

Chapter 4

Ambient Noise Cross-Correlations

4.1 Introduction

In observational seismology, the major tools for imaging the Earth are earthquakes and explosives. Over more than a century, earthquakes have been recorded, located and used in understanding the Earth. The traveltimes of the body waves; P and S waves, the dispersion of the group and phase velocities of the surface waves and the information derived from the normal modes of the Earth gave information to delineate the structure of the Earth's interior. Similarly, explosion based seismology, or in other words controlled source seismology, made possible the imaging of Earth's relatively shallow parts with greater resolution than from earthquakes. Both of the imaging methods rely on the presence of a specific source which creates a response from the Earth. The recordings that are collected mainly on the surface contain the responses of the Earth which have been created by the interaction of waves along their propagation paths. The intensity and the detail of the imaging of the Earth has significantly increased in the last 30 years with the introduction of the digital seismology.

The Earth itself involves a complex system of interactions between its climate, ocean and lithosphere which operate continuously. One of the results of these interactions is called the random seismic wavefield or with the common name as *seismic noise*. There are various causes of seismic noise that leave different marks on the spectrum. The most energetic component is the oceanic microseisms which is a result of the interaction of atmosphere, ocean and the coast. Perturbations in the atmosphere due to strong storms impact on the ocean to set up standing wave patterns which create continuous pressure on the sea bottom, with variable intensity. The disturbance of the sea bottom results in the emergence of the elastic waves as for an earthquake or an explosion. The major difference is that the random wavefield has a chaotic nature in contrast to earthquakes or explosive generated elastic waves. A number of studies have attempted to formulize and locate the

main sources of such microseisms; Darbyshire (1990); Kedar & Web (2005); Longuet-Higgins (1950); Rhie & Romanowicz (2004).

The spectrum of the microseisms has a distinct characteristics. To illustrate the spectrum of the microseisms, four different stations close to the coast were taken across Australia. With a chosen 30 days of averaging, the power spectral density for each station was calculated and given in figure 4.1. The dominating peaks between 0.09-0.18 Hz are associated with the influence of the microseisms which arise from the standing waves induced by strong storms in the deep ocean (Kennett, 2001, p 8). In addition to this, wave surf can be significant in the frequency range of 0.05-0.1 Hz. The variability of the spectrum of the main contributions is evident for the calculated stations. The highest peak for TL01 is close to 0.2 Hz, on the other hand at MG01, the highest peak is observed at 0.1 Hz. These subtle differences can be attributed to the differences in underlying processes of atmosphere ocean interaction which creates the microseisms for different locations in particular the water depth where the microseisms were generated.

The standing waves in the ocean also send pressure waves with the atmosphere as microbaroms that can couple to the Earth to produce seismic signals. Although weaker than the microseisms, the microbaroms can provide local excitation.

In recent years, the extraction of Green's function from the ambient noise field has emerged as an important tool to understand the Earth. The resultant waveform from the cross-correlation of random waves recorded at two different stations corresponds to the Green's function as if an impulsive force is applied at the one station and recorded at the other station. For geophysics, it was first shown by Claerbout (1968) that the autocorrelation of the transmission response of the Earth corresponds to the superposition of the reflection response and its acausal counterparts. This approach has been successfully applied in helioseismology by Duvall et al. (1993). However, this explanation is strictly 1-D and is not enough to describe the success of the cross-correlation technique for large interstation distances. Recent work from Shapiro & Campillo (2004); Shapiro et al. (2005) showed that by using two stations, it is possible to extract information along the connecting path by just using the Earth's noise. This new idea created a an opportunity to use an important part of the recorded wavefield of the dense networks on the Earth which is normally neglected and classified as *ineffective*.

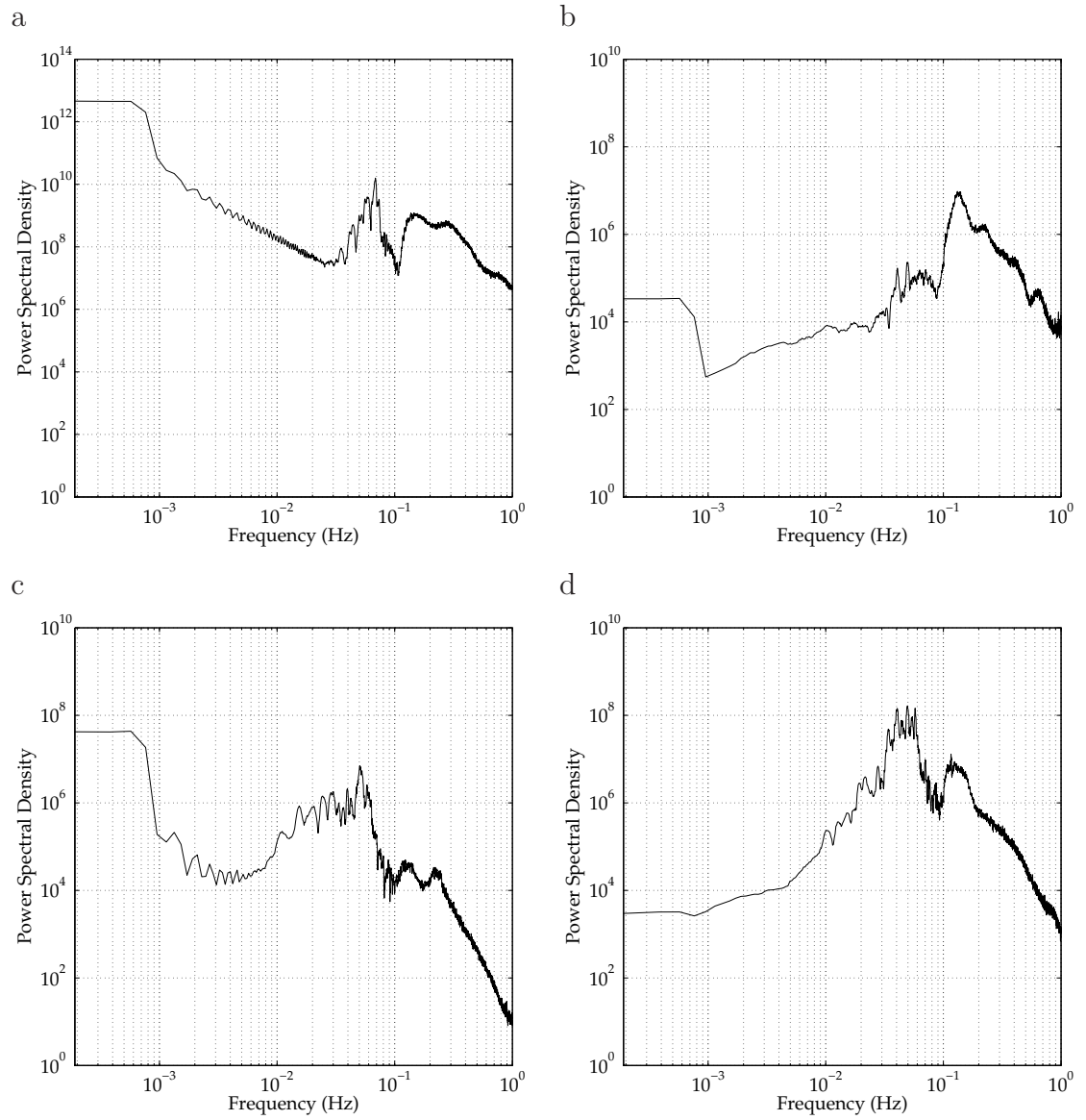


Figure 4.1: Power spectral density of microseisms recorded at 4 different stations across Australia for 30 days of averaging. a) LP01 (Northwestern Australia). b) TL01 (Northern Queensland). c) ARMA (Northern NSW). d) MG01 (South Australia).

In the following section, we show how the Green's function for an elastic wavefield is related to convolutional and correlational integrals over surfaces that can in certain circumstances be reduced to simpler, practical results.

4.2 Theoretical Background

4.2.1 Representation Theorem

We write the equation of motion for displacement as,

$$\frac{\partial \tau_{ij}}{\partial x_i} - \rho \frac{\partial^2 u_j}{\partial t^2} = -f_i, \quad (4.1)$$

where τ_{ij} is the stress tensor, x_i is the spatial variable, ρ is the density of the material, u_j is the displacement field, t is the time variable and f_j is the body force field. For convenience, the relation in eq.(4.1) can be written in frequency domain by taking the Fourier transformation in time domain,

$$\frac{\partial \tau_{ij}}{\partial x_i} + \rho \omega^2 u_j = -f_i, \quad (4.2)$$

where ω is the angular frequency.

The stress tensor is linked to the equation of motion in eq.(4.1) and eq.(4.2) by the generalized Hooke's law

$$\tau_{ij} = c_{ijkl} \frac{\partial u_k}{\partial x_l}, \quad (4.3)$$

where c_{ijkl} is the elastic modulus which has the following symmetries

$$c_{ijkl} = c_{jikl} = c_{ijlk} = c_{klij}. \quad (4.4)$$

4.2.2 Betti's Theorem

The dynamic extension of Betti's theorem on reciprocity can be shown as by following the derivations given in Aki & Richards (1980); Chapman (2004); Dahlen & Tromp (1998); Eringen & Şuhubi (1975); Kennett (2001).

For a homogeneous, isotropic body, let the displacement field $u^{(1)}$ will be the response due to stress $\tau^{(1)}$ and body force $f^{(1)}$. Also for the same body, let the displacement field $u^{(2)}$ will be the response due to stress $\tau^{(2)}$ and body force $f^{(2)}$. Then the following equality exist

$$\begin{aligned} \int_V (f^{(1)} - \rho \ddot{u}^{(1)}) \cdot u^{(2)} dV + \int_S u^{(2)} \tau^{(2)} \cdot \hat{n} dS = \\ \int_V (f^{(2)} - \rho \ddot{u}^{(2)}) \cdot u^{(1)} dV + \int_S u^{(1)} \tau^{(1)} \cdot \hat{n} dS, \end{aligned} \quad (4.5)$$

where ρ is material density, $\hat{n}$ is the normal vector from the plane and double dot denotes the second derivative of the field respect to time.

The equality in eq.(4.5) can be rewritten by converting the surface integrals into volume integrals by the divergence theorem, and using the relation of equation of motion given at eq.(4.1), then relation takes the form of

$$\int_V [(\nabla \cdot \tau^{(1)}) \cdot u^{(2)} - \nabla \cdot (u^{(2)} \cdot \tau^{(1)})] dV = \int_V [(\nabla \cdot \tau^{(2)}) \cdot u^{(1)} - \nabla \cdot (u^{(1)} \cdot \tau^{(2)})] dV. \quad (4.6)$$

If we switch to index notation for simplicity, then

$$\int_V \left[\frac{\partial \tau_{ij}^{(1)}}{\partial x_i} u_j^{(2)} - \frac{\partial}{\partial x_j} (u_i^{(2)} \tau_{ij}^{(1)}) \right] dV = \int_V \left[\frac{\partial \tau_{ij}^{(2)}}{\partial x_i} u_j^{(1)} - \frac{\partial}{\partial x_j} (u_i^{(1)} \tau_{ij}^{(2)}) \right] dV. \quad (4.7)$$

The second terms in left and right hand side can be expanded via chain rule and the symmetry of the stress tensors

$$\frac{\partial}{\partial x_j} (u_i^{(2)} \tau_{ij}^{(1)}) = \frac{\partial u_i^{(2)}}{\partial x_j} \tau_{ij}^{(1)} + u_i^{(2)} \frac{\partial \tau_{ij}^{(1)}}{\partial x_j} = \frac{\partial u_i^{(2)}}{\partial x_j} \tau_{ij}^{(1)} + u_j^{(2)} \frac{\partial \tau_{ij}^{(1)}}{\partial x_i} \quad (4.8)$$

$$\frac{\partial}{\partial x_j} (u_i^{(1)} \tau_{ij}^{(2)}) = \frac{\partial u_i^{(1)}}{\partial x_j} \tau_{ij}^{(2)} + u_i^{(1)} \frac{\partial \tau_{ij}^{(2)}}{\partial x_j} = \frac{\partial u_i^{(1)}}{\partial x_j} \tau_{ij}^{(2)} + u_j^{(1)} \frac{\partial \tau_{ij}^{(2)}}{\partial x_i}. \quad (4.9)$$

If eq.(4.9) and eq.(4.8) are substituted into eq.(4.7), then

$$\int_V \left[\frac{\partial u_i^{(2)}}{\partial x_j} \tau_{ij}^{(1)} \right] dV = \int_V \left[\frac{\partial u_i^{(1)}}{\partial x_j} \tau_{ij}^{(2)} \right] dV. \quad (4.10)$$

By using the symmetry of the elastic modulus c_{ijkl} given at eq.4.4), eq.(4.10) can be written as

$$\int_V \left[\frac{\partial u_l^{(2)}}{\partial x_k} c_{kl ij} \frac{\partial u_l^{(1)}}{\partial x_k} \right] dV = \int_V \left[\frac{\partial u_l^{(1)}}{\partial x_k} c_{lk ij} \frac{\partial u_l^{(2)}}{\partial x_k} \right] dV, \quad (4.11)$$

which proves the Betti's dynamic reciprocity theorem.

