phase velocity, most of these dispersion measurements are made by comparing synthetic and recorded seismograms either in the frequency domain (e.g. Laske & Masters, 1996; Montagner & Tanimoto, 1990) or in the time domain (e.g. Trampert & Woodhouse, 1995; Ekström et al., 1997). Because these methods examine one record at a time, automatic measurement techniques are needed and various objective choices of quality control criteria have to be included.

The first goal of this thesis is to develop an efficient technique to process the ever-increasing volume of seismic data. This technique is not fully automatic but requires minimal operator intervention and allows users greater control over

the quality of measurements.

The second goal of this thesis is to present a consistent picture of phase and group velocity of the globe. These phase/group velocity maps are useful in many ways. They provide the basic input data for the LITHO1.0 model which serves as an updated version of the CRUST2.0 model that has found a wide range of use in both the seismological and broader geophysical community (Pasyanos et al., 2014).

These maps are also useful by themselves. Phase velocity maps have been used to improve the accuracy of the Centroid Moment Tensor (CMT) solution (Hjörleifsdóttir & Ekström, 2010; Ekström et al., 2012) which is the standard reference for earthquake mechanisms. By incorporating existing surface wave phase/group velocity maps, the ability to locate glacier earthquakes (Ekström et al., 2003) and catastrophic land slides (Ekström & Stark, 2013) has also been demonstrated. This is normally below the detection level using the more traditional methods designed for body waves. The improved phase/group velocity maps with higher resolution produced in this thesis are expected to better facilitate these studies.

Seismic anisotropy has long been observed for both body waves and surface waves (e.g. Anderson, 1961; Aki & Kaminuma, 1963; Hess, 1964; Forsyth, 1975). In particular, surface wave data have been widely used to study global azimuthal anisotropy patterns of the Earth’s upper mantle (e.g. Tanimoto & Anderson, 1985; Laske & Masters, 1998; Trampert & Woodhouse, 2003; Debayle et al., 2005; Visser et al., 2008; Ekström, 2011).

However, the agreement among these different surface-wave derived models is not satisfactory. Various factors can affect the derived anisotropic structure, e.g. the data coverage, the wave propagation theory used (Sieminski et al., 2007), the details of the inversion techniques (Simons et al., 2002), etc. The third goal of this thesis is to demonstrate that uncertainties of earthquake locations, which have been generally ignored in most previous studies, also have significant impacts on the resulting azimuthal anisotropy.

Compared to the numerous models for the elastic properties of the Earth in the past decades, only a handful of studies attempt to resolve attenuation structure.

An improved understanding of the lithospheric attenuation is important in two aspects. The first is to better constrain the physical and chemical state of Earth’s interior by providing complementary information to elastic velocity (e.g. Karato, 2003; Faul & Jackson, 2005; Dalton et al., 2009; Zhu et al., 2013; Abers et al., 2014). The second is to improve our knowledge of the elastic properties themselves by providing better constraints on their small scale variations (Laske & Masters, 1996) and by taking into account the effect of physical dispersion (Ruan & Zhou, 2010).

The fundamental difficulty in inferring attenuation is that surface wave amplitudes are impacted by many things and intrinsic attenuation is a relatively minor contributor (eqn 1.12). The fourth goal of this thesis is to investigate the feasibility of applying a new theory to handle the focusing-defocusing terms. A set of 2D attenuation maps for Rayleigh waves between 5 and 25 mHz are also presented.

A byproduct of this thesis is to provide the seismic observations needed to constrain the evolution of the oceanic lithosphere. The overall trend of sea floor topography and heat flow measurements can be explained quite well by the cooling of the oceanic lithosphere after it forms at the ridges (e.g. Parker & Oldenburg, 1973; Parsons & Sclater, 1977; Stein & Stein, 1992). This simple and elegant explanation is included in many textbooks (e.g. Turcotte & Schubert, 2002) and demonstrates the success of plate tectonic theory. This plate cooling model has received some modifications recently by reconsidering some of its details (e.g. Doin & Fleitout, 1996; McKenzie et al., 2005; Adam & Vidal, 2010) but its general success in explaining topography and heat flow data has never been questioned.

Such an age progression trend is also observed seismologically (e.g. Nishimura & Forsyth, 1989; Ritzwoller et al., 2004; Maggi et al., 2006; Goes et al., 2013; Burgos et al., 2014). However, explaining these seismic observations based on the simple plate cooling model is still not completely satisfactory. This is largely due to the need of applying an inelastic correction because of intrinsic attenuation (e.g. Stixrude & Lithgow-Bertelloni, 2005) and the attenuation value is not well understood. This thesis presents a self-consistent model including both the elastic and

inelastic observations, which can be used as a starting point for future modeling efforts.

Lastly, an improved knowledge of lithosphere structure is of great importance for monitoring the Comprehensive Nuclear-Test-Ban Treaty (CTBT). It provides the essential velocity model to identify and locate possible nuclear test events and a reasonable attenuation model is essential to estimate the yields of the tested weapons.

1.4 Thesis organization

Chapter 2, in full, is a reformatted version of a publication in Bulletin of the Seismological Society of America: Ma, Z. and G. Masters (2014), A new global Rayleigh and Love wave group velocity dataset for constraining lithospheric properties, Bull. Seism. Soc. Am. 104 (4), 2007-2026. DOI: 10.1785/0120130320. I am the primary investigator and author of the paper, which describes a new method to measure Rayleigh and Love wave group velocity globally.

Chapter 3, in full, is a reformatted version of a publication in Geophysical Journal International: Ma, Z., G. Masters, G. Laske, and M.E. Pasyanos (2014), A comprehensive dispersion model of surface wave phase and group velocity for the globe, Geophys. J. Int. 199 (1), 113-135. DOI: 10.1093/gji/ggu246. I am the primary investigator and author of the paper, which extends the method described in chapter 2 to measure phase velocity. A self-consistent dispersion model for both Love and Rayleigh wave group and phase velocity is presented.

Chapter 4, in full, has been submitted to Geophysical Journal International: Ma, Z. and G. Masters, Effect of earthquake locations on Rayleigh wave azimuthal anisotropy. I am the primary investigator and author of the paper, which points out the importance of earthquake locations in inverting Rayleigh wave azimuthal anisotropy.

Chapter 5, in full, has been submitted to Geophysical Journal International: Ma, Z. and G. Masters, 2D global Rayleigh wave attenuation model by accounting for finite-frequency focusing and defocusing effect. I am the primary investigator

and author of the paper, which utilizes the finite-frequency theory in correcting the focusing-defocusing effect and image the global lithosphere attenuation structure.

Chapter 6 gives a summary of this thesis. Prospects for future study are also discussed.

Appendix A includes various research notes compiled during my Ph.D. study. These notes may be useful for future students who are interested in long period global seismology and provide them useful manuals and clarifications on topics that are not easy to find in the standard literature.

The measurements and products from this thesis can be freely downloaded from <http://igppweb.ucsd.edu/~gabi/>. Codes related to this thesis can be requested by sending email to z1ma@ucsd.edu.

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Chapter 2

A New Global Rayleigh and Love Wave Group Velocity Dataset For Constraining Lithosphere Properties

Abstract

We present a new and efficient method to measure Rayleigh and Love wave group velocity over a broad frequency range. This technique starts in a similar fashion to the traditional frequency-time analysis, but instead of making measurements for all frequencies for a single source-station pair, we apply cluster analysis to make measurements for all recordings from a single event at a single target frequency. We also develop an inversion method with laterally varying smoothnesses to generate 2D group velocity maps with uniform errors. These maps match large scale geological features and fit our data very well. This dataset will be used to constrain lithospheric structure globally.

2.1 Introduction

Among various methods to study lithospheric structure, surface waves provide one of the most valuable constraints (e.g. Trampert & Woodhouse, 1995; Ekström et al., 1997; Shapiro & Ritzwoller, 2002; Nettles & Dziewoński, 2008; Ritsema et al., 2011; Ekström, 2011). In contrast to body waves, surface waves at different frequencies are sensitive to structures at different depths, leading to good resolving power in the radial direction (Nettles & Dziewoński, 2008). Surface waves also provide almost complete global coverage. Recently, studies of group velocity have increased dramatically, especially in areas where few earthquakes and seismic stations are available, e.g. the Arctic (Levshin et al., 2001), Antarctica (Ritzwoller et al., 2001), South America (Feng et al., 2004), Africa and Arabia (Pasyanos & Nyblade, 2007).

However, in these many studies, the measurement algorithms are based on the one first introduced by Dziewonski et al. (1969), or a modified version of it by Levshin et al. (1992). This traditional method measures the dispersion curve for each source-station pair at one time and requires analysts to accurately identify the surface wave signal and decide the appropriate frequency range over which reliable measurements can be made (Ritzwoller & Levshin, 1998). Although the global datasets used in Ritzwoller et al. (2002) already consist of more than 100,000 paths, the nature of such a technique makes it very time consuming to process the ever increasing quantities of global seismic data.

This chapter describes a different technique, which measures relative group arrival times of hundreds of envelope functions filtered to the same frequency at one time and allows us to process large datasets efficiently. This technique is not fully automatic but requires minimal operator intervention, which allows users greater control over the quality of measurements. We perform synthetic tests and compare our new dataset to other existing global datasets and find no bias. We also perform a ray-theory based inversion with laterally variable smoothness to test the internal consistency of our dataset. The data are surprisingly well fit by this simple inversion.

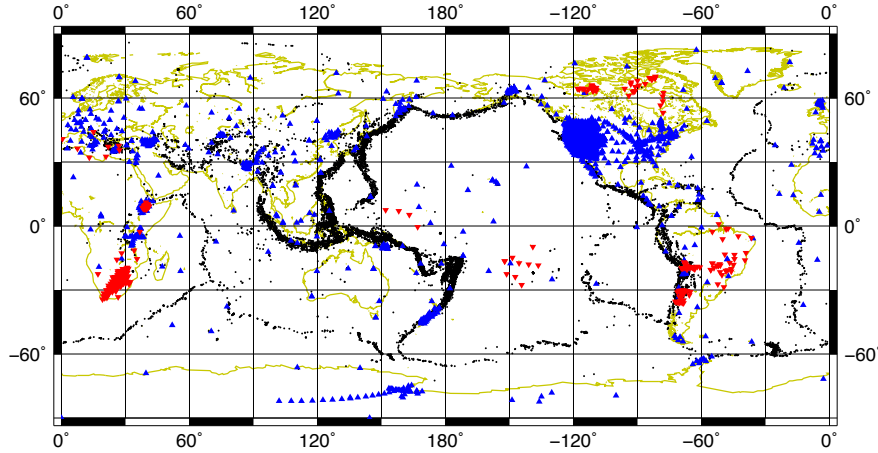


Figure 2.1: Distribution of earthquakes and stations analyzed in this chapter. Dots represent earthquakes with M_s or $m_b \geq 5.5$ from 1976 to 2007. Only earthquakes shallower than 200 km are considered. Triangles are stations with long period data available from IRIS. Inverse triangles are stations with broad band data from PASSCAL experiments and the POLARIS network, which improve data coverages in the southern hemisphere and the Arctic region.

2.2 Method

Our goal is to generate a large dataset of Rayleigh and Love wave group velocity at intermediate frequency range to constrain crust and uppermost mantle properties. Waveform data are obtained from IRIS and include all available long period data from permanent stations and temporary deployments, for all events with M_s or $m_b \geq 5.5$ from 1976 to 2007. We also include some broad band data from PASSCAL experiments in the southern hemisphere and the POLARIS network in northern Canada to improve data coverage (Figure 2.1). Earthquakes deeper than 200 km are not used in this chapter to reduce overtone contamination. Rayleigh wave measurements are made on the vertical components of the seismograms while Love wave are made on the transverse components.

2.2.1 Preprocessing

Our method starts in a similar fashion to the traditional frequency-time analysis (Levshin et al., 1992; Dziewonski et al., 1969). After correcting for instru-

ment response and for the source phase calculated according to the CMT catalog, we apply a narrow band Gaussian filter to the data and compute envelope functions. The difference is that we choose a single frequency band to band pass data from all stations for one event. We then apply cluster analysis techniques to measure the relative arrival time for all these envelope functions, the same way as Houser et al. (2008) measure body wave travel times.

The Gaussian bandpass filter we use is

$$H(\omega) = \exp[-\alpha(\frac{\omega - \omega_c}{\omega_c})^2] \quad (2.1)$$

where ω_c is the center frequency of the filter. We use $\alpha = 3./0.25^2$, so the band width for 20 mHz is about 5 mHz. Shapiro & Singh (1999) point out a systematic error in measuring group arrival time due to the discrepancy between the centroid frequency and target frequency. We mitigate this effect by iteratively varying the center frequency of the filter until the centroid frequency of the data spectrum matches the target frequency within an accuracy of 0.01 mHz.

We calculate the signal-to-noise ratio of each envelope function and we only keep traces with signal-to-noise ratio larger than 4.0. The peak of each envelope function is also identified and the arrival time of the peak with respect to the predicted arrival time from ak135 (Kennett et al., 1995) or PREM (Dziewonski & Anderson, 1981) is stored as t_{pi} for each trace.

2.2.2 Manual measurement

Cluster analysis technique

We manually make measurements for data filtered at 10, 20, 30 and 40 mHz for Rayleigh waves, and 10, 20 and 30 mHz for Love waves. We consider stations with epicentral distances ranging from 20 to 160 degrees. Traces are sorted by distance and plotted on the screen. An example is shown in Figure 2.2. This example is for Rayleigh waves at 20 mHz. Each trace represents the envelope function for one station for this event and we have 329 envelope functions for this example.

Figure 2.2: Example of all envelope functions from one single event (Jan 4th 2005). This example is for 20 mHz Rayleigh wave and there are 329 envelope functions. Instrument responses and source phases are corrected. Each trace is the envelope function for one station. Traces are ordered by epicentral distance and aligned on the predicted group arrival time from model ak135, which is marked by the center solid line. Users then use mouse to choose a window for cross-correlation.

Figure 2.3: Example of clusters and cluster tree diagram from the same event. The left hand side shows the envelopes which have been organized into clusters according to waveform similarity. The right hand side shows the cluster tree which governs the clustering. Users can click on the cluster tree to determine the degree of clustering. It also allows users to identify noisy and distorted records. The cross hair (only the vertical line matters) on the right indicates an appropriate level of clustering for this example.

Users then choose a time window for cross-correlation to align all traces. For this example, a window between -181 s and 351 s is used to include all clearly visible envelopes. Using the time domain method described in VanDecar & Crosson (1990), cross-correlation functions for every trace with every other trace are computed and the positive peaks in the cross-correlation function that contain the differential time and scaling information between traces are identified. According to the similarities between traces, the clustering algorithm described in Hartigan (1975) is used to generate clusters for pairs with correlation coefficients higher than 0.95.

An example of this clustering technique is shown in Figure 2.3. The left hand side shows the waveforms which have been organized into clusters according to waveform similarity. The right hand side shows the cluster tree which governs the clustering. This tree diagram allows the user to decide how finely the clustering needs to be done. A vertical cursor is used to do this. As the cursor is moved to the left of the tree diagram, the waveforms are divided into smaller and smaller clusters. For this example, the envelopes at the bottom are clearly distorted by noise while the envelopes at the center of the screen are clean and can be grouped together to form 3-4 big clusters. The cross hair (only the vertical line matters) on the right indicates an appropriate level of clustering for this example.

This method is not fully automatic, but it requires minimal amount of human interaction, allowing users to have greater quality control of the measurements. This method also allows users to isolate contaminated waveforms without pre-screening the data, which is essential when dealing with very large datasets. It is important to note that we keep only those envelopes with simple bell shapes. This may explain why simple ray-theory based inversions of the data are very successful at explaining the measurements (see below).

After users decide how finely the clustering needs to be done, optimal time shifts t_{ri} (with zero median) and errors σ_{1i} for traces within one cluster are then calculated again using the cross-correlation method and the aligned traces within each cluster are plotted for checking. Figure 2.4 plots every third trace in Figure 2.2 for clarity and shows all accepted traces with solid lines. We can see that the

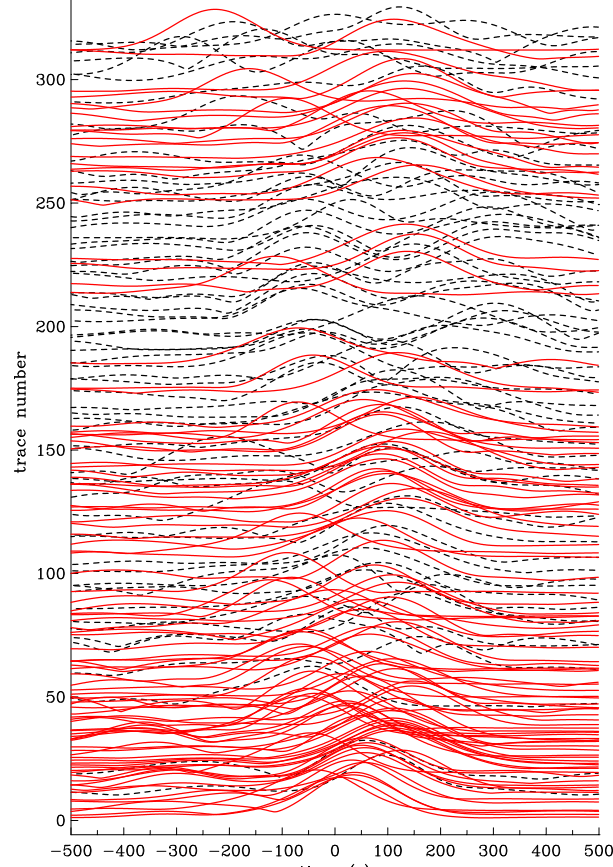


Figure 2.4: Re-plotting every third traces in Figure 2.2 for clarity. Dash lines represent all traces and the solid lines are the final accepted traces. Note we have successfully identified most of the envelope functions with simple bell shapes. The time window used for cross-correlation for this cluster is -181 and 351 s, which include all envelopes with simple shapes.

clustering technique can indeed identify most of the envelope functions with simple bell shapes and discard distorted ones.

The errors σ_{1i} estimated from cross-correlation methods are usually too optimistic. More reliable measurement errors for our dataset can be inferred by looking at nearby stations. We find all station pairs with inter-station distances less than d_m . The difference in group arrival times for each station pair is then calculated, and the standard deviations of these differences are plotted against d_m . (Figure 2.5) Station pairs within 5-10 km are mostly from PASSCAL experiments and may have poor quality. We find that a minimum error of 4 s is reasonable.

Therefore, we set the error of our measurement to be

$$\tilde{\sigma}_{1i} = \max(4, \sigma_{1i}) \quad (2.2)$$

Errors estimated from other frequencies can be slightly different (e.g. about 7.5 s for 10 mHz and 2.5 s for 30 mHz). However, we feel that assigning a minimum error of 4 s is sufficient for our purpose.

Converting relative time to absolute time

The times t_{ri} we measure in the last section are the relative arrival times among traces within the same cluster. On the one hand, although these relative arrival times can be directly used for tomography, it is beneficial to convert these relative times to absolute times. (Houser et al., 2008) On the other hand,

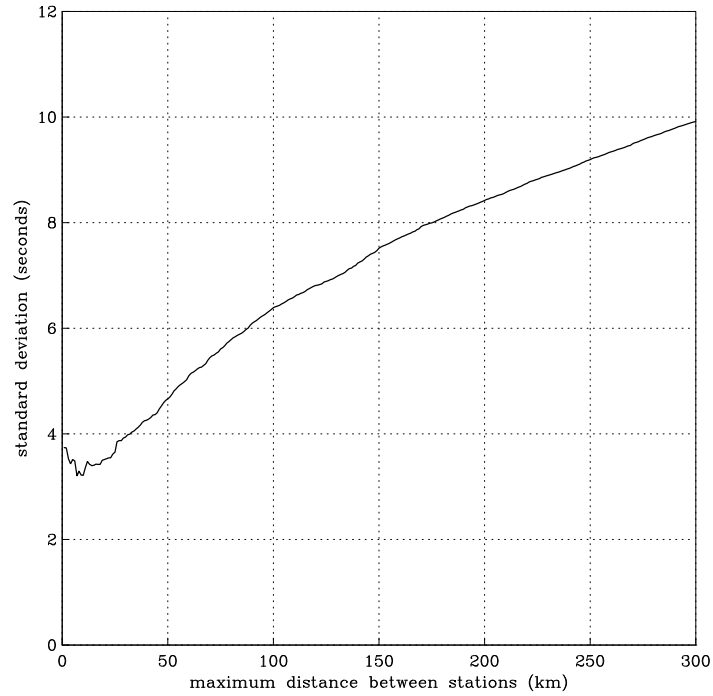


Figure 2.5: Standard deviations of the differences in arrival time measurements between station pairs as a function of the maximum distance between stations. This is for Rayleigh wave at 20 mHz.

the peak arrival times t_{pi} are the absolute group arrival times in the absence of noise. These times actually correspond to the amplitude ridges in the traditional FTAN-diagrams (Levshin et al., 1992), from which group velocities are estimated. However, peak arrival times are more likely to be affected by noise than relative arrival times that are inferred from cross-correlation methods, especially when the envelopes are broad in the case of low frequencies.

To make use of the advantages of both relative and peak arrival times, we introduce the cluster time t_m , which shifts the whole cluster by a constant amount of time to match the relative arrival times to peak arrival times. We find the best t_m that satisfies

$$t_{1Di} + t_{ri} + t_m = t_{1Di} + t_{pi}, \text{ for } i = 1, 2, 3, \dots n \quad (2.3)$$

where n is the number of records in each cluster, t_{1Di} is the theoretical arrival times for a 1D model. Because the 1D arrival times from both sides cancel each other out, a robust estimate of the cluster time and its error is simply

$$t_m = \text{median}(t_{pi} - t_{ri}) \quad (2.4)$$

$$\sigma_2 = \frac{\text{median}|t_{pi} - t_{ri} - t_m|}{0.6745\sqrt{n}} \quad (2.5)$$

The absolute arrival time t_{ai} and its error are then

$$t_{ai} = t_{ri} + t_m \quad (2.6)$$

$$\sigma_i = \sqrt{\tilde{\sigma}_{1i}^2 + \sigma_2^2} \quad (2.7)$$

Because of the sparsity of seismic stations in early years (earlier than around 1995), we normally do not have enough traces at 40 mHz from one single event to apply cluster analysis techniques. In this case, we find it convenient to include multiple events and process at least 200 traces at one time. The shape of the envelopes are still similar enough for cluster analysis. (Figure 2.6) (Note that we do not separate traces from a single event.)

Our technique is very efficient at generating large datasets of Rayleigh and Love wave group velocity measurements. For Rayleigh waves, the current dataset

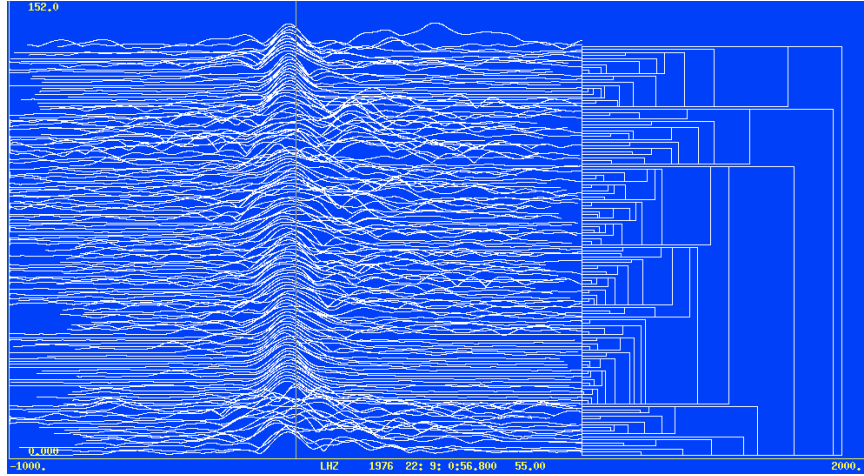


Figure 2.6: Example of clusters for 40 mHz in the early years. This example includes events from Jan 1 to Jan 22 in 1976, 23 events in total.

has over 300,000 measurements for 10 and 20 mHz, 200,000 measurements for 30 mHz and 110,000 for 40 mHz. The Love wave dataset is about half the size of the Rayleigh wave dataset, having about 170,000 measurements for 10 and 20 mHz, and 70,000 for 30 mHz. Because we can process hundreds of stations at one time, making group velocity measurements at a single frequency band for the whole dataset only takes 3-5 days.

At the high frequency end, the number of successful measurements decreases rapidly with frequency because of increased structural heterogeneities. While we can still obtain measurements at frequencies higher than 40 mHz in regions with dense seismic networks (e.g. Eurasia), parts of South America and Africa are not sampled at all. To avoid instability in the inversion for the unsampled regions, we choose to stop at 40 mHz for Rayleigh waves and 30 mHz for Love waves. At frequencies lower than 7.5 mHz, the envelopes are very broad. Reliable cross-correlation measurements are hard to make and the arrival times of the peaks are very likely distorted. To make use of the data at long periods, it is more desirable to modify our method to measure phase velocity instead, because cycle-skipping is less likely to happen at such long periods. We will discuss this modification in a future contribution.

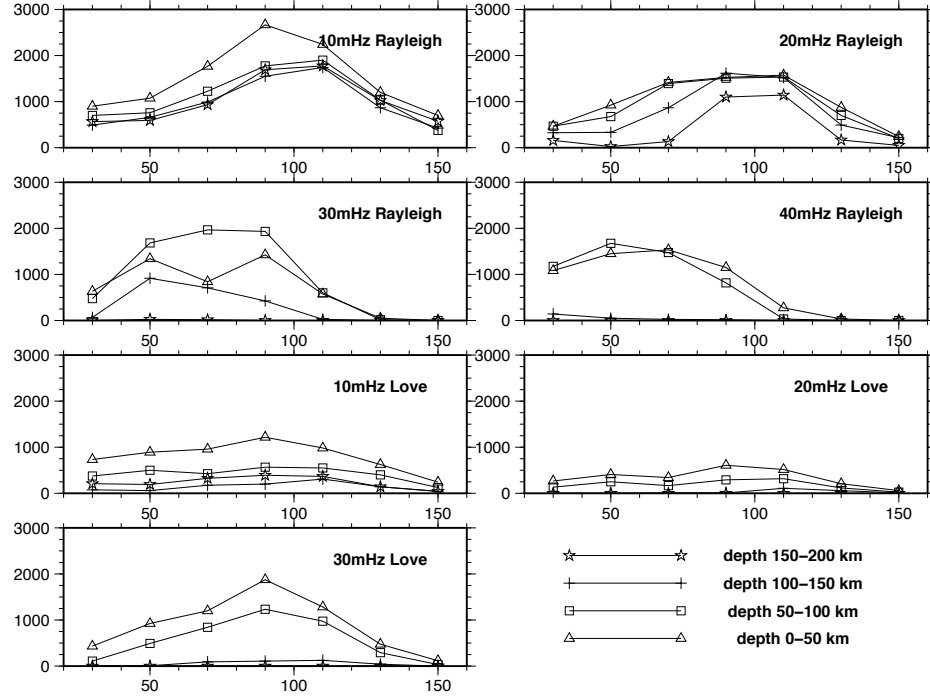


Figure 2.7: Number of measurements for all synthetic data, as function of distance and frequency. Notice that the number of measurements that can be made from deep earthquakes decreases rapidly as frequency increases.

2.2.3 Automatic measurement of intermediate frequencies

For intermediate frequencies, we can use the same clusters as determined from the measurements done at a nearby frequency. For example, measurements at 17.5 mHz are done automatically using the clusters determined from the 20 mHz measurements. The time window for cross-correlation is estimated manually by measuring a few months of data. Other steps are the same as described above and are processed fully automatically.

2.2.4 Synthetic test

We use synthetic seismograms calculated from ak135 model (Kennett et al., 1995) to test our method. We compute four sets of synthetic seismograms for earthquakes with depth range 0-50 km, 50-100 km, 100-150 km and 150-200 km. Station

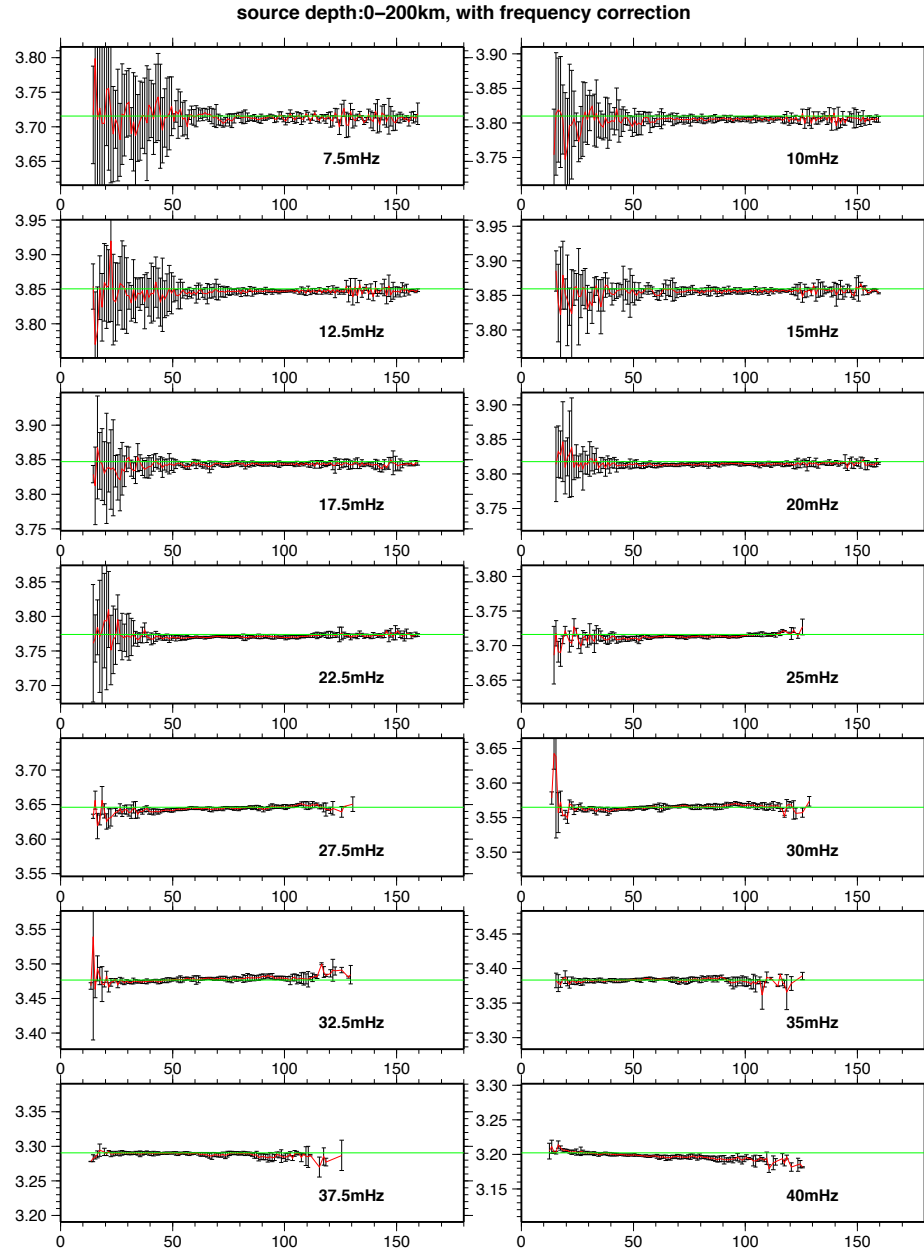


Figure 2.8: Synthetic test results of measured group velocity (km/s) as a function of distance (degree) for Rayleigh waves at different frequencies. We bin our measurements in every 1 degree bin. Error bars indicate the variance of measured group velocity for each bin. The thin horizontal line is the theoretical value for group velocity at each frequency. Variances in group velocity are rather small, except at short distances for low frequency measurements because of overtone interferences. Measured group velocity is slightly smaller than theoretical value at 40 mHz at long distance, though the bias is only 0.0044 km/s.

distributions are the same as for the real data. Each synthetic dataset consists of 15000 traces for Rayleigh waves and 10000 for Love waves. Measurements can be made from earthquakes shallower than 100 km for all frequencies. Large portions of the synthetics from deep earthquakes are distorted by overtones and are discarded during the measurement procedure. Note that the number of measurements from deep earthquakes decreases rapidly as frequency increases. (Figure 2.7)

Figure 2.8 and Table 2.1 show the comparison between the measured and theoretical values for Rayleigh waves. Variance in group velocity measurement is typically about 0.003-0.005 km/s. The largest variance happens at short distance for low frequency measurements, which is caused by overtone interference. The theoretical group velocity value at 32.5 mHz is 0.0015 km/s slower than our measured value but is 0.0044 km/s faster than our measured values at 10 and 40 mHz. However, both the variance and bias do not exceed 0.15% of the theoretical values, which is much smaller than the actual signal in our tomographic results, which can exceed 10% for most frequencies.

Similar to Nettles & Dziewoński (2011), we also find that overtone contamination has a larger effect on Love waves than Rayleigh waves (Figure 2.9), because of the smaller difference between the group velocities of fundamental and overtone branches. The effect of interference decreases with increasing epicentral distance and frequency. At 10 mHz where interference is the strongest, the scaled median absolute deviation (SMAD) of the errors caused by overtone interferences is about 0.035 km/s, which is smaller than the error (1.5% for Love waves, see below) we allow in our inversion. Applying the same technique to synthetic seismograms with fundamental modes only do not generate such oscillatory behavior, confirming that overtone interference is the cause. We further compare the actual signal from real data (the first 6000 measurements during the same time period) and find that the actual signal can be 2-3 times larger than the interference effect. (Figure 2.10) The apparent bias of 0.01 km/s is due to the fact that this dataset has more measurements around 90 degrees than other distances, where overtone interference tends to result in a higher than average velocity. While global tomography, which combines paths with different path lengths, is expected to further suppress such

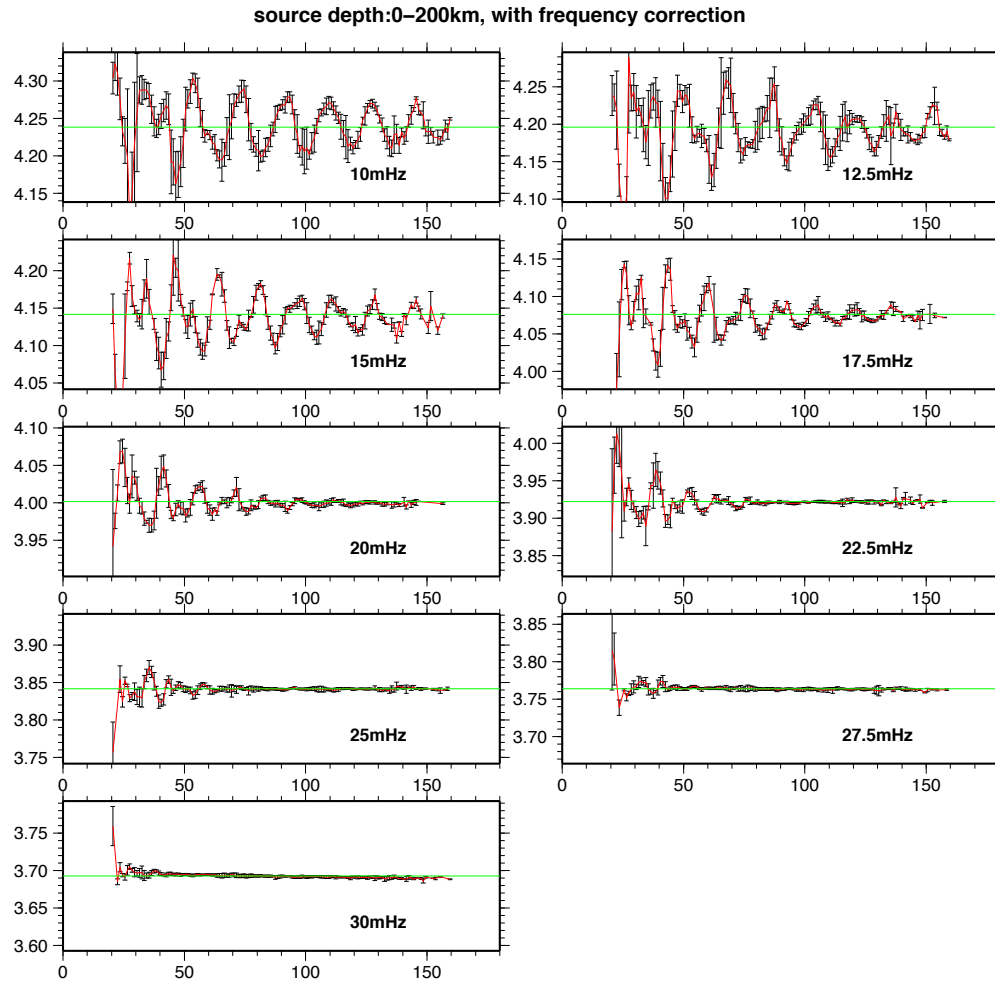


Figure 2.9: Same as Figure 2.8, but for Love waves.

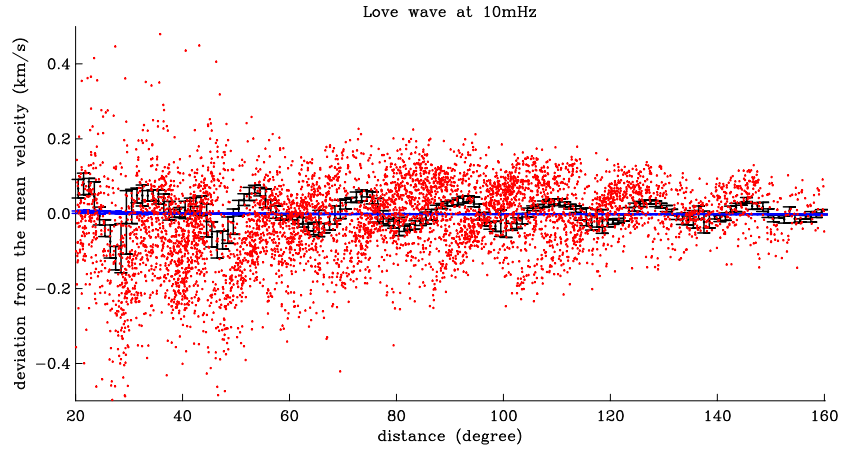


Figure 2.10: Comparison of signal level from measurements on real data and the errors caused by overtone contaminations for Love waves at 10 mHz. Error bars are the same as the ones in Figure 2.9 for 10 mHz. The horizontal line is for measurements from the same dataset, but the synthetics are calculated from fundamental modes only. Errors are smaller than the width of the line so are not plotted. Dots are the first 6000 measurements of real data during the same time period.

Table 2.1: Synthetic test for Rayleigh and Love waves with source depth 0-200 km. freq*: frequency. diff[†]: difference between the median of the measured group velocity and the theoretical value. SMAD[‡]: scaled median absolute deviation of the measured group velocity.

	Rayleigh		Love	
freq (mHz)*	diff (km/s) [†]	SMAD (km/s) [‡]	diff (km/s)	SMAD (km/s)
7.5	-0.0024	0.0092		
10.0	-0.0044	0.0057	0.0101	0.0349
12.5	-0.0036	0.0057	-0.0018	0.0399
15.0	-0.0033	0.0050	-0.0054	0.0293
17.5	-0.0036	0.0036	-0.0049	0.0162
20.0	-0.0038	0.0029	-0.0019	0.0066
22.5	-0.0032	0.0029	-0.0007	0.0031
25.0	-0.0034	0.0029	-0.0003	0.0026
27.5	-0.0021	0.0033	-0.0003	0.0021
30.0	0.0002	0.0039	-0.0004	0.0024
32.5	0.0015	0.0041		
35.0	0.0000	0.0031		
37.5	-0.0016	0.0028		
40.0	-0.0044	0.0038		

interference effects (Nettles & DzioŹowski, 2011), more sophisticated treatments are needed for regional studies with limited ranges of path lengths (e.g. Gaherty et al., 1996).

2.2.5 Dataset comparison

We also compare our new dataset with the one from the Colorado group (Ritzwoller and Levshin, personal communication - henceforth cub), which is measured using the traditional technique. Table 2.2 gives a summary of the results. We focus on measurements at 20 mHz for Rayleigh waves, because both datasets have good global coverage at this frequency. The number of measurements in the cub dataset is about 85,000, which is about 1/4 of the size of our dataset and only 23,000 of them overlap. The average offset between the two datasets is about 0.5 s, which is much smaller than our measurement precision.

The scaled median absolute deviation (SMAD) of the differences between

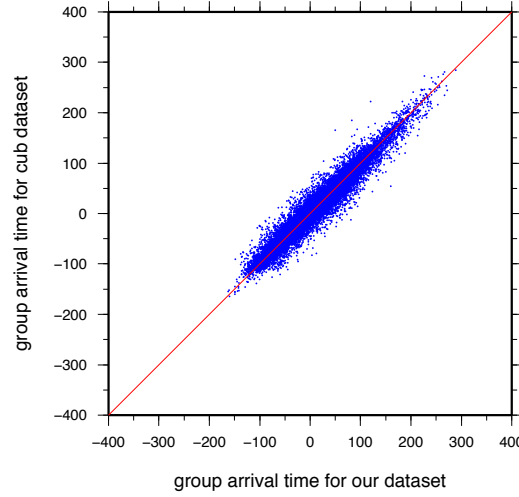


Figure 2.11: Scatter plot of group arrival times with respect to 1D predicted arrival time for our dataset v.s. the cub dataset at 20 mHz. Each dot represents one common measurement.

cub and our measurements is about 12 s. Internal uncertainties in our dataset, inferred from χ^2/N or the SMAD of residuals after inversion is about 12-16 s. The error of cub measurements can be roughly estimated from Figure 5 and 12 in Ritzwoller & Levshin (1998), and it is about 12 s ($6000 \text{ km}/3.8 \text{ km/s} - 6000 \text{ km}/(3.8+0.03) \text{ km/s} \approx 12 \text{ s}$). Therefore, the differences between the two datasets are close to the inherent uncertainties in each dataset. Such discrepancies are also much smaller than the actual travel time anomalies, which can exceed 200 s in some cases. (Figure 2.11) Therefore, we conclude that the two datasets are consistent with each other and we combine them in our inversions.

2.3 Inversion

We divide the surface of the earth into 41252 equal area blocks with the size of 1° by 1° at the equator. Sampling of the earth is quite non-uniform. Examples of ray coverage for Rayleigh waves are shown in Figure 2.12. All blocks have at least 100 hits for 20 mHz and 20 hits for 40 mHz. Coverage in most of the Eurasia

Table 2.2: Comparison between our dataset and cub dataset. freq*: frequency. mean[†]: the mean of the difference in travel time. median[‡]: the median of the difference in travel time. std[§]: the standard deviation of the difference in travel time. SMAD^{||}, the scaled median absolute deviation of the difference in travel time. common¹: the number of common measurements. ²cub, the number of measurements in the cub dataset.

type	freq (mHz)*	mean (s) [†]	median (s) [‡]	std (s) [§]	SMAD (s)	common ¹	cub ²
R	10.0	4.810	4.367	19.477	16.880	23461	53405
R	15.0	1.246	1.304	16.903	14.000	20245	68174
R	20.0	0.581	0.462	14.890	12.010	24019	84268
R	25.0	0.803	0.308	13.016	9.925	13606	91078
R	30.0	-1.014	-1.357	11.912	9.003	13046	82835
R	35.0	-1.163	-1.972	10.938	7.643	7081	75930
R	40.0	0.740	0.312	10.266	7.294	7907	83885
L	10.0	2.645	2.327	18.800	16.884	11033	22639
L	15.0	0.327	0.192	15.151	13.092	9488	40394
L	20.0	0.208	-0.118	12.496	10.019	10285	56131
L	25.0	-1.160	-0.840	13.889	9.810	11647	56568
L	30.0	0.248	0.265	10.180	6.768	5188	40400

and North America is excellent.

Our inversions are based on ray theory. Recently, finite frequency theories for surface wave delay times have been developed (e.g. Zhou et al., 2004; Dahlen & Zhou, 2006). To our knowledge, no study has applied such theories to group delay time measurements yet, probably due to the theoretical difficulty of having strong sideband sensitivity in the 2D kernels (Dahlen & Zhou, 2006). Recent global studies using phase delays based on ray theory (e.g. Ekström, 2011) are capable of generating high quality tomography results that fits the data very well. Finite frequency theory may have stronger effects on amplitudes (Ma & Masters, 2011) than arrival times, and including amplitude data in an inversion based on finite frequency theory will be presented in a later contribution.

In ray theory, the group arrival time perturbation can be written as

$$\delta t = s_{ref} \int_0^\Delta \frac{\delta s}{s_{ref}} d\gamma \quad (2.8)$$

where the integral is over the minor arc path from source to receiver, and $\delta s/s_{ref}$ is the relative perturbation of slowness of each block. The data are inverted using the LSQR method (Paige & Saunders, 1982) with a slight lateral smoothing constraint.

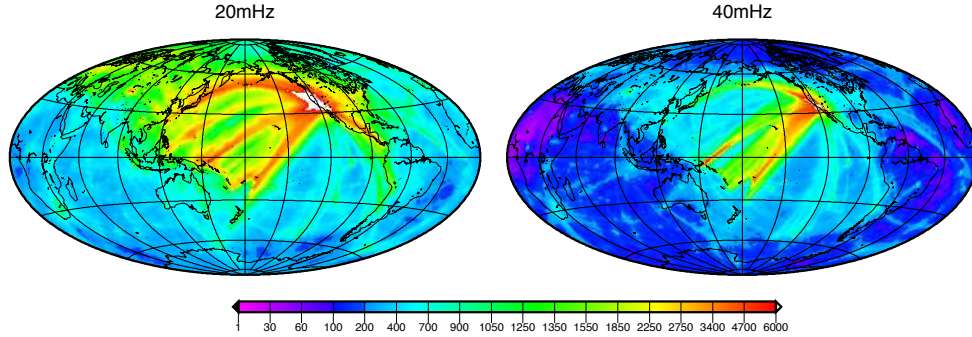


Figure 2.12: Hit count maps for 20 mHz and 40 mHz Rayleigh waves, including the cub dataset. Coverage is very non-uniform, but all cells have more than 100 hits for 20 mHz and more than 20 hits for 40 mHz.

The function we are minimizing is

$$f = (Am - d)^T (Am - d) + \lambda (Dm)^T Dm \quad (2.9)$$

where $m = \delta s/s$ is the perturbation in slowness, A is the data kernel matrix calculated assuming great circle geometry, λ is horizontal smoothness parameter and D is the numerical approximation to the 2D Laplacian. The reference slowness is estimated iteratively by doing inversions and calculating the global average slowness. The final reference slownesses can be found in Table 2.3. Weighting is according to the measurement error estimated from equation (7). Outliers with large residuals after inversion are removed. Such outliers do not exceed 3% of our data. Cutoff values for outliers are also shown in Table 2.3.

Because of the variability of ray coverage, various authors have developed different techniques to vary the smoothness parameter λ according to the hit count of each block (e.g. Pasyanos, 2005). While these methods work well in the continental regions, they are not suitable in our cases because the Pacific region is so densely sampled that these methods will result in a λ that is unreasonably small.

We find it more reasonable to vary λ according to the errors in the tomography maps, which can be estimated from a Monte Carlo error analysis technique (Houser et al., 2008). The goal is to generate models with uniform errors by assigning a different appropriate λ to each block. A flowchart of our procedure is shown

in Figure 2.13a. Firstly, we invert for tomography maps using a series of laterally uniform λ . We then add random Gaussian noise to the data, invert for a new map and repeat this process 50 times. The standard deviation of the Gaussian noise is the same as that of the data and is estimated from the χ^2/N after one inversion (e.g. for 20 mHz, it is about 4 times the error σ in equation (7)). The standard deviation of these 50 inverted maps is our estimated error. See an example for 20 mHz in Figure 2.13b.

Secondly, we choose a maximum acceptable value of model error. In this chapter, we use $e_{\max} = 1\%$ for Rayleigh waves and $e_{\max} = 1.5\%$ for Love waves at all frequencies. We choose the smallest possible λ for each block without making the error of this block exceed $e_{\max}$. We then use the resulting laterally varying λ to invert for our final tomography maps and use the Monte Carlo method again to estimate the errors associated with each block.

For better results, we need to perform a search for λ_{start} , corresponding to the most smooth tomography map we start with, and λ_{bad} , the smoothness parameter assigned to areas in the map where errors already exceeds $e_{\max}$. We try to find the best combination of λ_{start} and λ_{bad} to make the final error estimation as uniform across the map as possible. (Figure 2.13c)

A resulting map of λ is shown in Figure 2.13d, with $\lambda = 0.1$ in the western America, $\lambda = 1$ or 1.5 in most of the Eurasia and North America and $\lambda = 2.5$ in the southern Oceans. We can see that this method enables us to reach the highest possible resolution in densely sampled areas by assigning small λ to them, while maintaining stable results in poorly sampled places where a larger λ is used.

We should note that the Monte Carlo method used here is intended to test the sensitivity of our model to errors in the data. Errors from other sources (e.g. the use of simple ray theory, ignoring azimuthal anisotropy and etc) and the uniqueness of the final model are not examined.

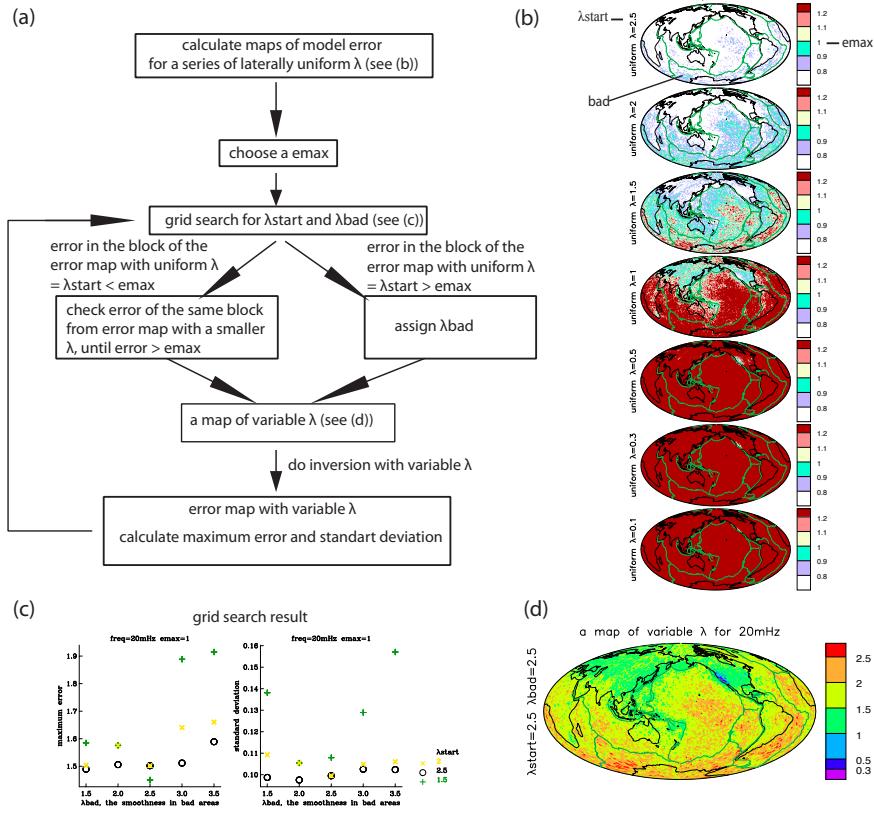


Figure 2.13: Flow chart for choosing variable smoothness and an example for 20 mHz. (a) The flow chart summarizes steps to choose variable smoothing parameter λ . (b) Error maps estimated by using a series of uniform λ . Maximum acceptable error $e_{\max}$ is taken to be 1%. In this example, we start with the error map calculated by using a uniform $\lambda_{\text{start}} = 2.5$. Areas with error larger than 1% is identified as ‘bad’ and λ_{bad} is assigned to these places. For other places, smaller λ (going downwards in the figure) can be used until error exceeds $e_{\max}$. (c) Grid search result. We try to minimize the maximum error and the variance of errors of the final model (see Figure 2.17). For this example, $\lambda_{\text{start}} = 2.5$ with $\lambda_{\text{bad}} = 1.5, 2.0, 2.5$ give acceptable and similar results. (d) A final map of variable λ . Note that the pattern is similar to that of the error maps, instead of hit count maps.

2.4 Results

Figures 2.14 and 2.15 show our Rayleigh and Love wave group velocity perturbation maps with respect to the reference velocity at each frequency. We define variance reduction (v.d.) and normalized variance reduction (n.v.d.) as

$$\text{v.d.} = 1 - \frac{\sum_i (t_i - \tilde{t}_i)^2}{\sum_i t_i^2} \quad (2.10)$$

$$\text{n.v.d.} = 1 - \frac{\sum_i (t_i/\sigma_i - \tilde{t}_i/\sigma_i)^2}{\sum_i (t_i/\sigma_i)^2} \quad (2.11)$$

where t_i is the measured group arrival time with respect to 1D reference time, $\tilde{t}_i$ is the predicted arrival time and σ_i is the error estimated for each measurement. The sum is done for all measurements. These tomographic maps are extremely good representations of our group arrival time data. Variance reductions are well over 90% for frequencies higher than 20 mHz, and decreases slightly to 60-80% for low frequencies. (Table 2.3 and Figure 2.16). This suggests that using simple ray theory is capable of modeling most of the data.

Our algorithm is capable of making a model with uniform model errors globally, except some places in the Atlantic and Indian Oceans at high frequencies (Figure 2.17). Even in these places, the maximum error does not exceed 1.5-1.7% for Rayleigh waves or 2-2.4% for Love waves. Compared to the amplitudes of the velocity perturbations in our tomography maps (Figure 2.14 and 2.15), these errors are rather small. This indicates that most of the signal in our tomography maps is reliable. Having a model with uniform errors enables us to ask what is the finest scale structure that can be resolved with an acceptable modeling error in different regions of the world. Checkerboard tests are used to examine the resolutions of our models. Input velocity perturbations are 10%. We consider the square of the checkerboard is resolved when 60% of the input amplitude is recovered, which is slightly stricter than the criterion in Ritzwoller & Levshin (1998) (50%). We find that, with a 1% modeling error globally for Rayleigh waves, the resolution for 20 mHz is 7.2 degrees globally and 5.1 degrees in Eurasia (Figure 2.18), which is consistent with a regional study of this area (Ritzwoller & Levshin, 1998). The

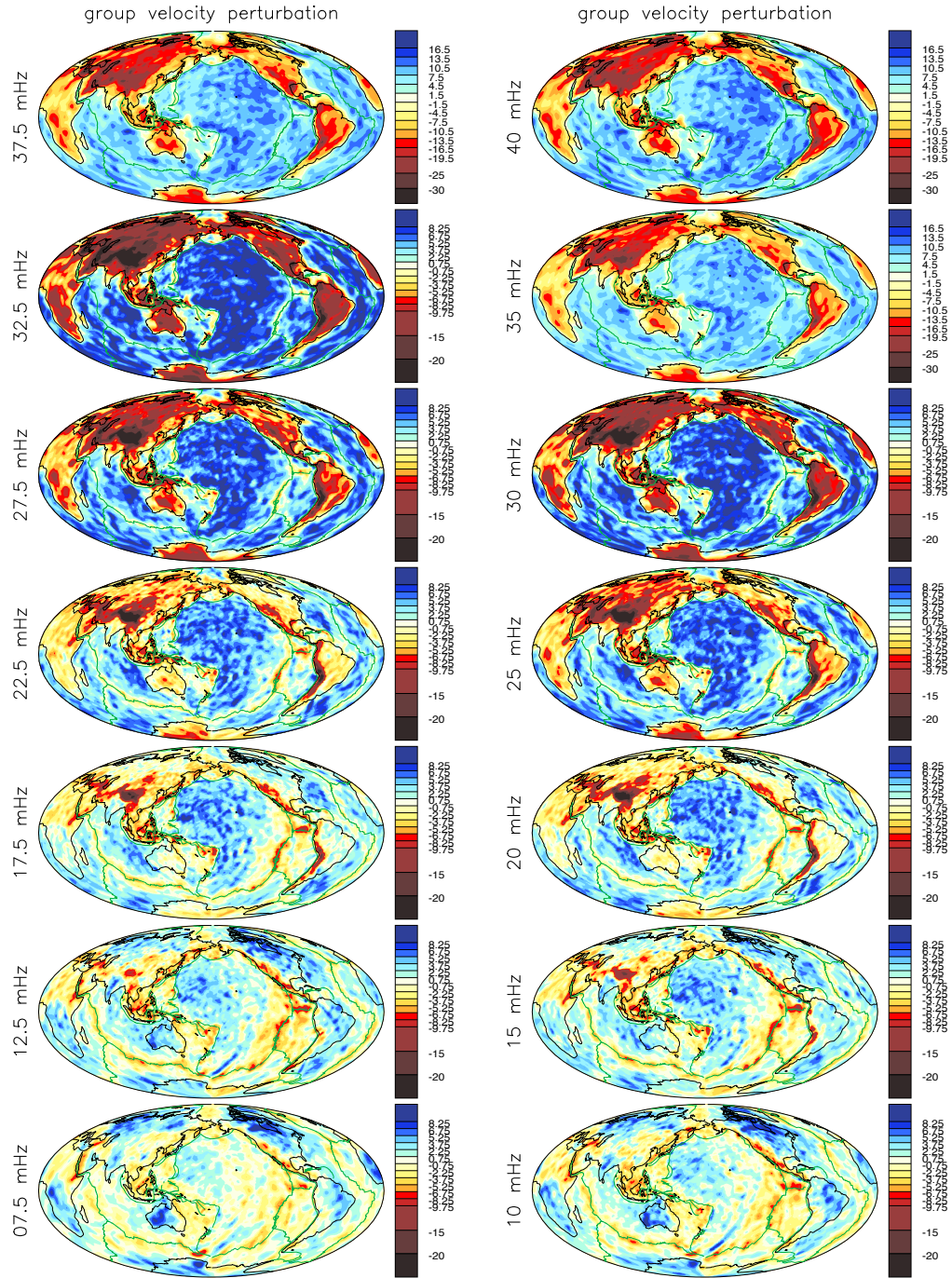


Figure 2.14: Rayleigh wave group velocity perturbations for all frequencies. Notice that the scale changes at 32.5 mHz. Reference group velocity at each frequency can be found in Table 2.3.

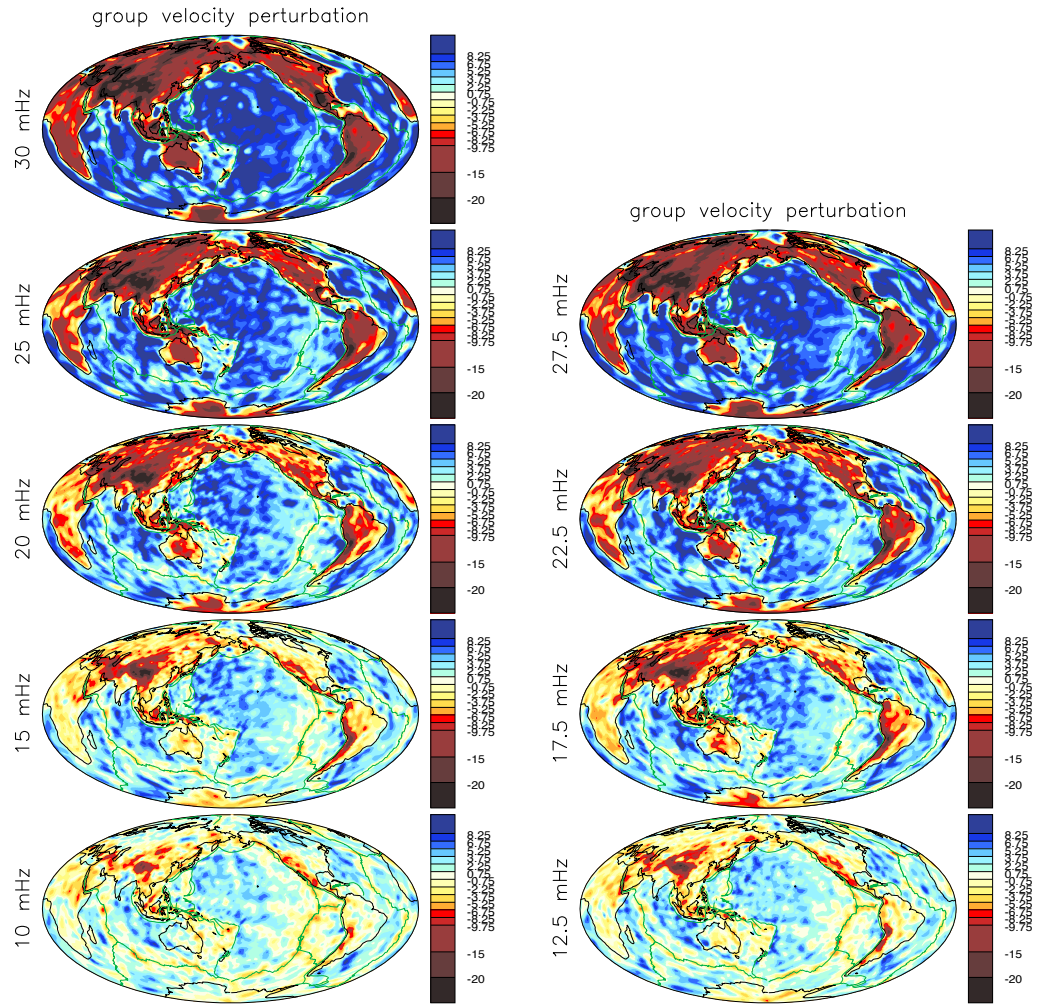


Figure 2.15: Love wave group velocity perturbations for all frequencies. Average group velocity at each frequency can be found in Table 2.3.

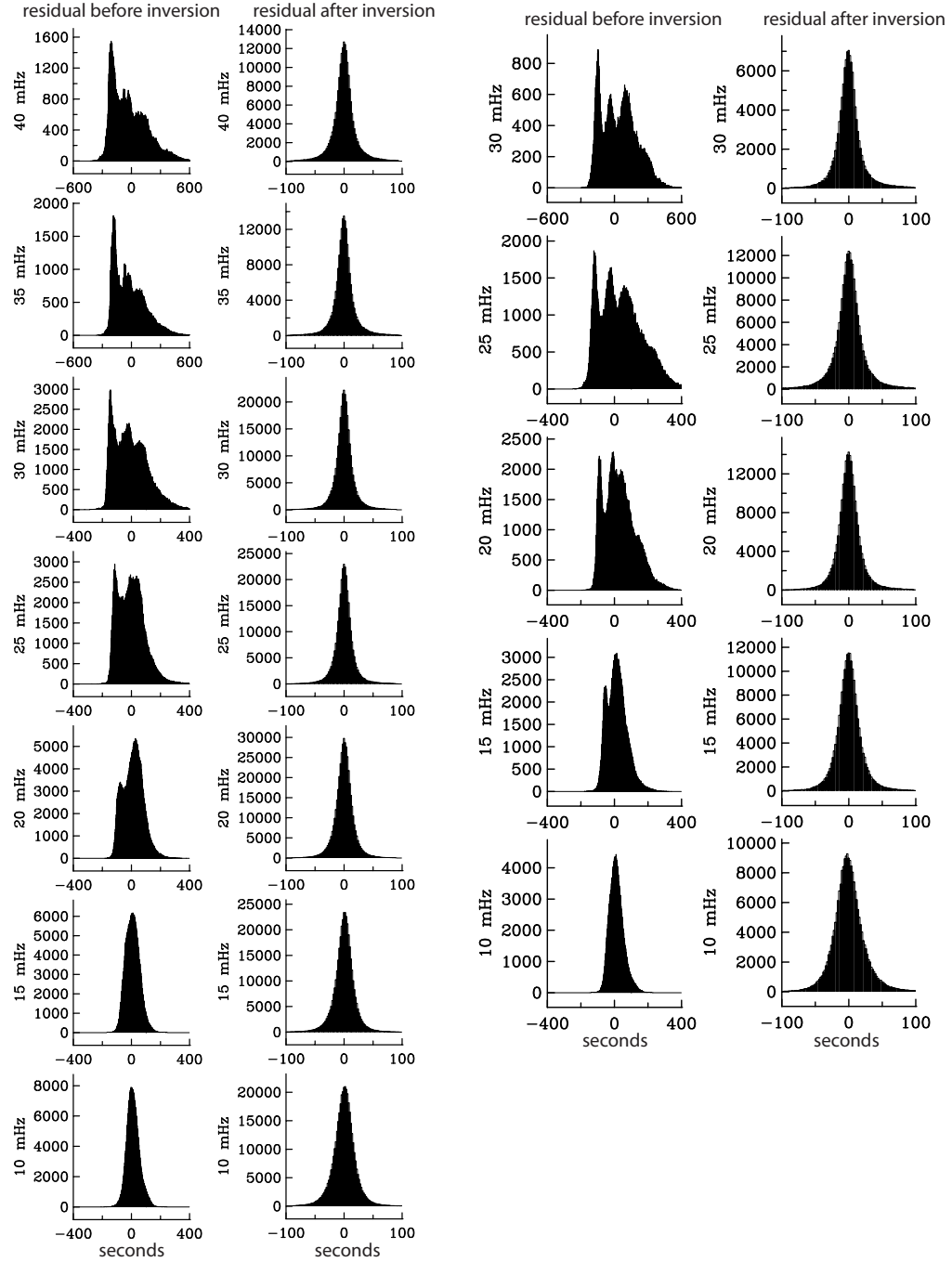


Figure 2.16: Comparison of group arrival time residuals before and after inversion. Left two panels are for Rayleigh waves and the right two panels are for Love waves. The reference group velocity can be found in Table 2.3. Notice the difference of the x scale before and after inversion. Bin width is 2 s.

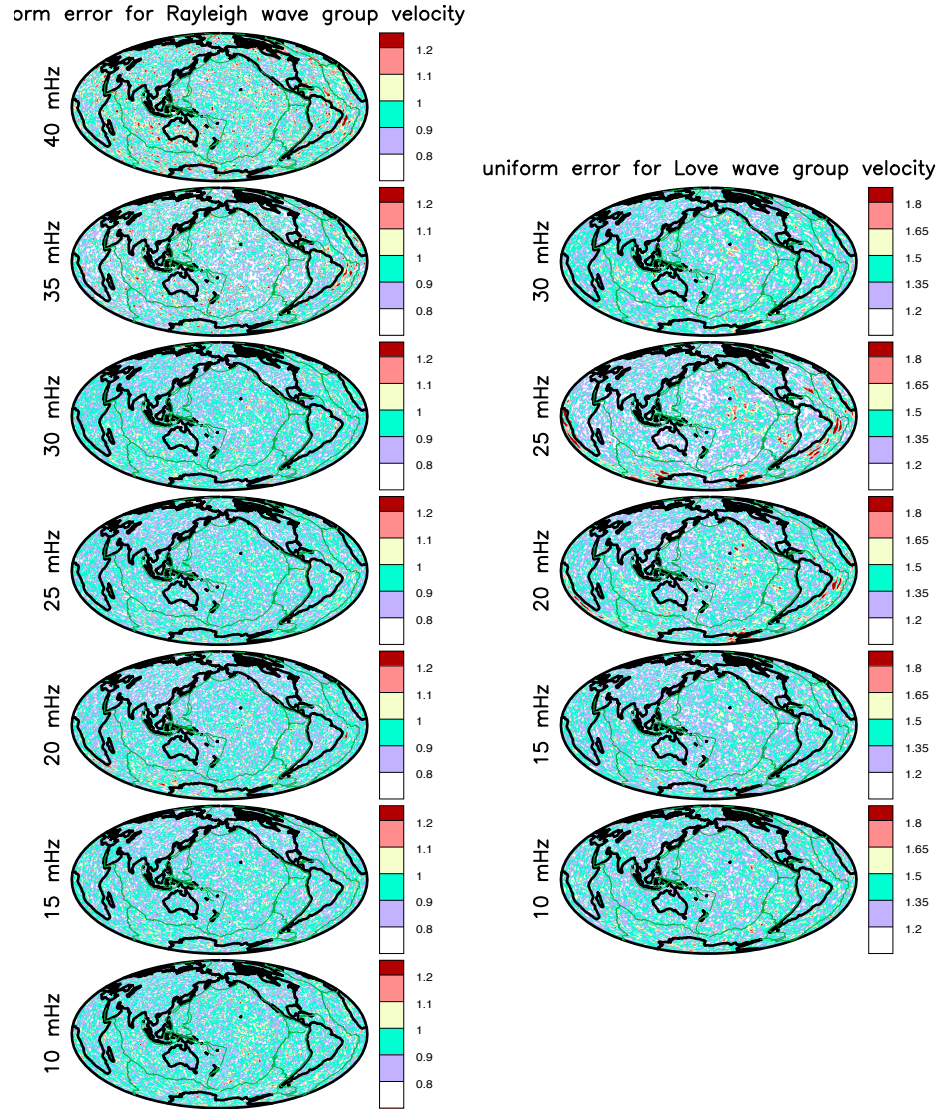


Figure 2.17: Error maps for different frequencies. Errors are nearly uniform globally, except some places in the Atlantic and Indian Oceans at high frequencies.

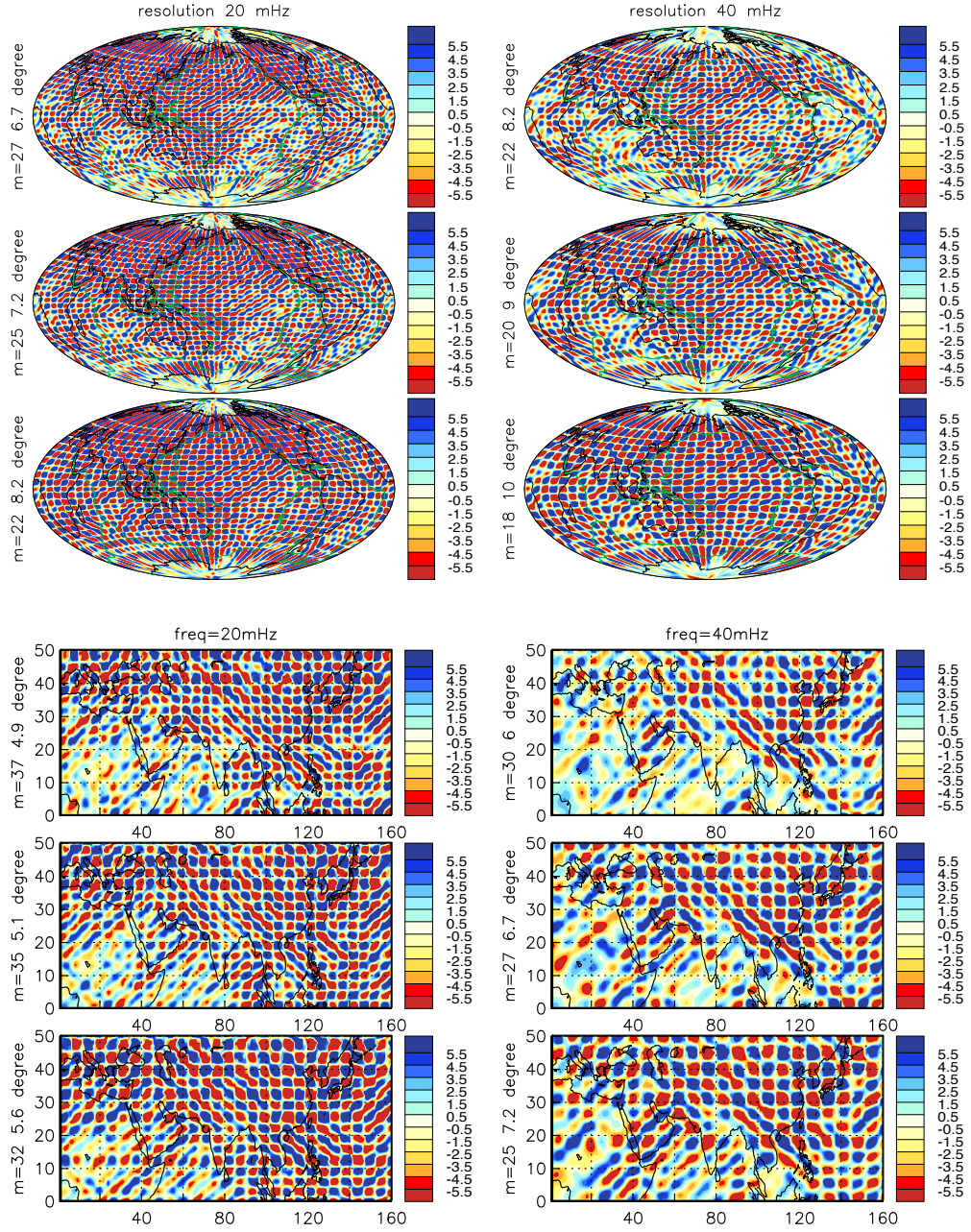


Figure 2.18: Checker board test for 20 mHz and 40 mHz Rayleigh waves, both globally (top) and in Eurasia (bottom).

resolution for higher frequency is slightly worse, though even our 40 mHz dataset can still resolve structure with the scale of 9 degrees globally and 6.7 degrees in Eurasia. (Figure 2.18) Resolutions for other frequencies can be found in Table 2.3.

2.5 Discussion

Figures 2.14 and 2.15 contain many very interesting features. Maps for Love waves are similar to the ones for Rayleigh waves with frequency 2.5-5 mHz higher. This is not surprising due to the fact that Love waves are more sensitive to shallow structure than Rayleigh waves at the same frequencies. It also indicates that these maps, which are made independently at different frequencies, are quite compatible with each other, and it is feasible to combine Love and Rayleigh wave maps together to solve for transverse anisotropy.

Maps at low frequency are more sensitive to deeper structure than maps at high frequency. Figure 2.19 shows the group velocity perturbation for 15 mHz Rayleigh waves in the continents, with boundaries of Archean and early to middle Proterozoic regions of the world (crustal types G and F in CRUST1.0 (Laske et al., 2013)). Most of the fast anomalies we see in the group velocity map coincides with these regions, indicating high velocity roots beneath these geological provinces. It also demonstrates that the resolution of our dataset is high enough to distinguish large geological structures on a global scale. We also compare this group velocity perturbation map with Figure 2 in Artemieva & Mooney (2001). Most of the cratons and shields can be easily seen in the group velocity map (e.g. Canadian Shield, West African Craton, Congo Craton, Kaapvaal Craton, Indian Shield, Siberia Craton, Russian Platform and etc). Small cratons (e.g. Yangtze Craton and Tarim Craton) can also be identified. The only craton that appears to be missing in our tomography map is the Sino-Korean Craton, suggesting that this area may have undergone active tectonics in a later time (e.g. Griffin et al., 1998).

Consider the map at 20 mHz for Rayleigh waves (Figure 2.14). Clearly, the biggest signal in continental regions is due to variations in crustal thickness. Most

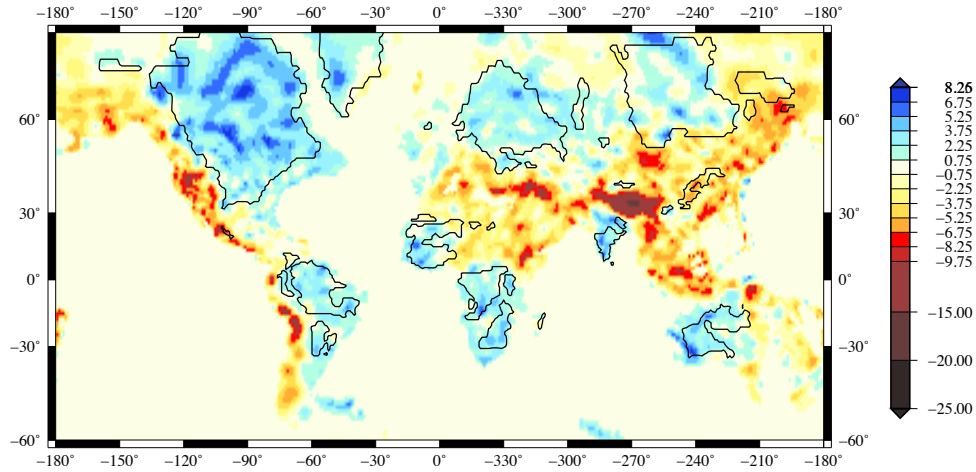


Figure 2.19: Group velocity perturbations for 15 mHz Rayleigh wave in the continents. Solid lines show the boundaries of Archean and early to middle Proterozoic regions of the world. Most of the fast anomalies in the group velocity perturbation map coincide with these regions.

extreme group velocity variations occur under Tibet ($> 25\%$) and under the Andes. This is not surprising since the sensitivity of 50 second Rayleigh waves peaks at 70-100 km so we are actually seeing a crust-mantle signal. Other continental signals seem associated with hot spots (East Africa). The signal in oceanic regions is also interesting. Back-arc basins are clearly seen as slow anomalies, e.g. Lau basin, Mariana trough and Andaman sea back-arc basins. The East Pacific Rise is also clearly very slow.

Higher frequency maps are sensitive to shallower structure, which is dominated by the difference between oceans and continents. Even though the checkerboard tests suggest a resolution of 8-10 degrees for high frequency inversion results, it is surprising that the sharp contrasts between oceans and continents are preserved in our tomography maps. (Figure 2.14 and 2.15) It is also interesting to compare our tomography map at 30 mHz with the map predicted by the most recent crust model CRUST1.0 (Laske et al., 2013) in the continents. (Figure 2.20) At this frequency, group velocity is mostly sensitive to crustal structure in the

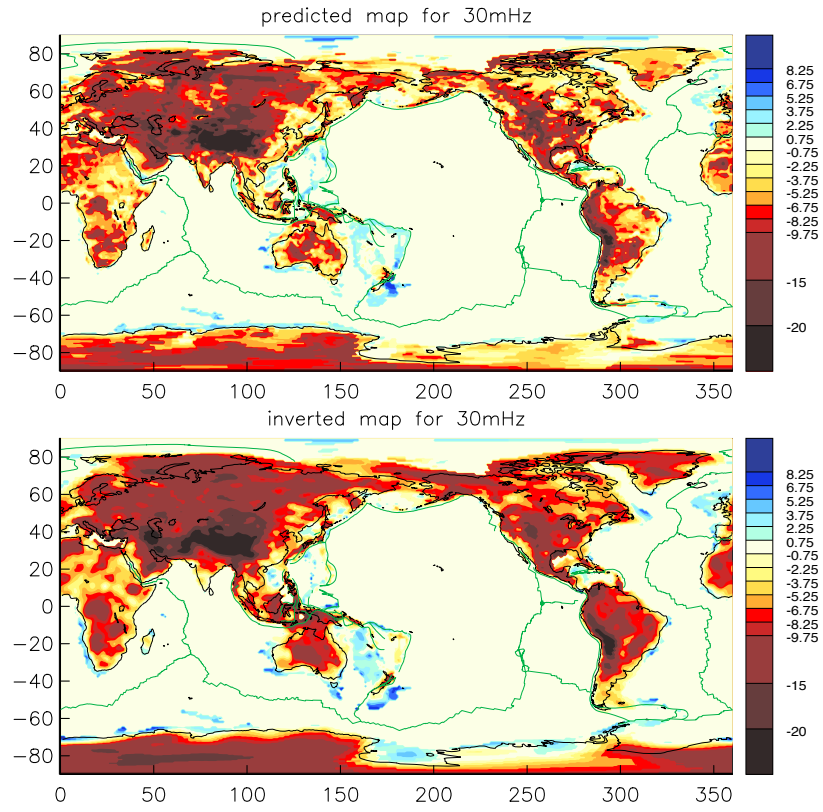


Figure 2.20: Comparison between the inverted group velocity perturbation map from our dataset (bottom) and predicted map from CRUST1.0 (top) in the continent for Rayleigh waves at 30 mHz. Units are percentages.

continental regions, which should be well characterized by CRUST1.0. The predicted map and our inverted map are indeed very close to each other, suggesting CRUST1.0 is a good starting crustal model for a future inversion for the entire lithospheric structure.

2.6 Conclusion

In this section, we demonstrate a new and efficient way of measuring Rayleigh and Love wave group velocity globally. For a single event, instead of making measurement for a broad frequency range for a single source-station pair, we filter all traces from all stations using a single narrow frequency band Gaussian filter and measure the relative arrival times among traces at once. Although Love waves are more likely to be contaminated by overtone interference than Rayleigh waves, synthetic tests show signals from the actual data can still be reliably recovered. Comparison with other existing datasets show no bias in our measurement techniques.

We also develop an inversion method with variable smoothness parameters, which produces a model with uniform errors globally. Our tomography maps match the pattern of large scale geological provinces. These maps provide a very good representation of our measurements, with well over 90% variance reduction for high frequencies and 80% for low frequencies. A future contribution will look at how well such data can resolve lithospheric structure globally.

Table 2.3: Parameters used in our inversions (see text for discussion). freq*: frequency. num[†]: number of measurements. ref[‡]: reference group velocity. cut[§]: cut off value for outliers. SMAD^{||}: the scaled median absolute deviation of the residuals after our inversion. Globe¹: the resolution for the globe. Eurasia²: the resolution for Eurasia.

type	freq (mHz)*	num [†]	ref (km/s) [‡]	cut (s) [§]	$\epsilon_{\max}$	λ_{start}	λ_{bad}	χ^2/N	SMAD (s)	v.d.	n.v.d.	Globe (°) ¹	Eurasia (°) ²
R	7.5	315693	3.7378	150	1.0	2.5	3.5	35.634	21.161	0.614	0.647	9.0	7.5
R	10	353021	3.8114	110	1.0	2.5	3.0	23.157	15.796	0.808	0.812	7.5	6.0
R	12.5	331292	3.8569	110	1.0	2.5	3.0	21.916	14.997	0.840	0.846	7.5	5.6
R	15	299080	3.8827	100	1.0	2.5	3.0	17.958	13.631	0.866	0.872	7.2	5.1
R	17.5	316046	3.8896	90	1.0	2.5	3.0	16.021	12.671	0.918	0.920	7.2	5.1
R	20	334229	3.8853	90	1.0	2.5	2.5	16.047	12.414	0.945	0.946	7.2	5.1
R	22.5	322771	3.8711	90	1.0	3.0	3.5	17.731	12.870	0.958	0.958	7.2	5.1
R	25	199070	3.8443	90	1.0	2.5	2.5	16.048	11.382	0.967	0.967	8.2	6.0
R	27.5	209878	3.8098	90	1.0	2.5	2.5	16.157	11.316	0.976	0.976	8.2	6.0
R	30	203324	3.7731	90	1.0	2.5	3.0	17.117	11.610	0.979	0.979	8.2	6.0
R	32.5	214258	3.7370	100	1.0	2.5	3.0	21.860	12.771	0.979	0.979	8.2	6.0
R	35	108857	3.6852	100	1.0	2.0	3.0	23.002	12.157	0.985	0.985	9.0	6.7
R	37.5	110295	3.6397	100	1.0	2.5	3.0	24.498	12.631	0.986	0.986	9.0	6.7
R	40	113247	3.5992	100	1.0	3.0	3.5	27.518	13.559	0.986	0.986	9.0	6.7
L	10	188702	4.3997	120	1.5	2.5	3.0	30.518	19.14	0.711	0.721	9.0	6.9
L	12.5	181562	4.3824	120	1.5	3.0	2.5	28.217	17.39	0.819	0.825	9.0	6.9
L	15	166457	4.3510	110	1.5	3.0	2.5	22.021	15.15	0.892	0.897	9.0	6.9
L	17.5	179941	4.3141	110	1.5	2.5	2.5	20.449	14.13	0.942	0.943	9.0	6.9
L	20	170804	4.2779	110	1.5	2.0	2.0	20.809	13.70	0.962	0.962	9.0	6.9
L	22.5	176282	4.2413	110	1.5	1.5	2.5	24.631	14.65	0.966	0.967	10.0	6.9
L	25	166160	4.2023	120	1.5	1.5	2.5	31.133	15.80	0.967	0.969	10.0	7.5
L	27.5	77741	4.1526	110	1.5	1.5	2.5	23.151	13.06	0.982	0.983	11.3	8.2
L	30	76300	4.1164	110	1.5	3.5	2.5	26.709	13.74	0.983	0.983	12.0	8.2

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Chapter 3

A comprehensive dispersion model of surface wave phase and group velocity for the globe

Abstract

A new method is developed to measure Rayleigh and Love wave phase velocities globally using a cluster analysis technique. This method clusters similar waveforms recorded at different stations from a single event and allows users to make measurements on hundreds of waveforms, which are filtered at a series of frequency ranges, at the same time. It also requires minimal amount of user interaction and allows easy assessment of the data quality. This method produces a large amount of phase delay measurements in a manageable time frame. Because there is a strong tradeoff between the isotropic part of the Rayleigh wave phase velocity and azimuthal anisotropy, we include the effect of azimuthal anisotropy in our inversions in order to obtain reliable isotropic phase velocity. We use b-splines to combine these isotropic phase velocity maps with our previous group velocity maps to produce an internally consistent global surface wave dispersion model.

3.1 Introduction

Surface waves are the dominant signals in long period seismograms for shallow earthquakes. They can be easily identified, and they allowed early investigators to identify large scale geological structures (e.g. Brune, 1969; Dorman, 1969; Knopoff, 1972). With the rapid growth of digital seismic data, surface wave dispersion can now be routinely measured on a global scale. Unlike body waves, which travel nearly vertically at shallow depths, surface wave dispersion measurements and the resulting phase and group velocity maps become essential datasets to constrain shallow upper mantle structure in recent global mantle models. (e.g. Masters et al., 2000; Kustowski et al., 2008; Ritsema et al., 2011)

For global studies, most of these dispersion measurements are made by comparing synthetic and recorded seismograms either in the frequency domain (e.g. Laske & Masters, 1996; Montagner & Tanimoto, 1990) or in the time domain (e.g. Trampert & Woodhouse, 1995; Ekström et al., 1997). Because these methods examine one record at a time, automatic measurement techniques are needed and various objective choices of quality control criteria have to be included. In contrast, our earlier paper (Ma & Masters, 2014) presents a new method to measure group velocity by comparing band-pass filtered seismograms recorded at different stations from different parts of the globe. This method allows analysts to visually examine each record while maintaining high efficiency in making measurements. The first goal of this chapter is to further develop this method to measure phase velocity.

The second goal of this chapter is to present reliable isotropic phase velocity maps over a broad frequency range for both Rayleigh and Love waves. Even though our final purpose is to use these maps to constrain 3D crust and mantle structures, these 2D surface wave dispersion maps are useful by themselves (e.g. for estimating earthquake moment tensors (Ekström et al., 2012)). However, as demonstrated in Ekström (2011), there are strong trade-offs between isotropic and azimuthal anisotropic signals. We show again that this is true and azimuthal anisotropy must be modeled together with isotropic signals in order to obtain reliable phase velocity maps.

The third goal is to present a dispersion model parameterized with equally

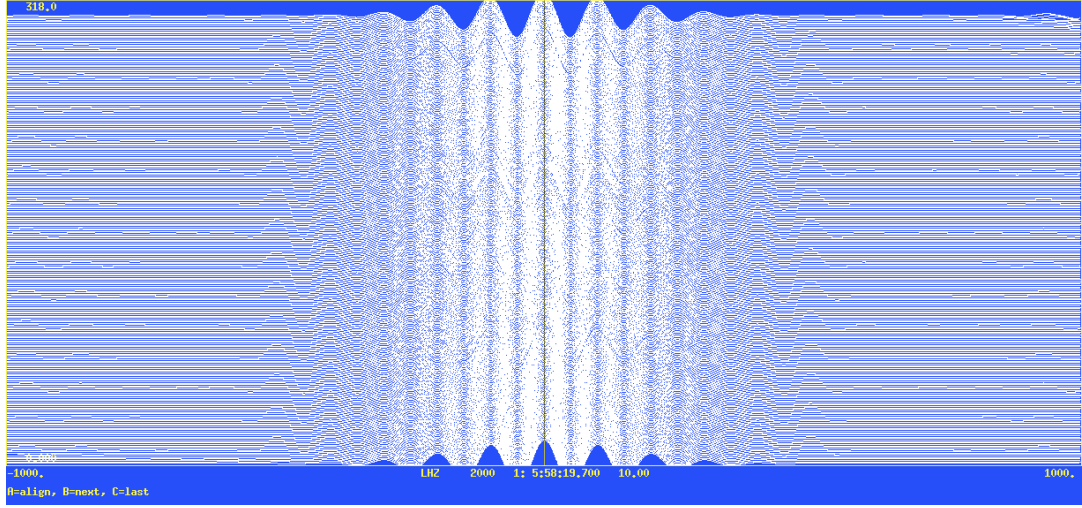


Figure 3.1: An example of synthetic waveforms after the preprocessing stage for one single event (01/05/2000). This example is for 10mHz Rayleigh waves and there are 318 waveforms. Synthetics are calculated using fundamental modes only. The horizontal axis is the time axis from -1000 sec to 1000 sec. The center of the screen, which is marked by a vertical yellow line, is the predicted arrival time from the PREM model. Each trace is the waveform for one station and sorted by epicentral distance. Note that all traces have almost identical shapes, which makes clustering and cross-correlation feasible. The largest peaks from all traces align well at the predicted arrival time.

spaced b-splines in the frequency domain, to explain both our group and phase velocity maps, which are made separately and independently at different frequencies. Such a spline model provides an internally consistent description of the dispersion characteristics of surface waves globally.

Because the main purpose of this chapter is to provide reliable estimates of isotropic phase and group velocity maps, the scope of this chapter is limited. We will only discuss phase delay measurements, although amplitude perturbations can also be measured using our methods,. We use ray theory throughout this chapter but other theory, e.g. finite frequency theories (Zhou et al., 2004), can be used to model our dataset. The pattern and reliability of azimuthal anisotropic structure will be discussed in more detail in a future paper.

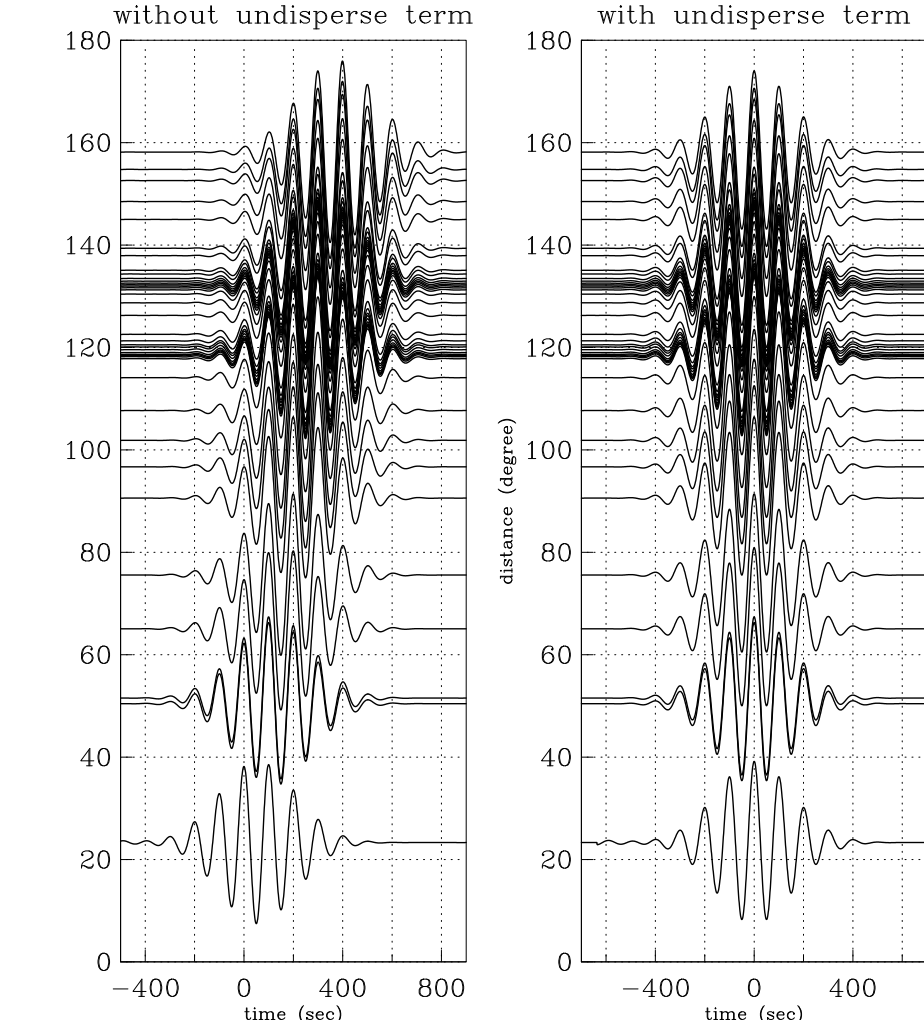


Figure 3.2: The effect of the ‘undisperse’ term on synthetic seismograms. Synthetics are calculated for a single event (event ID 010100B (globalcmt.org)) using fundamental modes only. Seismograms are filtered at 10mHz and processed as described in the chapter, with and without the ‘undisperse’ terms. Each trace represents one record from one station, sorted by epicentral distance. Time zero represents the predicted arrival time using phase velocity from PREM.