Betti's theorem does not involve any conditions for $u^{(1)}$ or $u^{(2)}$. Also the relation given at eq.(4.5) is only valid if it is evaluated on different times t_1 and t_2 . If we define two time variables t and τ and two different times are chosen as $t_1 = t$ and $t_2 = \tau - t$, the eq.(4.5) can be written in time domain as

$$\begin{aligned} & \int_{-\infty}^{\infty} \rho [\ddot{u}^{(1)}(t) u^{(2)}(\tau - t) - \ddot{u}^{(2)}(\tau - t) u^{(1)}(t)] dt \\ &= \rho \int_{-\infty}^{\infty} \frac{\partial}{\partial t} [\dot{u}^{(1)}(t) u^{(2)}(\tau - t) - \dot{u}^{(2)}(\tau - t) u^{(1)}(t)] dt \\ &= \rho [\dot{u}^{(1)}(\tau) u^{(2)}(0) + u^{(1)}(\tau) \dot{u}^{(2)}(0) - \dot{u}^{(1)}(0) u^{(2)}(\tau) - u^{(1)}(0) \dot{u}^{(2)}(\tau)]. \end{aligned} \quad (4.12)$$

If there is some time before τ_0 , $u^{(1)}$ and $u^{(2)}$ and their derivatives $\dot{u}^{(1)}$, $\dot{u}^{(2)}$ are zero throughout the V , then the convolution integral becomes

$$\rho \int_{-\infty}^{\infty} [\ddot{u}^{(1)}(t)u^{(2)}(\tau - t) - \ddot{u}^{(2)}(\tau - t)u^{(1)}(t)] dt = 0. \quad (4.13)$$

Then the time integrated Betti's theorem becomes

$$\int_{-\infty}^{\infty} dt \left[\int_V f^{(1)}(\mathbf{x}, t)u^{(2)}(\mathbf{x}, \tau - t)dV + \int_S u^{(2)}(\mathbf{x}, \tau - t)\tau^{(1)}(t)\hat{n}dS \right] = (4.14)$$

$$\int_{-\infty}^{\infty} dt \left[\int_V f^{(2)}(\mathbf{x}, \tau - t)u^{(1)}(\mathbf{x}, t)dV + \int_S u^{(1)}(\mathbf{x}, t)\tau^{(2)}(\tau - t)\hat{n}dS \right],$$

where $\mathbf{x}$ is the space variable.

If we introduce the convolution operator $*$, then

$$\int_V [f^{(1)}(\mathbf{x}, t) * u^{(2)}(\mathbf{x}, t)] dV + \int_S [u^{(2)}(\mathbf{x}, t) * \tau^{(1)}(t)] dS = \quad (4.15)$$

$$\int_V [f^{(2)}(\mathbf{x}, t) * u^{(1)}(\mathbf{x}, t)] dV + \int_S [u^{(1)}(\mathbf{x}, t) * \tau^{(2)}(t)] dS.$$

4.2.3 Green's Function

For a body force field in the p th direction at $\boldsymbol{\xi}$ $f_i = \delta_{ij}\delta(\mathbf{x} - \boldsymbol{\xi})\delta(t - t_1)$, the equation of motion for determining the displacement field in the l th direction at $\mathbf{x}$ can be written as

$$\frac{\partial}{\partial x_j} (c_{ijkl} \frac{\partial G_{kp}}{\partial x_l}) - \rho \frac{\partial^2 G_{jp}}{\partial t^2} = -\delta_{ij}\delta(\mathbf{x} - \boldsymbol{\xi})\delta(t - t_1), \quad (4.16)$$

where $G_{lp}(\mathbf{x}, \boldsymbol{\xi}, t)$ will be the Green's function response for a point source f_i .

Similarly, we can define the stress tensor for same force field as

$$H_{ijp}(\mathbf{x}, \boldsymbol{\xi}, t) = c_{ijkl} \frac{\partial}{\partial x_k} G_{lp}(\mathbf{x}, \boldsymbol{\xi}, t). \quad (4.17)$$

4.2.4 Seismic Reciprocity in Space and Time

The seismic reciprocity can be defined for the Green's function in time and space by using the relations derived from Betti's Theorem.

Consider two unit forces affecting a body for two different times t_1 , t_2 and spatial locations $\boldsymbol{\xi}_1$, $\boldsymbol{\xi}_2$ as

$$f_i^{(1)} = \delta(\mathbf{x} - \mathbf{x}_1)\delta(t - t_1)\delta_{ip} \quad (4.18)$$

$$f_i^{(2)} = \delta(\mathbf{x} - \mathbf{x}_2)\delta(t - t_2)\delta_{il}. \quad (4.19)$$

Then the related Green's function of displacement for each of the body force fields will be

$$u_l^{(1)} = G_{lp}(\mathbf{x}, t; \boldsymbol{\xi}_1, t_1) \quad (4.20)$$

$$u_p^{(2)} = G_{pl}(\mathbf{x}, t; \boldsymbol{\xi}_2, t_2). \quad (4.21)$$

If the boundary conditions on the surface S are homogeneous, then either the displacement field or traction on the surface is zero which means the surface integral will vanish. If we insert the Green's functions and point forces into the eq.(4.14), then

$$\int_{-\infty}^{\infty} dt \int_V \delta(\mathbf{x} - \boldsymbol{\xi}_1) \delta(t - t_1) G_{pl}(\mathbf{x}, \tau - t; \boldsymbol{\xi}_2, t_2) dV \quad (4.22)$$

$$= \int_{-\infty}^{\infty} dt \int_V \delta(\mathbf{x} - \boldsymbol{\xi}_2) \delta(\tau - t - t_2) G_{lp}(\mathbf{x}, t; \boldsymbol{\xi}_1, t_1) dV. \quad (4.23)$$

From the sifting property of the δ function and integrating over time and setting $t_1 = t_2 = 0$

$$G_{pl}(\boldsymbol{\xi}_1, \tau; \boldsymbol{\xi}_2, 0) = G_{lp}(\boldsymbol{\xi}_2, \tau; \boldsymbol{\xi}_1, 0). \quad (4.24)$$

Eq.(4.24) shows spatial reciprocity which means that by changing the source and receiver locations, one can get same seismic displacement field for isotropic homogeneous body as long as the same body force field is excited. Also this result holds for the associated Green's function for the velocity field.

The time reciprocity of the Green's function for displacement field G_{ij} can be shown by reversal of the time variable $t \rightarrow -t$ in equation of motion,

$$\frac{\partial}{\partial x_j} \left(c_{ijkl} \frac{\partial G_{kp}}{\partial x_l} \right) - \rho \frac{\partial}{\partial(-t)} \left(\frac{\partial}{\partial(-t)} (G_{jp}) \right) = -\delta_{ij} \delta(\mathbf{x} - \boldsymbol{\xi}) \delta(t_1 - t), \quad (4.25)$$

where

$$\frac{\partial^2}{\partial(-t)^2} = \frac{\partial^2}{\partial t^2}. \quad (4.26)$$

By time reversal, the Green's function for displacement is also a reversed solution for the equation of motion, the minus signs will cancel for the second derivative and the sifting property of the δ function requires

$$G(\mathbf{x}, t; \mathbf{x}, t_1) = G(\mathbf{x}, t; \mathbf{x}, -t_1). \quad (4.27)$$

It has to be noted that the Green's function for the velocity field component $\bar{G}(\mathbf{x}, t)$, will have an odd symmetry for the time reversal operation

$$\bar{G}(\mathbf{x}, t; \mathbf{x}_1, t_1) = -\bar{G}(\mathbf{x}, t; \mathbf{x}_1, -t_1). \quad (4.28)$$

This result shows that it is possible to retrieve same Green's function of displacement field by using the acausal part of time. In the velocity field case, there will be an inherent odd symmetry for the Green's function.

4.3 Green's Function Retrieval for Excitation Field Near Station

The idea of the extraction of Green's function or impulse response of a medium from the cross-correlation of the ambient noise field has been previously proposed from the areas outside of seismology. In statistical physics, it comes directly from the Fluctuation-Dissipation theorem of Kubo (1966). An electrical engineering translation of a similar concept was made for measuring the behaviour of a linear time invariant system in the case of an input of white noise by Lee (1960).

We can provide a simple illustration of the way in which the transfer function between two points enters by taking the oversimplified case of local excitation near the stations.

Consider $\mathbf{x}_1$ and $\mathbf{x}_2$ as two points on the ground with the associated local velocity fields of $v_1(\mathbf{x}_1, t)$ and $v_2(\mathbf{x}_2, t)$. If a local excitation $f(\mathbf{x}_1, t)$ is applied at $\mathbf{x}_1$ then the response or ground velocity $v_{f1}(\mathbf{x}_2, t)$ of $\mathbf{x}_2$ can be written in convolutional form as

$$v_{f1}(\mathbf{x}_2, t) = \int d\tau H(\mathbf{x}_1, \mathbf{x}_2, t - \tau) f(\mathbf{x}_1, \tau), \quad (4.29)$$

where $H(\mathbf{x}_1, \mathbf{x}_2, t)$ is the appropriate Green's tensor or its derivative.

The local velocity field of $\mathbf{x}_2$ has also additional local velocity contributions. The overall local velocity field of $\mathbf{x}_2$ can be written as

$$v_2(\mathbf{x}_2, t) = v_{f1}(\mathbf{x}_2, t) + w_2(\mathbf{x}_2, t), \quad (4.30)$$

where $w_2(\mathbf{x}_2, t)$ represents the other contributions to the ground. Since we assume that the result of excitation at $\mathbf{x}_1$ and local contributions are independent, the overall velocity field is represented as their sum and the expected coherency is negligible.

Following the same experiment, the local velocity field at $\mathbf{x}_1$ can be represented as a sum of the local velocity $v_f(\mathbf{x}_1, t)$ and the local contributions $w_1(\mathbf{x}_1, t)$,

$$v_1(\mathbf{x}_1, t) = v_f(\mathbf{x}_1, t) + w_1(\mathbf{x}_1, t) = Cf(\mathbf{x}_1, t) + w_1(\mathbf{x}_1, t), \quad (4.31)$$

where C is a scalar.