3.2 Measurement technique

3.2.1 Preprocessing

We collect all long period waveform data available from IRIS from 1988 to 2007 for events with m_b or $M_s \geq 5.5$. Broadband data from temporary deployments in the southern hemisphere, the Hawaiian PLUME experiment and the POLARIS array in northern Canada are also included to improve coverage. We measure Rayleigh wave dispersion on vertical components and Love waves on transverse components. Only minor arc waveforms are used.

We first transform our data to the frequency domain and apply a narrow bandpass Gaussian filter:

$$H(\omega) = \exp[-\alpha(\frac{\omega - \omega_m}{\omega_m})^2] \quad (3.1)$$

where ω_m is the center frequency of the filter and α can be expressed in terms of the relative bandwidth BAND and the decay rate β through $\alpha = \beta/\text{BAND}^2$ (equation (2.5) in Dziewonski et al. (1969)). We set BAND = 0.25 so that the bandwidth at 20 mHz is 5 mHz. And we use $\beta = 3$ in this chapter.

After filtering, the surface wave spectrum $u(\omega)$ can be written as

$$u(\omega) = H(\omega)A(\omega)\exp[i\phi(\omega)] \quad (3.2)$$

where $A(\omega)$ and $\phi(\omega)$ are the amplitude and phase at frequency ω respectively. Unlike Ma & Masters (2014) which deals with the envelopes of $u(\omega)$, in this chapter we are mainly interested in the part of the phase information $\phi(\omega)$ that is caused by the propagation of surface waves in the 3D Earth. However, $\phi(\omega)$ is affected by other factors, and we introduce a phase correction, ϕ_c , to correct for these effects:

$$\phi_c = \frac{\pi}{4} - \phi_r(\omega) - \phi_s(\omega) + \omega\Delta[\frac{1}{c(\omega)} - \frac{1}{c(\omega_m)}] \quad (3.3)$$

where $\pi/4$ is the static phase for minor arc surface waves (e.g. see equation (11.23) in Dahlen & Tromp (1998)), ϕ_r is the instrument response, ϕ_s is the source phase

for each source-station pair, Δ is the distance between this source-station pair and $c(\omega)$ is the average phase velocity along this path at frequency ω .

After filtering and correcting for ϕ_c , we transform $u(\omega)$ back to the time domain, shift the time series according to the predicted arrival time

$$t_0 = \frac{\Delta}{c(\omega_m)} = \int_0^\Delta \frac{dl}{c} \quad (3.4)$$

and plot them on the screen. Figure 3.1 shows an example of the synthetic data for 10 mHz Rayleigh waves. Note that all traces have almost identical shapes in this example, which makes the clustering and cross-correlation method feasible. For real data at frequencies higher than 10 mHz, existing 2D phase velocity maps (Ekström, 2011) are needed for this shifting in order to avoid cycle skipping.

Various terms in equation (3.3) need further explanations. The source phase ϕ_s for each source-station pair is calculated using PREM (Dziewonski & Anderson, 1981), centroid locations and centroid depths published on the global CMT website (globalcmt.org). Because we are comparing real waveforms recorded at different stations instead of comparing them with synthetics, a subtlety arises when calculating the source phase correction. This problem arises from the fact that using a first order approximation to the Associated Legendre functions can actually generate a small but observable bias in our measurements at frequencies lower than 10 mHz. As far as we know, this subtlety is not described in any standard textbooks, so we include a more detailed discussion about it in the Appendix.

We call the last term in equation (3) the ‘undisperse’ term, which corrects for the effect of dispersion so waveforms from different epicentral distances have similar shapes. Its meaning can be clearly seen when we process synthetic waveforms with and without it (Figure 3.2). Waveforms are filtered at 10 mHz and aligned at the predicted arrival time t_0 using the phase velocity from PREM. When measuring phase, we are not tracking the arrivals of the peak energies (those correspond to group velocity), but the arrivals of peaks and troughs. For waveforms processed without the undisperse term, the amplitudes of the correct peaks that we are measuring (the ones at time 0) become smaller and smaller as epicentral distance increases, especially when compared with the maximum peak in each

trace. For waveforms processed with the undisperse term, the correct peaks have the largest amplitudes and are easy to identify. More importantly, because the waveforms now have almost identical shapes, it makes the clustering and cross-correlation technique feasible.

For frequencies higher than 20 mHz, we find it necessary to calculate $c(\omega)$ in equation (3.3) for each path using a 3D Earth model or existing 2D phase velocity maps (Ekström, 2011). Figure 3.3 shows one example for the real data. This example is for a single event (event ID 010500C) and only a small number of stations are plotted to demonstrate the process. Waveforms are aligned according to the 2D phase velocity map. Waveforms processed with the ‘undisperse’ terms with 2D phase velocity maps are more likely to have large peaks centered on the screen, compared to the ones calculated by using 1D PREM. This makes it easier to group waveforms into one large cluster instead of several small ones, which makes cluster analysis faster and the estimation of cluster times more reliable (see below). We should also note that the arrival times of the peaks and troughs remain the same, indicating that the phase arrival times are not sensitive to the dispersion model we use. The only exception is station KRIS and the amplitude of the desired peak is so small that it is hard to identify and to be used for later analysis. The reason is that its ray path crosses the slowest anomaly (Himalayas) in the map and PREM is too far away from the real structure in this region. We also notice that for real seismograms, the correct peaks do not necessarily have the largest amplitudes as shown before for synthetics due to noise and the uncertainty of the real dispersion curve $c(\omega)$, but the enhancements are always effective and allow us to identify the correct peaks easily. This idea of correcting for a good estimation of starting models before making measurements is similar to Dziewonski et al. (1972), which helps to reduce measurement error.

Stations known to have incorrect polarities and orientations are corrected. Love waves are not measured on the PLUME dataset because of the uncertainties of the alignments of ocean bottom instruments. Stations with known timing errors are also excluded at this stage.

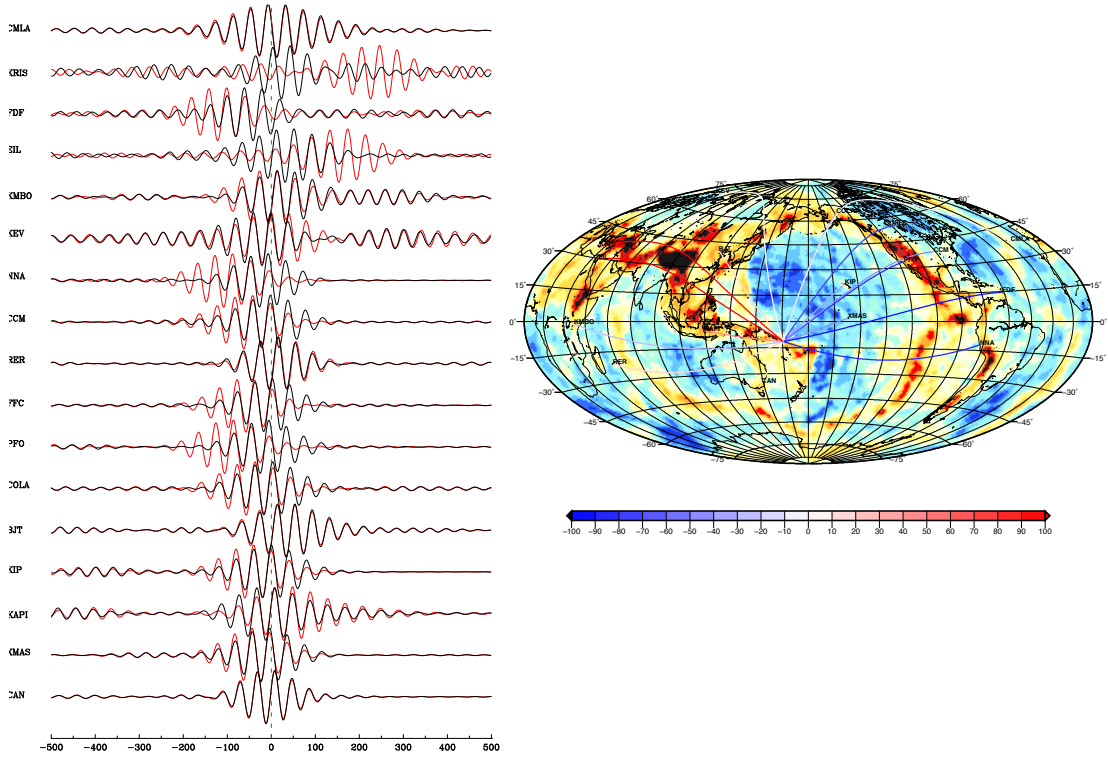


Figure 3.3: Importance of the ‘undisperse’ term at high frequency. The comparison between waveforms processed using a 2D starting model (Ekström, 2011) (black) and 1D PREM (red) for the ‘undisperse’ term on the left. The center of the figure (marked by a green dash line) is the estimated arrival time from the 2D starting model. On the right shows the paths for all traces and the color of each path indicates the arrival time of the maximum peak of each trace processed with PREM (units are seconds). Background color map is a phase velocity map for 25mHz Rayleigh wave, with red indicating slow and blue indicating fast velocity perturbations. Note that the traces calculated from the 2D starting model are more likely to have the peaks of waveforms center on the screen and this makes it easier for the cluster analysis. However, the arrival times of the peaks and troughs of each trace do not change regardless of the starting model we use, indicating that our measurements of phase arrival times are not sensitive to the starting model. The only exception is station KRIS, whose path crosses the slowest anomaly (the Himalayas) on the map.

3.2.2 Cluster analysis

We first manually choose windows for cross-correlation for a few months of data for each frequency on screens similar to Figure 3.1. The windows we choose for each frequency are very consistent, and typically cover 4 to 5 cycles. Then the same windows are applied to all other events.

Cross-correlation functions for every trace with every other trace with the ratio of the signal power to the noise power greater than 9 are then calculated and the maximum positive peaks that represent the differential times are identified. It is important to limit how much each record can be time-shifted in the cross-correlation analysis. We have found that we must limit time shifts to less than 0.6 of a cycle to avoid cycle skipping. This is why it is important to pre-shift records using an existing phase velocity map (equation 3.4).

A time domain method as described in VanDecar & Crosson (1990) is then used to determine the relative travel times among all traces for one event. This method tends to underestimate the real errors in the measurements. By examining similar paths for which the starting and ending points are less than 4° apart (roughly 1 wavelength), we determine an error of about 4 seconds. Although the long period data have slightly larger errors than short period data (e.g. 4.0 s for 10 mHz and 3.2 s for 20 mHz), we find that using an error of 4 s for all frequencies is sufficient for our purpose. According to the similarities between traces as measured by the cross-correlation coefficients, the clustering algorithm in Hartigan (1975) is used to generate clusters and cluster trees.

Figure 3.4 shows one example of clustering of waveforms. Waveforms are shifted according to their relative arrival times and plotted on the left. The corresponding cluster tree is plotted on the right. Users can then use the mouse to choose an appropriate cut off value as described in Ma & Masters (2014). The vertical yellow line indicates an appropriate choice for this example. Distorted and noisy records can also be identified and discarded at this stage. The fact that we can process hundreds of waveforms at one time makes this method very efficient and we have very good quality control on our data. After clusters are identified, relative travel times among traces within the same cluster are then calculated again

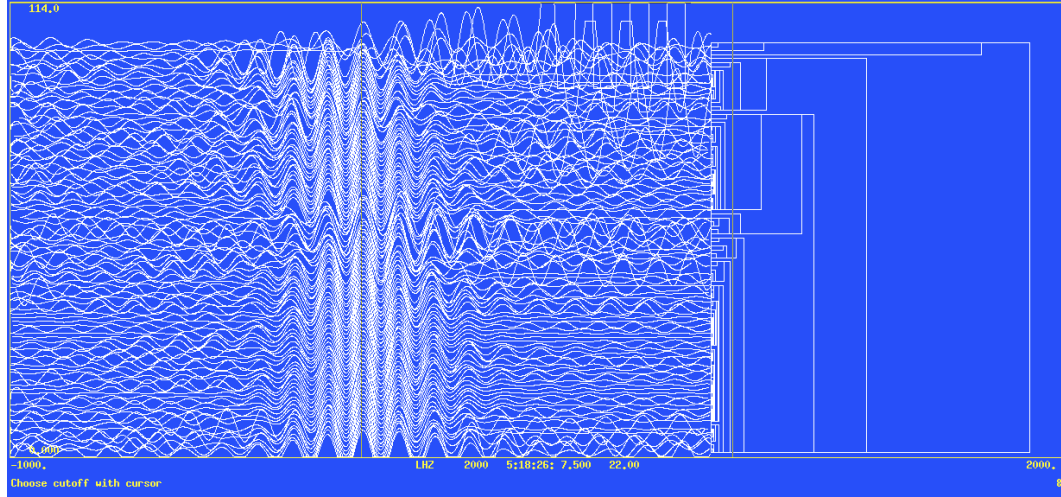


Figure 3.4: An example of the clustering of waveforms which are shifted according to their relative arrival times and the corresponding cluster tree. There are 114 waveforms in this example. Users can use their mouse to click on the cluster tree on the right to choose an appropriate cut off value. The vertical yellow line indicates a reasonable value for this example, which will generate three separate large clusters of clean records. Note that the distorted and noisy records can also be easily rejected at this stage without prescreening the data.

and stored.

We use this technique to make clusters manually between 10 mHz and 35 mHz for Rayleigh and Love waves, in steps of 5 mHz. In addition, we make measurements for Rayleigh waves at 5 and 7.5 mHz, and for Love waves at 7 mHz. Other frequencies are made automatically using clusters from nearby frequencies. For example, measurements at 12.5 mHz are made using clusters from 10 and 15 mHz. We have nearly 600,000 measurements for low frequency Rayleigh waves and 200,000 for Love waves. Measurements at high frequencies are harder to make, but we still have 280,000 measurements for Rayleigh waves at 35 mHz and 100,000 for Love waves at 30 mHz. (Table 3.1)

3.2.3 Converting relative travel time to absolute travel time

Unlike making group velocity measurements where clear peaks can be used as proxies for the absolute times, a more sophisticated approach is needed to convert relative times to absolute times when measuring phase. We introduce the

‘cluster time’ t_{cluster} for each cluster to shift the relative time t_{rel} we get in the previous section to obtain absolute time t_{abs} by using the following equation

$$t_{\text{rel}} - t_{\text{cluster}} = t_{\text{abs}} \quad (3.5)$$

If we further set $t_{\text{abs}} = t_{3D} + t_{1D}$, where t_{3D} and t_{1D} are the contributions to travel times from the 3D and 1D parts of Earth structures respectively, equation (3.5) can be rewritten as

$$\begin{aligned} t_{\text{rel}} - t_{3D} &= t_{1D} + t_{\text{cluster}} \\ &= d \cdot \bar{s} + t_{\text{cluster}} \end{aligned} \quad (3.6)$$

where t_{3D} can be estimated from phase velocity maps inverted from relative times. We can then solve for a single unknown parameter $\bar{s}$ (mean slowness for each frequency) and t_{cluster} for each cluster. Errors introduced in this step are

$$\sigma = \frac{\text{median}|t_{\text{rel}} - t_{3D} - d \cdot \bar{s} - t_{\text{cluster}}|}{0.6745\sqrt{n}} \quad (3.7)$$

where n is the number of traces in the cluster. The coefficient 0.6745 is used so that the scaled median absolute deviation is a consistent estimation of the standard deviation. Cluster times for small clusters are not well resolved because of the trade off between t_{3D} and t_{cluster} . We discard clusters with less than 5 traces.

We compare the phase delay data collected by this method with the data collected in Ekström et al. (1997) and Ekström (2011). Figure 3.5 shows an example for 10 mHz Rayleigh and Love waves. For Rayleigh waves, the standard deviation of the differences in phase delay times is about 4.5 s, slightly larger than that for Love waves, which is 3.8 s. Both of these are close to our estimated measurement errors (4 s) and are much smaller than the signals we are modeling, which can exceed 80 s in some cases. We do notice that there is an average offset of about 1.5 s for Rayleigh waves. This may lead to 0.06% discrepancy in the 1D average phase velocity (assuming the dominant path length is 90°), which is small and has negligible effect on the resulting 3D tomography maps. Because of the consistency among these datasets, we include the data collected by Ekström et

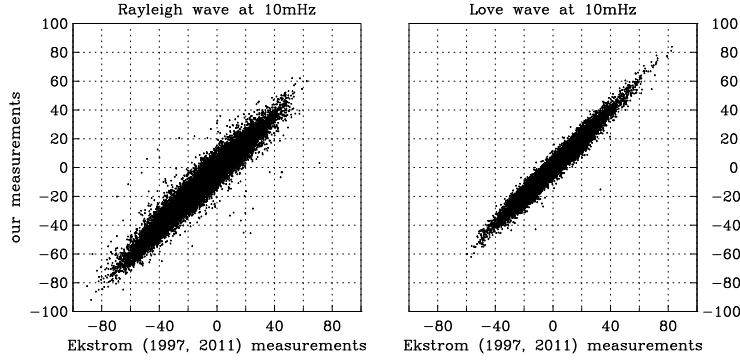


Figure 3.5: Comparison of the phase delay measurements collected in this study and the measurements collected in Ekström et al. (1997) and Ekström (2011). The units are in seconds.

al. (1997) and Ekström (2011) at the overlapping frequencies (5, 10, 20, 25 mHz) in our inversions below. Inversions done with our dataset alone give very similar results.

3.3 Inversions

3.3.1 Theory

Our model is parameterized as 41252 equal area blocks with the size of 1° by 1° at the equator. In ray theory, the arrival time anomaly δt is

$$\delta t = s_{ref} \int_0^\Delta \frac{\delta s}{s_{ref}} d\gamma \quad (3.8)$$

where the integral is over the minor arc path connecting each source-station pair, and $\delta s/s_{ref}$ is the relative perturbation in slowness for each block.

In a slightly anisotropic medium, δs includes both an isotropic term and azimuthal anisotropic terms (Smith & Dahlen, 1973)

$$\begin{aligned}
\delta s &= \delta s_0 + A_1 \cos(2\phi) + A_2 \sin(2\phi) \\
&+ A_3 \cos(4\phi) + A_4 \sin(4\phi)
\end{aligned} \tag{3.9}$$

where ϕ is the azimuth along each path. There are numerous debates on which terms can be resolved for Rayleigh and Love waves. Laske & Masters (1998) suggest their phase velocity dataset is not sensitive enough to resolve any azimuthal anisotropy, while Trampert & Woodhouse (2003) argue both the 2ϕ and 4ϕ terms are required for Rayleigh waves and the 4ϕ term is required for Love waves. In this chapter, because we focus mainly on isotropic velocity and only the 2ϕ terms for Rayleigh waves have significant impacts on the resulting isotropic velocity maps (Ekström, 2011), we will only consider the 2ϕ terms for Rayleigh waves in our inversions and neglect all other anisotropic terms.

In our inversions, we use two different lateral smoothness parameters λ_0 and λ for the isotropic and anisotropic parts respectively, and we minimize the function:

$$\begin{aligned}
f &= (\mathbf{A}\mathbf{m} - \mathbf{d})^T (\mathbf{A}\mathbf{m} - \mathbf{d}) + \lambda_0 (\mathbf{D}\mathbf{m}_0)^T \mathbf{D}\mathbf{m}_0 \\
&+ \lambda (\mathbf{D}\mathbf{m}_1)^T \mathbf{D}\mathbf{m}_1 + \lambda (\mathbf{D}\mathbf{m}_2)^T \mathbf{D}\mathbf{m}_2
\end{aligned} \tag{3.10}$$

where $\mathbf{A}$ is the matrix that corresponds to the numerical integral of equation (3.8) for all source-station pairs, D is the numerical approximation to the 2D Laplacian, m_0 , m_1 and m_2 correspond to δs_0 , A_1 and A_2 terms for all blocks. Because we consider azimuthal anisotropy only for Rayleigh waves, the last two terms, m_1 and m_2 , are set to be 0 for Love waves. The standard LSQR method (Paige & Saunders, 1982) is used for the inversion.

3.3.2 Trade off between isotropic and anisotropic signals for Rayleigh waves

Because there are strong trade offs between isotropic and anisotropic signals for Rayleigh waves (Ekström, 2011), we have to include azimuthal anisotropic terms

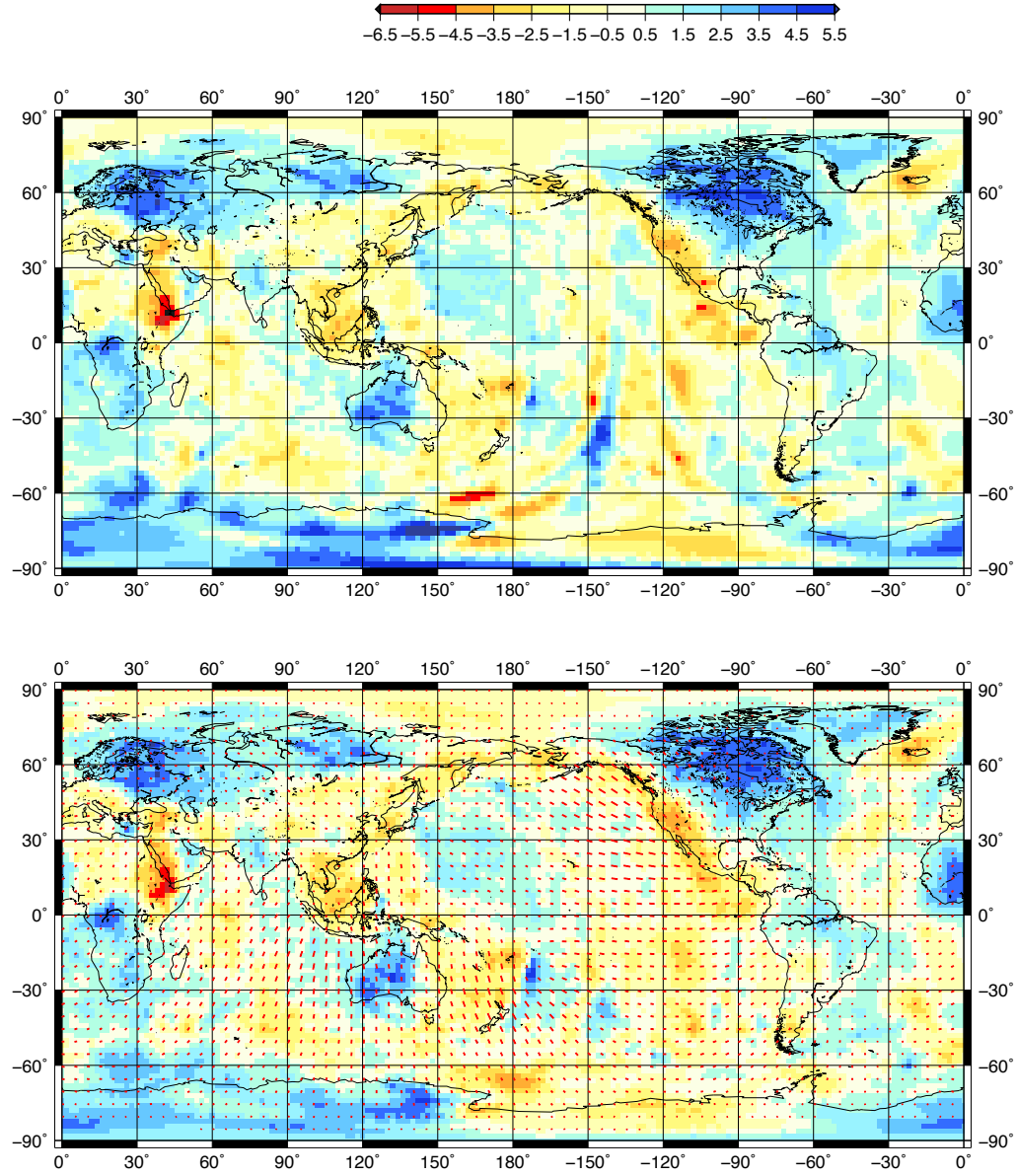


Figure 3.6: An example of inverted Rayleigh wave isotropic and azimuthal anisotropic maps at 100 seconds. Background colors represent the perturbations of isotropic velocity in percents. Directions and lengths of the red bars represent the fast directions and amplitudes of azimuthal anisotropy. (top): Inversion result for $\lambda = 10000$. This is the same as not including azimuthal anisotropy in the inversion. A distinct streak can be seen in the middle of the Pacific ocean. Other possible artifacts can also be identified in the southern oceans. (bottom): Inversion result for $\lambda = 200$. The maximum amplitude of anisotropy (the longest bar) is 0.63%. Note that all artifacts caused by ignoring azimuthal anisotropy in the top panel are successfully removed.

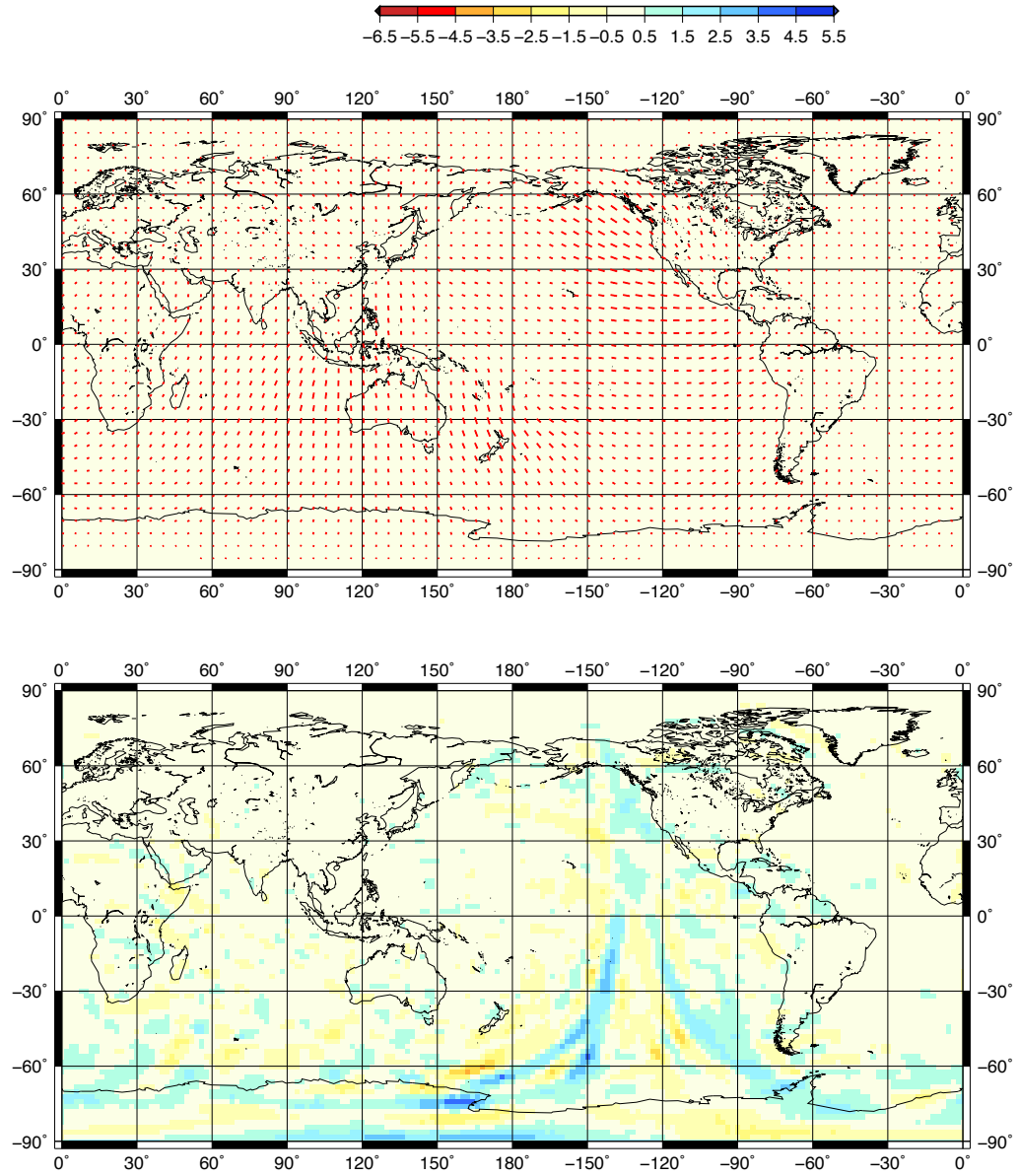


Figure 3.7: The result of a synthetic test using an anisotropic map for input and the resulting isotropic map. Top panel shows the anisotropic map from Figure 3.6, but the isotropic perturbation is set to be zero. We use it as an input model to calculate synthetic data using equations (3.8) and (3.9). Bottom panel shows the inversion result from these synthetic data, but only the isotropic part is inverted for. Compared with Figure 3.6, we reproduce the artifacts at exactly the same places and almost identical amplitudes.

in order to obtain reliable isotropic velocity maps. An example is shown in Figure 3.6. In this example, we are solving for Rayleigh wave phase velocity at 100 seconds using our dataset. The top panel shows the result of setting $\lambda = 10000$, which is equivalent to ignoring anisotropy in our inversion. A distinct streak can be seen in the Pacific, which we suspect is an artifact resulting from incomplete ray coverage, and other possible artifacts are found in different places in the southern oceans. These types of structures can also be found in other studies using surface wave datasets (e.g. Figure 4 in Ritsema et al. 2011 and Figure 8 in Houser et al. 2008), so they are unlikely to be caused by certain bad measurements in our dataset or the details of inversion methods.

Dramatic changes can be seen by simply adding azimuthal anisotropy. The bottom panel in Figure 3.6 shows the inversion result for the same dataset using the same smoothness for the isotropic parts, but setting $\lambda = 200$ for azimuthal anisotropy. Most of the artifacts we see in the top panel are successfully removed. Note that the maximum amplitude of anisotropy ($\sqrt{A_1^2 + A_2^2}$) in the figure is only 0.63%. The fact that such a small amount of anisotropy can cause significant bias in isotropic maps indicates that the azimuthal anisotropy pattern in the Pacific is likely to be coherent over very large distances, so that the contributions from anisotropy can be effectively accumulated to a significant level.

To further demonstrate this trade off, we use the anisotropic part of the map we get (bottom panel of Figure 3.6) as an input model, calculate synthetic data from this anisotropy model using equations (3.8) and (3.9), and invert them for an isotropic map. Figure 3.7 shows the result of this experiment. We successfully reproduce the artifacts we find in Figure 3.6 and conclude that those artifacts are caused by ignoring azimuthal anisotropy in the inversion.

3.3.3 Regularization

The choice for λ_0 follows the same procedure as described in Ma & Masters (2014), with the goal of obtaining uniform model errors throughout the globe. The error allowed for Rayleigh waves is 0.4% and for Love waves is 0.5%. This method then tries to apply the smallest possible smoothness for each block. An

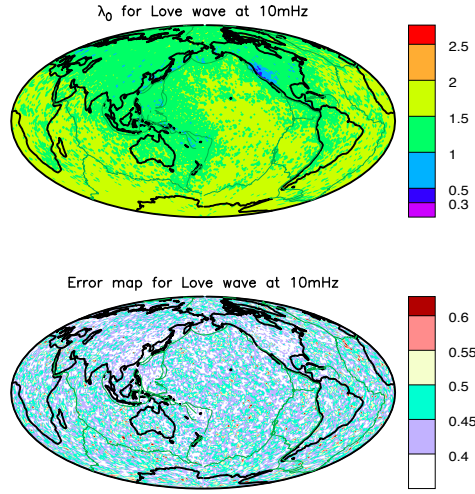


Figure 3.8: Smoothness parameters λ_0 used for Love wave at 10mHz (top) and the resulting error map (bottom). Units of the error map are in percent.

example of the resulting smoothness λ_0 and error map is shown in Figure 3.8.

The choice for λ requires more careful considerations because the actual amplitude of azimuthal anisotropy of the Earth is still not well known. Montagner & Anderson (1989) estimate the amplitude can be as large as about 4% for Rayleigh waves at 10 mHz for a pyrolite model. However, various global studies using surface wave datasets estimate maximum amplitudes to reach only 1-2%. (see comparison in Figure 1 in Becker et al. (2007)) We experiment with different smoothness λ for different frequencies and calculate the L curves between the fit to the data and the mean square amplitude of azimuthal anisotropy. Overlapping frequency (5, 10, 20, 25 mHz) data from Ekström et al. (1997) and Ekström (2011) are also included. (Inversions done with our dataset alone give L-curves with similar shapes.) We seek the models that best balance the fit to the data and the complexities of models themselves. The result is shown in Figure 3.9 and it indicates that smoothness between $\lambda = 15$ and $\lambda = 200$ seem to be reasonable choices for all frequencies.

While the amplitudes of azimuthal anisotropy are different when we use different values of smoothness for inversions, isotropic velocity perturbations remain rather stable and only a small differences are found in limited parts of the oceans. (Figure 3.10) In the maps inverted using $\lambda = 200$, the maximum amplitude for az-

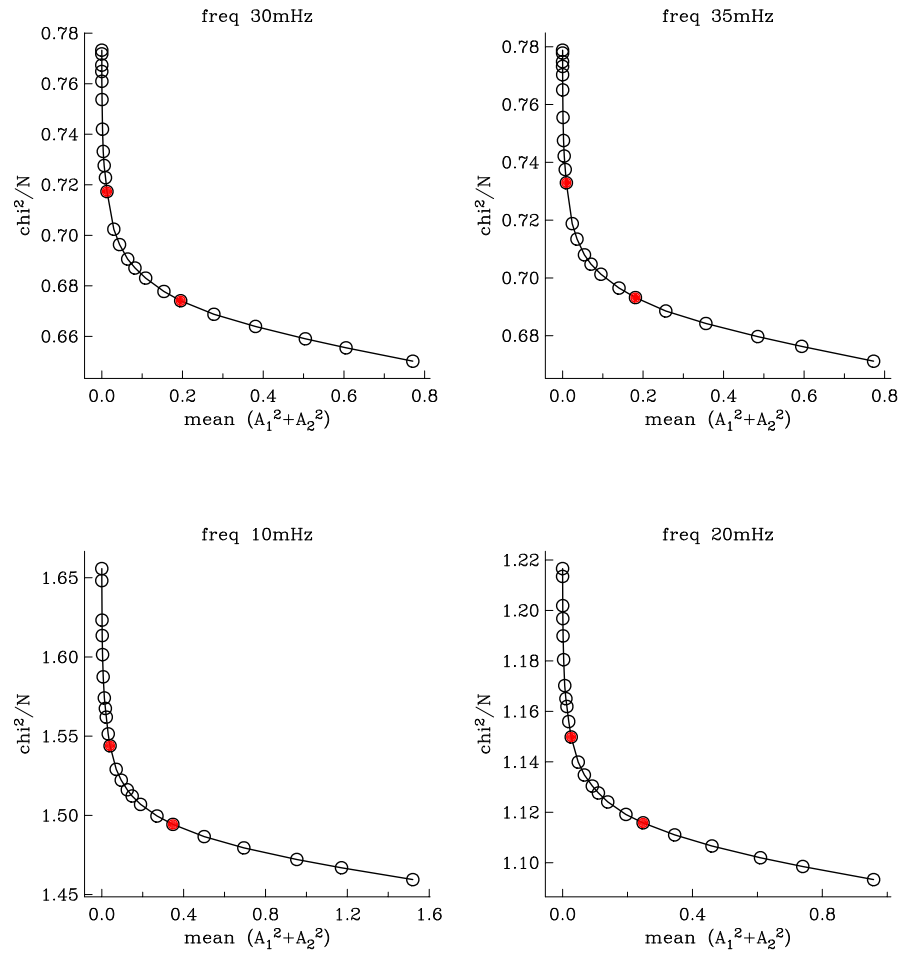


Figure 3.9: L-curves between χ^2/N and the mean square amplitudes $A_1^2 + A_2^2$ for different smoothnesses λ and frequencies. Red circles correspond to $\lambda = 15$ and $\lambda = 200$, representing a reasonable range of λ .

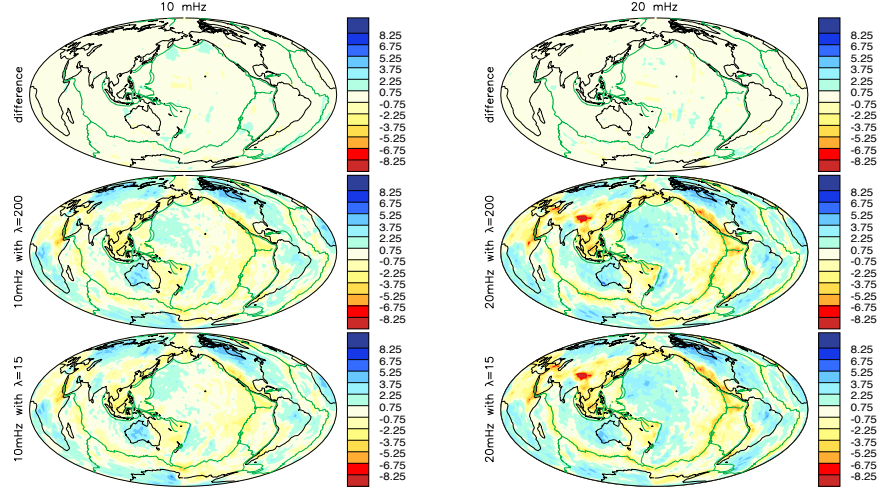


Figure 3.10: Comparison between the isotropic maps obtained by using different smoothness for azimuthal anisotropy. These examples are for 10mHz and 20mHz Rayleigh waves. Colors represent perturbations in percent.

imuthal anisotropy is only 0.63% for 10 mHz and 0.58% for 20 mHz. This indicates that while anisotropy should be included to obtain reliable isotropic maps, only a small amount of anisotropy is needed to remove artifacts and choices of smoothness do not have substantial effects on the resulting isotropic maps. Because we focus on isotropic maps in this chapter, we will use $\lambda = 200$ for our inversions. Effects of regularization on azimuthal anisotropic maps, especially on amplitude, will be discussed in a future paper.

3.3.4 Tomography results

Figure 3.11 and Figure 3.12 show the inverted isotropic maps for both Rayleigh and Love waves. These tomographic maps are excellent representations of our large datasets and produce over 90% variance reduction for most frequencies for both Rayleigh and Love waves. Details can be found in Table 3.1. The low variance reduction at 5 mHz is due to the fact that velocity perturbations at this frequency are rather small. Large scale features are well resolved in both the Rayleigh and Love wave cases. Slow anomalies are found beneath orogenic zones and other regions with thick crust, e.g. Himalayas and Andes. The Basin and Range province in North America, mid-ocean ridges, and back-arc basins where

the upper mantle is expected to be warmer than the surrounding regions also show up as slow anomalies. Cratons can be seen in the low frequency maps as regions of anomalously high phase velocity, e.g. Canadian Shield, West African Craton, Congo Craton, Kaapvaal Craton, Indian Shield, Siberia Craton, Russian Platform, etc. It is also surprising to find that the largest signal in maps with frequencies lower than 10 mHz is observed in the east African rift, suggesting a very deep origin (at least 200-300 km) for the anomaly.

We use checkerboard tests to estimate the resolutions of our maps. We focus on isotropic velocity perturbations in this chapter so only isotropic inversions are used in these experiments. Figure 3.13 shows the results for Rayleigh waves at 20 mHz, which is highly sensitive to crustal structure. An input pattern is generated using a spherical harmonic with different degrees of l and m (and we set $l = 2m$) with a 10% velocity perturbation. 60% of the perturbation can be recovered on a global scale for 6 degree pixels, and 90 % of the perturbation can be recovered for 6.7 degree pixels. We also notice that perturbations near the edges of each pixel are not well resolved in the oceans due to limited ray coverage and smoothing, but sharp changes of velocity in the ocean on such short scales are not likely to happen in the real Earth. Recovery of the input signal in Eurasia is better, with pixel recovery as small as 5-5.5 degrees because of dense ray coverage so that less smoothing is required in this region.

An (2012) suggests another way to estimate the resolution of a general inverse problem. He uses a Gaussian function to parameterize every row of the resolution matrix and uses random numbers to probe this matrix. The half width at the half maximum of the Gaussian function is denoted as w_i in his study and we consider $4w_i$ as equivalent to the size of our checkerboard. It is interesting to compare the results from his method with the more traditional checkerboard tests here. An application using our dataset for 20 mHz Rayleigh waves is shown in Figure 3.14. The pattern of the resolution estimated from An's method matches the pattern from our checkerboard tests very well (see the recovered patterns for $m=30$ and 27 for the globe in Figure 3.13). Unlike traditional checkerboard test, An's method computes the resolution consistently for every part of the globe and gives

a compact way to display the global resolution on a single map. The consistency of the resolutions estimated by using An's method and checkerboard test suggests that the more traditional checkerboard test is probably not as bad as L  v  que et al. (1993) suggest.

3.4 A b-spline model to explain both phase and group velocities at different frequencies

The work by Larson & Ekstr  m (2001) is among the earliest efforts to relate phase and group velocities on a global scale. They sample the derivatives of the full dispersion curves, $c(\omega)$, which are measured using the method described in Ekstr  m et al. (1997), to derive group velocity measurements. Ekstr  m (2011) further improves this method and relates phase and group slowness (instead of velocities) through a simpler linear equation.

Here, we design a model that is constrained by both phase and group velocity maps which are inverted independently. Because our measurements and inversions are done separately for phase and group velocities and independently at different frequencies, this method also examines the internal consistency of our tomographic maps. The isotropic part of phase velocity c and group velocity U can be related to each other by using

$$\frac{1}{U(\omega)} = \frac{1}{c(\omega)} + \omega \cdot \frac{d}{d\omega} \frac{1}{c(\omega)} \quad (3.11)$$

If we use b-spline functions, $B_i(\omega)$, to parameterize either the phase or group slowness, we can calculate the other using

$$\begin{aligned} \frac{1}{U(\omega)} &= \sum_i B_i(\omega) a_i \\ \frac{\omega}{c(\omega)} &= \sum_i \left[\int_{\omega_0}^{\omega} B(\omega') d\omega' \right] a_i + a_0 \end{aligned} \quad (3.12)$$

or

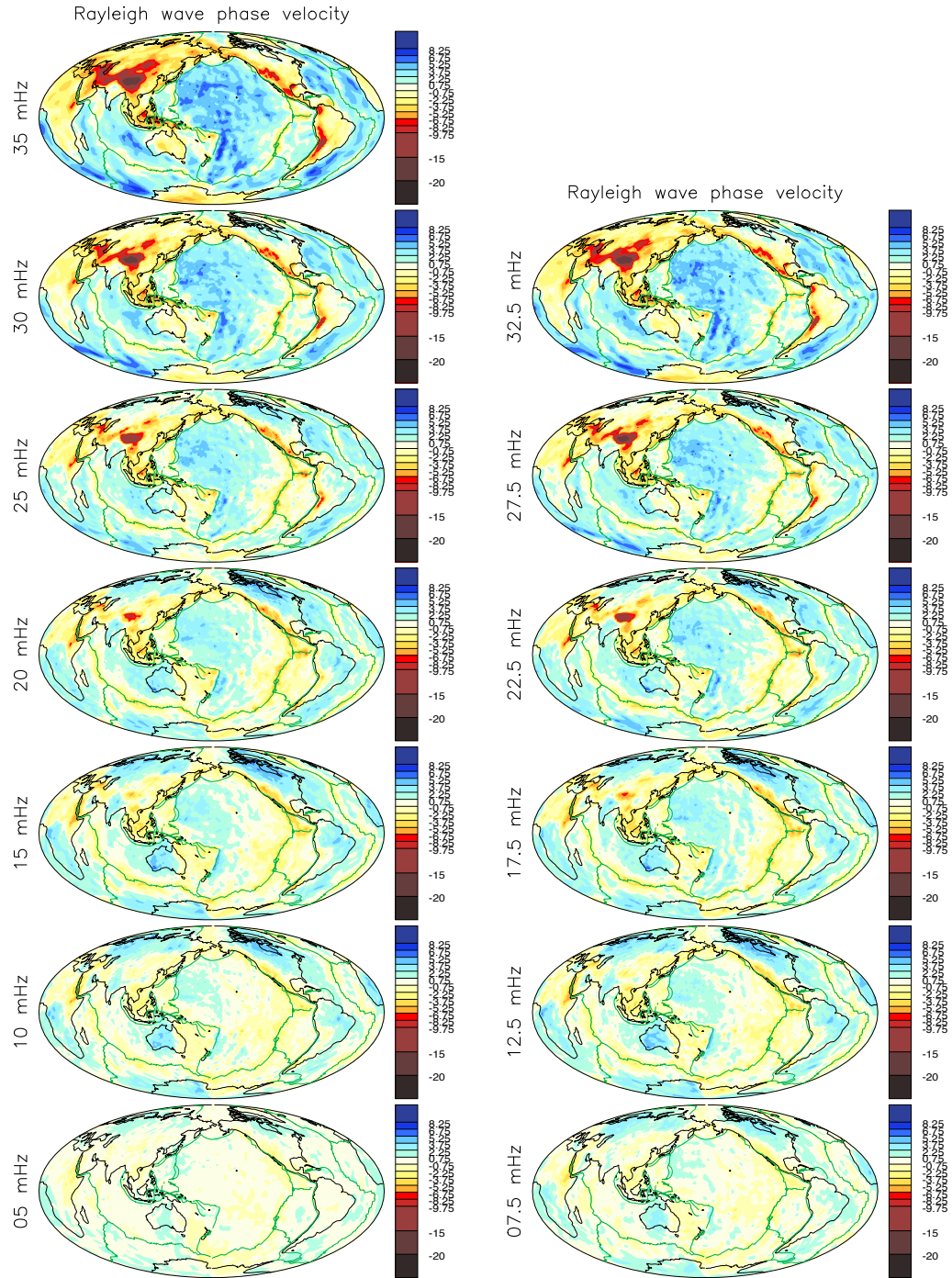


Figure 3.11: Rayleigh wave isotropic phase velocity with $\lambda = 200$ for azimuthal anisotropy. Colors indicate isotropic velocity perturbations with respect to the 1D average of each frequency.

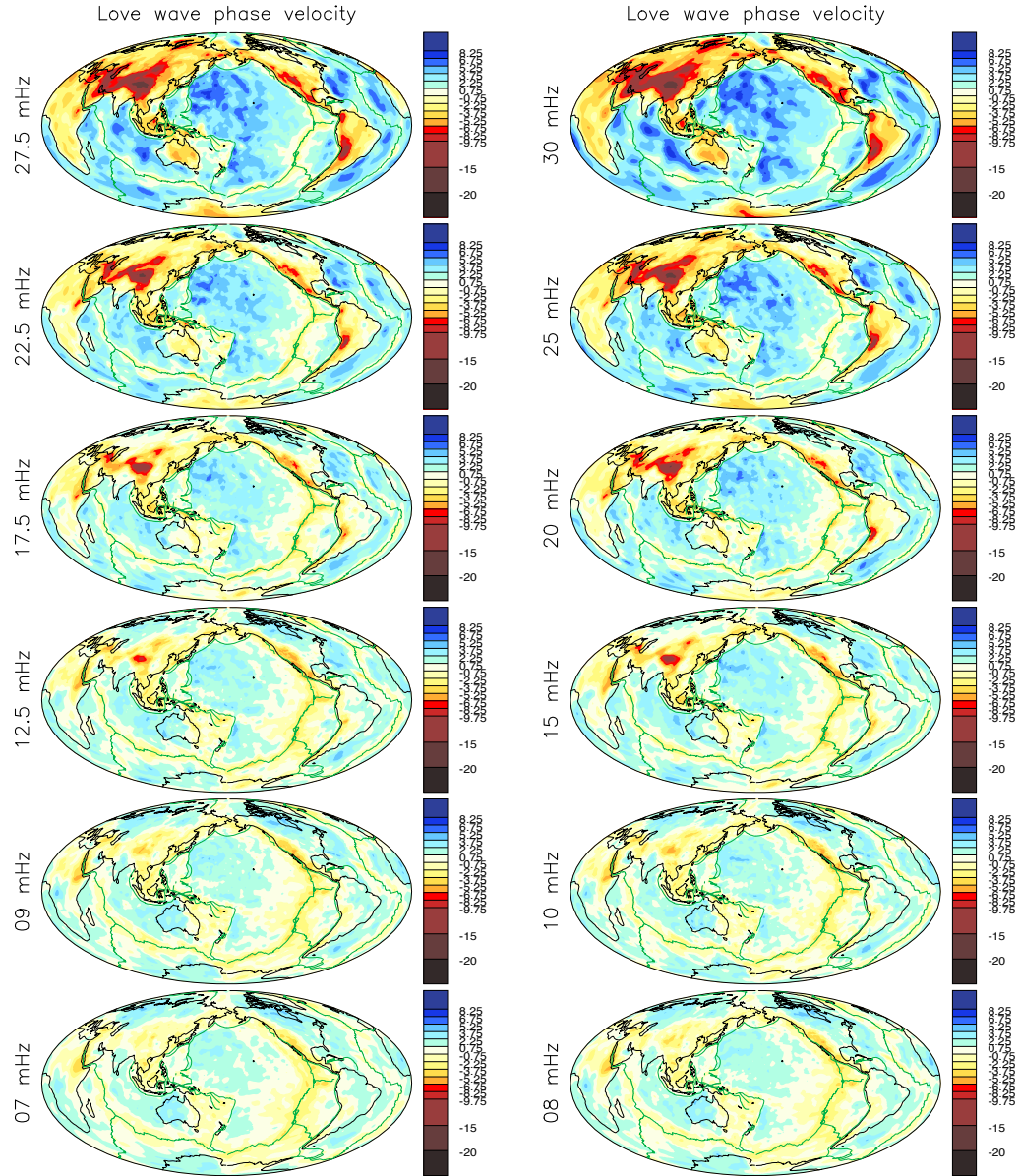


Figure 3.12: Love wave phase velocity. Colors indicate isotropic velocity perturbations with respect to the 1D average of each frequency. Note that no azimuthal anisotropy is included in the inversion.

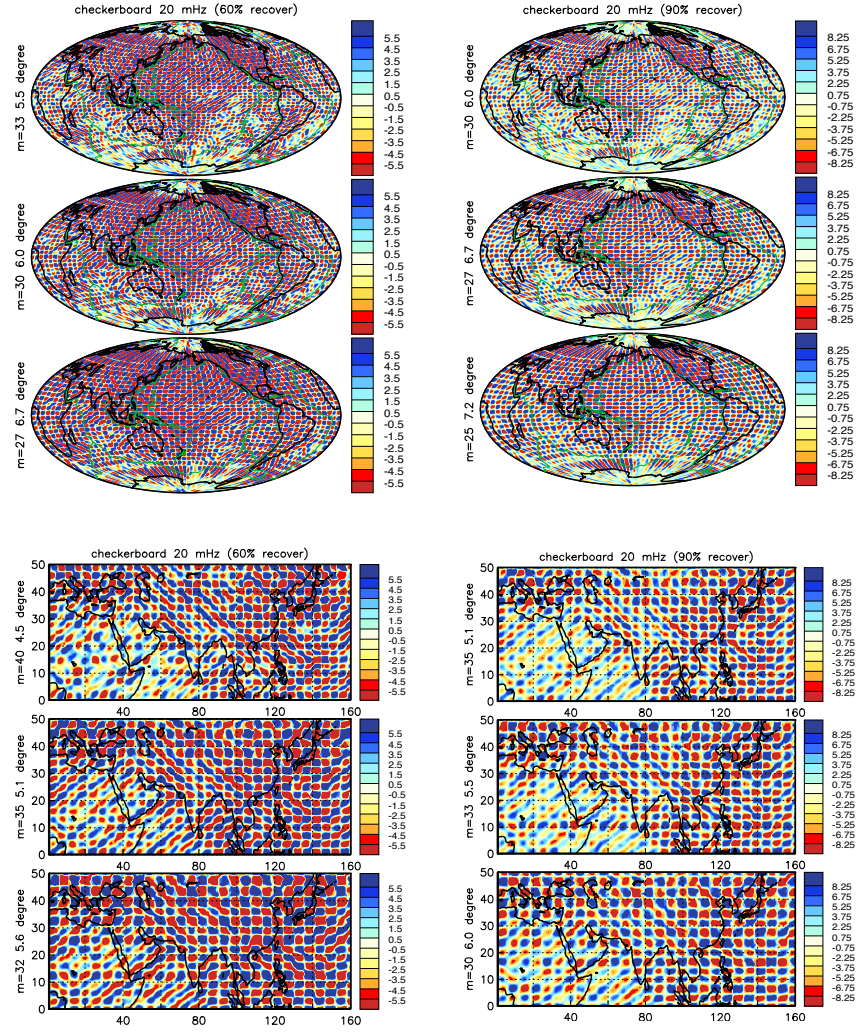


Figure 3.13: Checkerboard test for Rayleigh waves at 20mHz. The amplitude of the input checkerboard is 10%. Left panel shows the recovered pattern where 60% of the amplitude is recovered both globally (top) and in Eurasia (bottom). Right panel shows the recovered pattern where 90% of the amplitude is recovered.

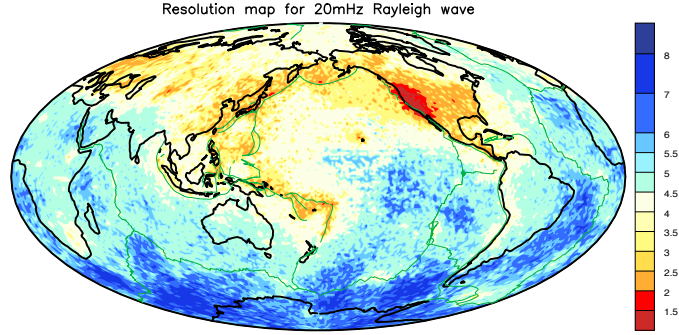


Figure 3.14: Resolution map for Rayleigh wave at 20mHz. The units are in degrees.

$$\begin{aligned}\frac{1}{c(\omega)} &= \sum_i B_i(\omega) a_i \\ \frac{1}{U(\omega)} &= \sum_i \left[\omega \frac{dB_i(\omega)}{d\omega} + B_i(\omega) \right] a_i\end{aligned}\quad (3.13)$$

These two sets of parameterizations give the same results and we use equation (3.12) in this chapter. We also constrain the smoothness of $1/U$ by constraining

$$\int \left[\frac{d^2}{d\omega^2} \frac{1}{U(\omega)} \right]^2 d\omega = \int \left[\sum_i \frac{d^2 B_i(\omega)}{d\omega^2} \cdot a_i \right]^2 d\omega = \sum_{i,j} a_i a_j H_{i,j} \quad (3.14)$$

where

$$H_{i,j} = \int \frac{d^2 B_i(\omega)}{d\omega^2} \frac{d^2 B_j(\omega)}{d\omega^2} d\omega \quad (3.15)$$

Similar formulations can be found in Trampert & Woodhouse (1995), where the authors use this parameterization to facilitate their measurement process. Here we use it to analyze our inverted maps. It is also more convenient to model $\delta(1/U)$ and $\delta(1/c)$, which are the perturbations of group and phase slowness with respect to PREM, to avoid the problem of the sharp rise of phase velocity at long periods. We use equally spaced cubic b-splines (Lancaster & Salkauskas, 1986) with knot spacing of 2mHz. Tests with a model with CRUST1.0 (Laske et al., 2012) on top of PREM show that this parameterization is sufficient.

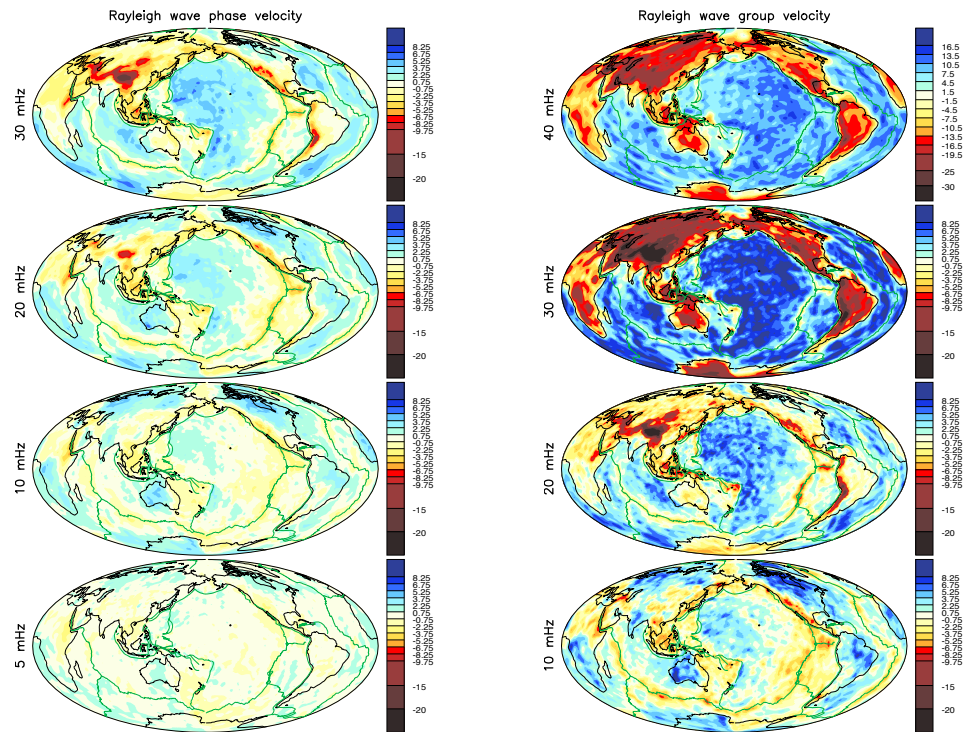


Figure 3.15: Rayleigh wave phase and group velocity from the spline model.

We calculate the spline model independently for each of the 41252 blocks in our model. The group velocity maps in Ma & Masters (2014) are used here. However, because they do not include azimuthal anisotropy for Rayleigh waves in their work, an artificial streak can be seen in their low frequency maps (as we have shown in section 3.2). We find that re-inverting their datasets using a $\lambda = 1000$ for azimuthal anisotropy is sufficient to remove this artifact. The re-inverted group velocity maps are then combined with the phase velocity maps to calculate the spline model. The relative weightings for group slowness, phase slowness and smoothness are 1.6, 1.0 and 0.1 for Rayleigh waves, and 2.0, 1.0 and 0.05 for Love waves. While not absolutely necessary, we also constrain the resulting mean velocities to be close to the values from individual maps:

$$\begin{aligned}\overline{1/U(\omega)} &= \sum_i B_i(\omega) \overline{a_i} \\ \overline{\omega/c(\omega)} &= \sum_i \left[\int_{\omega_0}^{\omega} B(\omega') d\omega' \right] \overline{a_i} + \overline{a_0}\end{aligned}\tag{3.16}$$

where the averaging is done over all the blocks. We set the relative weighting between phase and group velocity to be 10 to 1, because the mean phase velocity is estimated independently using equation (6). Figures 3.15 and 3.16 show a set of the predicted Rayleigh and Love maps from this spline model. Because the model smooths across different frequencies, these maps are slightly smoother than individual maps from inversions (Figures 3.11 and 3.12).

Figure 3.17 shows the spline model for nine different locations on the Earth. Errors for group velocity are 1.0%. Errors for phase velocity are 0.6%, which is slightly larger than the estimated error (0.4%) from tomography, but consistent with errors estimated in Shapiro & Ritzwoller (2002). As a comparison, predictions from the GDM52 model (Ekström, 2011) are also plotted. In most cases, the phase velocity values match each other well. The group velocities from GDM52 are more oscillatory. This is probably because the derivatives of the dispersion curves, from which group velocities are derived, are not as well constrained as the dispersion curves themselves. Compared to the GDM52 model, our spline model is able to explain both the phase and group maps more consistently. Group velocity for the

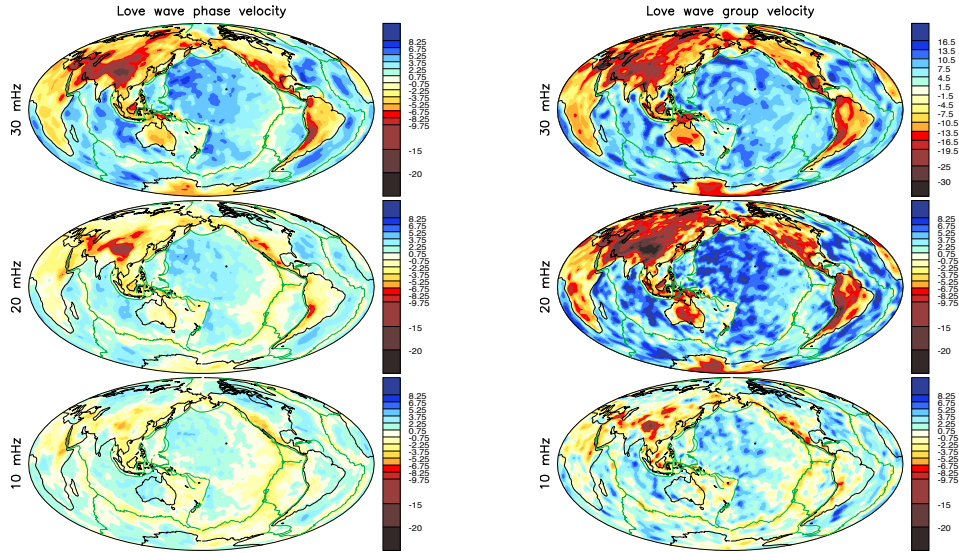


Figure 3.16: Love wave phase and group velocity from the spline model.

Andes is slightly slower than the prediction from the spline model. This suggests that either our simple ray theory may not work properly or our use of lateral smoothing produces overshoot near places where sharp velocity contrasts between ocean and thick continents exist. Figure 3.18 is an example for Love waves for the same locations. The error for phase velocity is 0.75%, slightly larger than the one estimated previously in tomography. Error for group velocity is 1.5% and is the same as the one from tomography. We should point out that the high frequency end (≥ 27.5 mHz) of Love wave phase velocity is not always well predicted from our model in some places. This may indicate that a larger error should be assigned to these measurements.

3.5 Discussion

This chapter extends the cluster analysis technique described in Ma & Masters (2014) to measure phase delays for Rayleigh and Love waves at frequencies ranging from 5 mHz to 35 mHz. The main differences are in the preprocessing

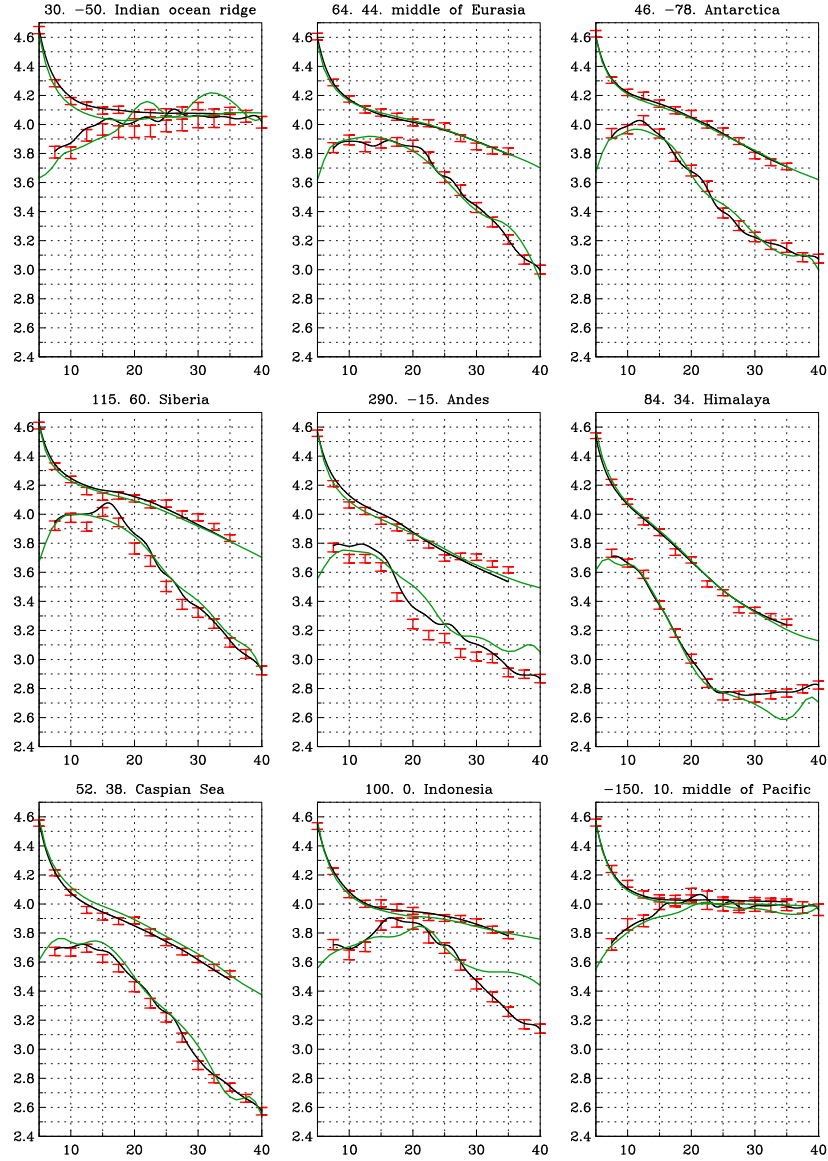


Figure 3.17: The spline model for Rayleigh waves for different locations on the Earth. Red bars are our inverted phase and group velocities. Black lines are estimates from our spline model. Green lines are the GDM52 model (Ekström, 2011). Numbers on top of each figures are the corresponding longitudes and latitudes. In most cases, our phase velocity values are similar to the GDM52 model. While the group velocities from the GDM52 model tend to be oscillatory, our spline model can explain both phase and group velocities consistently. See text for possible explanations for the misfit in the Andes.

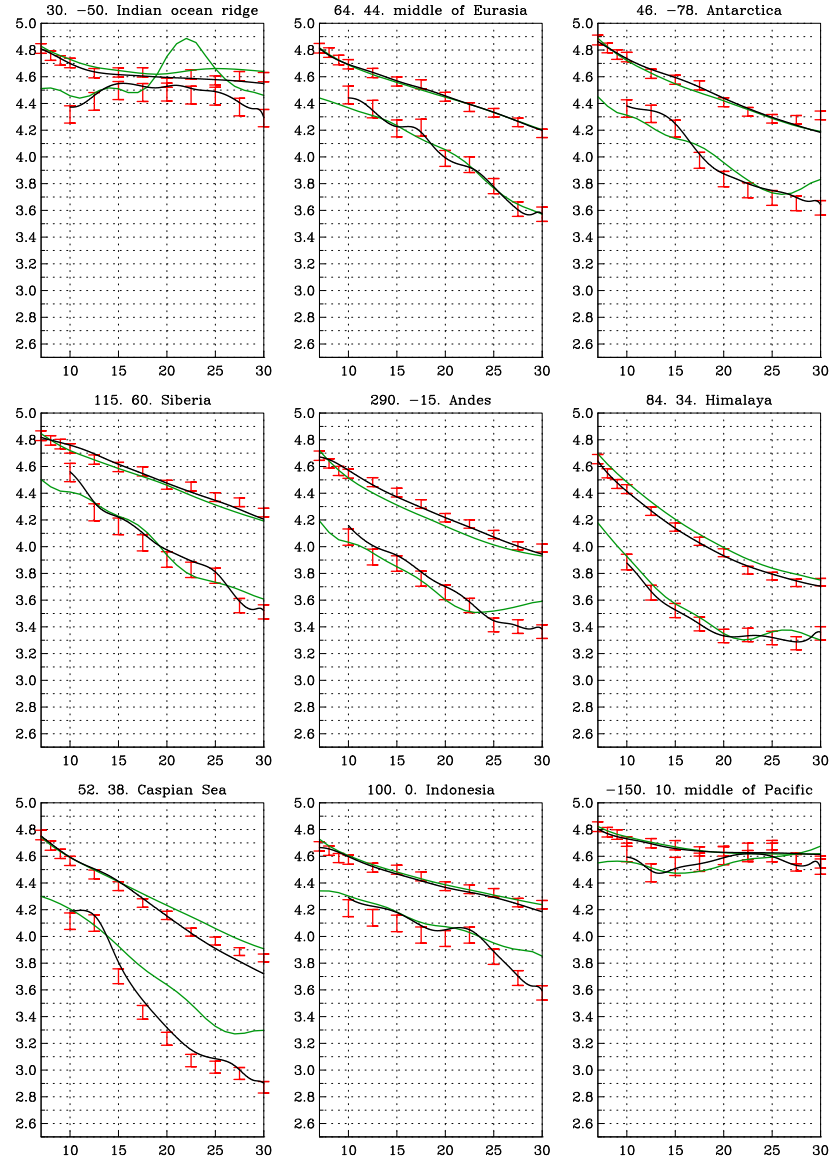


Figure 3.18: Same as Figure 3.17, but for Love waves.

stage where various terms that affect phase delays need to be corrected for. This method is very efficient and makes relative phase delay measurements from hundreds of stations at one time. It only takes about three to four days to make a dataset for one frequency. Surprisingly, the resulting phase delay dataset is larger than our group delay dataset. This is likely because of our use of a good starting 2D model, which allows us to pre-shift our waveform data to be better aligned and increase the chance to include records from different stations to form large clusters. In addition, at low frequencies, the envelope functions of surface waves are so broad that the envelopes are likely to be distorted by overtones and not included in our measurements.

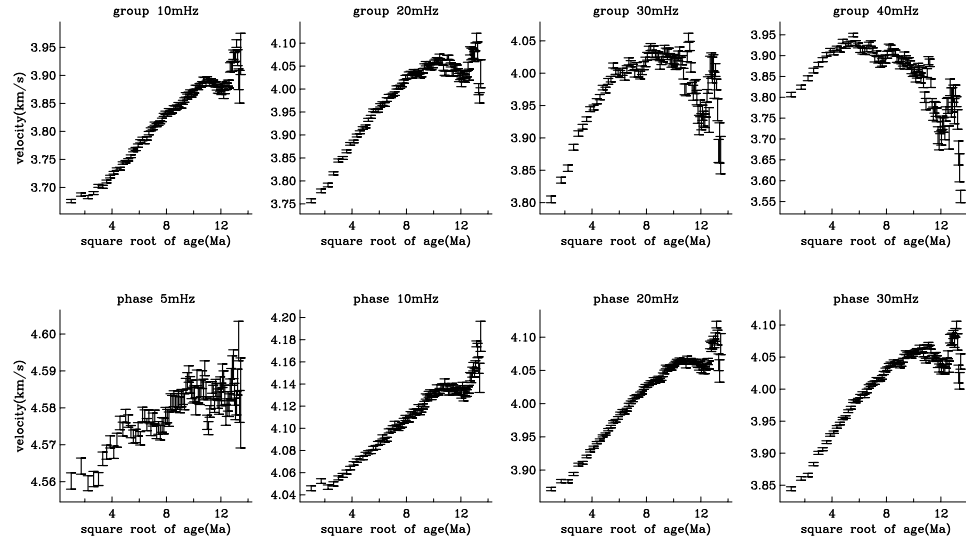


Figure 3.19: Rayleigh wave velocities as a function of the square root of seafloor ages. Data are for all ocean basins. The bin width is 2Ma. Error bars are the estimates of the errors of the mean velocities in each bin.

Similar to the findings in previous surface wave studies (e.g. Nishimura & Forsyth, 1989; Ritzwoller et al., 2004; Maggi et al., 2006; Burgos et al., 2014), an age-progression trend of both phase and group velocities in the ocean basins can be clearly seen in our dataset. We bin our Rayleigh wave velocities in all ocean basins from our spline model for every 2 million years of age and plot them as a function of the square root of age. (Figure 3.19) The seafloor age model is taken

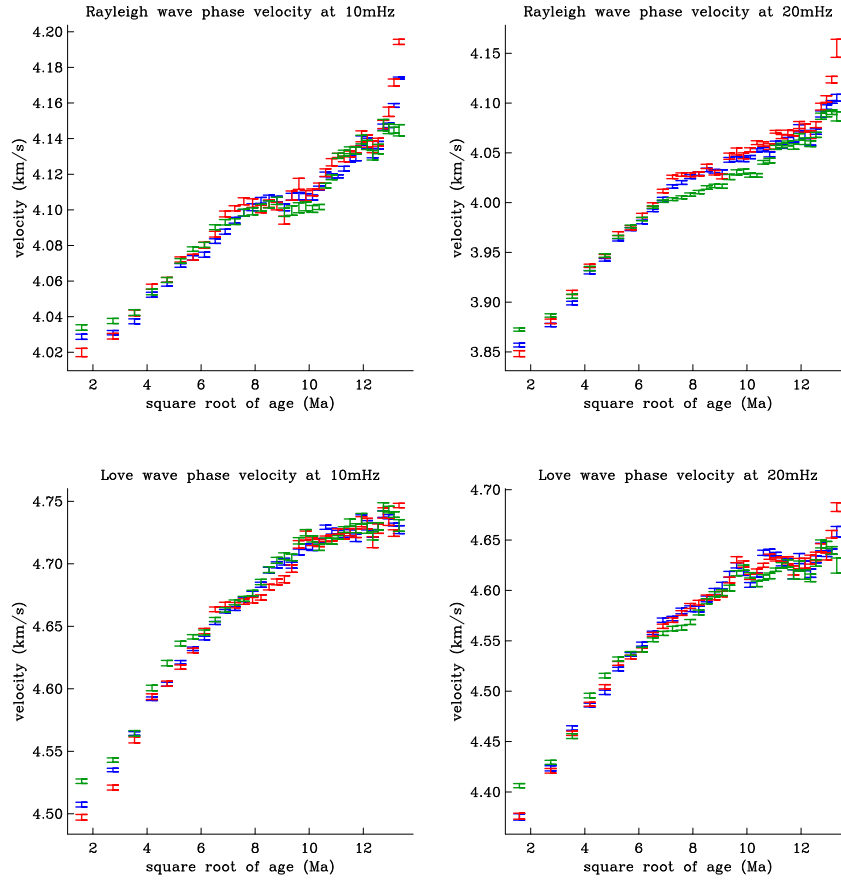


Figure 3.20: Comparisons of the age progression trend in the Pacific from different studies. Red: maps from this study; Blue: maps from Ekström (2011); Green: maps from Visser et al. (2008).

from Müller et al. (2008). The progression as a function of the square root of age is a prediction from the half-space cooling model (Turcotte & Schubert, 2002). Error bars are the estimates of the errors of the mean velocities in each bin. The small overall changes in 5 mHz phase velocities indicates that the cooling of the sea floor has a small effect below a depth of about 250 km (the peak sensitive depth for 5 mHz phase velocity). Note that the breakdown of the simple half-space cooling model (the deviation from a straight line in the figures) seems to happen earlier and earlier as frequencies increase. This can be seen more clearly in group velocities, which are more sensitive than phase velocities to shallow structure.

Binning data from the Pacific only gives the same trend, but a flattening of

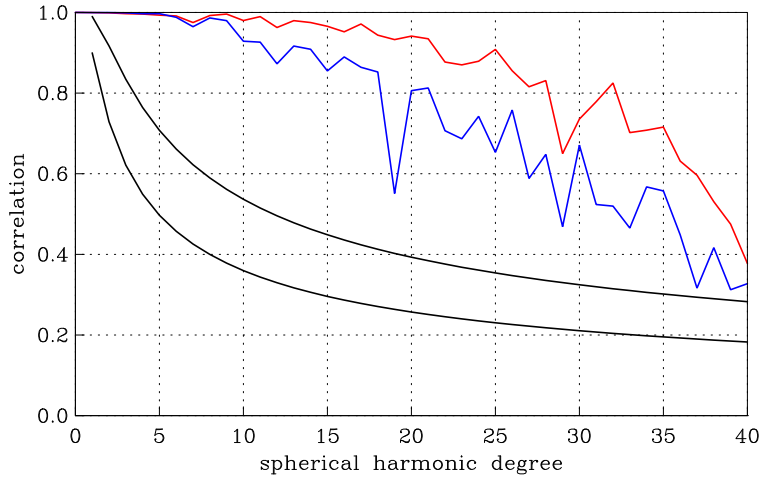


Figure 3.21: Correlations between maps shown in this chapter and the GDM52 model for group (blue) and phase (red) velocity for Rayleigh waves at 30mHz. The black lines correspond to the 90% and 99% confidence levels.

velocity in the central Pacific is present around 70 Ma. Figure 3.20 compares our results with the phase velocity maps from Ekström (2011) and Visser et al. (2008). While the velocity flattenings for Love waves are less consistent, the flattening for Rayleigh waves can be clearly seen in all three studies. The cause for this flattening may be a reheating process (Ritzwoller et al. 2004) or the variation of anisotropy structure in the Pacific (Ekström & Dziewonski 1998). The absence of such flattening in Maggi et al. (2006) is probably due to the difference in the inversion technique (invert directly for V_s vs. invert for phase velocity) and requires further study. The discussion of a physical model that explains our seismic observations requires sophisticated thermal dynamic calculations (e.g. Stixrude & Lithgow-Bertelloni, 2005; Goes et al., 2013) and is clearly beyond the scope of this chapter. But we believe that adding seismic constraints to seafloor topography and heat flow measurements will give us a more complete understanding of the seafloor cooling process.

The GDM52 model (Ekström 2011) is an excellent description of surface wave dispersion over a broad frequency range. The calculation of the ‘undisperse’

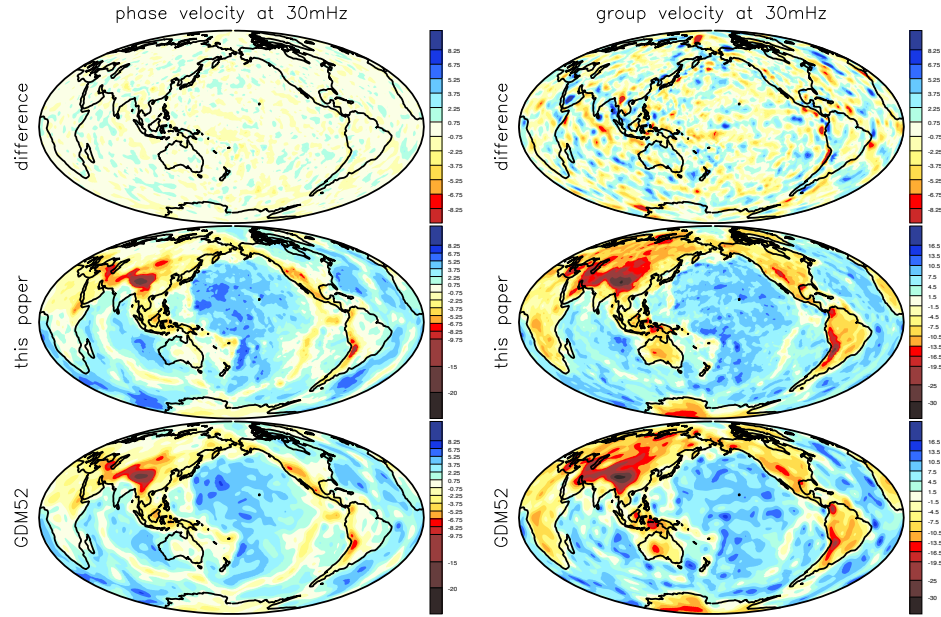


Figure 3.22: Direct comparisons between maps shown in this chapter and the GDM52 model for group and phase velocity for Rayleigh waves at 30mHz.

term in equation (3) requires a high quality starting dispersion model and GDM52 serves very well for this purpose. We compare our resulting maps presented in this chapter with the GDM52 model. An example for Rayleigh waves at 30 mHz is shown in Figure 3.21. Comparisons made at other frequencies share similar features. The correlation for phase velocity is well above the 99% confidence level for spherical degree l less than 38 (the nominal resolution for GDM52) and is above 0.9 for l less than 20 (corresponding to the resolution for the globe with 90% of the amplitude recovered from the checkerboard test). The correlation for group velocity is worse, as shown in Figure 3.17 and 3.18, but still above the 99% confidence level for all l . A direct comparison is shown in Figure 3.22, which highlights the differences between the two models and the improvement of resolution achieved by the work in this chapter.

Finally, it is interesting to compare the 1D averages from our spline models with PREM predictions (Figure 3.23). Such a comparison shows that the isotropic PREM model predicts our Rayleigh wave 1D averages very well but not the Love

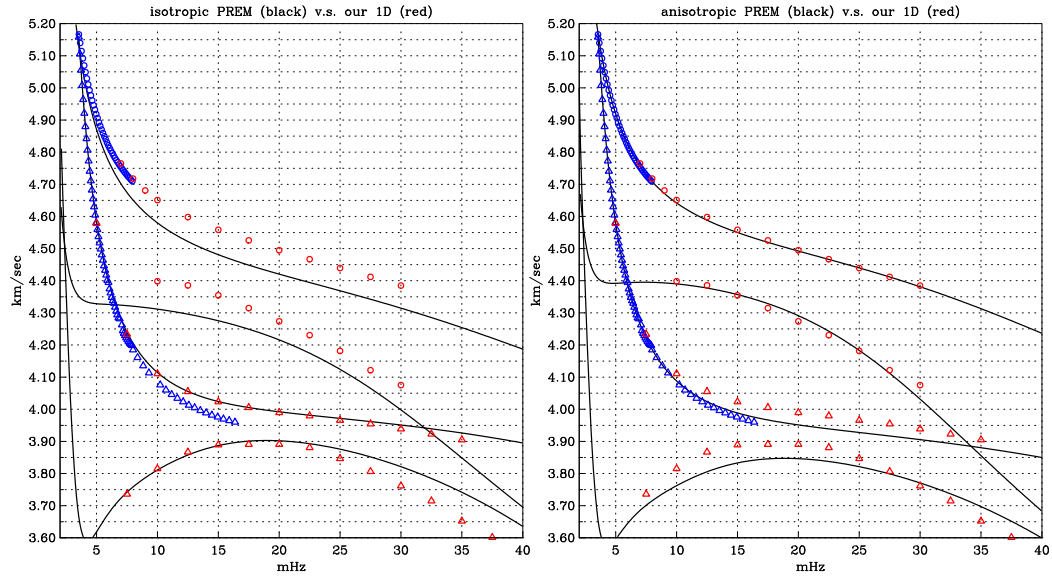


Figure 3.23: Comparison between the 1D averages from our model and PREM predictions. Red circles (for Love waves) and triangles (for Rayleigh waves) are our 1D averages. Blue circles and triangles are the data used to produce PREM (Dziewonski & Anderson, 1981). Black lines on the left panel show the predictions for both Love and Rayleigh group and phase velocities from the isotropic PREM model. Black lines on the right panels are the predictions from the anisotropic PREM model. We can see that the isotropic PREM model predicts our Rayleigh wave 1D averages very well but not the Love wave, while anisotropic PREM predicts Love waves very well but not the Rayleigh waves. It indicates that the transverse anisotropy in the shallow part of PREM is likely overestimated.

waves, while anisotropic PREM predicts Love waves very well but not the Rayleigh waves. This implies that the amplitudes of the transverse anisotropy in the shallow part of the Earth is probably only about half of the values given in PREM.

3.6 Conclusions

We extend our cluster analysis technique to measure phase delays and compile a large dataset of phase velocity measurements for Rayleigh and Love waves over a broad frequency range. While discussion of the details of azimuthal anisotropy is beyond the scope of this chapter, we demonstrates that including az-

Table 3.1: Detailed information for our phase velocity datasets and inversions. Each column is: wave type, frequency (mHz), number of measurements, 1D reference velocity (km/sec) used in inversion ($1/\bar{s}$ in equation 3.6), cut off value for outliers (seconds), χ^2/N for isotropic inversions, variance reduction, resolution for the globe with 60% of the amplitude recovered, resolution for the globe with 90% of the amplitude recovered, resolution for Eurasia with 60% of the amplitude recovered, resolution for Eurasia with 90% of the amplitude recovered, χ^2/N for anisotropic inversions and 1D averages from our spline model. Measurements with a symbol p are the frequencies we manually pick, while those without the symbol p are calculated automatically from nearby frequencies.

	freq	num	1D ref	cut	χ^2_{iso}/N	vred	G _{60%}	G _{90%}	E _{60%}	E _{90%}	χ^2_{ani}/N	1D avg
R	5	282861(p)	4.5791	40	2.14	0.663	9.0	10.	6.7	7.2	2.01	4.5781
R	7.5	478104(p)	4.2333	40	1.89	0.876	7.2	8.2	5.5	6.0	1.80	4.2334
R	10	555513(p)	4.1110	40	1.66	0.910	6.0	6.7	4.9	5.5	1.54	4.1106
R	12.5	609265	4.0548	40	1.73	0.945	6.7	7.2	4.9	5.5	1.58	4.0550
R	15	603048(p)	4.0230	40	1.84	0.956	6.7	7.8	5.1	5.5	1.71	4.0231
R	17.5	594520	4.0049	35	1.46	0.967	6.7	7.8	5.1	5.5	1.39	4.0052
R	20	400178(p)	3.9901	30	1.22	0.940	6.0	6.7	5.1	5.5	1.15	3.9896
R	22.5	525158	3.9773	30	1.08	0.979	6.7	7.2	5.5	6.0	0.95	3.9793
R	25	414328(p)	3.9670	30	1.33	0.970	5.6	6.7	5.5	6.0	1.24	3.9655
R	27.5	489399	3.9537	25	0.87	0.990	6.7	8.2	5.5	6.0	0.79	3.9543
R	30	320101(p)	3.9387	25	0.77	0.994	7.8	9.0	6.7	7.8	0.72	3.9387
R	32.5	403377	3.9229	25	0.77	0.995	7.8	9.0	6.7	7.8	0.70	3.9226
R	35	282349(p)	3.9049	25	0.78	0.996	9.0	10.	7.8	9.0	0.73	3.9048
L	7	136201(p)	4.7651	45	2.65	0.861	10.0	12.0	7.2	9.0		4.7650
L	8	176961	4.7177	40	2.16	0.900	9.0	10.6	6.7	7.8		4.7179
L	9	199698	4.6808	40	2.04	0.917	8.2	9.0	6.7	7.8		4.6811
L	10	194015(p)	4.6511	40	1.84	0.912	7.2	9.0	6.7	7.2		4.6512
L	12.5	240299	4.5980	40	1.75	0.941	7.8	9.0	6.7	7.2		4.5981
L	15	171613(p)	4.5580	35	1.60	0.953	10.0	10.6	7.2	7.8		4.5584
L	17.5	215801	4.5252	30	1.21	0.970	9.0	10.0	6.7	7.2		4.5254
L	20	145058(p)	4.4955	30	1.35	0.969	8.2	9.0	6.7	7.8		4.4948
L	22.5	219064	4.4668	30	1.09	0.988	9.0	10.0	6.7	7.8		4.4669
L	25	151528(p)	4.4399	30	1.47	0.989	9.0	10.6	7.8	10.0		4.4393
L	27.5	194450	4.4120	30	1.18	0.994	10.0	12.0	7.8	10.0		4.4118
L	30	108185(p)	4.3848	30	1.08	0.996	12.0	13.8	10.0	12.0		4.3847

imuthal anisotropy is necessary to invert for reliable isotropic phase velocity maps. Within a reasonable range, the choices of smoothness for anisotropy do not have a significant impact on the resulting isotropic maps. These maps provide a very compact way to represent our datasets and give over 90% of variance reduction for most frequencies for both Love and Rayleigh waves. A simple b-spline model can be used to simultaneously model both phase and group velocity perturbations, with uncertainties of 1.0% and 0.6% for Rayleigh wave group and phase velocity respectively, and 1.5% and 0.75% for Love waves. Large scale geologic features and clear age progression trends of both group and phase velocities in the ocean basins can be resolved. These maps provide useful constraints on the velocity of the crust and uppermost mantle and are currently being used to produce a new global lithosphere model. (Pasyanos et al., 2014)

The main product of this chapter, our phase delay measurements and the spline model for both Rayleigh and Love waves, can be downloaded from <http://igppweb.ucsd.edu/~gabi/litho1.0.html>.

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Appendix - Asymptotic forms for source phase corrections

This appendix explains the details of making source phase corrections. We follow the normal mode theory proposed in Gilbert & Dziewonski (1975) (henceforth GD). We use the same definitions of surface harmonics X_l^m and constants b_l^m as in GD equations (2) and (21)

$$X_l^m = (-1)^m \left[\frac{2l+1}{4\pi} \frac{(l-m)!}{(l+m)!} \right]^{1/2} P_l^m \cos(\theta)$$

$$b_l^m = \frac{(-1)^m}{2^m m!} \left[\frac{2l+1}{4\pi} \frac{(l+m)!}{(l-m)!} \right]^{1/2}$$

The second order asymptotic approximation to X_l^m is (equation (10) in Romanowicz & Roult (1986))

$$X_l^m = \frac{1}{\pi} \frac{1}{\sqrt{\sin\theta}} \cos \left[(l+1/2)\theta + m\pi/2 - \pi/4 + (m^2 - 1/4) \frac{\cot\theta}{2l+1} \right] \quad (3.17)$$

Therefore, for $m = 0, 1, 2$:

$$\begin{aligned} b_l^0 X_l^0 &= \frac{1}{\pi} \frac{1}{\sqrt{\sin\theta}} \sqrt{\frac{2l+1}{4\pi}} \cos \left[(l+1/2)\theta - \pi/4 - \frac{1}{4} \frac{\cot\theta}{2l+1} \right] \\ b_l^1 X_l^1 &= \frac{1}{\pi} \frac{1}{\sqrt{\sin\theta}} \sqrt{\frac{2l+1}{4\pi}} \frac{1}{2} \sqrt{(l+1)l} \sin \left[(l+1/2)\theta - \pi/4 + \frac{3}{4} \frac{\cot\theta}{2l+1} \right] \\ b_l^2 X_l^2 &= \frac{1}{\pi} \frac{1}{\sqrt{\sin\theta}} \sqrt{\frac{2l+1}{4\pi}} \frac{-1}{8} \sqrt{(l+2)(l+1)l(l-1)} \cdot \\ &\quad \cos \left[(l+1/2)\theta - \pi/4 + \frac{15}{4} \frac{\cot\theta}{2l+1} \right] \end{aligned}$$

For Love waves, the seismogram for each individual mode can be calculated as (equations (24), (31) in GD)

$$s_{\text{Love}} = -e_2 c_{12} b_l^2 X_l^{2'} - e_1 c_{11} b_l^1 X_l^{1'} \quad (3.18)$$

where

$$\begin{aligned}
e_1 &= W/r - W' \\
e_2 &= -4W/r \\
c_{11} &= 2(f_5 \cos \phi - f_4 \sin \phi) \\
c_{12} &= \sin 2\phi (f_3 - f_2) + 2 \cos 2\phi f_6
\end{aligned}$$

and W is the eigenfunction for each toroidal mode, f is the moment tensor solution and ϕ is azimuth of the receiver from an earthquake.

To compute the derivative of X_l^1 and X_l^2 , we use the following recursion relationship (which can be derived from Dahlen & Tromp (1998), eqn B.121 and the definition of b_l^m)

$$b_l^m X_l^{m'} = \frac{1}{2m} (l+m)(l-m+1) b_l^{m-1} X_l^{m-1} - m \cot \theta b_l^m X_l^m \quad (3.19)$$

Therefore, we have these exact identities:

$$\begin{aligned}
b_l^1 X_l^{1'} &= \frac{1}{2} l(l+1) b_l^0 X_l^0 - \cot \theta b_l^1 X_l^1 \\
b_l^2 X_l^{2'} &= \frac{1}{4} (l+2)(l-1) b_l^1 X_l^1 - 2 \cot \theta b_l^2 X_l^2
\end{aligned}$$

In most practices, the cot terms can be neglected for large l . However, for $l < 100$, we find that neglecting the cot terms can actually introduce a small but measurable bias.

$$\begin{aligned}
b_1 X_l^{1'} &= \left(\frac{1}{2} \sqrt{(l-1)(l+2)} X_l^2 - \frac{1}{2} \sqrt{(l+1)l} X_l^0 \right) \frac{-1}{2} \sqrt{\frac{2l+1}{4\pi}} \sqrt{(l+1)l} \\
&= -\frac{1}{4} \sqrt{\frac{2l+1}{4\pi}} \sqrt{(l+2)(l+1)l(l-1)} X_l^2 + \frac{1}{4} \sqrt{\frac{2l+1}{4\pi}} (l+1)l X_l^0 \\
&= \frac{1}{4} \sqrt{\frac{2l+1}{4\pi}} \frac{1}{\pi \sqrt{\sin \theta}} \sqrt{(l+2)(l+1)l(l-1)} \cdot \\
&\quad \cos \left[(l+1/2)\theta - \pi/4 + \frac{15}{4} \frac{\cot \theta}{2l+1} \right] + \\
&\quad \frac{1}{4} \sqrt{\frac{2l+1}{4\pi}} \frac{1}{\pi \sqrt{\sin \theta}} (l+1)l \cos \left[(l+1/2)\theta - \pi/4 - \frac{1}{4} \frac{\cot \theta}{2l+1} \right]
\end{aligned}$$

$$\begin{aligned}
b_2 X_l^{2'} &= \left(\frac{1}{2} \sqrt{(l-2)(l+3)} X_l^3 - \frac{1}{2} \sqrt{(l+2)(l-1)} X_l^1 \right) \\
&\quad \frac{1}{8} \sqrt{\frac{2l+1}{4\pi}} \sqrt{(l+2)(l+1)l(l-1)} \\
&= \frac{1}{16} \sqrt{\frac{2l+1}{4\pi}} \frac{1}{\pi \sqrt{\sin\theta}} \sqrt{(l+3)(l+2)(l+1)l(l-1)(l-2)} \cdot \\
&\quad \sin \left[(l+1/2)\theta - \pi/4 + \frac{35}{4} \frac{\cot\theta}{2l+1} \right] + \\
&\quad \frac{1}{16} \sqrt{\frac{2l+1}{4\pi}} \frac{1}{\pi \sqrt{\sin\theta}} (l+2)(l-1) \sqrt{(l+1)l} \cdot \\
&\quad \sin \left[(l+1/2)\theta - \pi/4 + \frac{3}{4} \frac{\cot\theta}{2l+1} \right]
\end{aligned}$$

One way to check these results is to use the fact that

$$\begin{aligned}
\frac{b_3}{b_2} &= -1/6 \sqrt{(l+3)(l-2)} \\
\frac{b_2}{b_1} &= -1/4 \sqrt{(l+2)(l-1)} \\
\frac{b_1}{b_0} &= -1/2 \sqrt{(l+1)l}
\end{aligned}$$

Therefore,

$$\begin{aligned}
b_1 X_l^{1'} &= \frac{1}{2} b_1 \sqrt{(l-1)(l+2)} X_l^2 - \frac{1}{2} \sqrt{(l+1)l} X_l^0 b_1 \\
&= -2b_2 X_l^2 + \frac{1}{4} l(l+1) b_0 X_l^0 \\
b_2 X_2^{1'} &= \frac{1}{2} b_2 \sqrt{(l-2)(l+3)} X_l^3 - \frac{1}{2} \sqrt{(l+2)(l-1)} X_l^1 b_2 \\
&= -3b_3 X_l^3 + \frac{1}{8} (l+2)(l-1) b_1 X_l^1
\end{aligned}$$

which match our earlier results for $b_1 X_l^{1'}$ and $b_2 X_2^{1'}$.