If we take the cross-correlation Φ_{12} of two local velocity fields,

$$\Phi_{12}(\mathbf{x}_1, \mathbf{x}_2, t) = \int_0^T d\tau v_1(\mathbf{x}_1, \tau) v_2(\mathbf{x}_2, t + \tau), \quad (4.32)$$

which will have a Fourier Spectrum of

$$\begin{aligned} \bar{\Phi}_{12}(\mathbf{x}_1, \mathbf{x}_2, \omega) &= \bar{v}_1^*(\mathbf{x}_1, \omega) \bar{v}_2(\mathbf{x}_2, \omega) \\ &= \bar{v}_{f1}(\mathbf{x}_2, \omega) \bar{v}_f^*(\mathbf{x}_1, \omega) + \bar{v}_{f1}(\mathbf{x}_2, \omega) \bar{w}_1^*(\omega) \\ &\quad + \bar{v}_f^*(\mathbf{x}_1, \omega) \bar{w}_2(\omega) + \bar{w}_2(\omega) \bar{w}_1^*(\omega), \end{aligned} \quad (4.33)$$

where $*$ denotes the complex conjugate.

If we calculate the cross-spectrum for many separate time windows and average, then

$$\langle \bar{v}_1^*(\mathbf{x}_1, \omega) \bar{v}_2(\mathbf{x}_2, \omega) \rangle \rightarrow \bar{v}_{f1}(\mathbf{x}_2, \omega) \bar{v}_f^*(\mathbf{x}_1, \omega) = \bar{H}(\mathbf{x}_1, \mathbf{x}_2, \omega) C |f(\mathbf{x}_1, \omega)|^2, \quad (4.34)$$

since the non-coherent parts will tend to cancel. We can transform the relation at eq.(4.34) to the time domain which will be the time averaged cross-correlation of the ground velocity fields,

$$\langle \Phi_{12}(\mathbf{x}_1, \mathbf{x}_2, t) \rangle = C |f(\mathbf{x}_1, t)|^2 * H(\mathbf{x}_1, \mathbf{x}_2, t), \quad (4.35)$$

where $*$ is the convolution operator.

The eq.(4.34) shows the emergence of the scaled Green's function between points $\mathbf{x}_1$ and $\mathbf{x}_2$ from the averaged cross-correlation of the excitation field at $\mathbf{x}_1$. One important point regarding the estimation of Green's function is the nature of $|f(\mathbf{x}, \omega)|^2$ which is the power spectrum of the excitation field. In the case of white noise input, the power spectrum will correspond to a constant value K ,

$$|f(\mathbf{x}, \omega)|^2 \rightarrow K, \quad (4.36)$$

which yields to a near-perfect reconstruction of Green's function.

4.4 Green's Function Retrieval with Cross-Correlation of Distributed Noise Wavefields

The velocity field due to a body force f_j in frequency domain can be written as similarly the relation at eq.(4.2)

$$\partial_i \tau_{ij} - i\rho\omega v_j = -f_j, \quad (4.37)$$

where i is the complex variable.

We can use the definition of Green's tensor in eq.(4.16), and the tensor field H_{ijp} , however this time they are written for velocity field

$$\partial_i \bar{H}_{ijp} - i\rho\omega \bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) = -\delta_{jp}\delta(\mathbf{x} - \boldsymbol{\xi}), \quad (4.38)$$

where

$$\bar{H}_{ijp}(\mathbf{x}, \boldsymbol{\xi}, \omega) = c_{ijkl}(\mathbf{x}, \omega) \frac{\partial}{\partial x_k} \bar{G}_{lp}(\mathbf{x}, \boldsymbol{\xi}, \omega). \quad (4.39)$$

The scalar product of the Green's tensor for velocity field $\bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega)$ with eq.(4.37)

$$\bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) \partial_i \tau_{ij} - \bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) i\rho\omega v_j = -\bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) f_j. \quad (4.40)$$

The scalar product of $v_j(\mathbf{x}, \omega)$ with eq.(4.38) will be

$$v_j(\mathbf{x}, \omega) \partial_i \bar{H}_{ijp} - i v_j(\mathbf{x}, \omega) \rho\omega \bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) = -v_j(\mathbf{x}, \omega) \delta_{jp}\delta(\mathbf{x} - \boldsymbol{\xi}). \quad (4.41)$$

On subtracting eq.(4.40) from eq.(4.41), then integrating over volume V , then

$$\begin{aligned} & \int_V d^3\mathbf{x} [\bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) \partial_i \tau_{ij}(\mathbf{x}, \omega) - v_j \partial_i \bar{H}_{ijp}(\mathbf{x}, \boldsymbol{\xi}, \omega)] \\ &= - \int_V d^3\mathbf{x} \bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) f_j(\mathbf{x}, \omega) + \int_V d^3\mathbf{x} \delta_{jp}\delta(\mathbf{x} - \boldsymbol{\xi}) v_j(\mathbf{x}, \boldsymbol{\xi}). \end{aligned} \quad (4.42)$$

By using the sifting property of the delta function and rearranging eq.(4.42), velocity field can be represented by

$$\begin{aligned} \Theta(\boldsymbol{\xi}) v_i(\boldsymbol{\xi}, \omega) &= \int_V d^3\mathbf{x} \bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) f_j(\mathbf{x}, \omega) \\ &+ \int_V d^3\mathbf{x} [\bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) \partial_i \tau_{ij}(\mathbf{x}, \omega) - v_j(\mathbf{x}, \boldsymbol{\xi}) \partial_i \bar{H}_{ijp}(\mathbf{x}, \boldsymbol{\xi}, \omega)], \end{aligned} \quad (4.43)$$

where

$$\Theta(\boldsymbol{\xi}) = \begin{cases} 1, & \mathbf{x} \in V, \\ 0, & \mathbf{x} \notin V. \end{cases}$$

If source point $\mathbf{x}$ is outside the volume, then the velocity field becomes zero.

The divergence theorem can be used to convert the volume integral to surface integral, then

$$\begin{aligned} & \int_V d^3\mathbf{x} [\bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) \partial_i \tau_{ij}(\mathbf{x}, \omega) - v_j(\mathbf{x}, \boldsymbol{\xi}) \partial_i \bar{H}_{ijp}(\mathbf{x}, \boldsymbol{\xi}, \omega)] \\ &= \int_{\partial V} d^2\mathbf{x} n_p [\bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) \tau_{ij}(\mathbf{x}, \omega) - v_j(\mathbf{x}, \boldsymbol{\xi}) \bar{H}_{ijp}(\mathbf{x}, \boldsymbol{\xi}, \omega)], \end{aligned} \quad (4.44)$$

where n_p is normal vector to the integration surface ∂V .

The representation of the velocity field $v(\mathbf{x}, \omega)$ can be written with the new integrand and by interchanging the roles of $\mathbf{x}$ and $\boldsymbol{\xi}$ for later convenience

$$\begin{aligned} \Theta(\mathbf{x})v_k(\mathbf{x}, \omega) &= \int_V d^3\boldsymbol{\xi} \bar{G}_{qk}(\boldsymbol{\xi}, \mathbf{x}, \omega) f_q(\boldsymbol{\xi}, \omega) \\ &+ \int_{\partial V} d^2\boldsymbol{\xi} n_p [\bar{G}_{qk}(\boldsymbol{\xi}, \mathbf{x}, \omega) \tau_{pq}(\boldsymbol{\xi}, \omega) - v_q(\boldsymbol{\xi}, \mathbf{x}) \bar{H}_{pqk}(\boldsymbol{\xi}, \mathbf{x}, \omega)]. \end{aligned} \quad (4.45)$$

The spatial reciprocity of the Green's tensor for the velocity field is

$$\bar{G}_{jp}(\mathbf{x}, \boldsymbol{\xi}, \omega) = \bar{G}_{pj}(\boldsymbol{\xi}, \mathbf{x}, \omega). \quad (4.46)$$

The representation of $v_k(\mathbf{x}, \omega)$ can be recasted for receiver at $\mathbf{x}$ and a source at $\boldsymbol{\xi}$ with the spatial reciprocity condition

$$\begin{aligned} \Theta(\mathbf{x})v_k(\mathbf{x}, \omega) &= \int_V d^3\boldsymbol{\xi} \bar{G}_{kq}(\mathbf{x}, \boldsymbol{\xi}, \omega) f_q(\boldsymbol{\xi}, \omega) \\ &+ \int_{\partial V} d^2\boldsymbol{\xi} [\bar{G}_{kq}(\mathbf{x}, \boldsymbol{\xi}, \omega) t_q(\boldsymbol{\xi}, \omega) - v_q(\boldsymbol{\xi}, \mathbf{x}) \bar{h}_{kq}(\mathbf{x}, \boldsymbol{\xi}, \omega)], \end{aligned} \quad (4.47)$$

where the tractions on ∂V are given by

$$t_q = n_p \tau_{pq}, \quad \bar{h}_{kq} = n_p \bar{H}_{kpq}. \quad (4.48)$$

The scalar product of v_j^* with the equation of motion given in eq.(4.37) for velocity field v_i can be constructed

$$v_j^* \frac{\partial}{\partial x_j} \tau_{ij} - i\rho\omega v_j^* v_j = -f_j v_j^*, \quad (4.49)$$

where $*$ denotes the complex conjugation. Similarly, by complex conjugation and multiplying with velocity field v_j the equation of motion becomes

$$v_j \frac{\partial}{\partial x_j} \tau_{ij}^* + i\rho\omega v_j^* v_j = -f_j^* v_j. \quad (4.50)$$

On adding the representations in eq.(4.49) and eq.(4.50),

$$v_j^* \frac{\partial}{\partial x_j} \tau_{ij} + v_j \frac{\partial}{\partial x_j} \tau_{ij}^* = -f_j v_j^* - f_j^* v_j. \quad (4.51)$$

Consider the body force field f_j as delta function at $\mathbf{x}_A$ and observe at $\mathbf{x}_B$, then

$$\begin{aligned} \bar{G}_{kl}(\mathbf{x}_B, \mathbf{x}_A, \omega) &= - \int_V d^3\boldsymbol{\xi} \bar{G}_{qk}^*(\boldsymbol{\xi}, \mathbf{x}_B, \omega) \delta_{ql} \delta(\boldsymbol{\xi} - \mathbf{x}_A) \\ &+ \int_{\partial V} d^2\boldsymbol{\xi} n_p [\bar{G}_{qk}^*(\mathbf{x}_B, \boldsymbol{\xi}, \omega) \bar{H}_{pql}(\boldsymbol{\xi}, \mathbf{x}_A, \omega) \\ &+ \bar{G}_{ql}(\boldsymbol{\xi}, \mathbf{x}_A, \omega) \bar{H}_{kpq}^*(\mathbf{x}_B, \boldsymbol{\xi}, \omega)], \end{aligned} \quad (4.52)$$

where now the products correspond to correlations in the time domain. By comparing eq.(4.52) with eq.(4.47) and using spatial reciprocity results

$$\begin{aligned} \bar{G}_{kl}(\mathbf{x}_B, \mathbf{x}_A, \omega) + \bar{G}_{lk}^*(\mathbf{x}_A, \mathbf{x}_B, \omega) = \\ \int_{\partial V} d^2\xi \eta_p [\bar{G}_{kq}^*(\mathbf{x}_B, \boldsymbol{\xi}, \omega) \bar{H}_{pql}(\mathbf{x}_A, \boldsymbol{\xi}, \omega) \\ + \bar{G}_{ql}(\mathbf{x}_A, \boldsymbol{\xi}, \omega) \bar{H}_{pkq}^*(\mathbf{x}_B, \boldsymbol{\xi}, \omega)], \end{aligned} \quad (4.53)$$

which is equivalent to the results of Wapenaar & Fokkema (2006).