We further define $\text{corr} = \frac{\cot\theta}{2l+1}$, $\text{const} = \frac{1}{4} \sqrt{\frac{2l+1}{4\pi}} \frac{1}{\pi}$ and the amplitudes and

phases of the cosine and sine terms in $b_1 X_l^{1'}$ and $b_2 X_2^{1'}$:

$$\begin{aligned}
s_1 &= \text{const} \times l(l+1) \times e1 \times c_{11} \\
s_2 &= \text{const} \times \sqrt{(l+2)(l+1)l(l-1)} \times e1 \times c_{11} \\
s_3 &= 0.25\text{const} \times (l+2)(l-1)\sqrt{(l+1)l} \times e2 \times c_{12} \\
s_4 &= 0.25\text{const} \times \sqrt{(l+3)(l+2)(l+1)l(l-1)(l-2)} \times e2 \times c_{12} \\
\phi_1 &= -1/4\text{corr} = -0.25\text{corr} \\
\phi_2 &= 15/4\text{corr} = 3.75\text{corr} \\
\phi_3 &= 3/4\text{corr} = 0.75\text{corr} \\
\phi_4 &= 35/4\text{corr} = 8.75\text{corr} \\
\phi &= (l+1/2)\theta - \pi/4
\end{aligned}$$

Insert these into the equation for s_{Love} , we get

$$\begin{aligned}
(-\sqrt{\sin\theta})s_{\text{Love}} &= s_1 \cos(\phi + \phi_1) + s_2 \cos(\phi + \phi_2) + s_3 \sin(\phi + \phi_3) + s_4 \sin(\phi + \phi_4) \\
&= s_1(\cos\phi \cos\phi_1 - \sin\phi \sin\phi_1) + s_2(\cos\phi \cos\phi_2 - \sin\phi \sin\phi_2) \\
&\quad + s_3(\sin\phi \cos\phi_3 + \cos\phi \sin\phi_3) + s_4(\sin\phi \cos\phi_4 + \cos\phi \sin\phi_4) \\
&= (s_1 \cos\phi_1 + s_2 \cos\phi_2 + s_3 \sin\phi_3 + s_4 \sin\phi_4) \cos\phi \\
&\quad - (s_1 \sin\phi_1 + s_2 \sin\phi_2 - s_3 \cos\phi_3 - s_4 \cos\phi_4) \sin\phi
\end{aligned}$$

If we define $\tilde{s}_1 = s_1 \cos\phi_1 + s_2 \cos\phi_2 + s_3 \sin\phi_3 + s_4 \sin\phi_4$ and $\tilde{s}_2 = s_1 \sin\phi_1 + s_2 \sin\phi_2 - s_3 \cos\phi_3 - s_4 \cos\phi_4$, seismogram

$$s_{\text{Love}} = \frac{-1}{\sqrt{\sin\theta}} (\tilde{s}_1 \cos\phi - \tilde{s}_2 \sin\phi) = -\frac{\sqrt{\tilde{s}_1^2 + \tilde{s}_2^2}}{\sqrt{\sin\theta}} \cos(\phi + \phi_0)$$

Therefore, Love wave seismogram can be treated as a cosine function with initial source phase $\phi_0 = \arctan(\tilde{s}_2/\tilde{s}_1)$.

For Rayleigh waves, the seismogram can be written as (equation 30 in GD)

$$s_{\text{Rayleigh}} = [e_1 f_1 + e_2 (f_2 + f_3)] b_0 X_l^0 + e_3 c_{11} b_1 X_l^1 + e_4 c_{12} b_2 X_l^2$$

where

$$\begin{aligned}
e_1 &= U' \\
e_2 &= \frac{1}{r}(U - \frac{1}{2}l(l+1)V) \\
e_3 &= V' + \frac{1}{r}(U - V) \\
e_4 &= \frac{2V}{r} \\
c_{11} &= 2(f_4 \cos \phi + f_5 \sin \phi) \\
c_{12} &= 2 \cos 2\phi (f_2 - f_3) + 4 \sin 2\phi f_6
\end{aligned}$$

For Rayleigh waves, the calculation for source phase is much easier than Love because we do not need to calculate derivatives. We define $\text{corr} = \frac{\cot \theta}{2l+1}$, $\text{const} = \sqrt{\frac{2l+1}{4\pi}} \frac{1}{\pi}$ and

$$\begin{aligned}
s_1 &= (e_1 f_1 + e_2 (f_2 + f_3)) \times \text{const} \\
s_2 &= e_4 c_{12} \frac{-1}{8} \sqrt{(l+2)(l+1)l(l-1)} \times \text{const} \\
s_3 &= e_3 c_{11} \frac{1}{2} \sqrt{(l+1)l} \times \text{const} \\
s_4 &= 0 \\
\phi_1 &= -0.25 \text{corr} \\
\phi_2 &= 3.75 \text{corr} \\
\phi_3 &= 0.75 \text{corr} \\
\phi_4 &= 0
\end{aligned}$$

The rest of the calculation is the same as for Love waves.

Using the second order term $(m^2 - 1/4) \frac{\cot \theta}{2l+1}$ in the approximation to X_l^m has minor but perceivable effects (1-2 seconds) at frequencies lower than 10mHz, or l less than about 100. We compute synthetics using only the fundamental modes, process them with the same preprocessing technique described in this chapter and identify the arrival times of the largest peak in each waveform. (Figure 3.24). Including the second order term reduces the scatters in Rayleigh wave measurements. For Love waves, it corrects for the dependence of arrival times on distance, which will generate a bias of about 0.1% ($5\text{sec}/(160\text{deg}/4.76\text{km/sec}) \approx 0.1\%$) in the average velocity. This correction is needed for our measurement technique, which

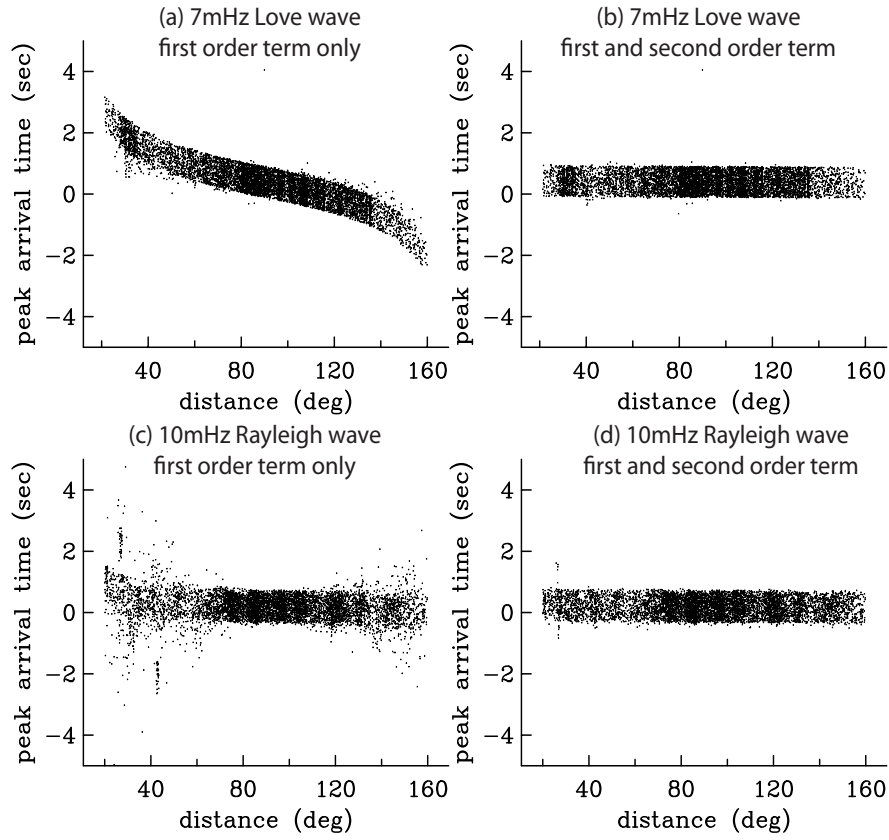


Figure 3.24: Synthetic test for our measurement technique. Synthetic seismograms are computed for fundamental modes only. Each dot represents one measurement. Including the second order term reduces the scatter for Rayleigh waves and corrects for the dependance of arrival times on distances for Love waves.

compares real waveforms measured at different stations. But methods based on comparing waveforms with synthetics do not need this correction, because accurate calculation of X_l^m is already part of the algorithm that calculates the synthetics.

Chapter 4

Effect of earthquake locations on Rayleigh wave azimuthal anisotropy

Abstract

Various factors need to be considered when inverting for surface wave azimuthal anisotropic structure. This chapter focuses on the 2ϕ terms for Rayleigh wave azimuthal anisotropy and show that the uncertainties of earthquake locations also have significant impacts on the resulting anisotropic structure. We use the global Rayleigh wave phase velocity dataset collected in a previous study to demonstrate this effect. The differences between azimuthal anisotropic patterns with and without source relocations are greatest near plate boundaries. Large differences around the South American plate are also identified. Although most of the earthquakes are shifted by less than 15 km from the CMT locations, earthquakes near the Andes can be systematically shifted by more than 30 km. Our final epicenters for earthquakes on ridge-transform fault systems better match the plate boundaries.

4.1 Introduction

Seismic anisotropy has long been observed in the Earth’s upper mantle. Radial anisotropy, which means seismic waves polarized in the radial direction travel at a speed different from waves polarized in the horizontal direction, was inferred from the discrepancy between the Rayleigh and Love wave speed (e.g. Anderson, 1961; Aki & Kaminuma, 1963) and has been widely accepted in 1D reference Earth models (e.g. Dziewonski & Anderson, 1981; Kustowski et al., 2008). Azimuthal anisotropy, which is another type of anisotropy and characterizes the variability of seismic wave speeds in terms of propagation directions, was later observed for both body waves (Hess, 1964) and surface waves (Forsyth, 1975). Smith & Dahlen (1973) and Montagner & Nataf (1986) first derived a consistent mathematical framework to interpret these two types of anisotropy and Montagner & Tanimoto (1991) showed the first global upper mantle tomographic maps of anisotropy. The more recent observations of shear-wave splittings (see reviews from Silver, 1996; Savage, 1999; Long & Silver, 2009) have further improved our understanding of azimuthal anisotropy and are the most conclusive evidence for an anisotropic mantle.

Azimuthal anisotropy is less well constrained than radial anisotropy. Most of the shear-wave splitting measurements can only be made directly beneath seismic stations and have very limited lateral coverage and depth resolution. In contrast, surface waves provide almost complete global coverages and have been widely used to study global azimuthal anisotropic patterns (e.g. Tanimoto & Anderson, 1985; Laske & Masters, 1998; Trampert & Woodhouse, 2003; Debayle et al., 2005; Visser et al., 2008; Ekström, 2011). However, the agreement among these different surface-wave derived models is not satisfactory. In addition, the compatibility between surface-wave derived anisotropy and shear-wave splitting observations is also poor (Montagner et al., 2000).

Various factors can affect the derived anisotropic structure, e.g. the data coverage, the wave propagation theory used (Sieminski et al., 2007), the details of the inversion techniques (Simons et al., 2002), etc. This chapter focuses on the 2ϕ terms of Rayleigh waves and shows that the uncertainties of earthquake

locations, which have been generally ignored in most previous studies, also have significant impacts on the resulting azimuthal anisotropy. We first present the theory explaining how to include the location correction terms in the inversion and how to implement the smoothness constraints on the azimuthal anisotropic terms under our block parameterization. We then use the datasets described in Ma et al. (2014) to invert for azimuthal anisotropy and earthquake relocations.

4.2 Theory

4.2.1 Including errors of earthquake locations

We divide the surface of the Earth into 41252 equal area blocks with the size of 1° by 1° at the equator. For each block, the slowness can be written as (Smith & Dahlen, 1973)

$$\delta s = \delta s_0 + A_1 \cos 2\phi + A_2 \sin 2\phi \quad (4.1)$$

where ϕ is the azimuth of the ray path direction, measured clockwise from the North. The total perturbation of slowness δs is the sum between the contributions from the isotropic term δs_0 and the azimuthal anisotropic terms A_1 and A_2 . The amplitude of the azimuthal anisotropy A and the fast direction α can be calculated using

$$A = \sqrt{A_1^2 + A_2^2} \quad (4.2)$$

$$\alpha = \arctan(A_2/A_1)/2 + \pi/2 \quad (4.3)$$

The $\pi/2$ term arises because we are working with slowness instead of velocity.

For Rayleigh waves, Ekström (2011) has demonstrated conclusively that including the 2ϕ terms is necessary to obtain reliable isotropic tomography maps. However, there is some disagreement on whether or not we can resolve the 4ϕ terms for Rayleigh waves (e.g. Laske & Masters, 1998; Trampert & Woodhouse, 2003; Ekström, 2011). We will focus on the 2ϕ terms for Rayleigh waves in this chapter

for simplicity and show how earthquake (mis)locations can affect the resulting azimuthal anisotropic patterns.

Errors of earthquake locations (latitudes and longitudes) affect the epicentral distance Δ . From trigonometry, we have

$$\cos \Delta = \cos \theta_0 \cos \theta_1 + \sin \theta_0 \sin \theta_1 \cos(\phi_1 - \phi_0) \quad (4.4)$$

where θ_0 and ϕ_0 are the colatitude and longitude of the source and θ_1 and ϕ_1 are the colatitude and longitude of the receiver respectively. Take the derivatives of both sides and we have

$$\frac{\partial \Delta}{\partial \theta_0} = \frac{-1}{\sin \Delta} [-\sin \theta_0 \cos \theta_1 + \cos \theta_0 \sin \theta_1 \cos(\phi_1 - \phi_0)] \quad (4.5)$$

$$\frac{\partial \Delta}{\partial \phi_0} = \frac{-1}{\sin \Delta} [\sin \theta_0 \sin \theta_1 \sin(\phi_1 - \phi_0)] \quad (4.6)$$

Therefore, to the first order, travel time anomalies δt including errors of earthquake locations can be written as

$$\delta t = s_{\text{ref}} \left[\int_0^\Delta \frac{\delta s}{s_{\text{ref}}} \delta \gamma + \frac{\partial \Delta}{\partial \theta_0} \delta \theta_0 + \frac{\partial \Delta}{\partial \phi_0} \delta \phi_0 \right] \quad (4.7)$$

where s_{ref} is the reference slowness at a certain frequency and the integral is done along the minor arc connecting the source and receiver.

4.2.2 Implementing smoothness constraints on azimuthal anisotropy

Various authors (Boschi & Woodhouse, 2006; Ekström, 2006) point out that applying smoothness constraints directly on the A_1 and A_2 terms generates undesirable results near the poles and they propose different methods to solve this problem. The key to their solutions is to find the appropriate combinations of A_1 and A_2 and apply smoothness constraints on these combinations, instead of on A_1 and A_2 themselves. Here, we adopt the concept of ‘locally parallel azimuth’ proposed in Ekström (2006) and apply it to our block parameterization. A schematic diagram is shown in Figure 4.1 to demonstrate the idea. Let Pc denote the center

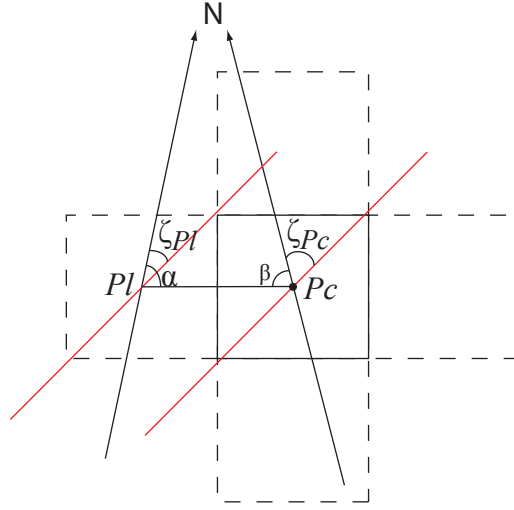


Figure 4.1: A schematic diagram explaining the concept of ‘locally parallel azimuth’. P_c and P_l are the center points of two adjacent blocks. The two black lines point to the geographical North directions. The two red lines are locally parallel in this diagram.

point of the block in the middle and P_l the center point of the block on its left. The solid lines point to the geographical North directions at these two points. The two red lines are ‘locally parallel’ with geographical azimuth ζ_{P_c} and ζ_{P_l} at points P_c and P_l respectively. α and β are the azimuths (measured from the North) from P_l to P_c and from P_c to P_l respectively. From geometry (note that β is negative in the figure), we have (Ekström, 2006, equation 6)

$$\zeta_{P_c} = \zeta_{P_l} - (\alpha - \beta - \pi) \quad (4.8)$$

Along the locally defined North direction at point P_c , we set $\zeta_{P_c} = 0$ and we have

$$\zeta_{Pl} = \alpha - \beta - \pi \quad (4.9)$$

$$\delta s_c(\phi = \zeta_{Pc}) = A_1 \cos(2\zeta_{Pc}) + A_2 \sin(2\zeta_{Pc}) \quad (4.10)$$

$$= A_1 \quad (4.11)$$

$$\delta s_l(\phi = \zeta_{Pl}) = A_1 \cos(2\zeta_{Pl}) + A_2 \sin(2\zeta_{Pl}) \quad (4.12)$$

$$= A_1 \cos[2(\alpha - \beta)] + A_2 \sin[2(\alpha - \beta)] \quad (4.13)$$

where δs_c and δs_l are the anisotropic perturbations at points Pc and Pl respectively. The smoothness of the velocity in the locally defined North direction is then calculated using the standard discretized Laplacian operator:

$$D\delta s_N = 4\delta s_c - (\delta s_l + \delta s_r + \delta s_a + \delta s_b) \quad (4.14)$$

where δs_a and δs_b and δs_r are the slowness perturbations of the blocks above, below and on the right of the centered block respectively, calculated in the locally defined North direction.

Similarly, in the locally defined Northeast direction, we set $\zeta_{Pc} = \pi/4$ and we have

$$\zeta_{Pl} = \alpha - \beta - 3\pi/4 \quad (4.15)$$

$$\delta s_c(\phi = \zeta_{Pc}) = A_2 \quad (4.16)$$

$$\delta s_l(\phi = \zeta_{Pl}) = -A_1 \sin[2(\alpha - \beta)] + A_2 \cos[2(\alpha - \beta)] \quad (4.17)$$

The smoothness can then be calculated as before. Constraining the smoothness in the East direction gives the same set of equations as constraining it in the North direction because of the 2ϕ periodicity.

In summary, we seek to minimize f in our inversion

$$f = ||Am-d||^2 + \lambda ||D\delta s_0||^2 + \lambda_1 ||D\delta s_N||^2 + \lambda_1 ||D\delta s_{NE}||^2 + \lambda_2 ||\delta\theta||^2 + \lambda_2 ||\delta\phi||^2 \quad (4.18)$$

where A is the numerical implementation of the first term in equation (4.7), m corresponds to δs , d corresponds to the data δt , D is the Laplacian operator and δs_N and δs_{NE} are the slowness perturbations in the North and Northeast directions.

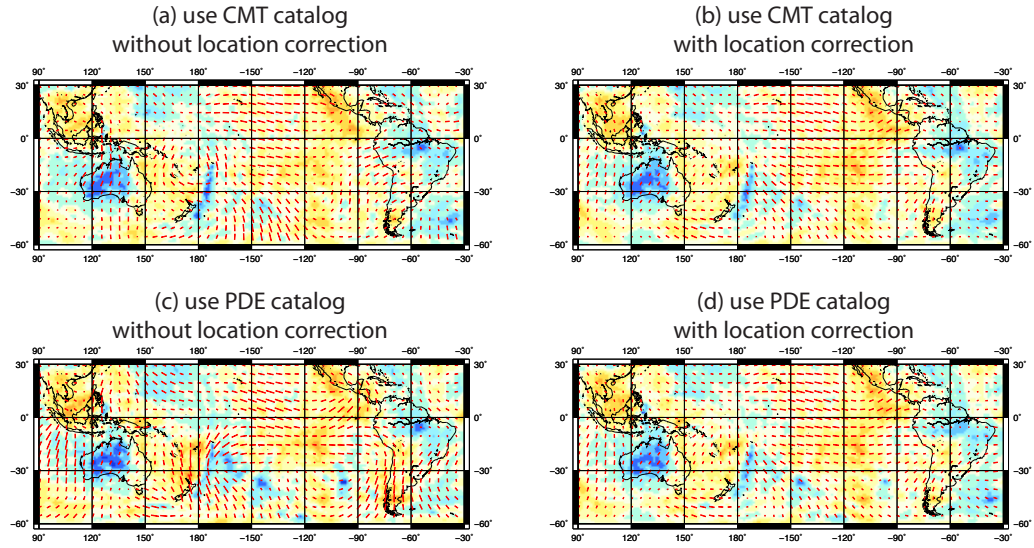


Figure 4.2: A test of our inversion algorithm including the corrections for earthquake locations. The data used here are the 10mHz Rayleigh wave measurements described in Ma et al. (2014). The original dataset uses CMT locations. We reprocess it using the PDE locations. The background color is the isotropic velocity perturbation. The direction and length of each bar represent the fast direction and the amplitude of anisotropy. Compare (a) and (c), the azimuthal anisotropy patterns are different when we use different catalogs. Compare (b) and (d), allowing changes in earthquake locations (latitudes and longitudes) in our inversion greatly reduces such discrepancies.

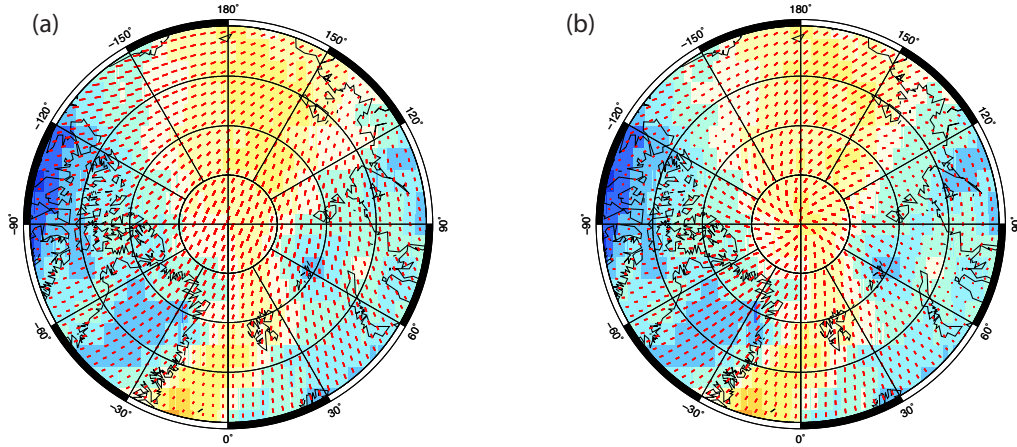


Figure 4.3: Azimuthal anisotropy patterns near the North Pole. (a) Inversion result with smoothness constraints on δs_N and δs_{NE} in the locally parallel direction. (b) Inversion result with smoothness constraints on A_1 and A_2 .

4.2.3 An example for 10mHz Rayleigh waves

We first test our algorithm with the measurements of 10 mHz Rayleigh wave described in Ma et al. (2014). The earthquake locations used in their paper are the centroid locations from the Global Centroid-Moment-Tensor (CMT) catalog (Dziewonski et al., 1981; Ekström et al., 2012). We first apply their automatic measuring technique to redo the measurements using the earthquake locations in the Preliminary Determination of Epicenters (PDE) bulletin (downloaded from <ftp://hazards.cr.usgs.gov/pde/>). We then compare the anisotropic patterns inverted from this new dataset, which uses PDE locations, with the original one, which uses CMT locations.

We fix $\lambda_1 = 30$ in this experiment (see below for this choice). In the first case, we do not allow changes in earthquake locations and set $\lambda_2 = 10000$. While the patterns of azimuthal anisotropy are quite similar away from earthquake epicenters, differences can be seen around the subduction zones, especially in the Tonga and southern Andes region (Figure 4.2 (a) and (c)). However, when we set $\lambda_2 = 0$ to allow earthquake locations to change freely, the resulting azimuthal anisotropic patterns are nearly identical (Figure 4.2 (b) and (d)). It is worth

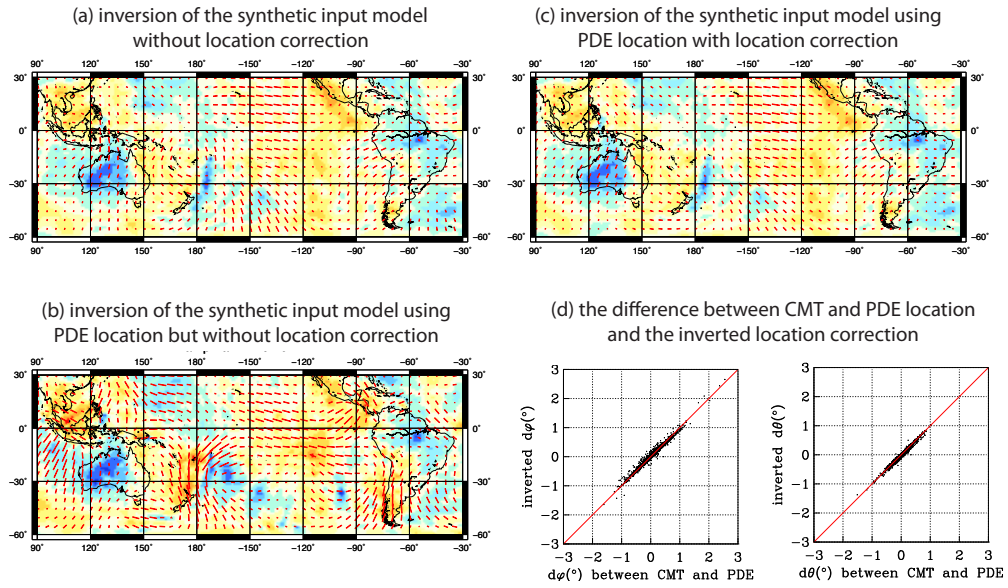


Figure 4.4: A test of our inversion algorithm including the corrections for earthquake locations using synthetic data. The input model is the inversion result for 10mHz Rayleigh wave using CMT location (the one in Figure 4.2a). Synthetic data are generated by forward calculating the phase arrival times from this model. Inverting this synthetic dataset gives a model that is almost identical to the input model (compare panel a and Figure 4.2a). Inversion using the same dataset but assuming PDE locations gives a model that is quite different from the input model (panel b). Adding location corrections to the inversion greatly reduces such discrepancies (panel c). Earthquake locations can also be recovered (panel d).

pointing out that the final results are closer to the result inverted from the dataset using the CMT catalog than the one using the PDE catalog. This reflects the fact that the CMT catalog is mostly constrained by long period data and is more suitable for surface wave analysis.

Results inverted from the dataset using the CMT catalog with location corrections are also plotted around the North Pole (Figure 4.3). Constraining the smoothness in the locally parallel directions δs_N and δs_{NE} gives reasonably smooth results around the North Pole. As a comparison, if we apply smoothness constraints directly on A_1 and A_2 , namely we minimize $\|DA_1\|^2$ and $\|DA_2\|^2$, we will get unphysical results.

To further demonstrate the effect of earthquake locations, we carry out a synthetic test (Figure 4.4). We use the model inverted from the dataset using CMT location but without location corrections (Figure 4.2a) as our synthetic input model. To generate the synthetic dataset, we then use equation (4.7) ($\delta\theta_0$ and $\delta\phi_0$ are set to be 0) to forward calculate the phase arrival time anomalies. The source and station distributions are the same as in the 10 mHz Rayleigh wave dataset. Inversion of this dataset without location corrections gives a model (Figure 4.4a) that is almost identical to the input model (Figure 4.2a), but with a slightly smaller amplitude of anisotropy due to smoothing. Inversion using the same dataset but assuming PDE locations gives a result (Figure 4.4b) that is quite different from the input model, but similar to the patterns seen in Figure 4.2c. Adding location corrections to the inversion greatly reduces such discrepancies (Figure 4.4c). In addition, the inverted location corrections are also close to the differences between the CMT and PDE catalogs (Figure 4.4d), which means that earthquake locations can be reliably recovered using our method.

4.3 Inversion and results

The dataset described in Ma et al. (2014) is used for our inversions. We use data from multiple frequencies at the same time in the inversion, so the earthquake locations can be constrained by data from different frequencies and hence have

smaller errors. We show here the results when we use the measurements at 10, 20 and 30 mHz for the inversions. The standard LSQR method (Paige & Saunders, 1982) is used to minimize f defined in equation (4.18). At each iteration, we do not allow earthquakes to shift too much from their starting locations by penalizing the size of $\delta\theta_0$ and $\delta\phi_0$. We calculate the L-curves (Parker, 1994) for different combinations of λ_1 and λ_2 to decide their appropriate values. Figure 4.5 shows the process of deciding these parameters for the first iteration. We do this in two steps. In the first step, we fix λ_1 to be 10 and 300, which is a reasonable range for λ_1 . We find that $\lambda_2 = 15$ gives a reasonable balance between the fit to the data and the mean size of the earthquake relocation vectors ($||\delta\theta||^2 + ||\delta\phi||^2$) for both $\lambda_1 = 10$ and 300. In the second step, we fix $\lambda_2 = 15$ and the L-curve indicates that using a λ_1 around 25 is the best. Because phase-skipping may occur for high-frequency measurements if changes of locations are too large, we redo our measurements as described in Ma et al. (2014) using the new locations and redo the inversion after each iteration. We find that earthquake locations converge very quickly and the changes of the locations are typically less than 5 km after three iterations. This is below the location error (see below) so we stop after three iterations.

Figure 4.6 shows the resulting isotropic and azimuthal anisotropic perturbations for each frequency. The isotropic patterns are explained elsewhere (Ma et al., 2014). The dominant anisotropic structure in the eastern Pacific, with a nearly East-West fast direction, is a common feature seen since the earliest studies (e.g. Montagner & Tanimoto, 1990; Laske & Masters, 1998; Trampert & Woodhouse, 2003; Ekström, 2011). A change in the fast directions to NE-SW at high frequencies in the central Pacific near the Hawaii seamount chain is also observed in more recent studies where high frequency data are used (e.g. Smith et al., 2004; Ekström, 2011), suggesting that the fast directions at high frequencies tend to align with the fossil seafloor spreading directions (Nishimura & Forsyth, 1988; Smith et al., 2004).

The results of earthquake relocations are shown in Figure 4.7(a) and (b). Most of the earthquakes are shifted by about 10-20 km. To investigate the systematic behavior of earthquake relocations at different places, we find it useful to

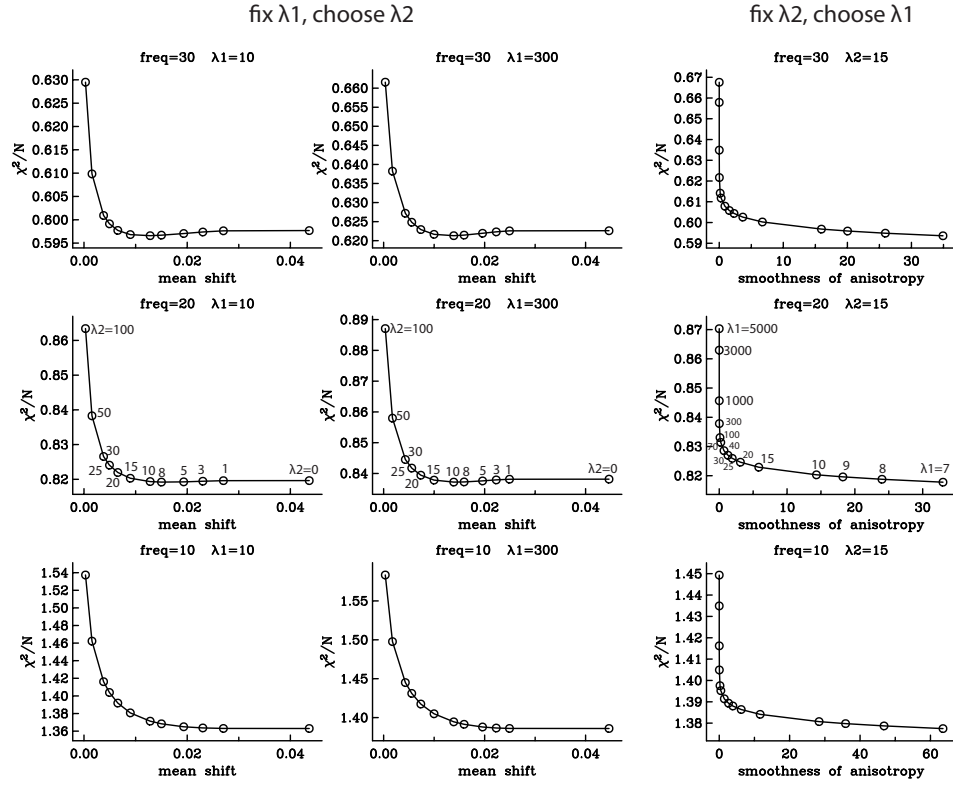


Figure 4.5: L-curves showing our process to determine the appropriate λ_1 and λ_2 in the inversion. Each row represents one frequency. The two left columns demonstrate the process of choosing λ_2 , which penalizes the size of earthquake relocations, for fixed $\lambda_1 = 10$ and 300 . The y-axis represents data misfit. The x-axis represents the mean size of earthquake relocation vector ($\|\delta\theta\|^2 + \|\delta\phi\|^2$). The column on the right demonstrates the process of choosing λ_1 after λ_2 is chosen. The x-axis represents the smoothness of the anisotropic terms ($\|Dx\|^2$)

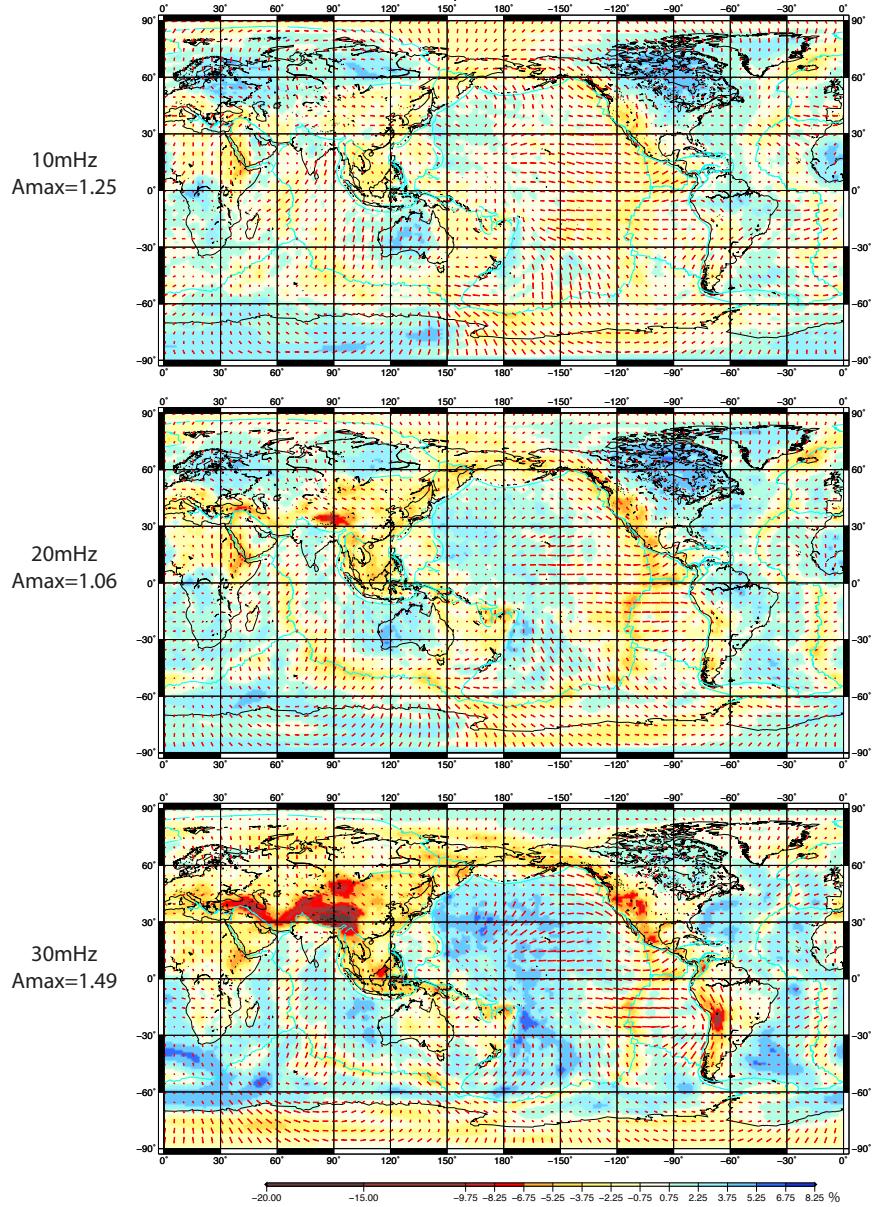


Figure 4.6: Inversion results for 10, 20 and 30 mHz Rayleigh waves. The background colors indicate the perturbations of isotropic phase velocity with respect to the global average value for each frequency. The lengths of the red bars indicate the amplitudes of azimuthal anisotropy ($A = \sqrt{A_1^2 + A_2^2}$). The directions of the bars show the fast directions. The maximum amplitude of azimuthal anisotropy, which corresponds to the longest bar at each frequency, is shown to the left of each plot.

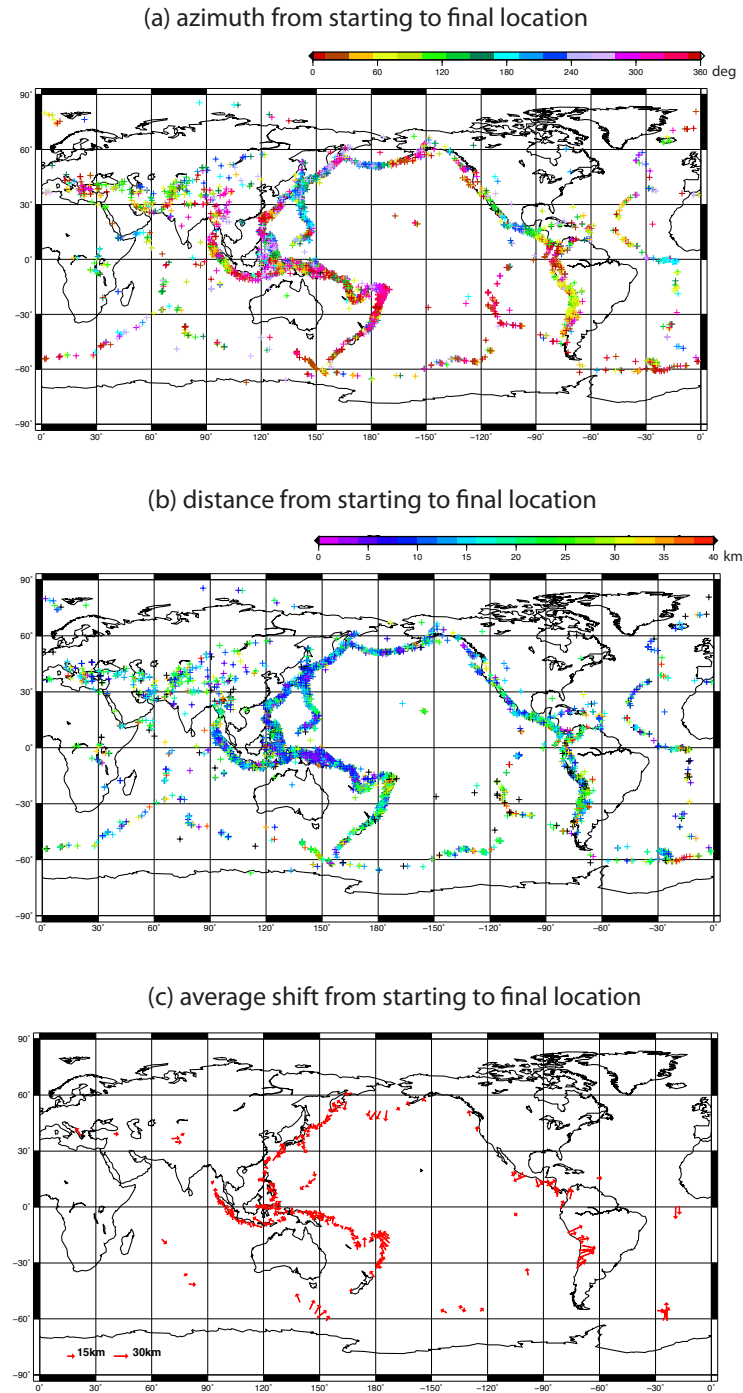


Figure 4.7: Results for source relocations. Each cross in (a) and (b) represents one earthquake. The arrows in (c) represent the average shifts of earthquakes within each 2° by 2° block.

also calculate the average shifts of earthquakes within every 2° blocks. We discard blocks having less than 5 earthquakes to avoid statistical biases. We see that the majority of the average shifts are less than 15 km (Figure 4.7(c)). This means systematic changes from the CMT locations are generally small in most places.

However, earthquakes in the Andes can be systematically shifted in the down-dip directions for over 30 km. A similar location error is also reported in Hjörleifsdóttir & Ekström (2010) as a result of the inaccuracy of the 3D Earth velocity structure. It is interesting to note that earthquakes are also systematically relocated to the down-dip regions in other subduction zones, e.g. Tonga, New Hebrides, Philippines and the northern section of the Kurile subduction zone, and to some lesser extent in the Indonesia and Mariana trenches. But not all subduction earthquakes share this behavior (e.g. earthquakes in the Japan and Aleutians trenches; Figure 4.7(a) and (c)). An explanation for this systematic behavior is that the isotropic velocity perturbations in the back-arc regions (e.g. the mountain belts and back-arc basins) are probably underestimated (i.e. not slow enough) in the Earth model used in the CMT analysis.

We also look at the changes of earthquake locations on ridge-transform fault systems, because earthquakes are expected to happen right on the plate boundaries in these regions. In general, the changes are small except for a few places in the Southern Oceans and one on the Atlantic ridge. Figure 4.8 shows places where such changes are the largest. Data for the plate boundaries are taken from DeMets et al. (1994) and Coffin et al. (1998). After our inversions, most earthquakes are now indeed shifted closer to the plate boundaries.

We have performed checkerboard tests to evaluate the resolution of azimuthal anisotropy. The input models have only the azimuthal anisotropic terms A_1 and A_2 . The isotropic velocity perturbations δs_0 and the earthquake relocation parameters $\delta\theta_0$ and $\delta\phi_0$ are all set to be zero. The recovered patterns for 10 mHz are shown in Figure 4.9. The leakage from anisotropy to earthquake locations is very small, mostly less than 3 km. The test also suggests possible trade-offs between isotropic and anisotropic components. This trade-off is best shown in Tanimoto & Anderson (1985). However, it does not seem as bad as this earlier

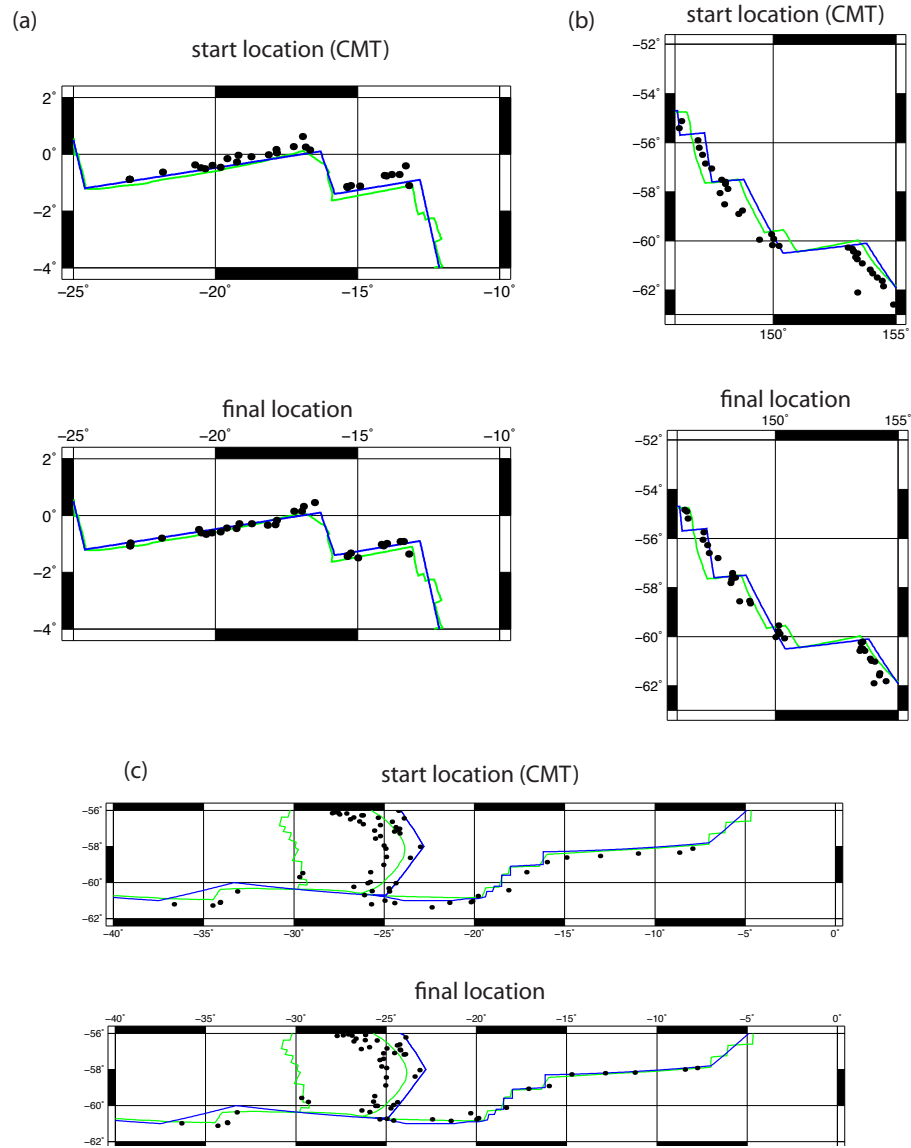


Figure 4.8: Comparisons between earthquake locations in the CMT catalog and our inverted locations in some ridge-transform fault systems. The plate boundaries are from Coffin et al. (1998) (green) and DeMets et al. (1994) (blue). Panel (a) shows a section of the central Mid-Atlantic ridge; Panel (b) shows a region south of Australia and west of Macquarie Island; Panel (c) shows a region near the South Sandwich Islands.

study because there are now much more data available. The leakage only occurs in places where outgoing ray paths always lie in the fast (or slow) directions of all the surrounding blocks. This scenario is possible but likely to be rare for the real Earth.