The left hand side of eq.(4.53) can also be expressed as $2\Re[\bar{G}_{lk}(\mathbf{x}_A, \mathbf{x}_B, \omega)]$ which is the Fourier transform of $\bar{G}_{lk}(\mathbf{x}_A, \mathbf{x}_B, t) + \bar{G}_{lk}(\mathbf{x}_A, \mathbf{x}_B, -t)$ and thus represents a two-sided Green's tensor for waves propagating between $\mathbf{x}_A$ and $\mathbf{x}_B$. The right hand side of eq.(4.53) represents the Fourier transform of the correlation fields generated by the superposition of sources all along the surface ∂V provided that this surface enclosed $\mathbf{x}_A$ and $\mathbf{x}_B$.

The constructive interference of the propagation from the sources along the surface ∂V to $\mathbf{x}_A$ and $\mathbf{x}_B$ is sufficient to allow the construction of the Green's function between $\mathbf{x}_A$ and $\mathbf{x}_B$. The most direct contributions come from the vicinity of $\mathbf{x}_A$ and $\mathbf{x}_B$ since the result is independent of the configuration of the surface ∂V . In the case where the surface ∂V is far from the points of $\mathbf{x}_A$ and $\mathbf{x}_B$, the integral of rather subtle interference effects is needed to represent the Green's function between these points. The Green's function $\bar{G}(\mathbf{x}_A, \mathbf{x}_B, \omega)$ given in eq.(4.53) will include all scattering and multipath effects. Wapenaar & Fokkema (2006) showed the approximations to ignore the effects from outside the ∂V in appropriate circumstances. If the volume V extends further than the twice the slowest propagation time between $\mathbf{x}_A$ and $\mathbf{x}_B$, then the any energy returning from outside the surface ∂V will have a sufficiently long delay to have a little practical effect.

Wapenaar (2004) has demonstrated that the expression eq.(4.53) can be significantly simplified if the contributions from ∂V come from spatially uncorrelated noise sources. In such circumstances, for a homogeneous region outside ∂V ,

$$2\Re [\bar{G}_{lk}(\mathbf{x}_A, \mathbf{x}_B, \omega)] S(\omega) = \frac{2}{\rho c} \langle v_l^*(\mathbf{x}_A, \omega) v_k(\mathbf{x}_B, \omega) \rangle, \quad (4.54)$$

where

$$S(\omega) = s^*(\mathbf{x}, \omega) s(\mathbf{x}, \omega) \quad (4.55)$$

with the $s(\mathbf{x}, \omega)$ the spectrum for the uncorrelated noise sources; c is the appropriate wavespeed. The multiplication of the velocity fields in frequency domain in eq.(4.54) corresponds to cross-correlation in time domain. The averaging is taken over a spatial ensemble of the stochastic sources. The spatial ensemble concept comes from the temporal variation of the Earth's seismic noise, which is interacting with the heterogeneous medium to produce a spatially distributed set of virtual sources. The extracted Green's function is enhanced via the coherent contributions over time. The situation for scattered surface waves from ambient noise can be represented through the use of an enclosing surface ∂V extending to substantial depth well into the evanescent tail of the ground velocity associated with the surface waves given in.

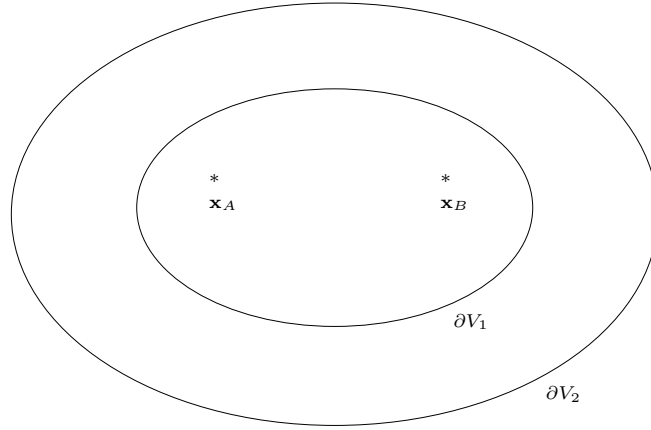


Figure 4.2: Plan view configuration of surfaces ∂V_1 and ∂V_2 for surface wave analysis.

The secondary sources from the scattering inside the medium can be associated with a ring around $\mathbf{x}_A$ and $\mathbf{x}_B$. The superposition of multiple surface integrals will sweep out further and further onto the medium as indicated by the multiple surfaces in figure 4.2. On the other hand, the interaction of the diverse range of sources will suppress the retrieval of coherent Green's function between $\mathbf{x}_A$ and $\mathbf{x}_B$. Again, the major contribution will come from the vicinity of receiver points $\mathbf{x}_A$ and $\mathbf{x}_B$. The interference of the scattered wavefields will reinforce any direct excitation in the vicinity of receiver points due to, e.g., atmospheric loading. This scattered wavefield viewpoint combines two different explanation of the correlation approach, the deterministic

approach of Wapenaar (2004); Wapenaar & Fokkema (2006) and the diffusive wave propagation of Shapiro & Campillo (2004).

The temporal averaging embedded in the processing scheme of the actual data processing for retrieving the Green's function works with the contribution of secondary noise wavefield distribution which is varying with time as, e.g., the primary noise wavefield moves with distant storms. The varying location of the primary sources provides a component of spatial averaging, but because some azimuths will be preferentially sampled the recovery of the two-sided Green's function is not complete. One side tends to be enhanced relative to the other.

4.5 Calculation Scheme

The ambient noise processing for the reconstructing the Green's function of the medium between the path of connecting two stations primarily relies on simultaneous recordings. In addition to this, sufficiently long times of recording at the two stations will improve the reliability, and the signal to noise ratio of the estimates.

We list the steps taken for the calculation of final estimates of Green's function between two stations as:

1. Prepare daily SAC files for each of the stations.
2. Remove the supirous days due to the instrument problem from seismic records.
3. Divide the full day record into 4 hour segments and compute the cross-correlations for the corresponding station pairs with 1 hour overlap and then average out the daily estimate.
4. Stack all of the averaged cross-correlations of the individual days to improve the signal to noise ratio of the final signal.

In this procedure we do not attempt to exclude seismic events but use long-term averaging to enhance the coherent contributions to the Green's function.

Another important feature of the cross-correlation is the stability of the estimates. For different time periods, the intensity of the noise field can

fluctuate related to the climate variability and its effect on the ocean. However, this should not change the shape of the final waveform. A test was conducted for TASMAL experiment by exploring the variability of the final estimates respect to the different months. In figure 4.3, for four different station pairs, three estimates of Green's function for a one-month period are given. Although the amplitudes have a temporal variability, the phase of the dispersed wavetrains is generally preserved for different monthly estimates. This demonstration shows the reliability of the measurements in the estimation of the phase and group velocities particularly when multi-month results are stacked.

Other studies proposed some operations on the amplitude of the seismic records in order to improve the signal to noise ratio such as one-bit normalisation (Larose et al., 2004) and amplitude threshold clipping (Sabra et al., 2005). In this study, no intervention on the amplitude of the daily seismic records has been carried out in order to preserve the frequency content of the records. In chapter 6 we discuss alternative ways of broadening the spectral response of the Green's function.

4.6 Group Velocity Extraction

The most energetic part of the interstation Green's function is associated with the surface waves and these emerge most clearly from the temporal averaging. Thus, the estimated Green's function or its tensor is an estimate of the surface wave propagation for the vertical-vertical ambient noise correlations. This can be deduced from the vertical coupling of the waves with the sea bottom. The interaction causes a Rayleigh wave type propagation. Hence the extracted component is Rayleigh wave component of Green's function between two stations.

4.6.1 Dispersion

The dispersion mechanism can be best described with the superposition of the two harmonic waves. Following the development of this idea given by Eringen & Şuhubi (1975); if we consider two harmonic waves propagating in the direction of $\mathbf{x}_1$ with an equal amplitude of A then the superposition field will be

$$\phi = A [\cos(k_1 \mathbf{x}_1 - \omega_1 t) + \cos(k_2 \mathbf{x}_1 - \omega_2 t)], \quad (4.56)$$

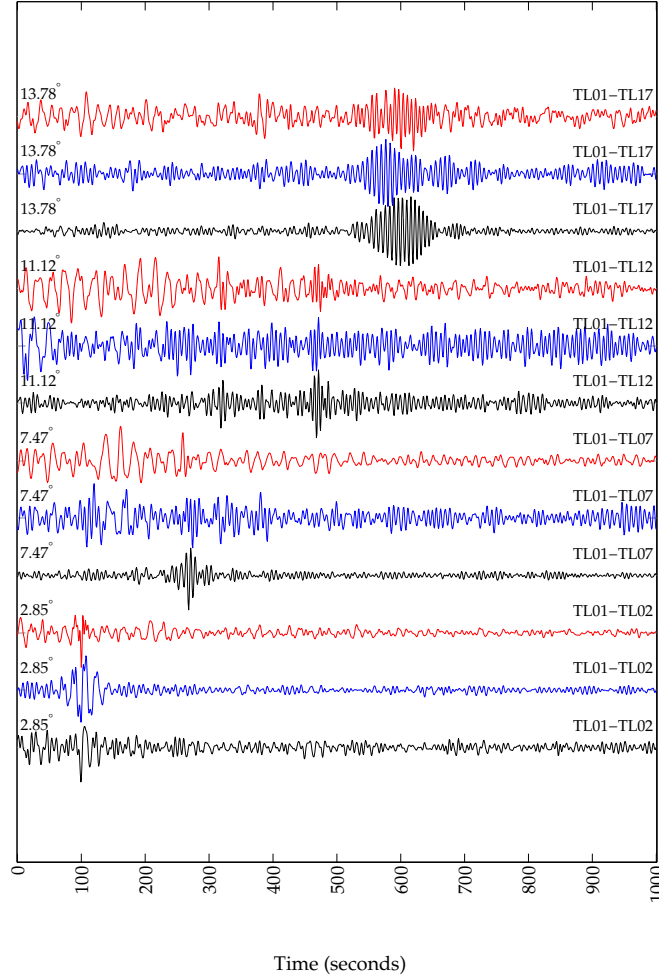


Figure 4.3: Monthly variation of Green's function estimates for four different station pairs; TL01-TL02, TL01-TL07, TL01-TL12, and TL01-TL17. Black: October 2003, Blue: January 2004, Red: January 2005. The amplitudes of the extracted Green's functions have a temporal variability but the phase of the dispersed wavetrains is preserved which marks the reliability of the results.

where k is the wave number, ω is the angular frequency and t is the time variable. By using the trigonometric summation eq.(4.56) can be also written as

$$\phi = 2A \cos \left(\frac{k_1 + k_2}{2} \mathbf{x}_1 - \frac{\omega_1 + \omega_2}{2} t \right) \cos \left(\frac{k_1 - k_2}{2} \mathbf{x}_1 - \frac{\omega_1 - \omega_2}{2} t \right). \quad (4.57)$$

The velocities of the two individual waves in eq.(4.56) are ω_1/k_1 and ω_2/k_2 . If we look at the parts of the superposed wave separately, then the first term

has a similar velocity as $(\omega_1 + \omega_2)/(k_1 + k_2)$. However, the second term has a slowly varying wave velocity of $(\omega_1 - \omega_2)/(k_1 - k_2)$. If the superposed wave ϕ is written as

$$\phi = \phi_0 \cos \left(\frac{k_1 + k_2}{2} \mathbf{x}_1 - \frac{\omega_1 + \omega_2}{2} t \right), \quad (4.58)$$

where

$$\phi_0 = 2A \cos \left(\frac{k_1 - k_2}{2} \mathbf{x}_1 - \frac{\omega_1 - \omega_2}{2} t \right). \quad (4.59)$$

In eq.(4.58), the wave ϕ will have an amplitude of ϕ_0 varying with time t and distance $\mathbf{x}_1$. As a result the waveform ϕ is composed of the wavepackets which is encapsulated by the amplitude curves of ϕ_0 . The rate of change ϕ_0 is

$$U = (\omega_1 - \omega_2)/(k_1 - k_2), \quad (4.60)$$

which is called as *group velocity*. Also the velocity of the faster moving wavepackets of the superposed wave ϕ is

$$c = (\omega_1 + \omega_2)/(k_1 + k_2), \quad (4.61)$$

which is referred to as the *phase velocity*. Therefore, we have a modulated wave ϕ , with a slowly varying modulation part ϕ_0 , and with a carrier wave as the remaining factor in eq.(4.58).

In figure 4.4, a dispersed wave train ϕ is given with its envelope curve ϕ_0 .

The group velocity U is related to the phase velocity c for all frequencies as follows

$$U(k) = d\omega/dk, \quad c(k) = \omega(k)/k \quad (4.62)$$

$$U = \frac{d(kc)}{dk} = c + k \frac{dc}{dk} = c - \lambda \frac{dc}{d\lambda}, \quad (4.63)$$

where λ is the corresponding wavelength. The effect of the dispersion on the phase and group velocities is clear at eq.(4.62). If the propagating wave was not dispersive, then the $dc/dk = 0$ and group and phase velocities would be equal.

Dispersion controls the frequency dependent propagation of surface waves. In the Earth, change in seismic velocity with depth is the main cause of the dispersion of the surface waves (Stein & Wysession, 2003).

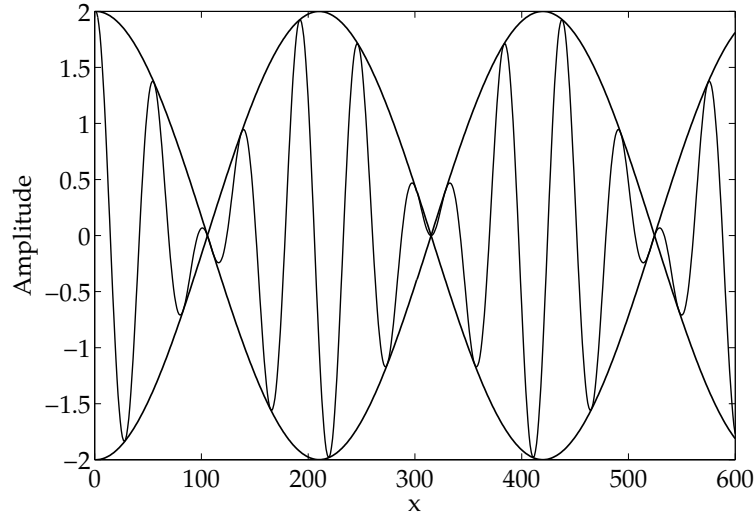


Figure 4.4: The dispersed wavetrain ϕ due to the superposition of two harmonic waves. The wavepackets propagate with the group velocity U and rapidly oscillating part with the phase velocity c inside the amplitude envelope in the direction of $\mathbf{x}$.

4.6.2 Measurement of Group Velocity

Each of the frequency components of the surface wave will sample with differing Earth radius. In general the seismic velocity of the Earth increases radially downwards so that the longer wavelength wave components which propagate deeper, will travel faster than the shallower ones. On the surface, the observed dispersed wavetrain will contain structural information which belongs to different depths of the Earth ‘collected’ along its travel path. It is crucial to distinguish the different frequency parts of the dispersed wavetrain to exploit the information carried within the surface wave.

Dziewonski et al. (1969) developed a narrow band multiple filtering scheme which operates on the frequency-time domain for estimating the group velocities of surface wave. According to this approach, a Gaussian narrow band filter is applied successively for different frequencies on the dispersed wavetrain. The position of the peak of the wavepacket is calculated and divided by the source-receiver distance to estimate the group velocity for that particular frequency.

The n th filter kernel of the Gaussian filter in Fourier domain is defined

as

$$H_n(\omega) = e^{-\alpha \left(\frac{\omega - \omega_n}{\omega_n} \right)^2}, \quad (4.64)$$

where ω is angular frequency, ω_n is the centre frequency and α is the width factor for the kernel. The centre frequency ω_n is dependent on the choice of the bandwidth parameter ω_Δ which is

$$\omega_\Delta = \sqrt{\frac{\beta}{\alpha}}, \quad (4.65)$$

where β describes the decay characteristics of the filter kernel. The bandwidth parameter determines the lowest $\omega_{l,n}$ and highest frequencies $\omega_{u,n}$ of the filter which is centred at ω_n . These are given as

$$\omega_{l,n} = (1 - \omega_\Delta) \omega_n \quad (4.66)$$

$$\omega_{u,n} = (1 + \omega_\Delta) \omega_n. \quad (4.67)$$

In figure 4.5, the Gaussian filter kernels are shown for multiple frequencies. As the centre frequency increases, the bandwidth grows.

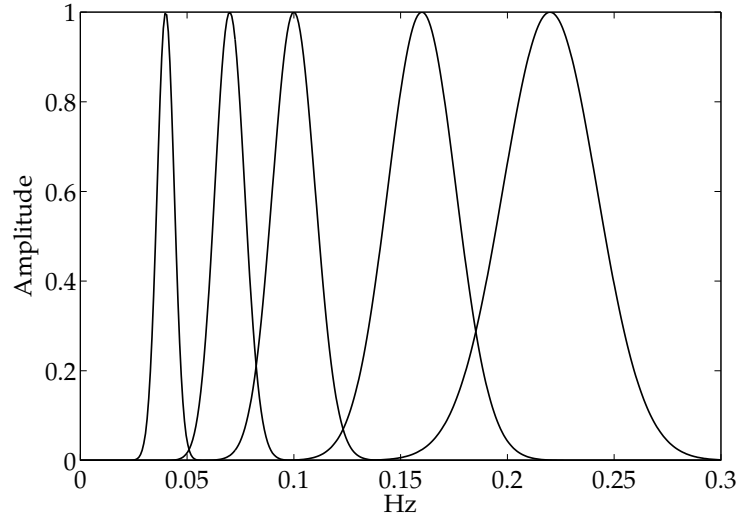


Figure 4.5: The Gaussian filter kernels $H_n(\omega)$ for $\alpha = 50.0$ and $\beta = 3.15$ in frequency domain for centre frequencies ω_n of 0.04 Hz, 0.07 Hz, 0.10 Hz, 0.16 and 0.22 Hz. As the centre frequency increases, the bandwidth grows. The rightmost filter kernel has a passband from 0.15 Hz to 0.3 Hz where the centre frequency is 0.22 Hz.

The group velocity of the dispersed waveform is defined as the rate of the change of the amplitude curve. In real data measurements, the dispersed

wave train does not contain any amplitude envelope. For picking the group velocities, a definition of the envelope should be introduced. By following Bracewell (1978) and Claerbout (1992), we can construct the analytic signal $v(t)$ which consists of a real valued signal $f(t)$ and its 90° phase shifted version, then

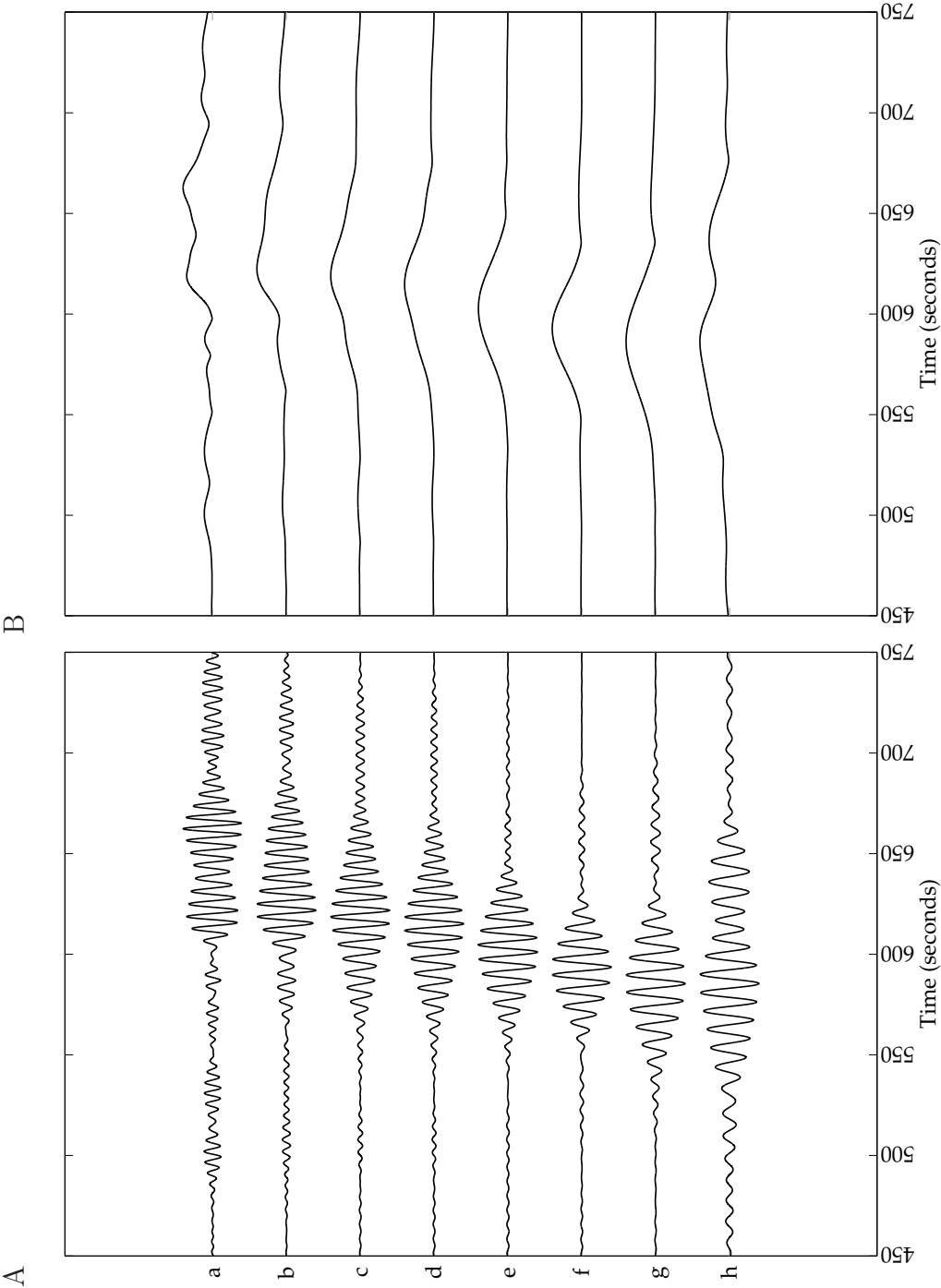
$$v(t) = f(t) + i f_{Hi}(t). \quad (4.68)$$

The imaginary part of the analytic signal $v(t)$ is the Hilbert transform of the real valued signal $f(t)$ and referred as the quadrature function of $f(t)$. The instantaneous amplitude of the analytic signal or the envelope of the input signal $f(t)$ is given as

$$|v(t)| = \sqrt{f(t)^2 + f_{Hi}^2(t)}. \quad (4.69)$$

After filtering out the dispersed wavetrain, the envelope of the signal is calculated to pick the arrival of the maximum amplitude which marks the arrival of the particular phase. In figure 4.6.2, the effect of narrow band filtering is given for an estimated Green's function. The frequency dependence of the arrivals shows sampled information at changing Earth radii.