From the checkerboard tests, we can see that, even though the dataset in Ma et al. (2014) has used all LH data available in the IRIS DMC from 1988-2007, it is still difficult to resolve features of azimuthal anisotropy with a scale less than 20° in the oceans. However, our inversion results show that oceanic regions actually have clear and large azimuthal anisotropy signals, especially in the East Pacific. This suggests that the fast directions in the oceans are likely to be coherent over very large distances ($> 20^\circ$). If changes of the fast directions exist on smaller scales, the real amplitude of the anisotropy needs to be much larger than our inverted value (about 1%) for us to see it in the tomographic maps. This scenario is probably less likely as only 3-4% of anisotropy is expected for perfectly aligned crystals based on a pyrolite model (Montagner & Anderson, 1989).

A Monte Carlo type error analysis technique (e.g. Houser et al., 2008; Ma & Masters, 2014) is used to estimate the errors in our analysis. This method estimates how sensitive the resulting models are to reasonable perturbations to the data. Random noise from a Gaussian distribution with a standard deviation the same as the errors in the measurements are added to the data (δt) and then we invert for a new model with the perturbed data. We repeat this process 50 times and the standard deviations of these 50 inverted results are taken to be the errors in our model. Calculating the standard deviations of the fast directions requires some thought because of circularity (359° and 1° are close to each other) and the periodicity (fast directions $\alpha = 0^\circ$ and $\alpha = 180^\circ$ are the same). For the circularity problem, we use Yamartino's method (Yamartino, 1984) which is the standard method to estimate wind directions in meteorology (equation 6.2.9 in Bailey, 2000). For the periodicity problem, we can estimate the errors of 2α (using Yamartino's method) first and then divide the errors by 2 to get the errors of α .

The results from this error analysis are shown in Figure 4.10 and 4.11. Because of the better sampling and azimuthal coverage in the continents, the errors

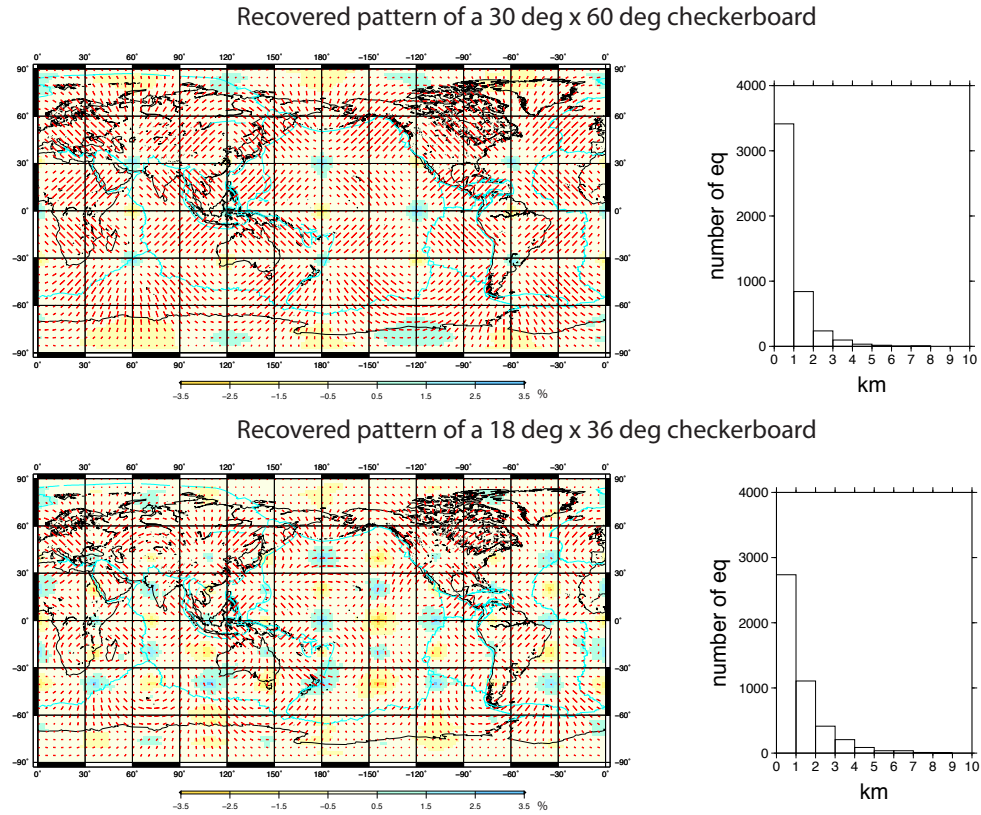


Figure 4.9: The recovered patterns of the checkerboard tests. The input pattern has azimuthal anisotropy only. Leakage from anisotropic signals to isotropic patterns can happen (panels on the left), but only occurs in places where outgoing ray paths always lie in the fast (or slow) directions of all the surrounding blocks. The leakage from anisotropic signals to earthquake locations is typically less than 5 km (panels on the right).

of anisotropic amplitudes are smaller in the continents than in the oceans (Figure 4.10). The errors of the fast directions are a strong function of the amplitude of azimuthal anisotropy (Figure 4.10), which is expected because we get the directions from the ratio between A_1 and A_2 . When the anisotropic amplitudes A are greater than 0.2%, errors in the fast directions are smaller than 10° . Errors in earthquake locations caused by measurement errors are smaller than 5 km and are strongly related to the number of measurements available (Figure 4.11). Errors for earthquakes from the more recent years are smaller than errors for the earlier years because of the expansion of the global seismic network.

4.4 Discussion

It is useful to find out how much and where the azimuthal anisotropy is changed when we include the effect of earthquake locations. We turn off the effect of earthquake relocations by setting $\lambda_2 = 10000$ and redo our inversions based on the same dataset we use in the previous section. We then compare the fast directions between the results with and without source locations (Figure 4.12). Keep in mind that when the anisotropic amplitudes are less than 0.2%, the errors of the fast directions can be larger than 10° (Figure 4.10). To be safe, we do not compare the fast directions in those regions and we set the color to be white in the figure.

Changes in the fast directions away from plate boundaries are small. The biggest change happens at the Andes, where the average changes of earthquake locations are the largest (Figure 4.7). We now see a consistent trench parallel pattern in fast directions at high frequencies throughout this region which seems to be consistent for all the frequencies we use. This may indicate a trench-parallel mantle flow pattern beneath this subduction zone proposed by Russo & Silver (1994). It is also interesting to note that the fast directions in the Nazca plate and the Scotia plate are more EW oriented after relocating earthquakes, which are closer to the expected plate motions in the hotspot frame (Gripp & Gordon, 2002, Figure 4).

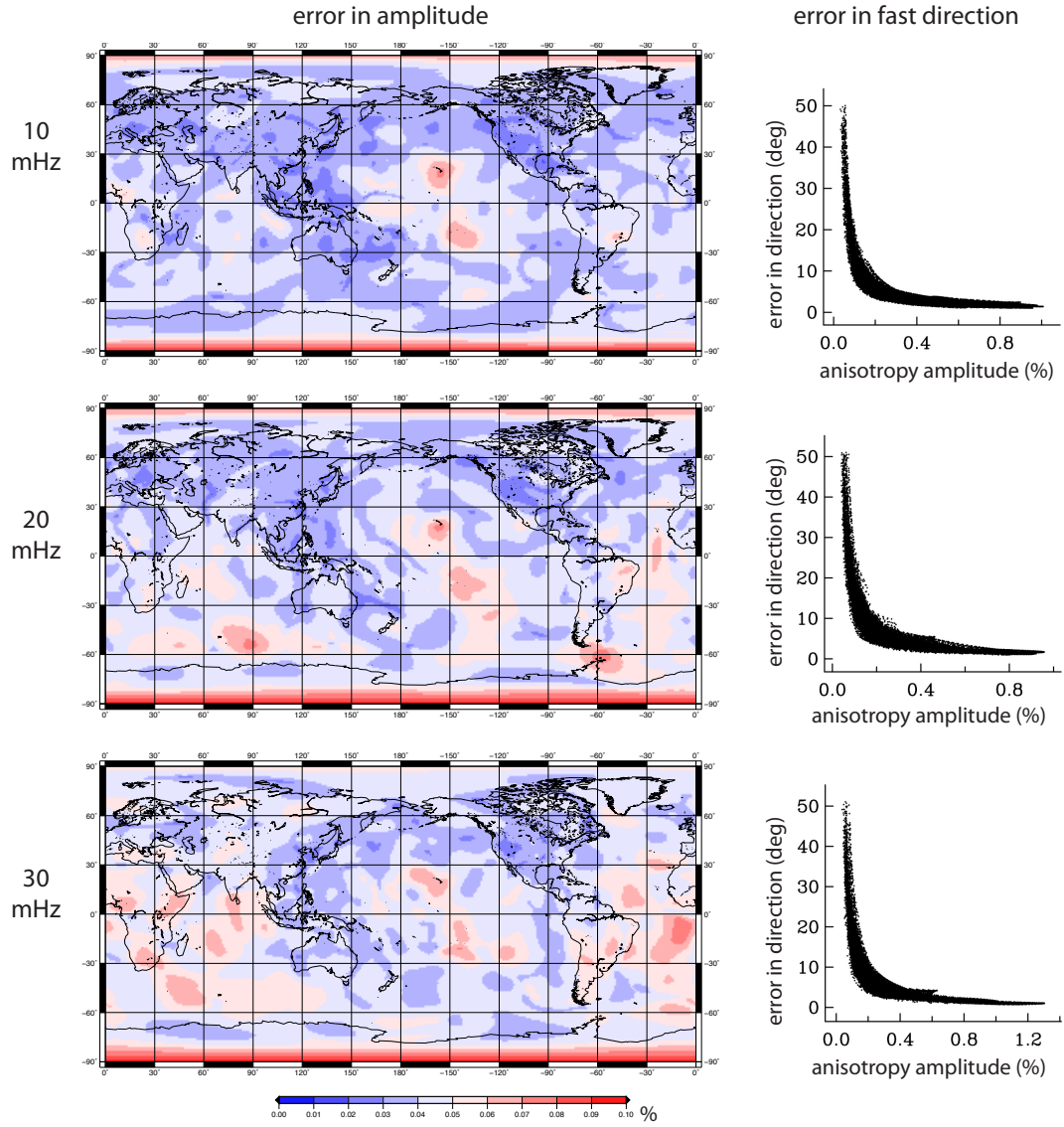


Figure 4.10: Estimated errors in our derived anisotropy structure at different frequencies. The left panels show the errors in the amplitudes. Compared with results shown in Figure 4.6, these errors are small. The right panels show the errors in the fast directions. Note that the errors of the fast directions are a strong function of the amplitude of the anisotropy. When the amplitude of the anisotropy is greater than 0.2%, errors in the fast directions are smaller than 10° .

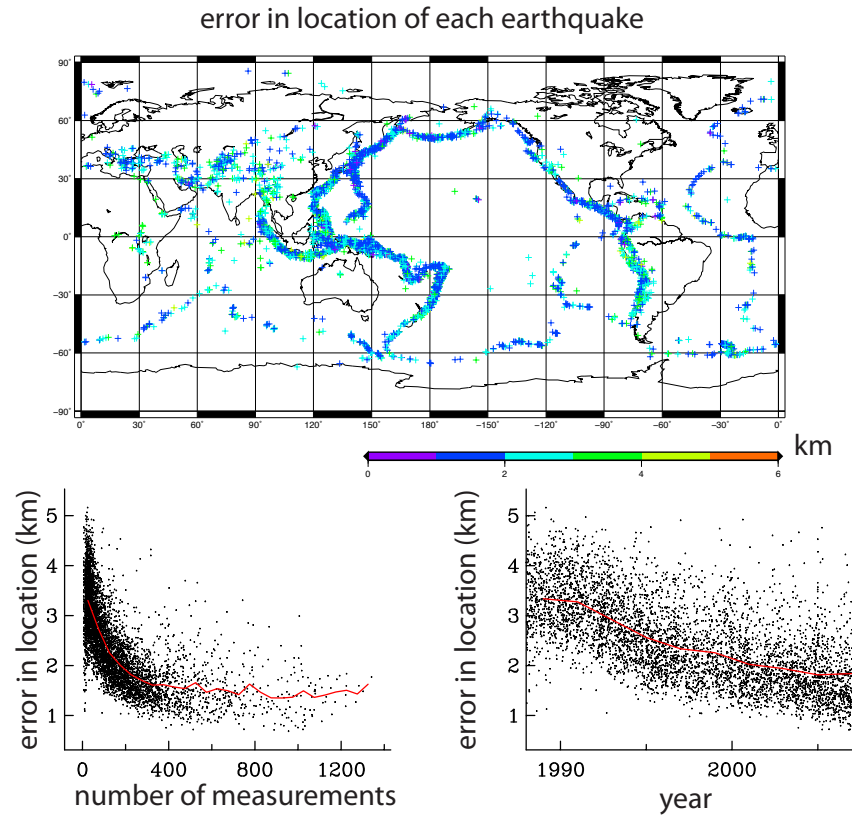


Figure 4.11: Errors in earthquake relocations. The top panel shows the location error for each earthquake. The location errors are generally smaller than 5 km. Note that only the errors caused by the measurement uncertainties in the original phase arrival time dataset are estimated here. The bottom two panels show that these estimated errors are a strong function of the number of measurements available. Errors for earthquakes from the more recent years are smaller than errors for the earlier years because of the expansion of global seismic network.

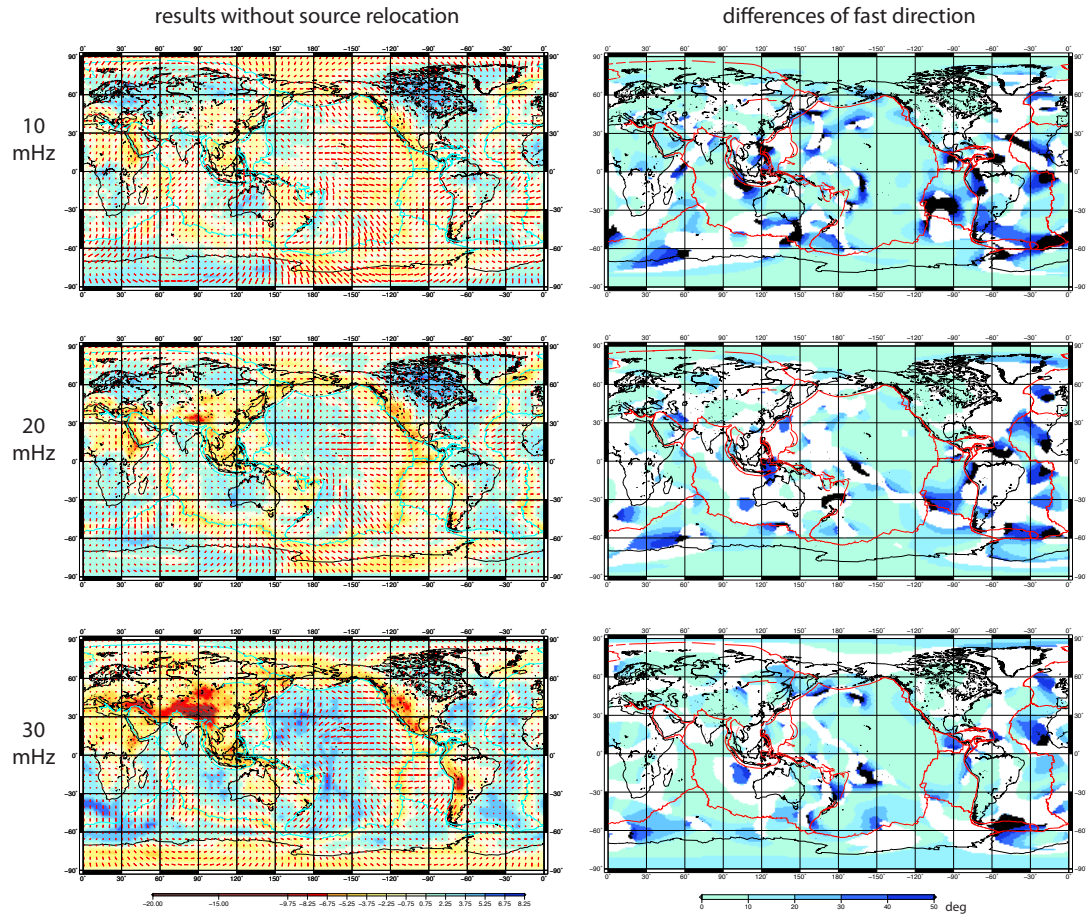


Figure 4.12: Comparisons between results with and without source relocations. Panels on the left show the inversion results without source relocation. Panels on the right compare the fast directions between the results shown in Figure 4.6 and the ones shown here in the left panels. Note that regions with anisotropic amplitudes less than 0.2% are colored white in the panels on the right.

A more direct and simpler approach to illustrate the importance of earthquake locations is to compare the signals from the isotropic and anisotropic velocity maps with the effects of relocating earthquakes. We use the maps shown in Figure 4.6 and forward calculate the contribution to δt from the isotropic term δs_0 and anisotropic terms A_1 and A_2 for all paths (Figure 4.13). If an earthquake is shifted by 10-20 km, it will contribute to a time shift of about 2.5-5 s, which is comparable to the signals from azimuthal anisotropy, but much smaller than the signals from the isotropic terms. This explains why it is reasonable to ignore earthquake mislocations when we are only interested in the isotropic velocity perturbations. But when we consider azimuthal anisotropy with signals that are only about 1/10 of the signals from the isotropic terms, including earthquake relocations in the inversion becomes necessary.

We further perform two experiments to test the stability of our inverted locations. The first is to redo the measurements and inversions starting with PDE locations. We then compare the results with the ones we obtain above which start with CMT locations. The second experiment is to use a dataset combining measurements from different frequencies (5, 15, 25 and 35 mHz). We find that the differences among inverted locations are small and of the order of 5-10 km (Figure 4.14). These experiments demonstrate that our inverted locations are quite stable regardless of the starting values and the dataset used.

Lastly, we compare our inverted results with the locations in the EHB catalog from the International Seismological Centre (Figure 4.15). The two dashed lines are according to the model in Smith & Ekström (1997) ($L = aM_w^{1/3}$) with $a = 2.79 \times 10^{-8}$ used in their paper and $a = 2.1 \times 10^{-8}$ which gives a better fit for our result. The discrepancies for earthquakes larger than magnitude 8 can be large, which is probably due to the finite length of the rupture and the fact that we only have about 15 earthquakes of these magnitudes (compared to a total of over 4000 earthquakes in the whole dataset).

For smaller events with magnitudes less than 7.5, the discrepancy of earthquake locations between the CMT and ISC (or PDE) catalog has been long recognized (e.g. Smith & Ekström, 1997; Kagan, 2003). The differences between the

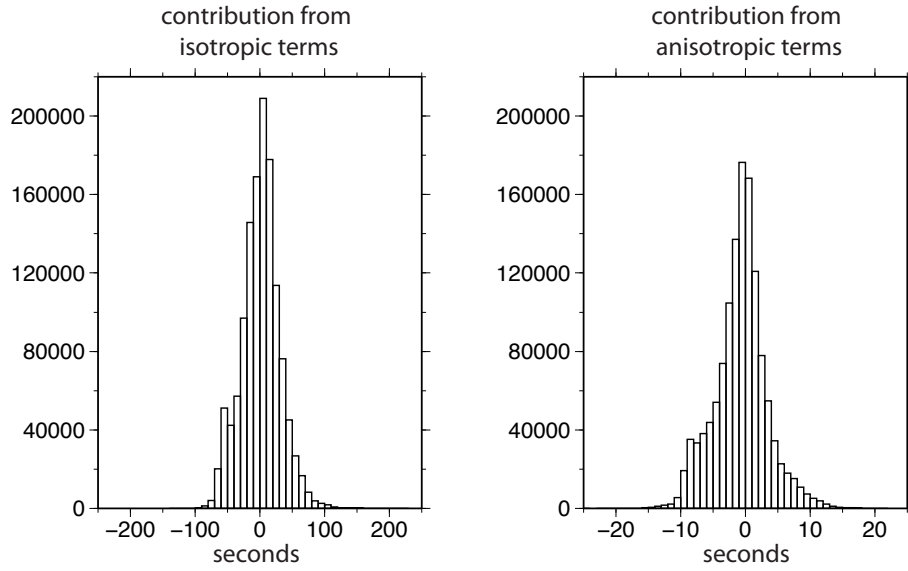
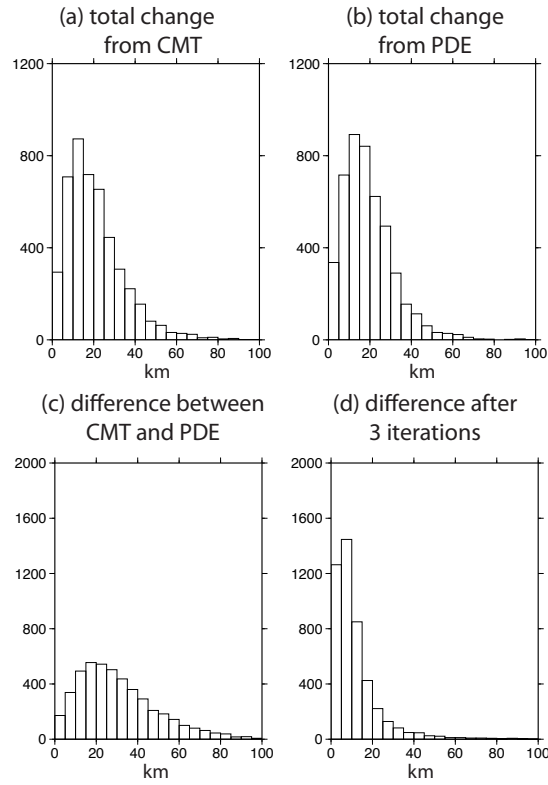


Figure 4.13: Comparison between the signals (phase arrival times) from the isotropic and anisotropic parts of the tomographic maps. This is for all three frequencies (10, 20 and 30 mHz). Note the different scales on the x-axes.

CMT and EHB catalog we find are smaller than the results reported before, which might be a result of including intermediate period surface waves in determining the CMT solutions since 2004 (Ekström et al., 2012) and the ever increasing number of global seismic stations. Relocating earthquakes using our method further reduces the discrepancy from about 27 km to 20 km. The error of the EHB location is about 10 km when the location algorithm is tested by using well-located and well-distributed earthquakes and explosions (Engdahl, 2006, Table 2) and is worse when the azimuthal coverage of stations is poor (Bondár et al., 2004). For our location method, if we combine all sources of errors identified in this chapter, the error is about 8-12 km. Other possible sources of errors (e.g. the use of simple ray theory, non-uniqueness of the solutions) may add another few kilometers. Therefore, some of the discrepancy we observe between our relocated catalog and the EHB catalog can be explained by combining the errors estimated separately for these two methods. However, there is still about 10-15 km that we have not accounted for. Whether such discrepancy is due to the usage of different types of data (body waves v.s. surface waves), the use of different Earth models (ak135 v.s.

start from CMT v.s. PDE, both are using 10+20+30mHz dataset



10+20+30mHz v.s. 5+15+25+35mHz, both start from CMT

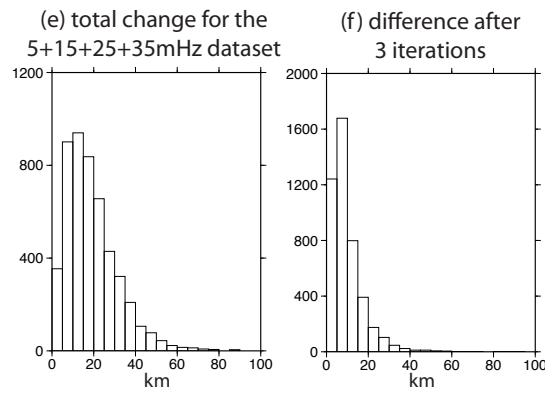


Figure 4.14: The stability of the inverted locations tested by two experiments. The first one compares the results from the same dataset but using two different catalogs (CMT v.s. PDE) for the starting locations. The second one compares the results from two different datasets, one includes measurements from 10, 20 and 30 mHz while the other one includes measurements from 5, 15, 25 and 35 mHz. The differences among the inverted locations are of the order of 5-10 km.

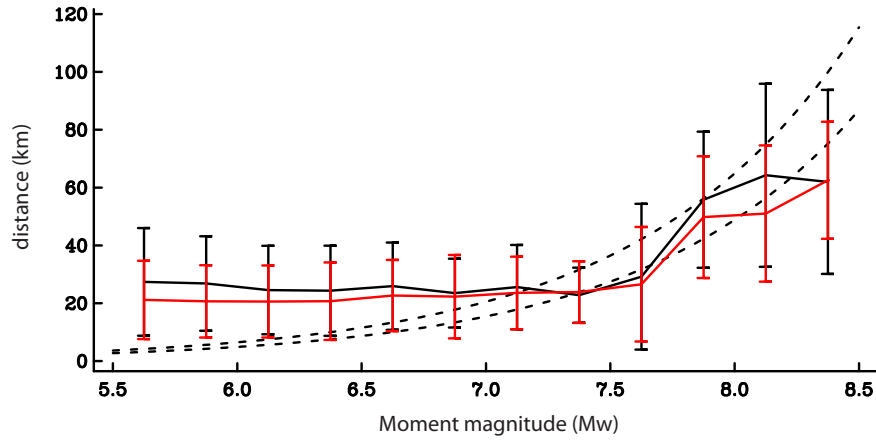


Figure 4.15: Comparison between our relocated locations and the ones in the EHB catalog. Data are binned with an interval of 0.25 magnitude. Black line shows the median of the distances between earthquakes in the CMT and EHB catalogs. Red line shows the median of the distances between our inverted earthquake locations and the ones in the EHB catalog. Error bars show the scaled Median-Absolute-Deviation (MAD) of the differences in each bin. The two dash lines show the model in Smith & Ekström (1997) ($L = aM_w^{1/3}$) with slightly different choices of a .

surface wave dispersion maps), or some other sources of mislocation, needs further investigation.

4.5 Conclusion

In this chapter, we demonstrate the important effect of earthquake (mis)-location on determining Rayleigh wave azimuthal anisotropy. We present the theory of including source relocations in the inversion for azimuthal anisotropy. The concept of ‘locally parallel azimuth’ is used to derive a simple way to implement smoothness constraints on azimuthal anisotropy to avoid the singularity problem around the poles.

We then use existing surface wave phase velocity datasets to invert for source (re)locations, isotropic and azimuthal anisotropic velocity perturbations. Our results suggest that earthquakes near the Andes can be systematically shifted by more than 30 km. Earthquake epicenters at the ridge-transform systems in the

oceans are now closer to the plate boundaries. The differences of the resulting azimuthal anisotropic patterns with and without source relocation are greatest near the plate boundaries and large changes can be found around the South American plate. Checkerboard tests suggest we can only resolve anisotropic structures on a scale of about 20° in the oceans, but the azimuthal anisotropic patterns in the Pacific are likely coherent over large distances. The discrepancy between our relocated earthquake locations and the locations in the EHB catalog is reduced to about 20 km for $M_w < 7.5$ earthquakes, but further investigation is still needed to explain the remaining difference.

Acknowledgments

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Chapter 4, in full, has been submitted for publication of the material as it may appear in *Geophysical Journal International*: Ma, Z. and G. Masters, Effect of earthquake locations on Rayleigh wave azimuthal anisotropy. I was the primary investigator and author of the paper.

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Chapter 5

2D global Rayleigh wave attenuation using finite frequency focusing and defocusing theory

Abstract

In this study, we obtain a set of 2D global phase velocity and attenuation maps for Rayleigh waves between 5 and 25 mHz. Correcting for the effect of focusing-defocusing is crucial in order to obtain reliable attenuation structure. Great circle ray theory, which has been used to date, can give useful predictions of this effect if careful attention is paid to how the phase velocity model is smoothed. In contrast, predictions based on the 2D finite frequency kernels are quite robust in this frequency range and suggest that they are better suited as a basis for inversion. We use a large dataset of Rayleigh wave phase and amplitude measurements to invert for the phase velocity, attenuation, source, and receiver terms simultaneously. Our models provide 60-70% variance reduction to the raw data though the source terms are the biggest contribution to the fit of the data. The attenuation maps show structures that correlate well with surface tectonics and the age progression trend of the attenuation is clearly seen in the ocean basins. We have also identified problematic stations and earthquake sources as a by-product of our data selection

process.

5.1 Introduction

Fundamental mode surface waves have been widely used to constrain upper mantle heterogeneities. Compared to the numerous models for the elastic properties in the past decades, only a handful of studies attempt to resolve attenuation structure. An improved understanding of upper mantle attenuation is important in two aspects. The first is to better constrain the physical and chemical state of Earth's interior by providing complementary information to elastic velocity (e.g. Karato, 2003; Faul & Jackson, 2005; Dalton et al., 2009; Zhu et al., 2013; Abers et al., 2014). The second is to improve our knowledge of the elastic properties themselves by providing better constraints on their small scale variations (Laske & Masters, 1996) and by taking into account the effect of physical dispersion (Ruan & Zhou, 2010).

The fundamental difficulty in inferring attenuation is that surface wave amplitudes are impacted by many things and intrinsic attenuation is a relatively minor contributor. Early studies used four consecutive orbits (R1, R2, R3 and R4) to remove the source, receiver and focus-defocusing effects (Romanowicz, 1990; Durek et al., 1993). But this is only feasible for large earthquakes recorded at long periods. The loss of sensitivity to odd order structure also limits its usage. Romanowicz (1994) developed a strict data selection criterium to remove paths with strong band-limited focusing effects and to correct for source magnitudes. She was then able to use R1 and R2 data in her modeling, which resulted in the first global attenuation model (Romanowicz, 1995). Selby & Woodhouse (2000) and Selby & Woodhouse (2002) concluded that the contamination from focusing-defocusing effects is not substantial for long wave length structure (below degree 8) though they cautioned that some features of their model may still be influenced by this effect. The uncertainties of earthquake sources were also shown to be important in their studies. Gung & Romanowicz (2004) used three-component long period waveforms to produce a degree-8 3D attenuation model. The authors did not include

the focusing-defocusing effect explicitly though the contamination was shown to be relatively small when considering long wave length structure. Billien et al. (2000) were the first to include the focusing-defocusing effects in the modeling explicitly. Most recently, Dalton & Ekstrom (2006b) revisited this problem. They demonstrated that it is crucial to include the source, receiver, focusing-defocusing and attenuation effects simultaneously in order to obtain reliable attenuation structure. We will follow their lead in this paper.

One motivation for revisiting the attenuation problem at this time is that we have developed an efficient technique to process the large amount of seismic data available to us from IRIS (Ma & Masters, 2014; Ma et al., 2014). This method generates a very large data set of both surface wave phase and amplitude measurements and the amplitude measurements themselves have not been utilized before. The second motivation is that a new theory incorporating the finite-frequency effects has been developed in a series of papers (Zhou et al., 2004; Zhou, 2009) to describe the surface wave amplitude variations. The implication of this new theory has only been discussed theoretically or by using synthetic seismograms (Dalton et al., 2014; Parisi et al., 2015) while its application to real data is still lagging.

The main goals of this chapter are to test the feasibility of applying this finite-frequency theory to the real data and to present a set of 2D attenuation maps for Rayleigh waves between 5 and 25 mHz. Section 2 describes the finite-frequency theory used in this chapter and compares it with the great circle ray theory which has been used to date. Section 3 discusses our data selection processes including a procedure to identify problematic stations and earthquake sources. Section 4 describes our inversion technique and presents our models. These inversion results and the current limitations of our method are further discussed in section 5.

5.2 Theory

In this chapter, we follow the basic procedure described by Dalton & Ekstrom (2006b). The observed surface wave amplitude anomaly $A(\omega)$ is expressed as

$$A(\omega) = A_S(\omega)A_R(\omega)A_F(\omega)A_Q(\omega) \quad (5.1)$$

where A_S , A_R , A_F and A_Q are corrections for earthquake sizes, corrections for instrument responses, focusing-defocusing effects and intrinsic attenuation effects respectively.

One of the major contributors to surface wave amplitudes is the effect of focusing and defocusing of the wavefront caused by the *elastic* velocity variations. In a very elegant paper, Woodhouse & Wong (1986) derived an expression for this effect based on great circle ray theory:

$$\ln A_F = 1/2 \operatorname{cosec} \Delta \int_0^\Delta \sin(\Delta - \phi) [\sin \phi \partial_\theta^2 - \cos \phi \partial_\phi] \frac{\delta c}{c_0} d\phi \quad (5.2)$$

where Δ is the epicentral distance, ϕ is the along-path coordinate, θ is the path-perpendicular coordinate, and δc is the perturbation in phase velocity. The dominant term in this expression is the second derivative ∂_θ^2 of the phase velocity perturbation. Note that, if the phase velocity perturbation is expanded in spherical harmonics, the second derivative term is of the order of l^2 , where l is the spherical harmonic degree. This implies that the focusing-defocusing predicted by ray theory is highly sensitive to the short-wavelength structure of a phase velocity map. Such structure is often poorly constrained and can be strongly dependent on the nature of damping used in the construction of the map.

Finite frequency effects will mitigate the effect of short wavelength structure. Here we consider the 2D approximation to the finite frequency kernel derived by Zhou et al. (2004) using Born scattering, and further neglect the effect of source radiation pattern:

$$\ln A_F = \int K_A^c \frac{\delta c}{c_0} d\Omega = \int -\frac{2k^2 \cos[k(\Delta' + \Delta'' - \Delta) + \pi/4]}{\sqrt{8\pi k |\sin \Delta'| |\sin \Delta''| |\sin \Delta|}} \frac{\delta c}{c_0} d\Omega \quad (5.3)$$

where k is the wave number, Δ' is the distance between the source and the scatterer, and Δ'' is the distance between the scatterer and the station. As a Gaussian band-pass filtering scheme is inherent in our measurement process, we can further average the kernel over a limited frequency band to suppress the side lobes. Figure

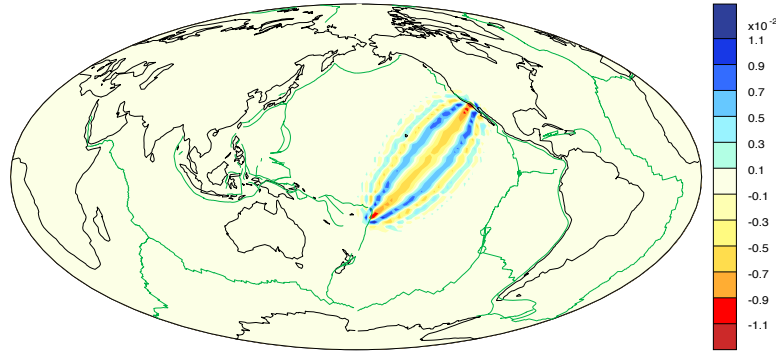


Figure 5.1: Finite frequency kernel for the focusing-defocusing term calculated at 10mHz for an event in Tonga (01/01/1996) and recorded at station PFO in California. This example uses a parameterization of 2 degree equal-area pixels and assumes that $\delta c/c$ is in percent.

5.1 shows an example of such a kernel for Rayleigh waves at 10 mHz. The details of the kernel computation are described in the Appendix.

We compare the predictions of great circle ray theory and finite frequency focusing-defocusing effects by doing forward calculations using an existing Rayleigh wave phase velocity map at 100 s (Ekström, 2011). Using other maps (e.g. Ma et al., 2014) leads to identical conclusions. To evaluate the effect of higher order structure, we perform a spherical harmonic analysis of the map and truncate the map at different spherical harmonic degrees l ($l = 20$ and $l = 40$). The two theories give similar results for the smooth map ($l = 20$, Figure 5.2a), but fail to agree with each other for the rough map ($l = 40$, Figure 5.2b). Finite frequency theory gives stable result regardless of the cut off value of l (Figure 5.2c). Ray theory predictions from $l = 20$ and $l = 40$ maps are quite different from each other, which indicates the strong dependence on the high order structure (Figure 5.2d). Figure 5.3 shows the same experiments but at different frequencies. At 25 mHz, the finite-frequency theory starts to share the same problems as ray theory (though less severe), so we choose to stop at 25 mHz in this chapter. At high frequencies, 3D kernels based on the same theory (Zhou et al., 2004) or based on the adjoint method (Sieminski et al., 2007) may be more appropriate. Exact ray theory may also be a better theory for the high frequencies (Dalton et al., 2014) but has a very

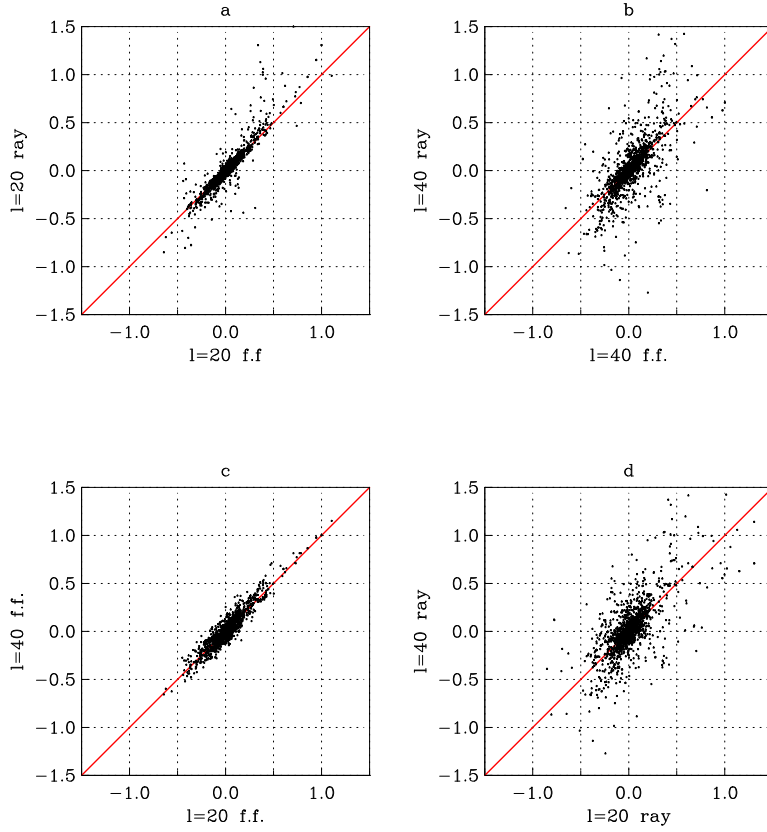


Figure 5.2: Comparison of the focusing-defocusing effect calculated by finite frequency theory (f.f.) and ray theory (ray). (a) The two theories give similar results for the smooth map ($l = 20$). (b) The two theories fail to agree with each other for the rough map ($l = 40$). (c) Finite frequency theory gives stable result regardless of the cut off value of l . (d) Ray theory predictions from $l = 20$ and $l = 40$ maps are quite different from each other, which indicates the strong dependence on the high order structure. In all cases, $\ln A$ is plotted.

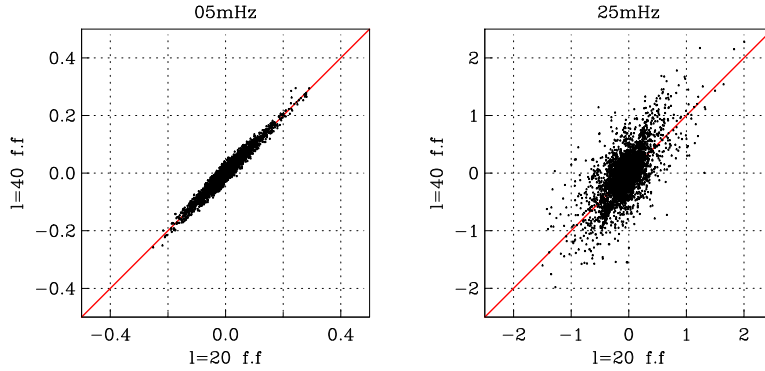


Figure 5.3: Same as Figure 5.2c but for 5 and 25 mHz. Notice the difference in scale. In all cases, $\ln A$ is plotted.

high computational cost for the actual inversions.

To conclude, since removing the focusing-defocusing effect (eqn 5.1) is essential in order to obtain reliable attenuation structure (Dalton & Ekstrom, 2006b), a robust theory that can reliably predict such effect and is insensitive to the details of making the phase velocity maps is preferred. Great circle ray theory can give useful prediction for the focusing-defocusing effect if careful attention is paid to how the phase velocity model is smoothed. However, the predictions of the finite frequency kernels are more robust at the low-intermediate frequency range (below 25 mHz) and suggest that they are better suited as a basis for inversion.

5.3 Data

5.3.1 The raw data

The raw amplitude measurements are from a dataset measured by a cluster analysis technique (Ma & Masters, 2014; Ma et al., 2014). This dataset includes all long period data available to us from the IRIS DMC for earthquakes with magnitudes greater than 5.5 from 1990 to 2007. The cluster analysis method groups waveforms from a single earthquake recorded by many seismic stations according to their waveform similarities. Note that the measurement process has already cor-

rected for the radiation pattern based on the CMT solutions (Ekström et al., 2012), geometrical spreading and the 1D attenuation effect based on PREM (Dziewonski & Anderson, 1981). In the original dataset, the amplitudes are relative amplitudes measured by a standard cross-correlation technique in given time windows. We find it more convenient to work with absolute amplitudes in the actual inversion. This also enables us to study the source terms (see below) for different frequencies. We have tested two methods to convert the relative amplitudes to absolute amplitudes. The first is to apply a static shift to each cluster based on the peak amplitudes of the envelopes, similar to the way we treat our group velocity data (Ma & Masters, 2014). The second is to simply use the peak amplitudes of the envelopes instead. The difference between the two are rather small (Figure 5.4). The distribution of the difference does have a longer tail on the negative side, indicating that a minor portion of our measurements can be affected by overtone contamination if we use the first method. In fact, inversion based on the peak envelope values indeed gives a slightly better fit (e.g. $\chi^2/N=1.34$ v.s. 1.47 for 10 mHz). Therefore, we choose to use the peak envelope values as our raw data in this chapter.

5.3.2 Data selection

Stations

Before our formal inversions in which various terms in eqn 5.1 are included in a joint inversion (see below), it is important to first identify as many problematic stations as possible. This step is necessary as some station gains can drift over time. We restrict ourselves to permanent stations only. The data quality from PASSCAL experiments is hard to quantify using our method as they tend to operate for a relatively short period of time.

The equation we are solving for is:

$$d_{ij}^k = \ln A_i - \ln A_j \quad (5.4)$$

where $\ln A_i$ and $\ln A_j$ are the station terms for station i and j respectively, d_{ij}^k are the relative amplitudes among all station pairs within the same cluster (hence the

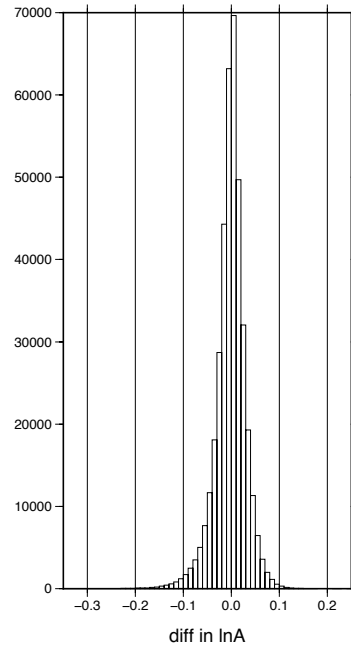


Figure 5.4: Difference between $\ln A$ measured as the peak values of the envelopes and the ones based on the relative amplitudes measured by cross-correlation. This example is for 10 mHz Rayleigh waves.

same event), k denotes different pairs. We solve this set of equations for each cluster and then plot the solution $\ln A$ as a function of time for visual examination (the median of $\ln A$ in each cluster is set to zero). Also plotted are the one year running median values (Figure 5.5).

This method is similar to a more sophisticated approach by Eddy & Ekstrom (2014) to infer the local amplification of USArray stations. One difference is that we are solving this equation for each cluster instead of the entire dataset. In addition, we do not limit our station pairs to be close to each other because we are dealing with a global network. These are not severe problems, because our purpose is to identify problematic stations and it is the drifting of the station terms that we are interested in at this step.