Figure 4.6.2: A) The effect of the Gaussian narrow band filter on the Green's function's surface wave component extracted from the station pair of TL02-TL18 of the TASMAL experiment is given. The interstation distance is 14.48° . With decreasing frequency, the wavepackets tend to arrive earlier which shows the increasing velocity with depth. The chosen centre frequencies ω_n are a-0.19 Hz, b-0.176 Hz, c-0.16 Hz, d-0.155 Hz, e-0.143 Hz, f-0.12 Hz, g-0.1 Hz, and h-0.09 Hz. B) The corresponding instantaneous amplitudes $|v(t)|$ (envelope) of the filtered signals which is used in the estimation of group velocities by determining the arrival time of the maxima of the envelope.



4.7 Nonlinear Traveltime Tomography of the Noise Field

4.7.1 Seismic Tomography Fundamentals

The Earth, to a first approximation, has a spherically symmetrical seismic velocity structure. Models such as AK135 (Kennett et al., 1995) allow traveltimes of different seismic waves to be predicted. Departures from such predicted traveltimes are known as ‘traveltime anomalies’. Seismic tomography addresses this problem; *is it possible to construct Earth’s 3-D structure by using the traveltimes?*. This challenge had been attacked in the past by Herglotz, Wiechert in order to construct a 1-D velocity model of the Earth. In reality, the Earth has a heterogeneous structure which creates traveltime anomalies in the propagation of the seismic waves. These anomalies can be used to derive useful information from the Earth’s itself.

With the increasing number of seismic receivers and precision of the records, it is possible to collect a large amount of data which covers an important portion of the Earth. By working relative to a suitable reference model, the seismic anomalies can be revealed, and therefore an inverse problem can be posed.

The seismic tomography has its roots from medical imaging technologies. Although the underlying theory is quite similar, in the medical imaging case, the source and receiver distribution is controlled by the operator unlike the seismology where we can only control the locations of the receivers.

Seismic tomography depends on the presence of contrast in seismic properties. These differences are reflected directly in the times of arrival of seismic phases or through the shape and amplitude of seismic waves (Kennett, 2002, p 426). There are various proposed techniques on estimating the structure from seismic waves. A detailed and elaborate coverage of crustal imaging with seismic tomography is given by Rawlinson & Sambridge (2003).

If we follow the formulation given by Rawlinson & Sambridge (2003), then the traveltime of a ray in a continuous velocity medium $v(\mathbf{x})$ can be given as

$$t = \int_{L(v)} \frac{1}{v(\mathbf{x})} dl, \quad (4.70)$$

where L is the raypath and $v(\mathbf{x})$ is the velocity field. The eq.(4.70) is non-linear. We assess the arrival time of a certain seismic phase by comparing

it to a reference velocity model. If we treat the observed arrival time as a result of perturbation to the reference velocity field $v_0(\mathbf{x})$ then we can rewrite eq.(4.70) as

$$t = \int_{L_0+\delta L} \frac{1}{v_0 + \delta v} dl, \quad (4.71)$$

where L is the new raypath and $t = t_0 + \delta t$ is the travelttime of a ray is due to the perturbation in the velocity field $v(\mathbf{x})$. By expanding the terms in eq.(4.71) by using the geometric series and ignoring the second-order terms, we can represent the travelttime field as

$$\begin{aligned} t &= \int_{L_0+\delta L} \left[\frac{1}{v_0} - \frac{\delta v}{v_0^2} \right] dl + \mathcal{O}(\delta v^2) \\ &= \int_{L_0} \left[\frac{1}{v_0} - \frac{\delta v}{v_0^2} \right] dl + \int_{\delta L} \left[\frac{1}{v_0} - \frac{\delta v}{v_0^2} \right] dl + \mathcal{O}(\delta v^2). \end{aligned} \quad (4.72)$$

By using the Fermat's principle on the stationarity of the travelttime of along the ray path due to the perturbations, the integrand can be set to 0 in eq.(4.72). Then the perturbation in the travelttime δt caused by the perturbation in velocity field along the raypath is

$$\delta t = - \int_{L_0} \frac{\delta v}{v_0^2} dl + \mathcal{O}(\delta v^2). \quad (4.73)$$

4.7.2 Tomographic inversion

Observation of a single perturbation of a travelttime due to a perturbation of the slowness field is not enough to recover the heterogenous body which causes this. In general, many source and receiver combinations are employed to create ray paths which sample the region to improve the resolution of the imaging. The underlying medium can be divided into cells in 2-D or blocks in 3-D to set up an inverse problem to reconstruct the slowness field of interest. If the travelttime integral is discretized for a single ray and the j th block, then

$$\delta t = \sum_j l_j \Delta u_j, \quad (4.74)$$

where Δu_j is the slowness of block j and l_j is the path length in the corresponding block.

For multiple rays with the index of i , the relation becomes

$$\delta t_i = \sum_j l_{ij} \Delta u_j. \quad (4.75)$$

The tomographic inversion is a good example of a *mixed-determined* inverse problem where the some part of the system can be over-determined and the other can be under-determined. The class of the problem has a direct impact on the solution set after the inversion. Initially, we can propose to solve the eq.(4.75) with generalized inverse as following

$$d = G\mathbf{m} \quad (4.76)$$

$$\mathbf{m}_{est} = [G^T G]^{-1} G^T G d, \quad (4.77)$$

where d is observed variables, $\mathbf{m}$ is the model parameters, G is the data kernel matrix and $\mathbf{m}_{est}$ is the estimated model parameters after inversion. However, this solution does not guarantee a unique solution since some part of the problem is under-determined. A priori information related to model parameters can be introduced to the inverse problem such as

$$\Phi(\mathbf{m}) = (d - G\mathbf{m})^T (d - G\mathbf{m}) + \epsilon(\mathbf{m} - \mathbf{m}_0)^T (\mathbf{m} - \mathbf{m}_0), \quad (4.78)$$

where ϵ is the damping parameter to control the contribution of under-determined part on the solution and $\mathbf{m}_0$ is the priori information about the model as a reference model.

In practice the collected data, our measurements, are far from perfect and involve some uncertainty. An uncertainty measure is generally introduced to the inverse problem formulation via an inverse data weighting matrix W_d^{-1} . Also some part of the priori information (model parameters) may have greater significance than the others; the relative weight can be given to the parameters by an inverse model parameter covariance matrix $W_{\mathbf{m}}^{-1}$ which is included in the formulation. As a final term, we may impose that our solution space should vary smoothly, a smoothness/flatness matrix D can be attached to the model parameters to control the outcome of the model parameter variations. The final equation to minimize can be written as

$$\begin{aligned} \Phi(\mathbf{m}) = & (d - G\mathbf{m})^T W_d^{-1} (d - G\mathbf{m}) \\ & + \epsilon(\mathbf{m} - \mathbf{m}_0)^T W_{\mathbf{m}}^{-1} (\mathbf{m} - \mathbf{m}_0) + \eta \mathbf{m}^T D^T D \mathbf{m}. \end{aligned} \quad (4.79)$$

By solving eq.(4.79) in travelttime tomography problem, one can get the slowness function attributed to the each block therefore an image related to the region of interest can be constructed.

For a known source and receiver points, a scheme is needed between these points to predict the traveltimes and path of seismic energy within a laterally heterogeneous 2-D or 3-D medium (Rawlinson & Sambridge, 2005). Traditionally, this has been achieved by employing ray tracing schemes. Shooting and bending methods are the common ones where the shooting method relies on calculating the ray equation along the given ray path. The bending method adjusts the given connecting ray path between the source and receiver until it satisfies Fermat's Principle.

There are major drawbacks of ray tracing schemes such as robustness, speed, and uniqueness of the derived ray path (Rawlinson & Sambridge, 2005). To overcome these problems, a new method; *Fast Marching Method* has been introduced by Rawlinson & Sambridge (2004a).

4.7.3 The Fast Marching Method

Fast marching method (FMM) was originally developed by Sethian (1996) for tracking advancing interfaces. It found application in seismology for modeling the wavefront propagation in heterogeneous media by Rawlinson & Sambridge (2004a) for the tomography problem. Rather than using ray tracing based methods, the forward problem is set via FMM and then the inversion step is carried out. This section follows the derivations from Rawlinson & Sambridge (2004a,b, 2005).

The eikonal equation states that the magnitude of the traveltime gradient at any point along the wavefront is equal to the inverse of the velocity at that point. It can be written as

$$|\nabla_{\mathbf{x}}T| = s(\mathbf{x}), \quad (4.80)$$

where $\nabla_{\mathbf{x}}$ is the gradient operator, T is the traveltime, and $s(\mathbf{x})$ is the slowness for the position $\mathbf{x}$. An important limitation of the finite difference scheme is that it can not solve the eikonal equation in the presence of the gradient discontinuities. This limitation can be overcome by seeking weak solutions which will result in continuous $T_{\mathbf{x}}$ but not necessarily a continuous $\nabla_{\mathbf{x}}T$. One way obtaining a weak solution is to solve the 'viscous' version of the eikonal equation

$$|\nabla_{\mathbf{x}}T| = s(\mathbf{x}) + \epsilon \nabla_{\mathbf{x}}^2 T, \quad (4.81)$$

where ϵ controls the smoothness of the solution. The limit of smooth solutions

is that they correspond to the first arriving wavefront. The FMM method approach can be likened to the advance of a firefront under the influence of wind. FMM method enforces an entropy condition for the propagation of the first arriving wavefront as *once a point burns, it stays burnt*. The wavefront evolves since it can only pass from a point once. The unconditional stability of FMM comes from the application of this condition. The upwind entropy condition for 2-D is given as

$$\left[\max(D_a^{-x}T, -D_b^{+x}T, 0)^2 + \max(D_c^{-y}T, -D_d^{+y}T, 0)^2 \right]_{i,j}^{1/2} = s_{i,j}, \quad (4.82)$$

where (i, j) are grid increment variables in (x, y) , T is the traveltime and the integer variables a, b, c, d define the order of accuracy of the upwind finite-difference operator used in each of the four cases.

For a spherical shell the upwind gradient operators can be defined in spherical coordinates as

$$D_1^{-\theta}T_{i,j} = \frac{T_{i,j} - T_{i-1,j}}{r\delta\theta} \quad (4.83)$$

$$D_2^{-\theta}T_{i,j} = \frac{3T_{i,j} - 4T_{i-1,j} + T_{i-2,j}}{2r\delta\theta} \quad (4.84)$$

$$D_1^{-\phi}T_{i,j} = \frac{T_{i,j} - T_{i,j-1}}{r \cos \theta \delta\phi} \quad (4.85)$$

$$D_2^{-\phi}T_{i,j} = \frac{3T_{i,j} - 4T_{i,j-1} + T_{i,j-2}}{2r \cos \theta \delta\phi}, \quad (4.86)$$

where r , θ , and ϕ are the spherical coordinates. The eq.(4.82) describes how to calculate new traveltimes using the known traveltimes from adjacent grid points. The traveltimes are computed in downwind fashion from the already computed traveltimes. The employed scheme is a *narrow band* approach. The concept is visualized in figure 4.6 for 2-D. Alive points correspond to the points where the traveltimes are already correctly assigned. The close points form the narrow band which have some assigned trial values and far points are the ones without the computed traveltimes. The traveltimes of the close points are calculated via using eq.(4.82). Then the narrow band evolves by finding the close point with the minimum traveltime. Once it is found, it will be tagged as an alive point. For the remaining points inside the narrow band, the calculation scheme will be repeated till all of the far points become alive points hence the propagation of the wavefront is tracked completely.

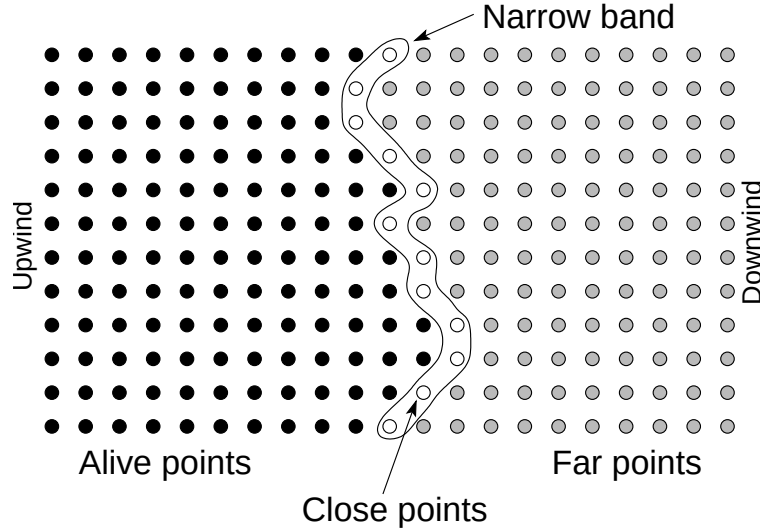


Figure 4.6: Tracking the evolution of the first arrival wavefront. The alive points have the correct traveltimes, the close points forms the narrow band with trial traveltimes and the far points do not have traveltimes values (Rawlinson & Sambridge, 2005).

4.7.4 Subspace Inversion

The tomographic inversion is utilized in order to find a model satisfying the data where the forward problem is solved via FMM. In contrast to the conventional tomographic problem set up for, the observables are the group traveltimes for a propagating Rayleigh wave between two stations as one station behaves as point source and the other receives. As a result of inversion, the velocity gradient is found by using combinations of many source and receiver points which cover the study area.

A nonlinear, iterative inversion scheme is used in inverting the traveltimes for solving the imaging problem. The subspace method solves the linearized inverse model successively, until the difference between observed and calculated traveltimes becomes significantly small. Subspace method relies on projecting n dimensional problems onto smaller systems to reduce computational effort.

The details of the subspace method is given in Kennett et al. (1988) and Rawlinson & Sambridge (2003). If the objective function in eq.(4.79) is used,

then the perturbation $\delta \mathbf{m}$ is

$$\delta \mathbf{m} = -A[A^T(G^T W_d^{-1} G + \epsilon W_{\mathbf{m}}^{-1} + \eta D^T D)A]^{-1} A^T \hat{\gamma}, \quad (4.87)$$

where $A = [a^j]$ is the $M \times n$ projection matrix, G is the Fréchet derivatives matrix, $\hat{\gamma}$ is the gradient vector ($\gamma = \partial S / \partial \mathbf{m}$ and $\gamma = C_{\mathbf{m}} \hat{\gamma}$), S is the objective function to minimize and $C_{\mathbf{m}}$ is the model covariance matrix.

The subspace inversion method operates by projecting eq.(4.79) onto an n -dimensional subspace of model space. The search directions are given by a^j where the first search direction a^1 corresponds to steepest ascent. The combination of FMM for forward problem calculation and subspace inversion offer stable and robust tomographic imaging.

4.8 Results

The extracted group velocities between 5 sec (0.2 Hz) and 12.5 sec (0.08 Hz) from the ambient noise Rayleigh type surface waves are used in the tomographic inversion for imaging the Australian crust. Shorter and longer periods were not been included in the study due to the nonuniform ray-path coverage of these ranges, where a seismic tomography problem can not be posed for the continent. Although there is a direct relation of the frequency of the surface wave and the group velocities therefore the depth, in this range of frequencies, it is acceptable to assume that the vertically sampled Earth is averaged velocity of the crust.

For the Australian continent, we have used two different velocity models during the interpretation of the depths sampled by the surface waves. In the first case, a thin sedimentary layer with the velocity of 2.5 km/sec was set in model. According to this analysis, the different components are going to sample mostly the midcrustal depths of 10-15 km. This is seen better in figure 4.7 which shows the phase velocity sensitivity of surface waves with a given crustal thickness of 30 km.

In the second case, we introduced a thicker sedimentary layer (≈ 3 km) instead of a thin layer in the same model. The effect of the thick sediment on the phase velocity sensitivities is more dominant than the first case. In figure 4.8, the derivatives of the phase velocities consist two peaks with varying amplitude for each frequency. In the higher frequencies, it is obvious that the surface waves will sample more of the thick sedimentary layer. With the

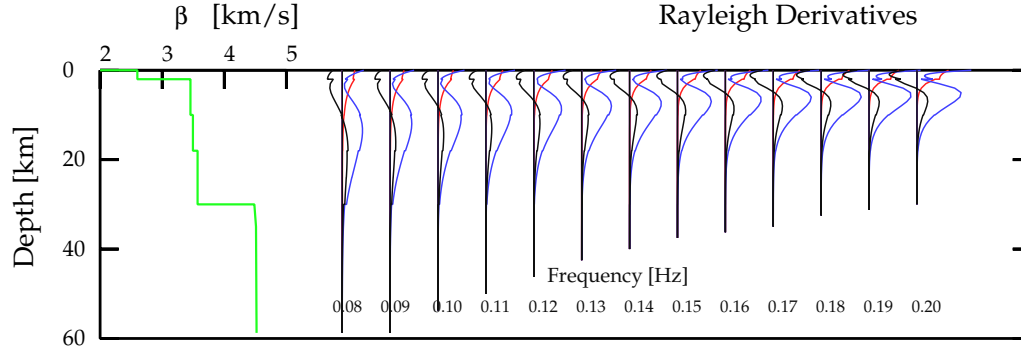


Figure 4.7: Phase velocity sensitivity for P wavespeed (red), S wavespeed (blue), and density (black) for the different range of frequencies. The given crustal thickness in the model is 30 km.

decreasing frequencies, the sampling area will be midcrustal depths of 10-15 km as in the first case.

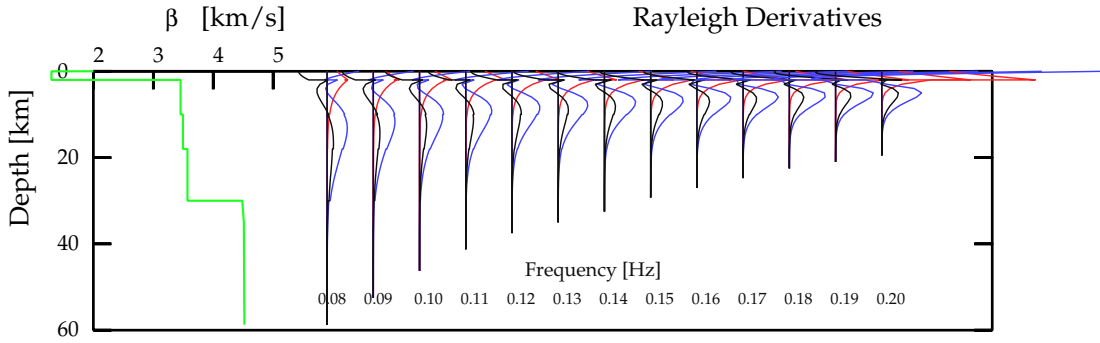


Figure 4.8: Phase velocity sensitivity for P wavespeed (red), S wavespeed (blue), and density (black) for the different range of frequencies with the presence of thick sedimentary layer. The given crustal thickness in the model is 30 km.

It is important to investigate the effect of the thick sedimentary layer on the surface waves for Australian continent since a thick sediment cover exists for the majority of the continent.

4.8.1 Extraction of Green's Functions for TASMAL Experiment

Initially, the idea of the Green's function extraction from ambient noise was tested with the data from TASMAL experiment. Green's function between

the stations were created for a 100 day period.

Examples of the extracted Rayleigh wave responses are shown in figure 4.9 and figure 4.10 for a number of stations, illustrating the clear emergence of the Rayleigh wave train to large distances and distinct timing variations for different paths with similar interstation separation. The variations are a clear indicator of the presence of lateral heterogeneity and with the intersecting paths between stations in the deployment, there is the possibility of localizing the regions of variation. For exploiting the group wavespeed properties of the surface waves, Gaussian narrow band filtering was applied on the signals in a number of frequency bands. In figure 4.11, for four different frequencies, the extracted wavespeeds on the great circle paths are plotted with different colors. The different range of the results for a limited area show that the extracted waveforms are sensitive to the gradients along the propagation path and carries structural information. Also the frequency dependency of the group wavespeeds is clear from the velocity maps. Relatively low wavespeed raypaths in the middle of the array became faster with the increasing periods.

These results demonstrate that it is possible to use continuous data from portable broadband experiments to extract the surface wave portion of the Green's function for separations up to 2000 km. Since all the portable data RSES broadband stations from 1992 on has been recorded continuously, these results indicate that there is the potential for achieving continental coverage particularly when permanent stations are used to link different temporary deployments.

The TASMAL experiment with its long duration of simultaneous recording at many stations provided an excellent test bed for the technique development and testing of data handling procedures. The approach developed in this study is now applied to the whole of the continent.

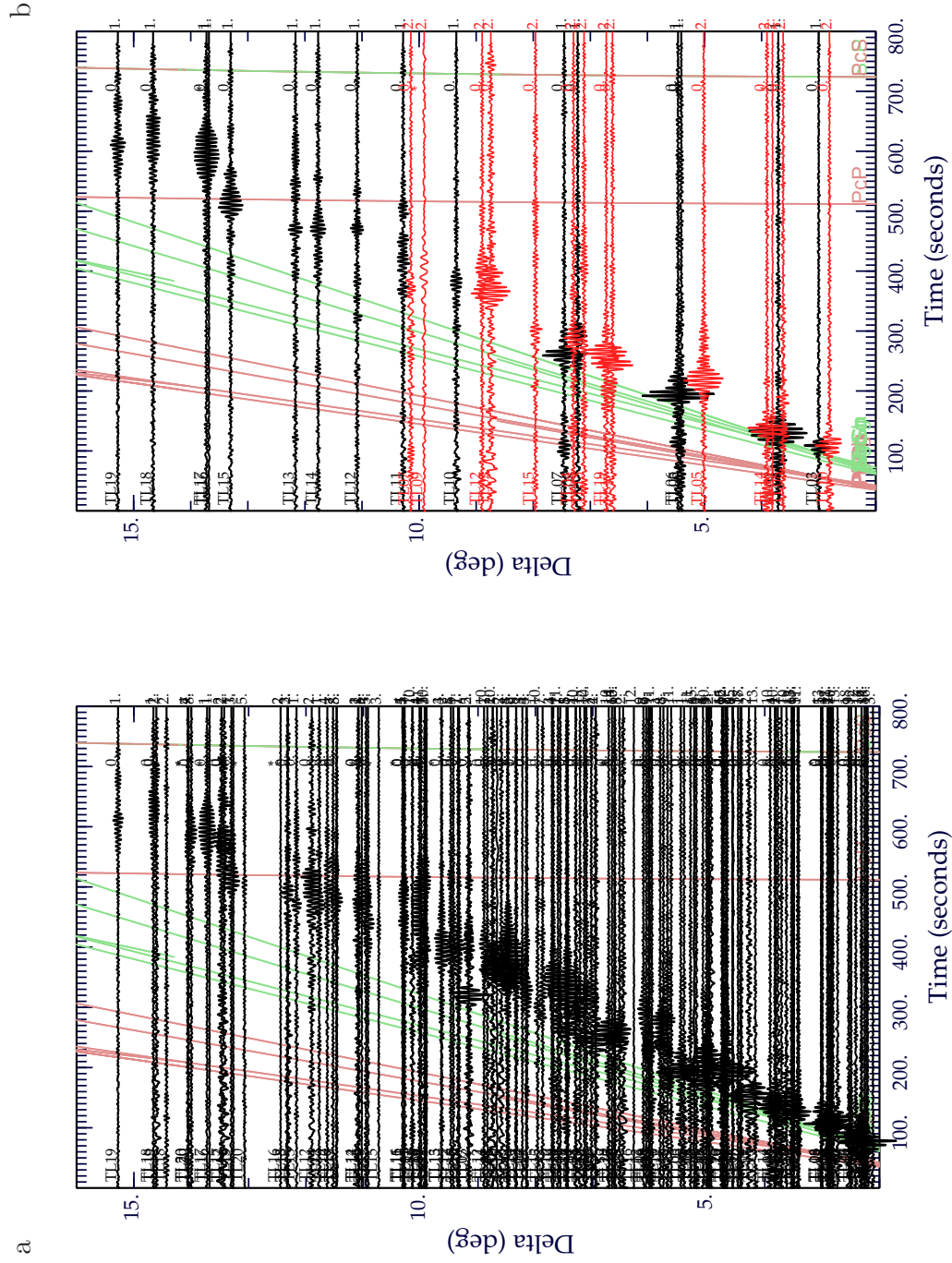


Figure 4.9: The extracted Green's functions of Rayleigh waves from cross-correlations of TASMAL experiment for all stations. a) Record section of recovered full wavefield. b) Record section between TL01 and rest of the stations (black) and TL11 and rest of the stations (red).