While the majority of seismic stations have been continuously generating high quality data, we indeed identify some problematic stations. Some examples are shown in Figure 5.5. For stations which behave questionably only for a certain time period, we discard data for that period (e.g. ALE 90-94, RER 96-02). For

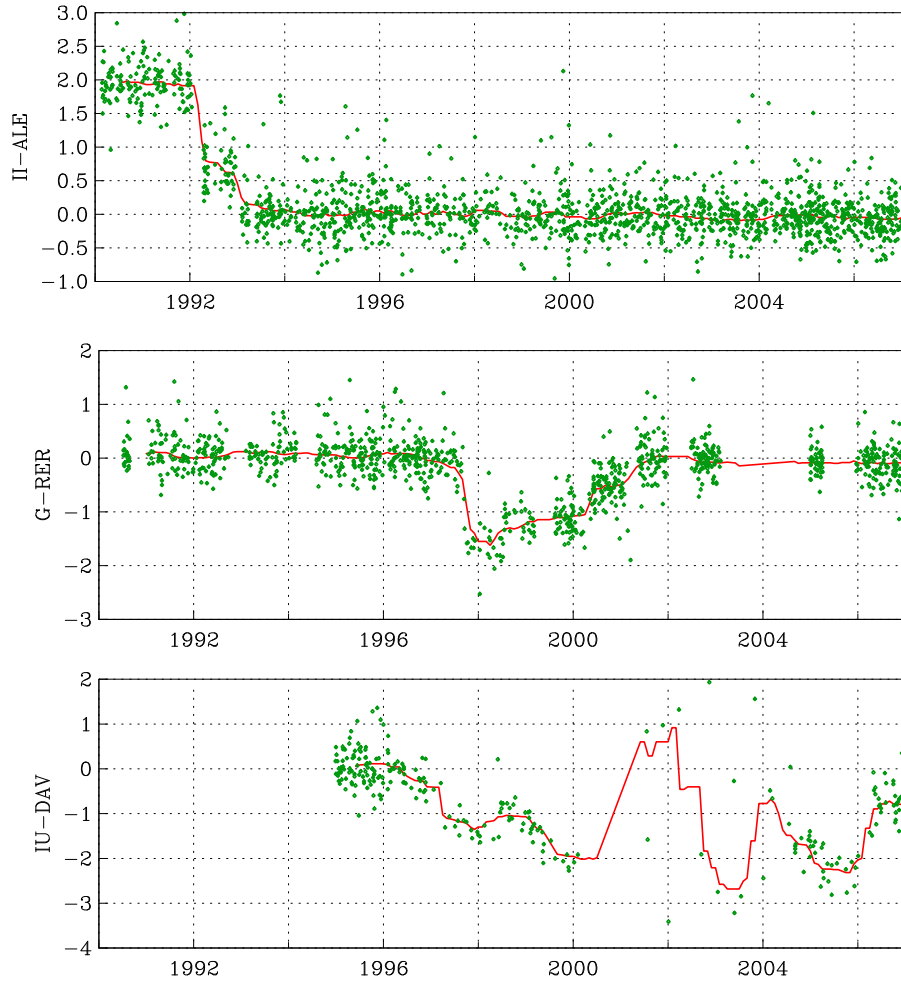


Figure 5.5: Examples of problematic stations. The green dots represent the $\ln A$ in eqn 5.4. The red lines represent the one year running median values. The questionable behavior of station ALE has been noticed before (Ekström & Nettles, 2010). The IDA network operators were able to locate the cause of this problem and the meta data are now fixed at the IRIS DMC. The cause for the malfunctioning at other stations is still not clear to us at the time of writing.

stations that have drifting station terms for the entire operational period, we discard records from these stations entirely (e.g. DAV). The questionable behavior of station ALE has been noticed before (Ekström & Nettles, 2010) and the meta data are now fixed at the IRIS DMC. The cause for the malfunctioning at other stations is still not clear to us at the time of writing. We provide a list of these stations accompanying this chapter for those investigators who are interested.

Sources

We first discard measurements that are near-nodal in the radiation pattern, which is defined here as less than 0.4 of the peak radiation amplitude. In addition, we manually scan through all earthquakes to compare the measurements with the predicted radiation pattern. Two examples are shown in Figure 5.6. Note that our measurements have been corrected for the theoretical radiation pattern so they should be close to 0 in the ‘ideal’ cases. Figure 5.6a shows an example of an earthquake at the Pacific Antarctic ridge. A number of stations with azimuth between about 5° and 25° (in this case, the stations in the Western US) record unusually high amplitudes (even when the 3D propagation effect is taken into account). In fact, a $\sim 4^\circ$ shift in the radiation pattern is sufficient to bring these measurements back to ‘normal’. We suspect that this is caused either by a small error in the CMT solution or the fact that the effect of take-off angle variations in the presence of heterogeneities (Um & Dahlen, 1992) is not included in our analysis. We have tested the latter possibility by doing ray-tracing through existing phase velocity maps and it gives a correction of about 2° . As there is still significant room to improve the fit to surface wave arrival angle data (Larson & Ekstrom, 2002; Laske & Masters, 1998), it is not surprising that the take-off angles can be underestimated. Therefore in this chapter, instead of correcting for the take-off angles, we choose to remove the affected stations in the limited azimuthal range.

Figure 5.6b shows another example for an earthquake ($M_w=6.6$) from the southern Iran region. For this earthquake, our measurements are inconsistent with CMT predictions for all azimuths. In fact, the NEIC catalog gives a different moment tensor solution for this earthquake (<http://earthquake.usgs.gov/>

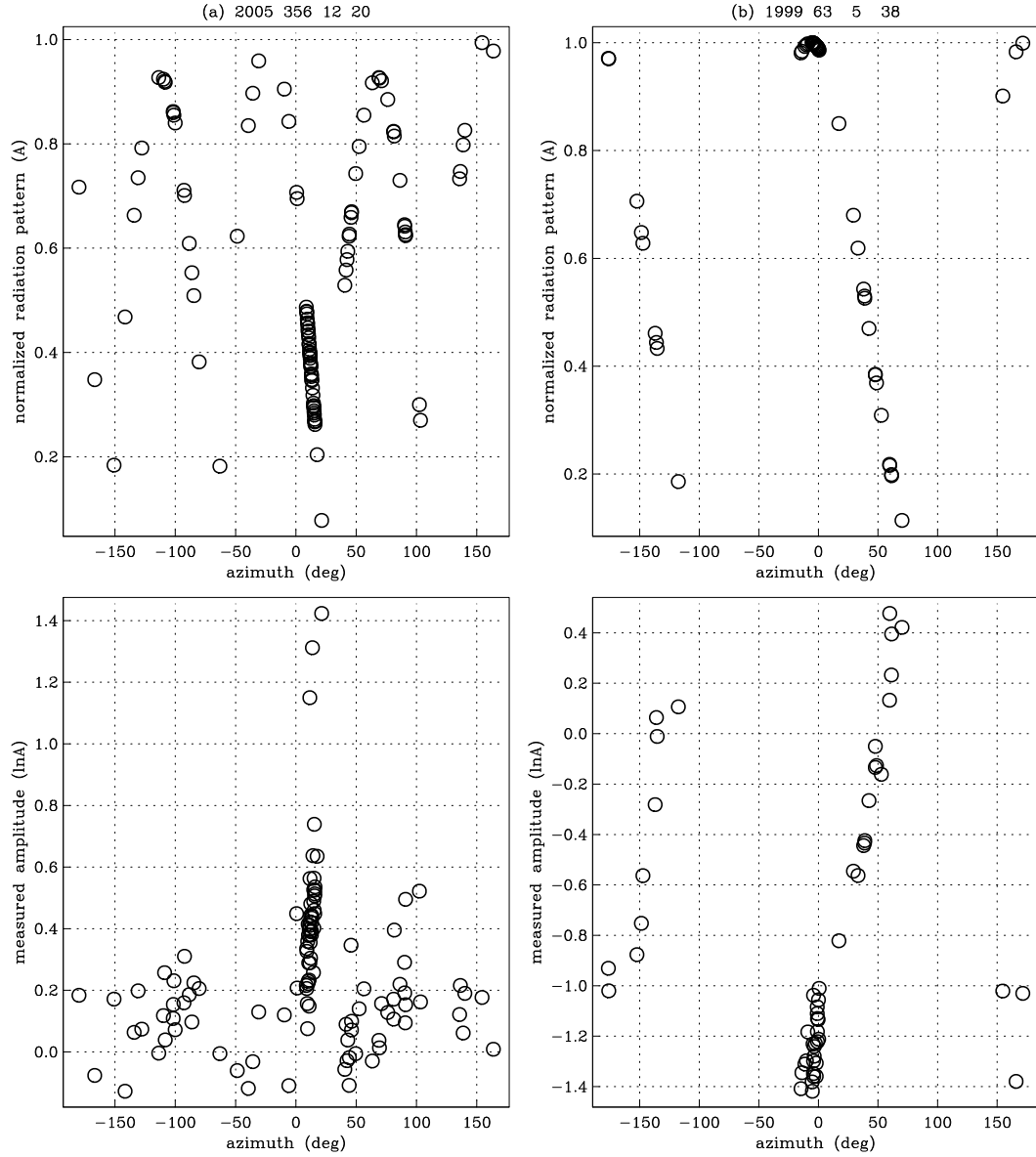


Figure 5.6: Examples of problematic earthquake sources. The top panels show the predicted radiation patterns. The bottom panels show the raw measured amplitudes (lnA). The left two panels show an earthquake on the Pacific Antarctic Ridge (origin time is shown in the title). This example shows that a number of stations with azimuth between about 5° and 25° (in this case stations in the Western US) record unusually high amplitudes (even when the 3D propagation effect is taken into account). The right two panels show another example for an earthquake (Mw=6.6) in the southern Iran region. For this earthquake, our measurements appear to be inconsistent with CMT predictions for all azimuths.

earthquakes/eventpage/usp00093tj#scientific_tensor). As a systematic study for the cause is not the focus of this chapter, we choose to remove records from these earthquakes when we recognize them. It is worth pointing out that only a small fraction of the raw data are affected (see Table 5.1). Again, a list of these problematic earthquake sources is provided with this chapter.

Path deviations far from minor arcs

After some experiments, we find it necessary to remove paths that deviate more than 5° from the minor arcs that connect earthquake sources and seismic stations. The ray tracing is based on the Rayleigh wave phase velocity maps from Ma et al. (2014) and uses the method described in Woodhouse & Wong (1986). To reduce the computational cost, we first bundle paths using $5^\circ \times 5^\circ$ blocks and perform ray tracing on these ‘summary rays’. In addition, the actual calculation is performed using maps truncated at degree $l = 40$. The truncation level changes the details of the shape of each path but does not significantly affect whether a path is removed or not.

One example is shown in Figure 5.7. Note that in this example, some paths deviate so far from the minor arc path that they are in the part of the kernel which has the opposite sign from the central lobe. This example also suffers from multipathing effect which is not included in the theory we use in this chapter (Ruan & Zhou, 2012).

These paths are mainly along the boundaries between oceans and continents (Figure 5.8). Keeping these paths in the inversion generates distinct artifacts in the attenuation maps. The artifacts are mostly obvious along the northern rim of the Pacific Ocean, from Indonesia and Japan to the Western US, for frequencies above about 10 mHz. While high attenuation structure is observed in inactive back arc basins from body wave studies (e.g. Barazangi et al., 1975; Flanagan & Wiens, 1994), the area of high attenuation structure we observe is much more extended and the attenuation is too strong. Removing paths that deviate far from the minor arcs effectively suppresses these artifacts (Figure 5.8). The necessity of removing these paths is further discussed below.

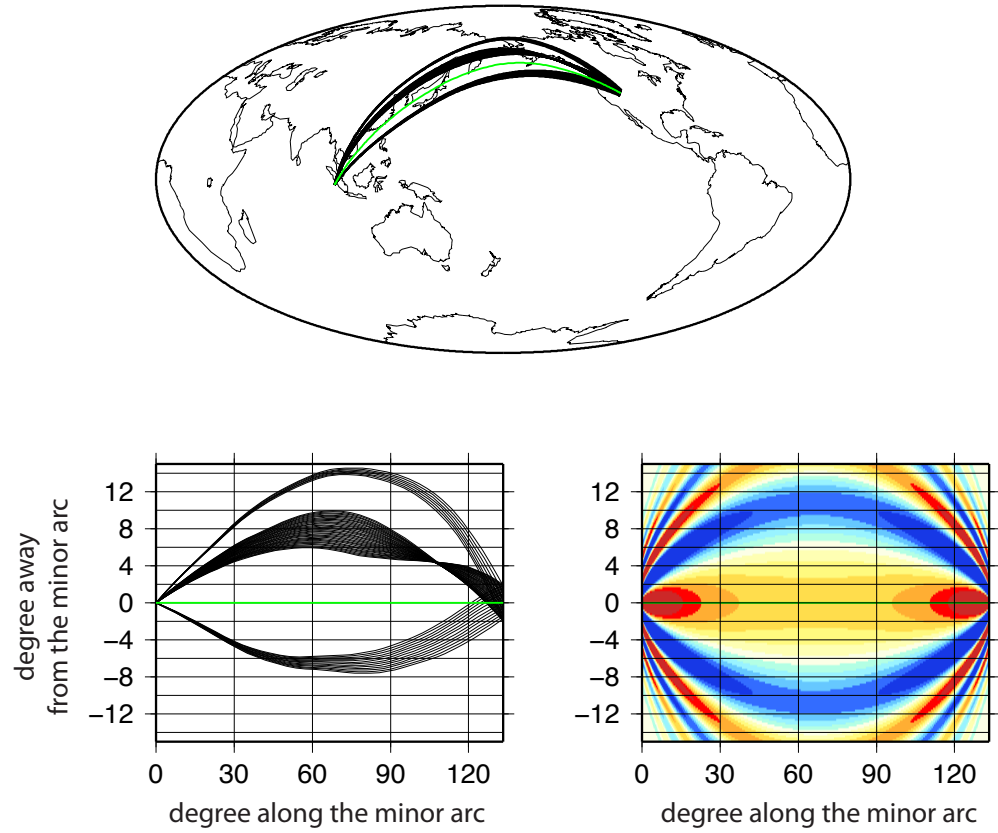


Figure 5.7: An example of ray paths that deviate far from the minor arc. This is for an earthquake that happened in Indonesia and recorded by a TA station in the US. The top panel shows the paths on a geographic map. The bottom left panel shows the same paths but in the epicentral coordinate system. Also plotted to its right is the corresponding focusing-defocusing kernel. Note that some paths deviate so far from the minor arc that they are in the part of the kernel which has the opposite sign from the central lobe. In both panels, the green line shows the minor arc connecting the earthquake and the station. The black lines show all possible paths that lie within 2° of the receiver. The ray tracing is based on the 15 mHz Rayleigh wave map from Ma et al. (2014) and uses the method described in Woodhouse & Wong (1986).

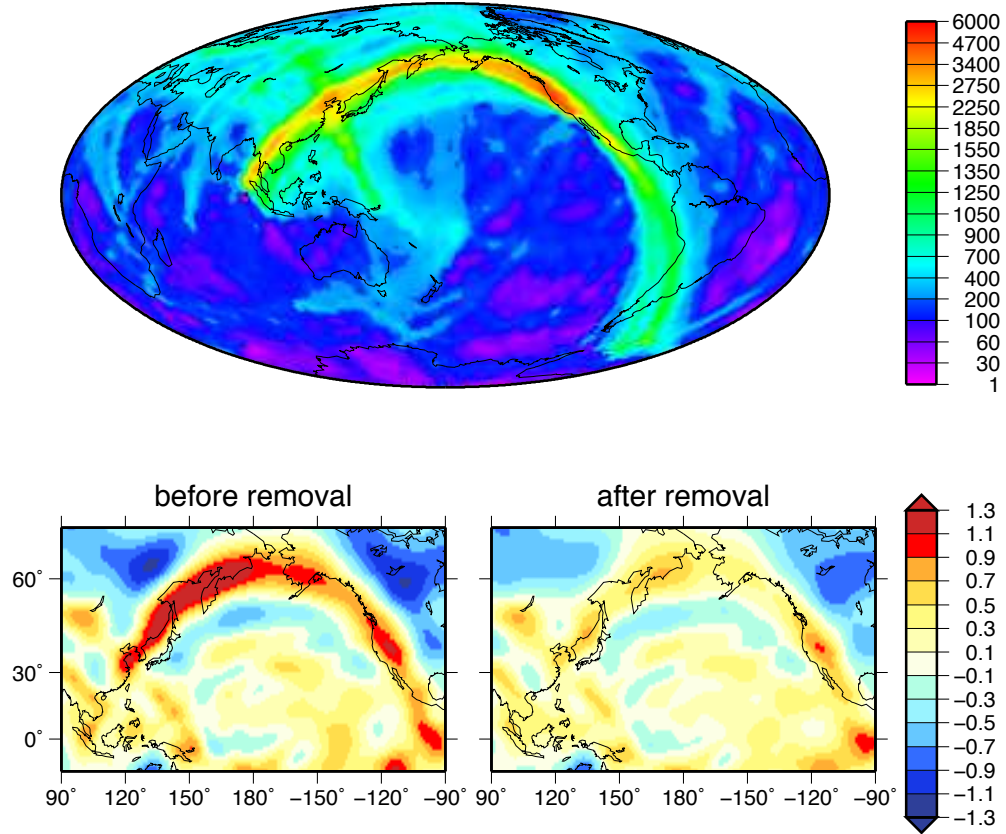


Figure 5.8: Effect of removing paths that deviate far from the minor arcs. The top panel is the hit count map, showing the number of these ray paths passing through each of the 2° by 2° equal area block. These paths are mainly along the boundaries between oceans and continents, especially along the northern rim of the Pacific Ocean. The bottom panels show the attenuation maps inverted before and after the removal of these paths. The example given is for 15 mHz Rayleigh waves.

Table 5.1: Summary of the data selection process. The first column shows the criteria for data selection. The second column shows the percentage of the raw data that are removed. This example is for 10 mHz but datasets from other frequencies share similar statistics.

peak arrival times too large or small	<0.5%
raw $\ln A$ too large or small	<0.5%
PASSCAL stations	16%
problematic stations	6%
paths near nodal	14%
paths with problems shown in Figure 5.6a	1%
paths with problems shown in Figure 5.6b	1%
paths deviated by more than 5° from the minor arcs	8%
earthquakes with too few records	8%

Other criteria

Other criteria for the data selection include: 1) the peak of the envelope arrives too late or too early, which is defined as the half width of the theoretical Gaussian wave packet for each frequency; 2) raw amplitude measurements $|\ln A|$ are too large; 3) earthquakes have less than 30 measurements. Table 5.1 summarizes the criteria we use for our data selection process.

5.4 Inversions

5.4.1 Theory

As stated above, the focusing-defocusing kernels K_A^c in our inversion are calculated based on 2D finite frequency theory (Zhou et al., 2004). The phase delay kernels K_ϕ^c are (Zhou et al., 2004):

$$\delta\phi = \int K_\phi^c \frac{\delta c}{c_0} d\Omega = \int -\frac{2k^2 \sin[k(\Delta' + \Delta'' - \Delta) + \pi/4]}{\sqrt{8\pi k} |\sin \Delta'| |\sin \Delta''| |\sin \Delta|} \frac{\delta c}{c_0} d\Omega \quad (5.5)$$

and can be computed in a similar fashion to the computation of K_A^c .

In the presence of attenuation, we have

$$\delta\phi = \int K_\phi^c \frac{\delta c}{c} d\Omega + \int K_\phi^Q \frac{\delta Q^{-1}}{Q^{-1}} d\Omega \quad (5.6)$$

$$\delta \ln A = \int K_A^c \frac{\delta c}{c} d\Omega + \int K_A^Q \frac{\delta Q^{-1}}{Q^{-1}} d\Omega \quad (5.7)$$

where the attenuation kernels K_ϕ^Q and K_A^Q can be computed from the elastic kernels (Zhou, 2009):

$$K_\phi^Q = \frac{c}{2UQ} \left[\frac{2}{\pi} \ln \left(\frac{\omega}{\omega_0} \right) K_\phi^c - K_A^c \right] \quad (5.8)$$

$$K_A^Q = \frac{c}{2UQ} \left[\frac{2}{\pi} \ln \left(\frac{\omega}{\omega_0} \right) K_A^c + K_\phi^c \right] \quad (5.9)$$

where c and U are the reference phase and group velocities. The reference frequency ω_0 is typically taken to be 2π rad/s. Note that the effect of physical dispersion has been included in the equations above. Zhou (2009) finds that 15 to 20% of the observed phase anomalies can come from anelastic effects so a joint inversion is needed. In addition, we use $K_\phi^c = \omega K_t^c$ to convert the phase kernel (K_ϕ^c) to the arrival time kernel (K_t^c).

We invert for phase velocity perturbations, attenuation perturbations, source terms and receiver terms simultaneously (see also eqn. 5.1). The system we solve for is:

$$\begin{pmatrix} K_t^c & \frac{c}{2UQ} [\frac{2}{\pi} \ln(\frac{\omega}{\omega_0}) K_t^c - \frac{K_A^c}{\omega}] & 0 & 0 \\ K_A^c & \frac{c}{2UQ} [\omega K_t^c + \frac{2}{\pi} \ln(\frac{\omega}{\omega_0}) K_A^c] & 1 & 1 \\ \lambda_1 D & 0 & 0 & 0 \\ 0 & \lambda_2 D & 0 & 0 \\ 0 & \lambda_3 I & 0 & 0 \end{pmatrix} \begin{pmatrix} \delta c/c \\ \delta Q^{-1}/Q^{-1} \\ \ln A_S \\ \ln A_R \end{pmatrix} = \begin{pmatrix} \delta t \\ \delta \ln A \\ 0 \\ 0 \\ 0 \end{pmatrix} \quad (5.10)$$

where D is a smoothing operator (a discrete approximation to the 2D Laplacian). λ_1 and λ_2 are smoothing parameters which control the smoothness of the phase velocity and attenuation maps respectively. λ_3 is a damping parameter which constrains the size of the attenuation perturbations. The relative weighting between phase and amplitude measurements (δt and $\delta \ln A$) is chosen according to the

measurement precision, which is determined by comparing records from close-by stations (see Table 5.2 for the actual values).

We can further simplify the equation above by putting back the physical dispersion:

$$\frac{\widetilde{\delta c}}{c} = \frac{\delta c}{c} + \frac{c}{\pi U} \ln \left(\frac{\omega}{\omega_0} \right) \delta Q^{-1} \quad (5.11)$$

and solving:

$$\begin{pmatrix} K_t^c & -\frac{c}{2UQ} \frac{K_A^c}{\omega} & 0 & 0 \\ K_A^c & \frac{c}{2UQ} \omega K_t^c & 1 & 1 \\ \lambda_1 D & 0 & 0 & 0 \\ 0 & \lambda_2 D & 0 & 0 \\ 0 & \lambda_3 I & 0 & 0 \end{pmatrix} \begin{pmatrix} \widetilde{\delta c/c} \\ \delta Q^{-1}/Q^{-1} \\ \ln A_s \\ \ln A_r \end{pmatrix} = \begin{pmatrix} \delta t \\ \delta \ln A \\ 0 \\ 0 \\ 0 \end{pmatrix} \quad (5.12)$$

Except for the effect of smoothness on phase velocities, eqn 5.10 and eqn 5.12 are equivalent. The difference between the attenuation maps inverted from these two methods is indeed quite small. Interestingly, when we examine possible trade-offs between phase velocity and attenuation maps (between $\delta c/c$ and $\delta Q^{-1}/Q^{-1}$ in the first method and between $\widetilde{\delta c/c}$ and $\delta Q^{-1}/Q^{-1}$ in the second method), the trade-offs from the second method are roughly 10% less than the first method (although these trade-offs are already small for both methods, see sections below). In addition, $\widetilde{\delta c/c}$ is the quantity measured in most previous surface wave studies. To make the comparison with previous studies more straightforward, we will present results from inverting eqn 5.12 in the following sections. For simplicity, we will drop the tilde symbol $\sim$ for the rest of this chapter.

5.4.2 Regularization

We need to specify the three regularization parameters in eqn. 5.12. Figure 5.9 shows an example of the procedure for selecting these values for the 15 mHz data. We compute L-curves and aim to balance the data misfit (described by χ^2/N) and the model roughness (described by $\|Dx\|^2$). In step (a), we vary λ_1 for a combinations of λ_2 ($\lambda_2 = 1, 20, 300$) and λ_3 (between 0 and 20). For the

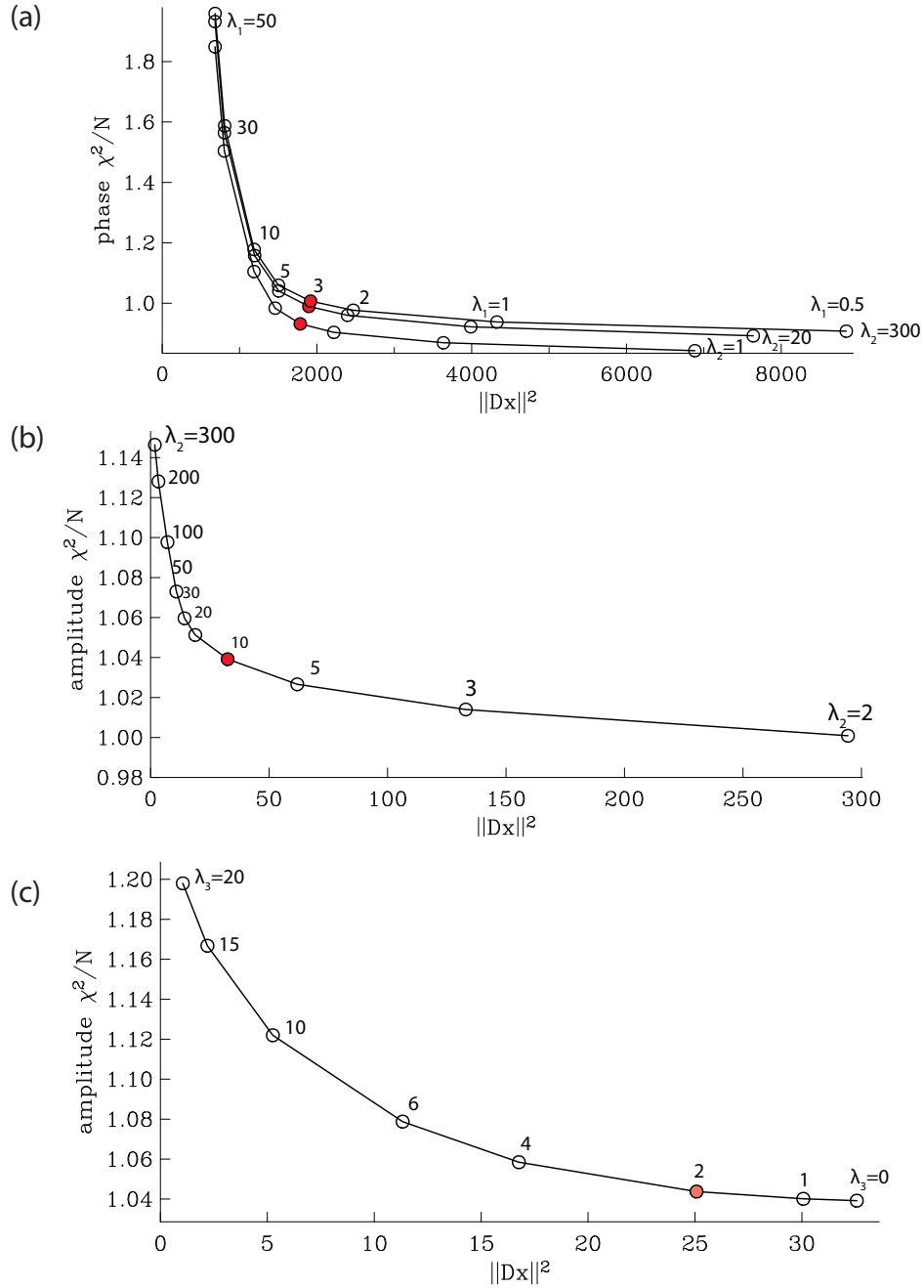


Figure 5.9: An example of choosing the regularization parameters for 15 mHz. We aim to balance the data misfit (described by χ^2/N) and the model roughness (described by $\|Dx\|^2$). (a) vary λ_1 for different λ_2 and λ_3 . These L-curves are plotted for $\lambda_3 = 0$ only for clarity. (b) fix $\lambda_1 = 3$ and vary λ_2 . Again, only the L-curve for $\lambda_3 = 0$ is plotted. (c) fix $\lambda_1 = 3$ and $\lambda_2 = 10$ with varying λ_3 . The red circle indicates our preference at each step.

Table 5.2: Detailed information for our datasets and inversions. Each column is: frequency (mHz); number of measurements; the cut off value for raw amplitude $|\ln A|$, over which the raw measurements are discarded; error of raw travel time data (t in sec) by looking at close-by stations 3° apart; error of raw amplitude data ($\ln A$) by looking at close-by stations 3° apart; λ_1 , λ_2 and λ_3 used in the inversion; χ^2/N for travel time data; χ^2/N for amplitude data; variance reduction for the travel time data; variance reduction for the amplitude data; the average 1D Q (computed as $1/\overline{Q^{-1}}$).

freq	num	lnA	error t	error lnA	λ_1	λ_2	λ_3	χ_t^2	$\chi_{\ln A}^2$	v.r. t	v.r. lnA	1D Q
5	140133	1.5	3.7	0.10	2	5	2	1.8	2.1	75%	63%	157
7.5	222651	2.0	3.7	0.13	2	10	2	1.4	1.4	91%	65%	119
10	247445	2.0	3.8	0.14	2	10	2	1.4	1.3	95%	70%	108
15	251993	2.0	4.5	0.19	3	10	2	1.0	1.0	97%	70%	119
20	174252	2.0	4.4	0.20	3	10	2	0.7	0.9	99%	77%	150
25	192535	2.0	4.4	0.24	3	10	2	0.8	0.8	99%	78%	190

purpose of clarity, only the L-curve for $\lambda_3 = 0$ is plotted. In this example, we choose $\lambda_1 = 3$. Decreasing λ_2 (i.e. increasing the contribution from attenuation) can indeed improve the fit to the phase arrival time data in the joint inversion, indicating the importance of physical dispersion when interpreting phase velocities (Karato, 1993; Ruan & Zhou, 2010). We can then fix $\lambda_1 = 3$ and select λ_2 and λ_3 in turn in step (b) and (c) (Figure 5.9).

5.4.3 Results

Our preferred model is shown in Figure 5.10. As in Dalton & Ekstrom (2006b), we obtain high attenuation structure along western America and the East Pacific rise, which indicates that the effect of focusing and defocusing is correctly accounted for (these authors show that these regions can appear to be high Q if the focusing terms are not handled correctly).

The largest signal in the long period phase velocity maps ($< 15\text{mHz}$) is the slow anomaly associated with the African rift system, suggesting that it has a deep origin. Low attenuation is associated with cratons in all continents. Large scale high attenuation regions correlate well with the mid-ocean ridge systems, including the East Pacific Rise, the Pacific-Antarctic Ridge, the Indian Ridges and the mid-

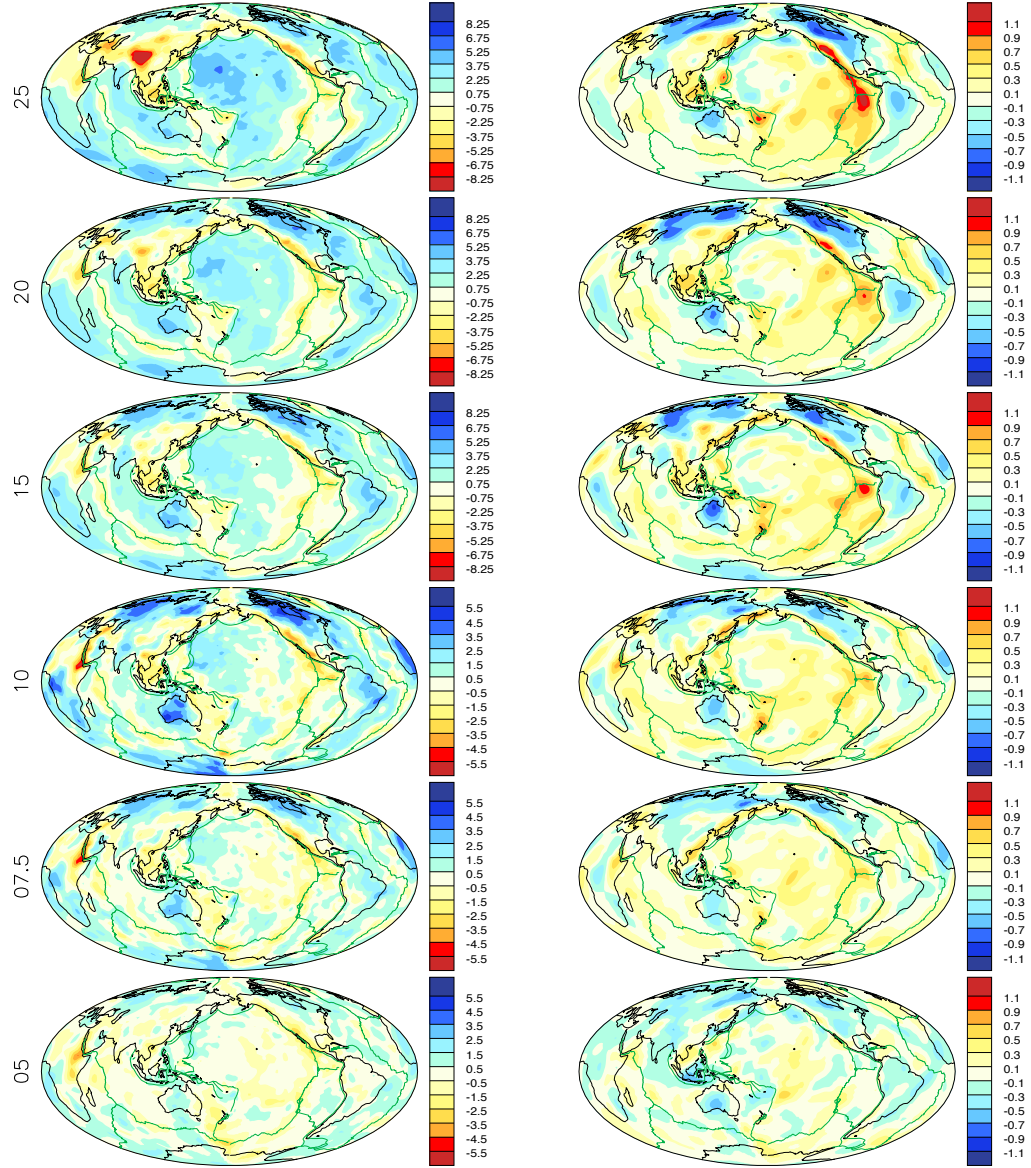


Figure 5.10: Results from the joint inversion for Rayleigh wave phase velocity (left) and attenuation perturbations (right) at different frequencies (mHz). The phase velocity perturbation is in $\delta c/c(\%)$ and attenuation is in $\delta Q^{-1}/Q^{-1}$. The reference value for each plot is taken to be the PREM prediction. Notice the change of color scales for the phase velocity maps.

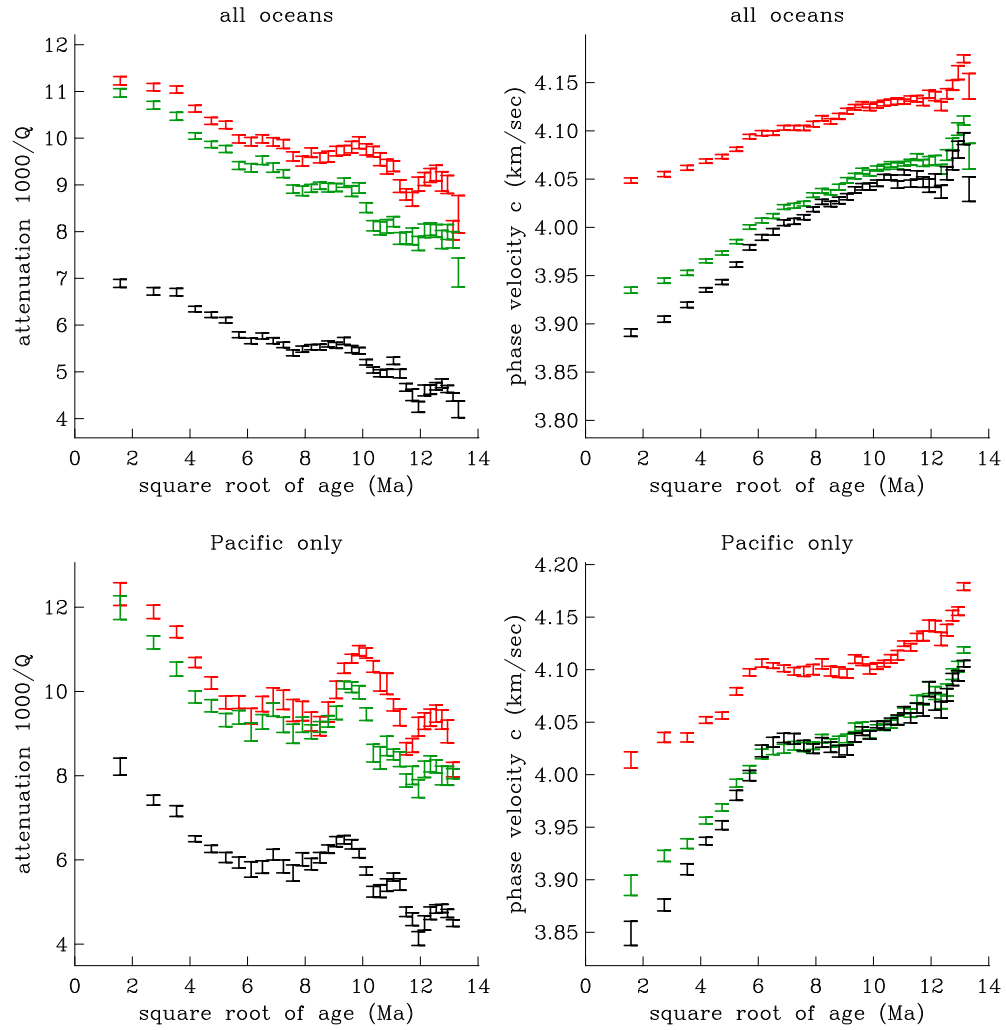


Figure 5.11: Attenuation and phase velocity variations in ocean basins. The bin width is 5 Ma. Note that the error bars are the errors of the mean value of each bin. Red: 10 mHz; green: 15 mHz; black: 25 mHz.

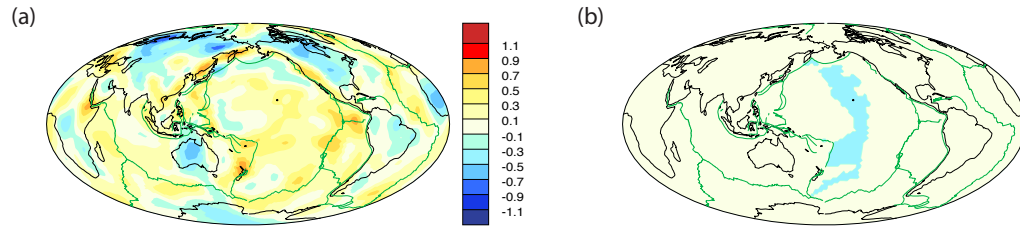


Figure 5.12: Location of the high attenuation structure in the Pacific seen in Figure 5.11. (a) The attenuation map for 10 mHz Rayleigh wave. (b) The region of the Pacific with sea floor ages between 70 and 110 Ma, which corresponds to the high attenuation structure seen in Figure 5.11. Sea floor age data is from Muller et al. (2008).

Atlantic Ridge. Some of the back-arc basins also have distinct high attenuation signals (e.g. Lau Basin, North Fiji Basin and Mariana Trough Basin). Iceland also has discernible high attenuation in our 10 and 15 mHz maps. It is interesting to see that part of the mantle wedge behind the Hellenic arc appears to have higher attenuation than its surrounding mantle. However, we are not able to recover the high attenuation structure beneath Central America (e.g. Rychert et al., 2008) in this global study.

Similar to elastic velocities (e.g. Nishimura & Forsyth, 1989; Ritzwoller et al., 2004; Maggi et al., 2006; Burgos et al., 2014; Ma et al., 2014), the attenuation structure also displays a clear age progression trend as a linear function of the square root of sea floor age although it appears that such a trend breaks down earlier than the trend for phase velocities (Figure 5.11). More interestingly, when we plot the data for the Pacific alone, a clear increase of attenuation can be seen around 70-110 Ma (see also Figure 5.12), slightly older than the region where a reheating process is proposed (Ritzwoller et al., 2004) and the region where a strong variation of anisotropy structure is observed (Ekström & Dziewonski, 1998). We believe that the observation of attenuation variations will provide complementary information to understand the thermal evolution of oceanic plates and the origin of the low velocity zone underneath (Stixrude & Lithgow-Bertelloni, 2005).

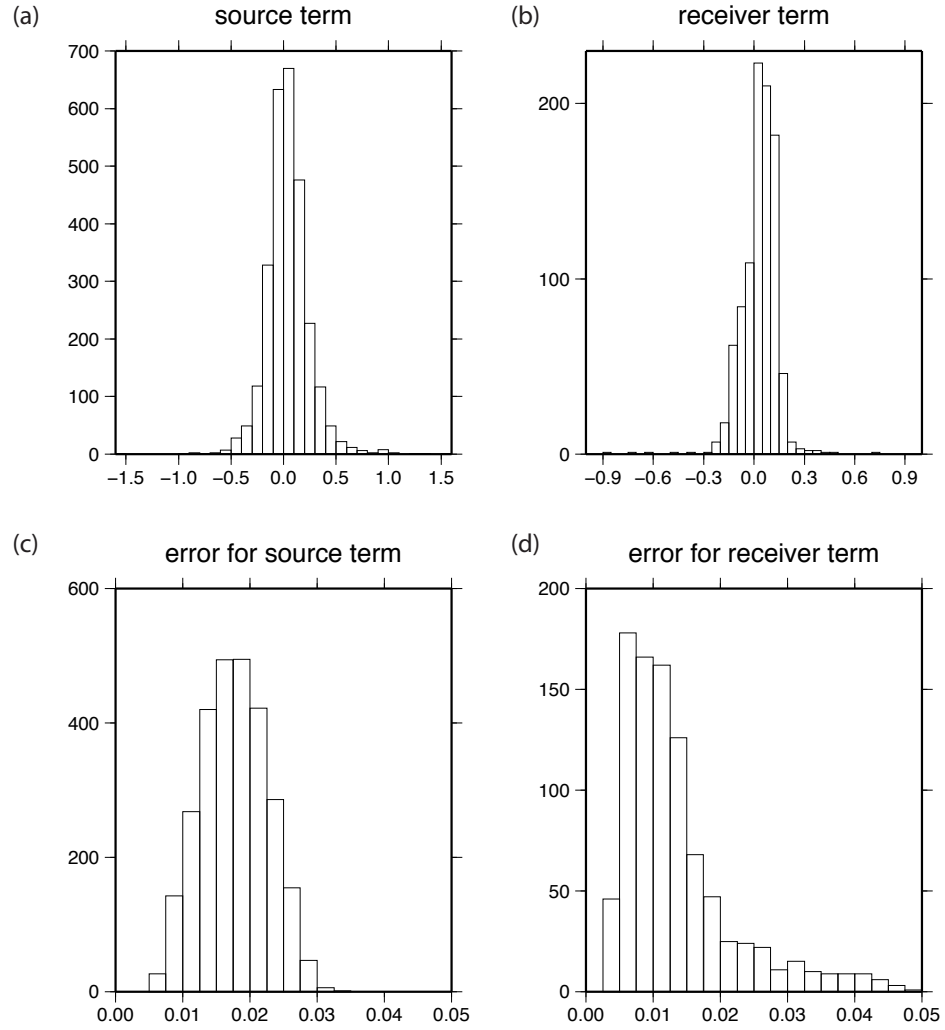


Figure 5.13: Distributions of the values and the estimated errors of the source and receiver terms. These are for 10 mHz Rayleigh waves.

Figures 5.13a and b show the distribution of the source and receiver terms for 10 mHz. The spread of the source terms is about 0.16. This is consistent with Dalton & Ekstrom (2006b) but larger than the effect of 3D Earth structure on the scalar moments estimated by Hjörleifsdóttir & Ekström (2010), which is of the order of 10%. This may be because our dataset contain earthquakes in more diverse tectonic settings with various crustal thickness. The station terms have smaller size though some stations require relatively large corrections. Some of these stations (GE-BOAB, TA-J17A, II-XPFO) have only a few months of data in our dataset so we cannot rule out other possible contaminations. Other stations (AZ-BDA2, AZ-HWB, US-MSO, CD-QIZ, CN-ULM) tend to have large errors over a few years. But unlike the problems shown in Figure 5.5, the errors for these stations are consistent throughout and can be corrected for.

5.4.4 Error analysis

We perform a Monte Carlo type error analysis (Houser et al., 2008) to determine how sensitive our models are to the noise in the data. Random Gaussian noise with a standard deviation the same as the estimated error in the measurements is added to the raw data. We then invert these perturbed data for a new model and repeat this process 50 times. The standard deviation of these 50 inverted models is taken to be the estimated error.

The errors for the source and receiver terms are plotted in Figure 5.13c and d. Compared to actual values of the observed source and receiver terms, the errors are small. However, we do observe that about 30 receiver terms (out of about 1000 examined) have slightly larger errors (> 0.05). These stations have less than 30 measurements associated with them and many of these are new TA stations that had only operated for a short period of time when their data were downloaded.

The errors for the phase velocity and attenuation maps are plotted in Figure 5.14a. The error for phase velocity is slightly smaller than half of the value estimated in our previous study (about 0.4%, see Ma et al., 2014) due to the fact that we are using a parameterization with coarser cells in this chapter. The error for attenuation is probably underestimated by this method as a damping parameter

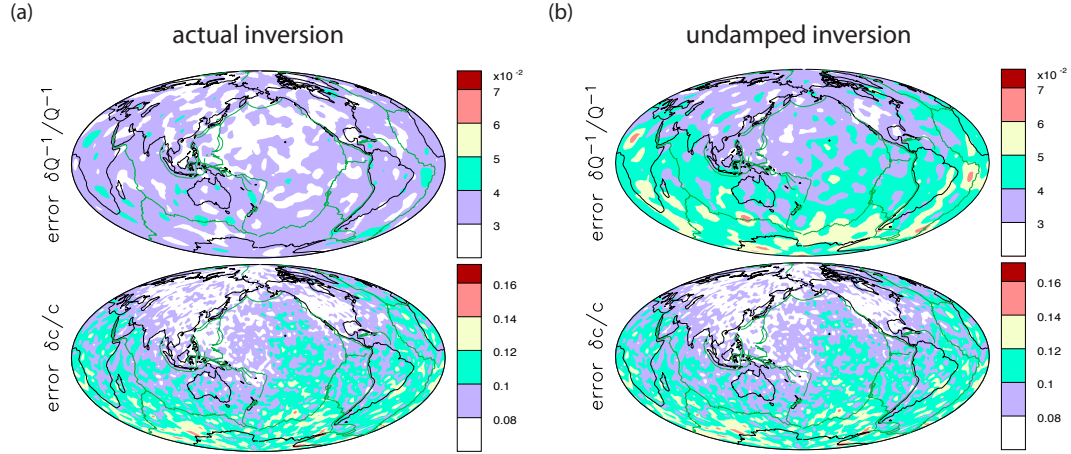


Figure 5.14: Results from an error analysis for 10 mHz Rayleigh waves. (left) results for the actual regularization parameters used in the inversion ($\lambda_1 = 2$, $\lambda_2 = 10$, $\lambda_3 = 2$); (right) results for an undamped version (i.e. by setting $\lambda_3 = 0$).

(λ_3 in eqn. 5.12) is used, although the error for an undamped inversion ($\lambda_3 = 0$) is still less than 0.08 (Figure 5.14b), which is much less than the signals we recover in the inversion (see Figure 5.10).

5.4.5 Resolution and trade-offs

A standard checker-board test is performed to investigate the resolution and the trade-offs in our inversions. Figure 5.15 shows one example for our 10 mHz data. Because of the damping and the use of a larger smoothing parameter, the resolution for attenuation is worse than the resolution for the phase velocity maps. However, structures at a scale of about 15° can still be well resolved in Eurasia and North America, while structures at a scale of about 30° (corresponding to $l = 12$) can be resolved globally.

The leakage to the source and station terms is about 0.01 and 0.008 respectively, which is negligible. We do observe leakages to elastic velocities from attenuation variations in areas with fewer ray paths, e.g. the Southern Oceans and Africa, but the leakages are generally very small (less than 0.5%, see Figure 5.15) compared to the actual signals we see in our inversion result.

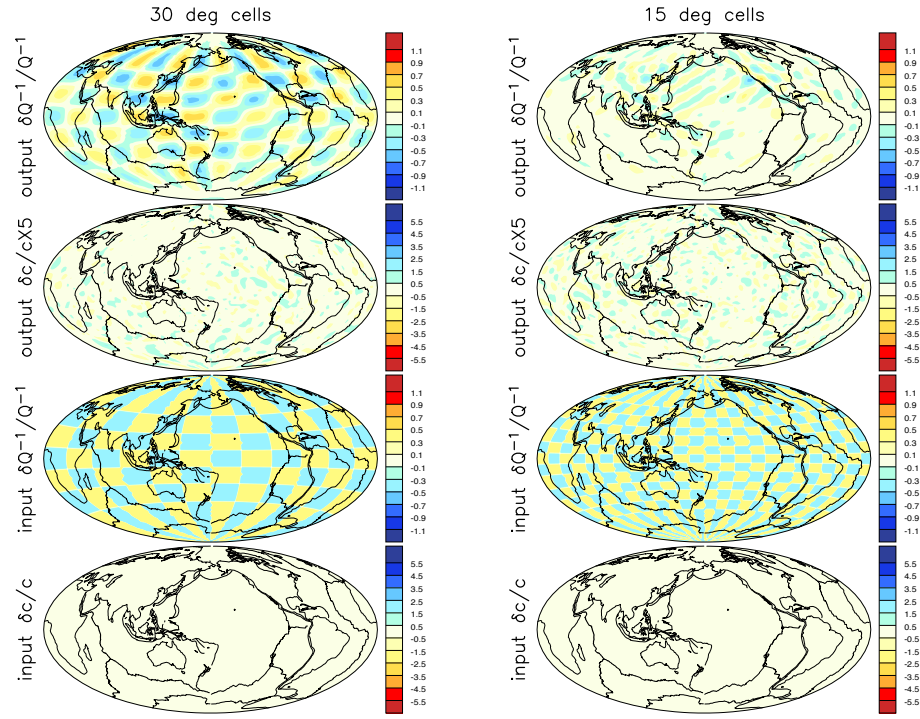


Figure 5.15: Checker-board tests for 10 mHz Rayleigh waves. We have also set the input $\delta c/c$ to be zero and check for possible cross contamination. Note that the output for $\delta c/c$ has been multiplied by five to make it more visible.

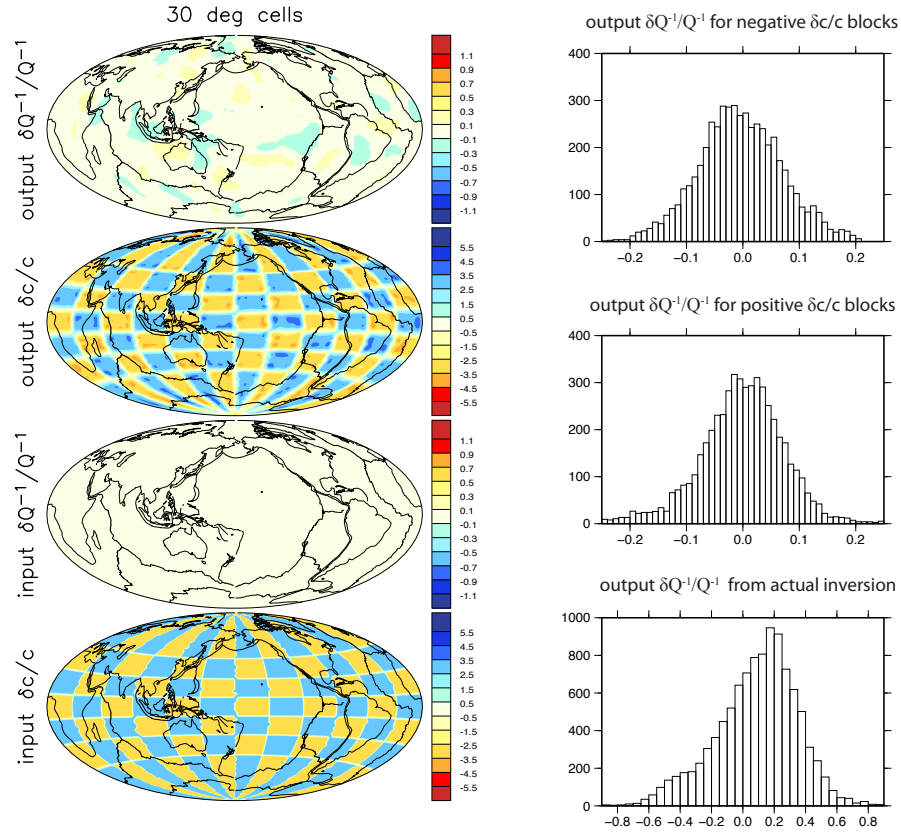


Figure 5.16: Checker-board tests for 10 mHz Rayleigh waves. The left panels show the checker-board patterns. In this case, we set the input $\delta Q^{-1}/Q^{-1}$ to be zero and check for leakages from $\delta c/c$. The right panels show the distribution of the leakage. As a comparison, the distribution of the $\delta Q^{-1}/Q^{-1}$ from the actual inversion is also plotted. Note that the distributions of the output $\delta Q^{-1}/Q^{-1}$ are centered at zero, regardless of the sign of the $\delta c/c$ in the input checker-board.

The leakage from elastic velocities to attenuation structure is stronger. The leakages have a standard deviation of 0.07, roughly 1/4 of the signals we observe (Figure 5.16). However, most of the leakages happen near sharp velocity boundaries of the checker-board ($\pm 3\%$ in our test). This is possible but is likely limited to few regions in realistic Earth models. But users of our models should still be cautious about interpreting signals below this level in our attenuation maps. It is worth pointing out that the distributions of the output $\delta Q^{-1}/Q^{-1}$ are centered at zero, regardless of the sign of the $\delta c/c$ in the input checker-board (Figure 5.16). This means that the high correlation between attenuation and phase velocity maps we see in Figure 5.10 is not caused by this trade-off.

5.5 Discussion

5.5.1 The average attenuation values

Figure 5.17 shows the global averaged Q values. Our averaged values fall between the prediction from PREM (Dziewonski & Anderson, 1981) and QL6 model (Durek & Ekstrom, 1996). They are slightly higher than the values from Selby & Woodhouse (2002) (see their Figure 5) but almost overlap with the values from Dalton & Ekstrom (2006b) (see their Figure 14) for periods longer than 50 s. We do observe a higher Q value at shorter periods. In general, compared to the QL6 model, our results suggest a higher value of Q (i.e. less attenuation) and that attenuation is more confined to the asthenosphere.

5.5.2 The importance of the source term

It is important to understand the relative contribution of each of the four terms in eqn. 5.1. Figure 5.18 shows the distribution of the raw data of 10 mHz Rayleigh wave and the residuals after accounting for different terms. The relative contribution of each term to the fit of the data is determined by the improvement of the variance reduction after adding each one. The variance reduction is computed as:

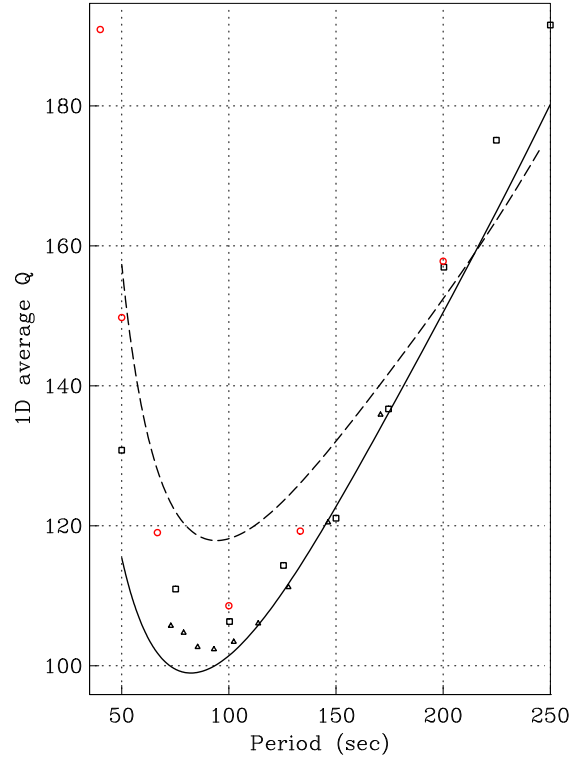


Figure 5.17: The 1D average values for attenuation (Q). Black solid line: prediction from QL6 (Durek & Ekstrom, 1996); Black dashed line: prediction from PREM (Dziewonski & Anderson, 1981); black triangles: values from Selby & Woodhouse (2002) ; black squares: values from Dalton & Ekstrom (2006b); red circle: values from this study.

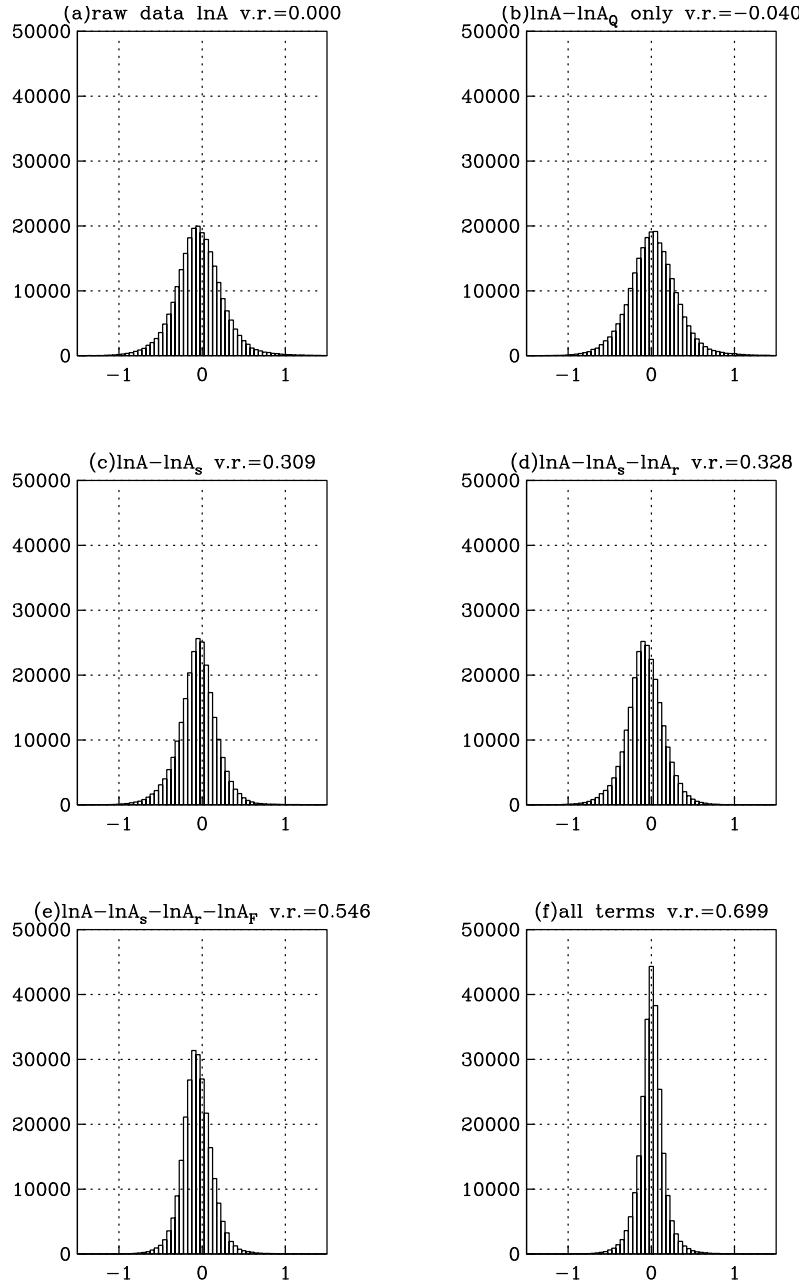


Figure 5.18: Distribution of raw amplitude data ($\ln A$) and residuals for 10 mHz. (a) Raw data. (b) Residuals after accounting for attenuation terms only – note the negative variance reduction. (c) Residuals after accounting for source terms. (d) Residuals after accounting for source and receiver terms. (e) Residuals after accounting for source, receiver and focusing terms. (f) Residuals after accounting for all four terms. v.r. means variance reduction.

$$\text{v.r.} = 1 - \frac{\sum (\ln A^{\text{obs}} - \ln A^{\text{pred}})^2}{\sum (\ln A^{\text{obs}})^2} \quad (5.13)$$

where $\ln A^{\text{obs}}$ is the raw amplitude measurement and $\ln A^{\text{pred}}$ is the predicted value after each term is added in the calculation.

The overall variance reduction for our 10 mHz dataset reaches 70%. However, the attenuation term alone gives negative variance reduction and does not improve the fit to the data at all (Figure 5.18b). It demonstrates again the importance of including the source, receiver and focusing-defocusing terms in order to obtain a reasonable attenuation structure (Dalton & Ekstrom, 2006b). It is in fact the source terms that provide the biggest contribution to the variance reduction.

We also observe an increasing spread in the range of the source terms as frequency increases (Figure 5.19). This is consistent with Dalton et al. (2014) (see their Figure 6) and Dalton & Ekstrom (2006b) (after adding back the U/ω terms in their Figure 10). Selby & Woodhouse (2002) also find that a single correction to the scalar moment of each event for all frequencies cannot explain their amplitude distributions. These observations suggest that local structure and source depths may have strong effect on surface wave amplitudes. As the global CMT project has recently included intermediate period surface waves (with a period range of about 50-150 s) in its routine analysis (Ekström et al., 2012), better velocity models for source regions and more robust estimates for source depths might be important in fitting waveforms in this period range.

5.5.3 Currently known limitations

With our current dataset, we are not able to reproduce the result of Dalton & Ekstrom (2006a), who show that phase velocity structure can be obtained by using amplitude data alone. We suspect that this is because only minor arc data are used in our study. While the attenuation terms decrease the amplitudes in both directions, the focusing-defocusing terms have different signs depending on the orbital directions (Woodhouse & Wong, 1986). It appears that this information is needed in order to constrain the phase velocity from amplitude data alone. But the exact reason is still under investigation.

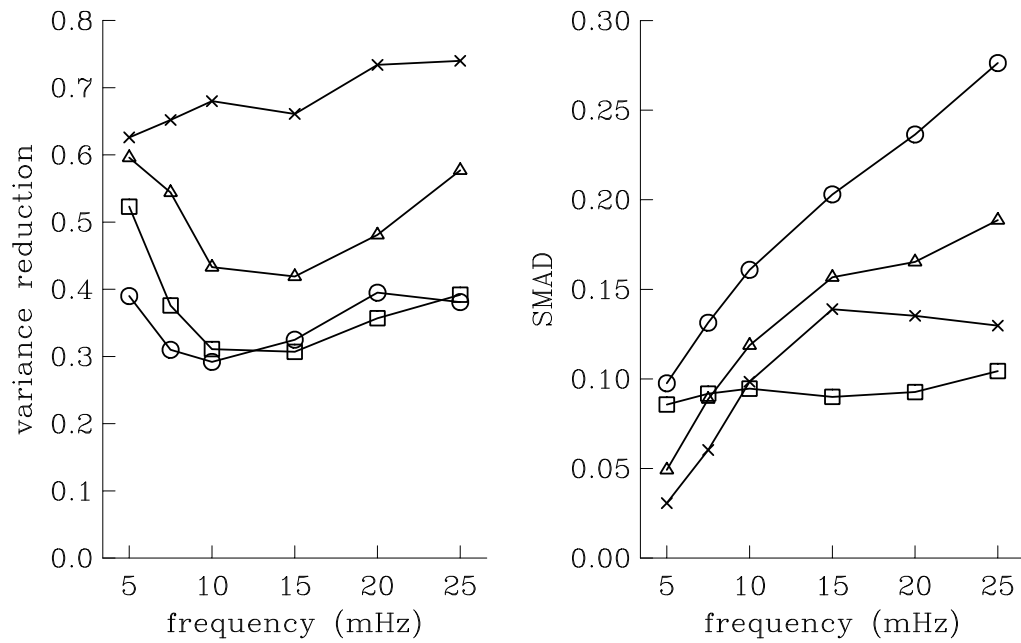


Figure 5.19: (Left) The relative contribution of each of the four terms to the fit of the data. Circle: source terms only; square: include both source and receiver terms; triangle: include source, receiver and focusing terms; cross: include all four terms. (Right) The scaled median absolute deviation of each of the four terms. Circle: source terms; square: receiver terms; triangle: focusing terms; cross: attenuation terms.

The need to remove heavily bent ray paths is another limitation. We have tried adding the source radiation pattern in the kernel calculation but the improvement (in terms of removing the artifacts seen in Figure 5.8) is rather minor. Dalton et al. (2014) show that exact ray theory outperforms finite frequency theory in predicting amplitude variations at short periods (50 s). Parisi et al. (2015) also suggest that, based on their numerical experiments, the first-order Born approximation may break down when velocity variation is large. While we have shown the capability of the current finite-frequency theory in explaining most of the data presented in this chapter, it is clear that further theoretical development is needed to include the ray bending effect.

5.6 Conclusion

This chapter utilizes the amplitude data we collected in our previous studies to infer the attenuation structure of the Earth’s uppermost mantle. We base our calculations for the focusing-defocusing effect on finite-frequency theory, which is less sensitive to high-order elastic structure and hence provides a better basis to invert for attenuation structure. A set of 2D phase velocity and attenuation maps for Rayleigh waves between 5 and 25 mHz are presented. These maps provide a good fit and give over 60-70% variance reduction to our amplitude data. This set of phase velocity and attenuation maps can be used to compute the amplitude variations for Rayleigh waves globally in a compact way. The attenuation maps show structures that correlate well with surface tectonics and the age progression trend of attenuation is clearly seen in the ocean basins. Our average Q values match previous results but are higher (i.e. less attenuation) at short periods. The source terms are the major contributions to the fit of the data and further work is still needed to better understand how the uncertainties from local structure and source depths can affect the resulting source parameters.

Acknowledgments

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Chapter 5, in full, has been submitted for publication of the material as it may appear in *Geophysical Journal International*: Ma, Z. and G. Masters, 2D global Rayleigh wave attenuation model by accounting for finite-frequency focusing and defocusing effect. I was the primary investigator and author of the paper.

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Appendix - Kernel calculation

This appendix describes details of calculating the finite frequency kernels used in this chapter.

The kernel for a single frequency is given by eqn. 5.3 in the main text. The wavenumber k is defined as $k = 2\pi\nu R/c$ where ν is the frequency (in Hz), R is the radius of the Earth (in km), c is the reference phase velocity (in km/s) at frequency ν and the wavenumber k has the unit rad^{-1} .

As a Gaussian band-pass filter is inherent in our measurement process, we further average the kernel over a frequency band:

$$\overline{K_A^c} = \frac{\int e^{-\alpha(\nu-\nu_0)^2} K_A^c d\nu}{\int e^{-\alpha(\nu-\nu_0)^2} d\nu} \quad (5.14)$$

where ν_0 is the center frequency of the Gaussian filter and its width is controlled by $\alpha = 3/(0.25\nu_0)^2$ (see Ma & Masters, 2014). The core of the integral,

$$\int e^{-\alpha(\nu-\nu_0)^2} \nu^{3/2} \cos[\nu \cdot \arg \cdot \frac{2\pi R}{c(\nu)} + \pi/4] d\nu \quad (5.15)$$

where $\arg = \Delta' + \Delta'' - \Delta$, can be pre-computed numerically and stored as a lookup table to facilitate later computations. Similar to Spetzler et al. (2002), we find that eqn 5.15 is changed negligibly if $c(\nu)$ is taken as a constant. Figure 5.20 compares the value of the cosine integral computed using different frequency averaging schemes. The blue line shows the value computed at a single frequency and truncated according to the width of the measurement window, which is between -250 s and 250 s (a total of five periods in our case). The green line shows the integral computed using a box-car function for the frequency averaging instead of the Gaussian filter. The box-car function is chosen so that it has the same area as the Gaussian function. The integral computed using the Gaussian filtering scheme indeed tapers to zero after a reasonable distance and this gives us a convenient way to choose the cut-off value. The cut-off values used are 0.6, 0.45, 0.35, 0.25, 0.20 and 0.15 for 5, 7.5, 10, 15, 20 and 25 mHz respectively. When $\arg = \Delta' + \Delta'' - \Delta$ is larger than the cut-off value, the integral is set to zero.

We use a 2° by 2° equal area blocks for our parameterizations. The kernel

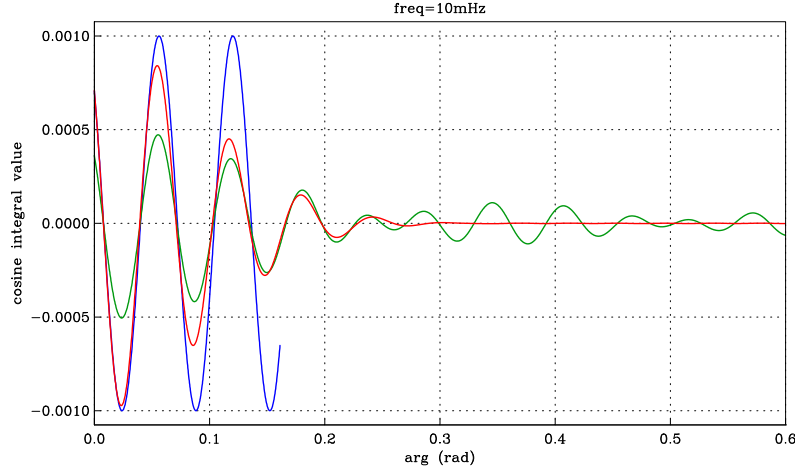


Figure 5.20: The value of the cosine integral computed using different frequency averaging schemes. blue: computed at a single frequency but truncated according to the width of our measurement window; green: computed using a box-car function for frequency averaging; red: computed using a Gaussian function for frequency averaging.

$\overline{K_A^c}$ is computed at a $0.2^\circ \times 0.2^\circ$ spacing and then averaged over each block. We note that eqn. 5.3 has singularities right at the source and receiver, as the asymptotic approximation $Y_l^m(\theta) \approx \cos[(l + 1/2)\theta + m\pi/2 - \pi/4]$ breaks down when θ is small. We choose to set the integral to zero when $\arg < 0.7/l$, where the asymptotic form deviates from the exact solution by 5%. This is a very small contribution even for the block right at the source or receiver (e.g. about 0.4° in the case for 10 mHz) and it has negligible effect on the final result.

Chapter 6

Conclusions and prospects for future study

6.1 Conclusions

Throughout this thesis, I have used long period surface waves to study the Earth's lithosphere on a global scale.

First of all, studying the global lithospheric structure requires a large dataset that can provide dense path coverage everywhere on the globe. To accomplish this goal, I have developed an efficient way to measure the dispersion properties and amplitude variations for both Rayleigh and Love waves. This method clusters similar waveforms recorded at different stations from a single event and allows users to make measurements on hundreds of waveforms, which are filtered at a series of frequency ranges, at the same time. It also requires a minimal amount of user interaction and allows easy assessment of data quality. This method allows hundreds of thousands of group/phase delay and amplitude variation measurements to be measured in a manageable time frame.

Secondly, these measurements are used to produce 2D phase and group velocity tomographic maps for the globe for periods between 25 s and 200 s. These maps provide a very compact way of representing the datasets and give over 90% variance reduction in the data for most frequencies for both Love and Rayleigh

waves. Large scale features of the lithosphere are well resolved in both the Rayleigh and Love wave cases. Slow anomalies are found beneath orogenic zones and other regions with thick crust, e.g. the Himalayas and Andes. The Basin and Range province in North America, mid-ocean ridges, and back-arc basins where the upper mantle is expected to be warmer than the surrounding regions also show up as slow anomalies. Cratons can be seen in the low frequency maps as regions of anomalously high phase velocity, e.g. the Canadian Shield, West African Craton, Congo Craton, Kaapvaal Craton, Indian Shield, Siberia Craton, Russian Platform, etc. It is also surprising to find that the largest signal in maps with frequencies lower than 10 mHz is observed in the east African rift, suggesting a very deep origin (at least 200-300 km) for the anomaly.

Thirdly, I have shown that the uncertainties of earthquake locations, which have been generally ignored, have significant impacts on the resulting azimuthal anisotropic structure. The differences of the azimuthal anisotropic patterns with and without source relocations are greatest near plate boundaries. Large differences around the South American plate are also identified. Although most of the earthquakes are shifted by less than 15 km from the CMT locations, earthquakes near the Andes can be systematically shifted by more than 30 km. The final epicenters for earthquakes on ridge-transform fault systems better match the plate boundaries.

Lastly, I use the amplitude variation data to infer the attenuation structure of the Earth's lithosphere. The calculation for the focusing-defocusing effect is based on the newly developed finite-frequency theory. Compared to ray theory, which has been used up to date, this new theory is less sensitive to the high-order elastic structure and hence provides a better basis for inverting the attenuation structure. A set of 2D phase velocity and attenuation maps for Rayleigh waves between 5 and 25 mHz are presented. These maps provide a good fit and give over 60-70% variance reduction to our amplitude data. The attenuation maps show structures that correlate well with surface tectonics and an age progression trend of the attenuation is clearly seen in the ocean basins. A combination of both the elastic phase/group velocity maps and the attenuation maps provides

complementary information to further improve our understanding of the structure and evolution of the Earth’s lithosphere.

6.2 Prospects for future study

In this thesis, I have studied many aspects of surface wave propagation and utilized such information to study the Earth’s lithosphere. One aspect that I have not studied is surface wave ray tracing, which describes how surface wave ray paths bend as they travel along the Earth surface. The bending of the ray paths can change the surface wave arrival angles at seismic stations. Measuring these arrival angles is useful as it can improve the resolution of both the isotropic and anisotropic structure (Laske & Masters, 1998). In Chapter 5 of this thesis, I suggest that ray bending can cause artifacts when modeling amplitude data. If we are to apply our method to infer attenuation in a more regional study where heterogeneities with large velocity contrasts exist, I believe including the effect of ray bending will be essential.

A more challenging topic is to extend the method developed in this thesis to measure surface wave overtones. This thesis focuses on studying the lithosphere. But if we are to study deeper Earth structure, using fundamental mode surface waves alone is not enough. The Earth structure below about 1000 km can be constrained by tele-seismic body waves and decades of studies have been devoted to that. However, the mantle transition zone, which is arguably one of the most important regions in understanding mantle dynamics, is still the least constrained part of the mantle in tomographic inversions. Overtone data, which have high sensitivity to structure between 300 km and 1000 km, would provide the crucial information to improve our understanding of this region (Ritsema et al., 2004).

We have experimented with a new technique, which is based on the cluster analysis technique described in this thesis and the concept of ‘mode stripping’ (van Heijst & Woodhouse, 1997), to measure overtone dispersion. One key aspect is to notice that, after correcting for dispersion, the waveforms of the fundamental mode surface waves are remarkably similar to each other (see Figures 2.3 and 3.4)

and can be removed effectively. This will reveal the higher mode signals, which typically have much smaller amplitudes. The second key aspect is to use many earthquake-station pairs that cover a large epicenter distance range and then turn them into a large aperture array. This can help us to separate mode branches that have similar arrival times.

The preliminary result looks promising and our second higher mode phase velocity map for 10 mHz Rayleigh wave compares well with the previous study (Visser et al., 2008) but with higher resolution. Currently, we have only analyzed the theoretically predicted dominant higher-mode in a given group velocity window at a particular frequency and event depth. Clearly, the next step is to develop a dispersion model for each mode branch to ultimately allow stripping of the dominant branch and to increase the number of measurements for the less dominant ones. The final goal is to combine these new overtone data, with the fundamental data I have collected in this thesis and existing body wave data to provide a new model of the whole mantle structure.

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Appendix A

Research notes

This appendix includes a number of research notes on various topics. The main purpose of these materials is to provide future students who are interested in long period global seismology with useful manuals and clarifications on topics that are not easy to find in the standard literature.

A.1 Building a dataset of long period seismic waveforms

This section provides a step-by-step guide to build a dataset of long period seismic waveforms.

1. Assemble an earthquake catalog. The Preliminary Determination of Epicenters (PDE) Bulletin can be found on <http://earthquake.usgs.gov/data/pde.php>. I have only used data format ‘ehdf’ during my Ph.D but two newer versions (‘isf2’ and ‘QuakeML’) are available now. The Centroid Moment Tensor (CMT) solutions can be downloaded from <http://www.globalcmt.org/CMTfiles.html>. The EHB bulletin from the International Seismological Centre (ISC) can be downloaded from <http://www.isc.ac.uk/ehbbulletin/>.
2. Request the SEED files from the Incorporated Research Institutions for Seismology (IRIS). The Standard for the Exchange of Earthquake Data

(SEED) is the standard data format to exchange seismological time series (<http://ds.iris.edu/ds/nodes/dmc/data/formats/>). I use `breq_fast` to send the request files (http://ds.iris.edu/ds/nodes/dmc/manuals/breq_fast/). Peter Shearer's group uses the software SOD, which makes the process more automatic. The software can be found in <http://ds.iris.edu/ds/nodes/dmc/software/downloads/sod/>.

3. Convert the SEED files to formats of your choice. I convert all seed files to the gfs format system. You can also convert them to efs (Peter's system).
4. Assemble a directory of instrument responses. The dataless SEED files that contain the instrument response can be downloaded from [ftp.iris.washington.edu/pub/RESPONSES/DATALESS_SEEDS](ftp://ftp.iris.washington.edu/pub/RESPONSES/DATALESS_SEEDS). Note that the OBS stations have their own SEED file (`_OBSIP.dataless`). You can then use the software 'evalresp' provided by IRIS to extract the information from these dataless SEED files.
5. Correcting instrument response from the raw data. The way I do this is to normalize all responses to station SR-GRFO, which is an old station and has one of the narrowest sensitivity band. You can also correct the instrument response by using your own subroutines or SAC. A useful tutorial for using SAC is <http://geophysics.eas.gatech.edu/people/jwalter/sacresponse.html>. Note that if you choose to use SAC, you should **NEVER** use the default command 'transfer from polezero ...' as this only corrects for the response from stage 1 (usually the Laplace transform analog response stage) and ignores all subsequent stages related to the data loggers.

A.2 Configure MPI

This section provides a step-by-step guide to configure the Message Passing Interface (MPI) environment, which is widely used in parallel computing. As an example, I will describe the steps to connect my local machine (using C-Shell) and a barnyard machine (using Bash shell) at IGPP (e.g. clyde.ucsd.edu). The package

I am using is Open-MPI <http://www.open-mpi.org/>. Running jobs on the Triton Shared Computing Cluster (TSCC) is a lot easier as most of the configurations have already been set up.

Most of the information described below is found in <http://www.open-mpi.org/faq/?category=running> (henceforth FAQ).

1. Download and install Open-MPI. You can download the source codes and find the instruction on <http://www.open-mpi.org/software/ompi/v1.8/> and <http://www.itp.phys.ethz.ch/education/hs11/pt/openmpi.pdf>. Note that the package should be installed on both your local machine and the barnyard machine. The `--prefix` parameter is the only thing that needs to be changed.
2. Set up the SSH. The goal of this step is to set up an SSH environment that does not require to type in passwords manually. The instructions are on <https://www.open-mpi.org/faq/?category=rsh>. After the setup, you need to type:

```
eval `ssh-agent -c`
```

```
ssh-add ~/.ssh/id_rsa
```

```
ssh-agent -k (to kill ssh-agent after you are done)
```

(I also find it convenient to let my computer remember the keychain so I do not need to retype it.)

3. Set up the PATH variables. You can follow item 1 and 4 in the FAQ. On the barnyard machine, include these in BOTH `~/.bashrc` and `~/.bash_profile` (remember to change them to the name of your home directory)

```
export PATH=$PATH:/home/zma/openmpi/bin:/home/zma/mpi.dir/
```

```
export LD_LIBRARY_PATH=/home/zma/openmpi/lib
```

On the local machine, include these in `~/.cshrc`

```
setenv PATH $PATH:/Users/zma/openmpi/bin
```

```
setenv LD_LIBRARY_PATH /Users/zma/openmpi/lib
```

All of my codes on both machines are in `$HOME/mpi.dir/` and the path `/home/zma/mpi.dir/` is needed for `mpirun` to find the executables.

4. Test run. Now you can follow item 11 in the FAQ to test the connections. The example codes are provided with the MPI package in the tar ball. Remember to setup the hostfile and to turn on the ssh-agent (described in step 2 above).

In addition, an excellent source for learning programming in MPI is <https://www.coursera.org/course/scicomp>.

A.3 Seismic anisotropy – some definitions

This note describes some definitions commonly used in the literature on seismic anisotropy. These definitions are spread in various textbooks and papers with different notations. I find it useful for future students if these definitions are organized into one common place. Two typos in the classic papers Montagner & Nataf (1986) and Montagner (2007) are also pointed out. This note is prompted by the discussion with Yi Cao at Seoul National University, South Korea.

A.3.1 Voigt notation

In general, the elastic stiffness tensor c_{ijkl} relates stress and strain in the following way:

$$\sigma_{ij} = c_{ijkl}\epsilon_{kl} \tag{A.1}$$

The Voigt notation means using these substitutes: $xx \rightarrow 1, yy \rightarrow 2, zz \rightarrow 3, yz, zy \rightarrow 4, xz, zx \rightarrow 5$, and $xy, yx \rightarrow 6$. Using the Voigt notation, the constitution relation is written as:

$$\begin{bmatrix} \sigma_{11} \\ \sigma_{22} \\ \sigma_{33} \\ \sigma_{23} \\ \sigma_{13} \\ \sigma_{12} \end{bmatrix} = \begin{bmatrix} c_{1111} & c_{1122} & c_{1133} & c_{1123} & c_{1113} & c_{1112} \\ c_{1122} & c_{2222} & c_{2233} & c_{2223} & c_{2213} & c_{2212} \\ c_{1133} & c_{2233} & c_{3333} & c_{3323} & c_{3313} & c_{3312} \\ c_{1123} & c_{2223} & c_{3323} & c_{2323} & c_{2313} & c_{2312} \\ c_{1113} & c_{2213} & c_{3313} & c_{2313} & c_{1313} & c_{1312} \\ c_{1112} & c_{1222} & c_{1233} & c_{1223} & c_{1213} & c_{1212} \end{bmatrix} \begin{bmatrix} \epsilon_{11} \\ \epsilon_{22} \\ \epsilon_{33} \\ 2\epsilon_{23} \\ 2\epsilon_{13} \\ 2\epsilon_{12} \end{bmatrix} \quad (\text{A.2})$$

Notice that the coefficient 2 on the strain is necessary to make eqn. (1) and (2) equivalent (because of the summation over k, l). The symmetry property of the elastic tensor c_{ijkl} is explained in details in the Appendix A.1.2 in Dellinger (1991).

The compliance tensor is usually defined as

$$\epsilon_{ij} = s_{ijkl} \sigma_{kl} \quad (\text{A.3})$$

Under the Voigt notation, this is actually:

$$\begin{bmatrix} \epsilon_{11} \\ \epsilon_{22} \\ \epsilon_{33} \\ 2\epsilon_{23} \\ 2\epsilon_{13} \\ 2\epsilon_{12} \end{bmatrix} = \begin{bmatrix} s_{1111} & s_{1122} & s_{1133} & 2s_{1123} & 2s_{1113} & 2s_{1112} \\ s_{1122} & s_{2222} & s_{2233} & 2s_{2223} & 2s_{2213} & 2s_{2212} \\ s_{1133} & s_{2233} & s_{3333} & 2s_{3323} & 2s_{3313} & 2s_{3312} \\ 2s_{1123} & 2s_{2223} & 2s_{3323} & 4s_{2323} & 4s_{2313} & 4s_{2312} \\ 2s_{1113} & 2s_{2213} & 2s_{3313} & 4s_{2313} & 4s_{1313} & 4s_{1312} \\ 2s_{1112} & 2s_{1222} & 2s_{1233} & 4s_{1223} & 4s_{1213} & 4s_{1212} \end{bmatrix} \begin{bmatrix} \sigma_{11} \\ \sigma_{22} \\ \sigma_{33} \\ \sigma_{23} \\ \sigma_{13} \\ \sigma_{12} \end{bmatrix} \quad (\text{A.4})$$

If we denote the stiffness and compliance matrixes in eqn (A.2) and (A.4) as C and S , we can easily see that

$$SC = CS = I \quad (\text{A.5})$$

A.3.2 Christoffel matrix

The Christoffel matrix B_{jl} , from which V_p and V_s can be computed, is defined as (Dahlen & Tromp, 1998, eqn 3.155)

$$\rho B_{jl} = c_{ijkl} \hat{k}_i \hat{k}_k \quad (\text{A.6})$$

where ρ is the density and $\hat{k}_i$ is one of the three components of the unit wavevector (i.e. the direction of the phase propagation vector $\hat{k}$). The polarization vectors and velocities are then computed by solving the following eigenvalue/eigenvector equation (Dahlen & Tromp, 1998, eqn 3.154)

$$B \cdot a = c^2 a \quad (\text{A.7})$$

where a gives the polarization vectors and c gives the wave speeds.

Dellinger (1991) gives an equivalent way to do the calculation using the Voigt notation (A.15 in his thesis)

$$DCD^T v_n = -\rho c^2 v_n \quad (\text{A.8})$$

where

$$D = \frac{1}{k} \begin{bmatrix} k_x & 0 & 0 & 0 & k_z & k_y \\ 0 & k_y & 0 & k_z & 0 & k_x \\ 0 & 0 & k_z & k_y & k_x & 0 \end{bmatrix} \quad (\text{A.9})$$

A proof for the equivalence between the two methods is simply to notice that

$$B_{jl} = D_{ji} C_{ik} D_{kl}^T = D_{ji} D_{lk} C_{ik} \quad (\text{A.10})$$

and

$$D_{ji} C_{i*} = \hat{k}_i c_{ji**} \quad (\text{A.11})$$

A.3.3 Computing Vp and Vs

Eqn A.7 gives a straightforward way to compute Vp and Vs propagating in any directions and gives the correct polarized vectors. This method should always give the correct answer, no matter how strong the anisotropy is.

It is interesting to compare the result with the classic study by Crampin (1984). Keep in mind that the eqn 10 in Crampin (1984) only works for weak azimuthal anisotropy.

For harzburgite, the difference in Vs is only about 0.1-0.2%. The difference between the two methods are large for antigorite, because of its strong anisotropy. In this case, we cannot use Crampin's method and should use the Christoffel equation. The data for harzburgite and antigorite are provided by Yi Cao.

A.3.4 Typos in two classic papers

I would also like to point out a few typos in Montagner & Nataf (1986) and Montagner (2007) in order to avoid future confusion:

(1) The equations for body waves in Montagner & Nataf (1986) (after eqn 5c) is the same as Crampin (1984). However, the definition of D and F in Montagner & Nataf (1986) are different from Crampin's paper but Montagner and Nataf forgot to change the notations accordingly. Montagner (2007) has made this correction.

(2) The definition of E_c (between eqn 6 and 7) in Montagner (2007) is incorrect. The E_c given in Montagner (2007) is

$$E_c = \frac{1}{8}(C_{11} + C_{22}) + \frac{1}{4}C_{12} - \frac{1}{2}C_{66} \quad (\text{A.12})$$

However, the coefficient in front of C_{12} should be $-1/4$ instead. This can be easily verified in the transverse anisotropic case with

$$C = \begin{bmatrix} A & A - 2N & F & 0 & 0 & 0 \\ A - 2N & A & F & 0 & 0 & 0 \\ F & F & C & 0 & 0 & 0 \\ 0 & 0 & 0 & L & 0 & 0 \\ 0 & 0 & 0 & 0 & L & 0 \\ 0 & 0 & 0 & 0 & 0 & N \end{bmatrix} \quad (\text{A.13})$$

Fortunately, the definition of E_c in Montagner & Nataf (1986) (or C_c in their notations) is correct.

A.4 A technical issue when inverting for surface wave azimuthal anisotropy

This note deals with a very technical issue when inverting for azimuthal anisotropy. The main conclusion is that for the G kernel, the results from Montagner & Nataf (1986) and Mochizuki (1986) are the same. However for the other kernels, these two results are the same only when $2U \ll l(l+1)V$. This condition may not be true for small l and for regions near the surface.

A.4.1 The concern

The theory of the azimuthal dependence of surface wave phase velocities in an arbitrarily stratified half-space was first given in Smith & Dahlen (1973) with a minor correction to it (Smith & Dahlen, 1975) (henceforth SD73). Montagner & Nataf (1986) (henceforth MN86) adopted a much simpler approach to the same problem and give essentially the same results. The main results are summarized in equations (16), (17), (20) and (21) in SD73 and equations (1)-(5) in MN86.

It is straight forward to show that these two methods are equivalent by converting the Cartesian components of the elastic tensor γ_{ijkl} in MN86 to the corresponding canonical harmonic components $\gamma_{\sigma}^{lm\phi}$ using the list in the appendix 1 in SD73.

MN86 also derived the partial derivatives (the kernels) of the azimuthal terms with respect to the elastic tensors (the B , G , H , C terms in their equation 5). In the case of a flat-layer model, these kernels can be computed using to kernels for the A , C , F , L , N terms, which are described in eqn A2 and A3 in Dziewonski & Anderson (1981) (henceforth DA81).

The theory is correct up to this point. However, these authors further assumed that the statement above still holds true for a spherical Earth. The purpose of this note is to investigate when and whether this assumption is valid by comparing the MN86 results with a more rigorous approach for the spherical earth proposed by Mochizuki (1986) (henceforth M86).

A.4.2 The verification

First, we substitute the contravariant components of the elastic tensor $C^{\alpha\beta\gamma\delta}$ in equation (13) in M86 with its corresponding Cartesian components, which are listed in the Appendix A in M86 (and they are equivalent to those listed in the Appendix 1 in SD73). We use these substitutions $\theta \rightarrow 1$, $\phi \rightarrow 2$ and $r \rightarrow 3$.

The 0ϕ term is:

$$\begin{aligned}
& \int \dot{U}^2 C_{3333} + \frac{1}{2} l(l+1)(l+2)(l-1) \frac{V^2}{r^2} \left(\frac{1}{4} C_{1111} + \frac{1}{4} C_{2222} - \frac{1}{2} C_{1122} + C_{1212} \right) \\
& + \frac{[2U - l(l+1)V]^2}{r^2} \left(\frac{1}{4} C_{1111} + \frac{1}{4} C_{2222} + \frac{1}{2} C_{1122} \right) \\
& + \frac{1}{r} [2U - l(l+1)V] \dot{U} (C_{1133} + C_{2233}) \\
& + \frac{1}{2} l(l+1) \left(\dot{V} + \frac{U-V}{r} \right)^2 (C_{1313} + C_{2323}) r^2 dr \\
& = \int \dot{U}^2 C + l(l+1)(l+2)(l-1) \frac{V^2}{r^2} N + \frac{[2U - l(l+1)V]^2}{r^2} (A - N) \\
& + \frac{2}{r} \dot{U} [2U - l(l+1)V] F + l(l+1) \left(\dot{V} + \frac{U-V}{r} \right)^2 L r^2 dr
\end{aligned}$$

We can see that this is exactly the same as equation (A2) in DA81, as expected.

The $+2\phi$ term for Rayleigh wave is:

$$\begin{aligned}
& l(l+1) \int \frac{V}{r} \dot{U} \left[\frac{1}{2} (C_{1133} - C_{2233}) - i C_{1233} \right] \\
& - \frac{1}{2} \left(\dot{V} + \frac{U-V}{r} \right)^2 \left[\frac{1}{2} (C_{1313} - C_{2323}) - i C_{1323} \right] \\
& - \frac{V}{r^2} [2U - l(l+1)V] \left[-\frac{1}{4} (C_{1111} - C_{2222} + \frac{i}{2} (C_{1112} + C_{1222})) \right] r^2 dr \\
& = l(l+1) \int \frac{V}{r} \dot{U} (H_c - i H_s) - \frac{1}{2} \left(\dot{V} + \frac{U-V}{r} \right)^2 (G_c - i G_s) \\
& - \frac{V}{r^2} [2U - l(l+1)V] \left(-\frac{1}{2} B_c + i \frac{1}{2} B_s \right) r^2 dr
\end{aligned}$$

The $+4\phi$ term for Rayleigh wave is:

$$\begin{aligned}
& l(l+1)(l+2)(l-1) \cdot \\
& \int \frac{1}{4} V^2 \left[\left(\frac{1}{4} C_{1111} + \frac{1}{4} C_{2222} - \frac{1}{2} C_{1122} - C_{1212} \right) - i (C_{1112} - C_{1222}) \right] dr \\
& = l(l+1)(l+2)(l-1) \int \frac{1}{4} V^2 [2C_c - 2iC_s] dr
\end{aligned}$$

Use equation (14) in M86 to combine the -2ϕ and -4ϕ terms, we have the following table:

kernel	M86	MN86
H	$-\frac{2V\dot{U}}{r}l(l+1)$	$-\frac{2\dot{U}}{r}[l(l+1)V - 2U]$
G	$\left(\dot{V} + \frac{U-V}{r}\right)^2 l(l+1)$	$\left(\dot{V} + \frac{U-V}{r}\right)^2 l(l+1)$
B	$\frac{[l(l+1)V - 2U]l(l+1)V}{r^2}$	$\frac{[2U - l(l+1)V]^2}{r^2}$
C	$\frac{V^2}{r^2}l(l+1)(l+2)(l-1)$	$\frac{[2U - l(l+1)V]^2}{r^2}$

Therefore, we can see that for the G kernel, the two methods are the same. However for the other kernels, the results from M86 and MN86 are the same only when $2U \ll l(l+1)V$. This condition may not be true for small l and for regions near the surface. To be safe, the kernels derived in M86 are recommended and this poses almost no extra computing cost.

A.4.3 Typos in SD73

There are a couple extra typos in SD73 that are not included in the original corrections. In eq (21) in SD73. The $(V'/k - V)^2$ terms should be $(V'/k - U)^2$. The coefficient of γ_S^{42c} (appendix 1) in the equation for γ_{1111} should be 1, instead of 6.

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