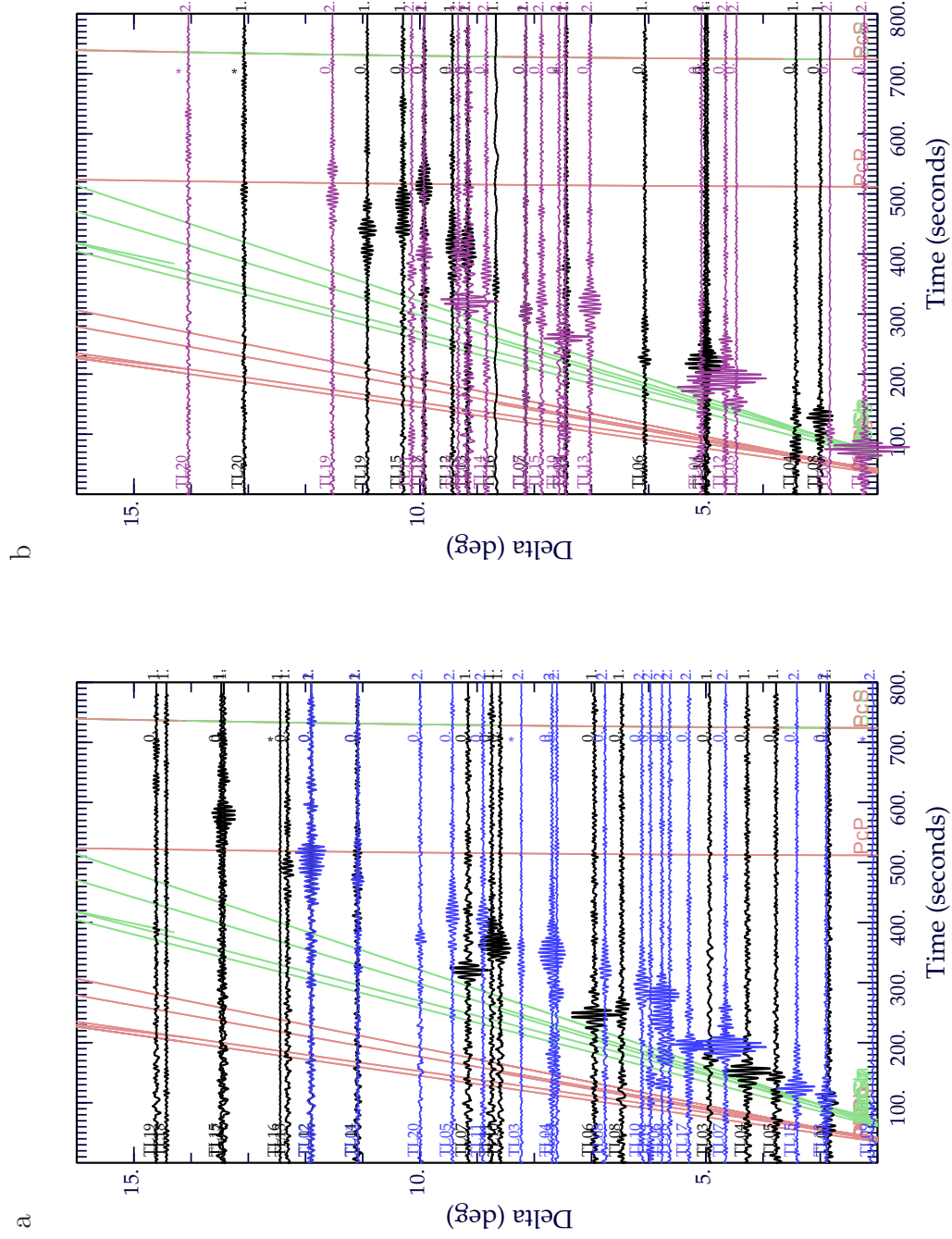


Figure 4.10: The extracted Green's functions of Rayleigh waves from cross-correlations of TASMAL experiment. a) Record section between TL02 and rest of the stations (black) and TL07 and rest of the stations (purple). b) Record section between TL05 and rest of the stations (black) and TL12 and rest of the stations (blue).

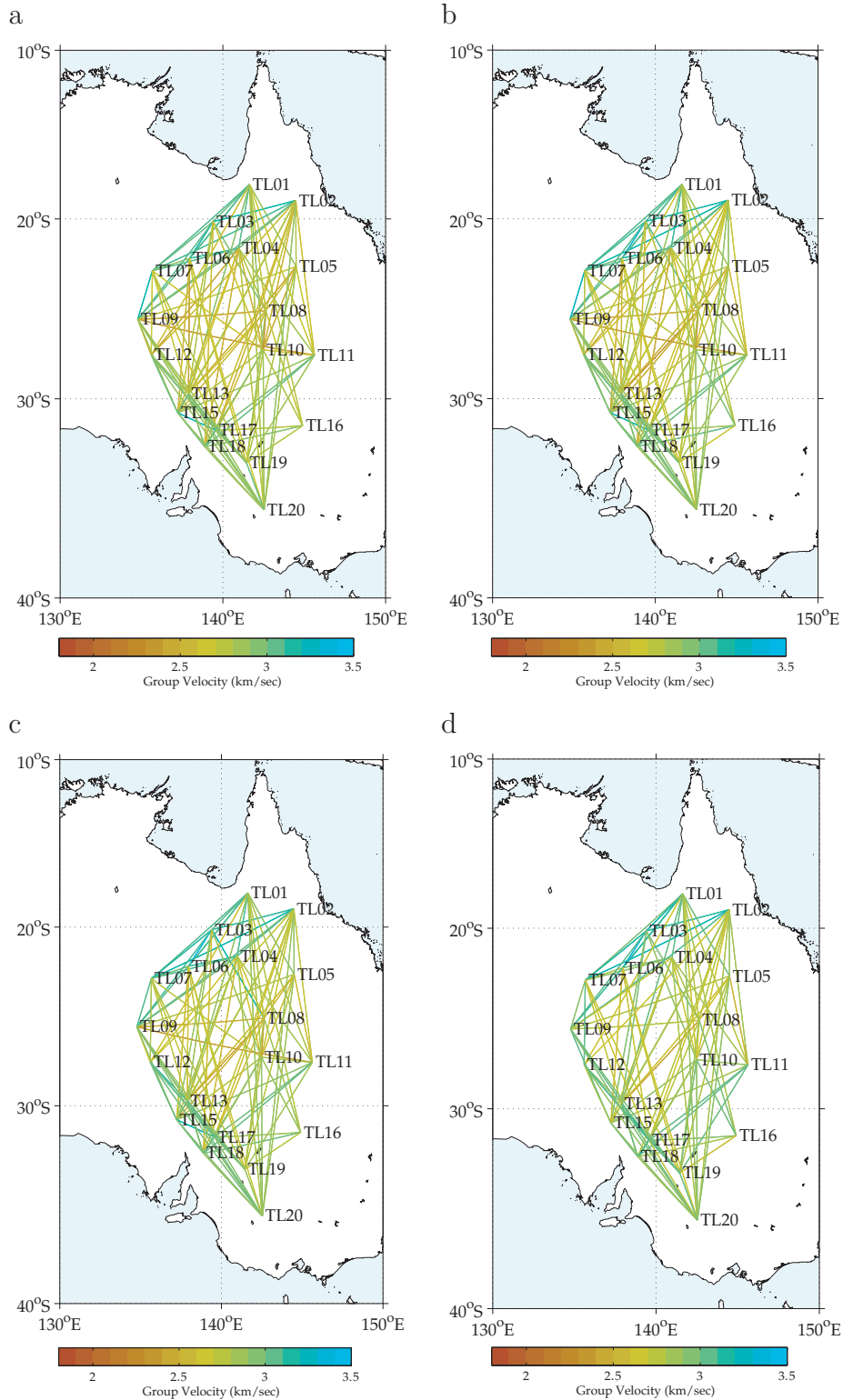


Figure 4.11: The group wavespeed of the cross-correlations between TASMAL stations for different frequencies. a) 0.2 Hz. b) 0.15 Hz. c) 0.12 Hz. d) 0.08 Hz. The effect of the structure on the wave propagation is visible from the northern and southern part of the array.

4.8.2 Extraction of Green's Functions for the Australian Continent

The compilation of the data from 1992 to 2006 was done including all of the temporary and permanent broadband stations across the Australia for the vertical component. All of the available data was used for the calculation scheme with a lower duration limit of 15 days up to several years. Overall 2299 individual cross-correlation section were calculated from the available data. Figure 1.1 indicates the extensive coverage of the continent using portable stations. Initially all the available station pairs for portable stations were used for cross-correlation and then the permanent stations were added as tie points for the temporary deployments. In table 4.1 and table 4.2, the duration of the data employed in this study is given.

Portable Deployments	Time Duration
BTL	1992
SKIPPY	1993-1996
KIMBA	1997-1998
QUOLL	1999
WACRATON-Phase 1	2000-2001
TIGGER	2001-2002
WACRATON-Phase 2	2002-2003
TASMAL	2003-2005
LINKAGE	2005-2006
MT. GAMBIER	2006-

Table 4.1: The time duration of the portable broadband experiments used in this study.

The handpicked arrivals of Rayleigh wave component of Green's functions were used for either the causal and acausal times depending on the shape and signal to noise ratio of the signal. Then the signals were bandpassed as it was given in previous section to extract the group velocities from the arrival times where the interstation distance is a known parameter. In figure 4.13, record sections of cross-correlations are presented for four permanent stations located in west Australia (NWA0), central Australia (WRAB), eastern Australia (CTAO), and southeast Australia (CAN) with all of the temporary

Permanent Stations	Time Duration
CTAO	1992-2006
NWAO	1992-2006
WRAB	1994-2006
TAU	1994-2006
CAN	1999-2005
MBWA	2002-2006
ARMA	2005-2006
BBOO	2005-2006
BLDU	2005-2006
COEN	2005-2006
EIDS	2005-2006
FITZ	2005-2006
FORT	2005-2006
KMBL	2005-2006
MCQ	2005-2006
MUN	2005-2006
STKA	2005-2006
TOO	2005-2006
YNG	2005-2006

Table 4.2: The time duration of the permanent broadband stations used in this study.

deployments. The extracted two-sided Green's function of the Rayleigh wave is clearly observable for all of the sections.

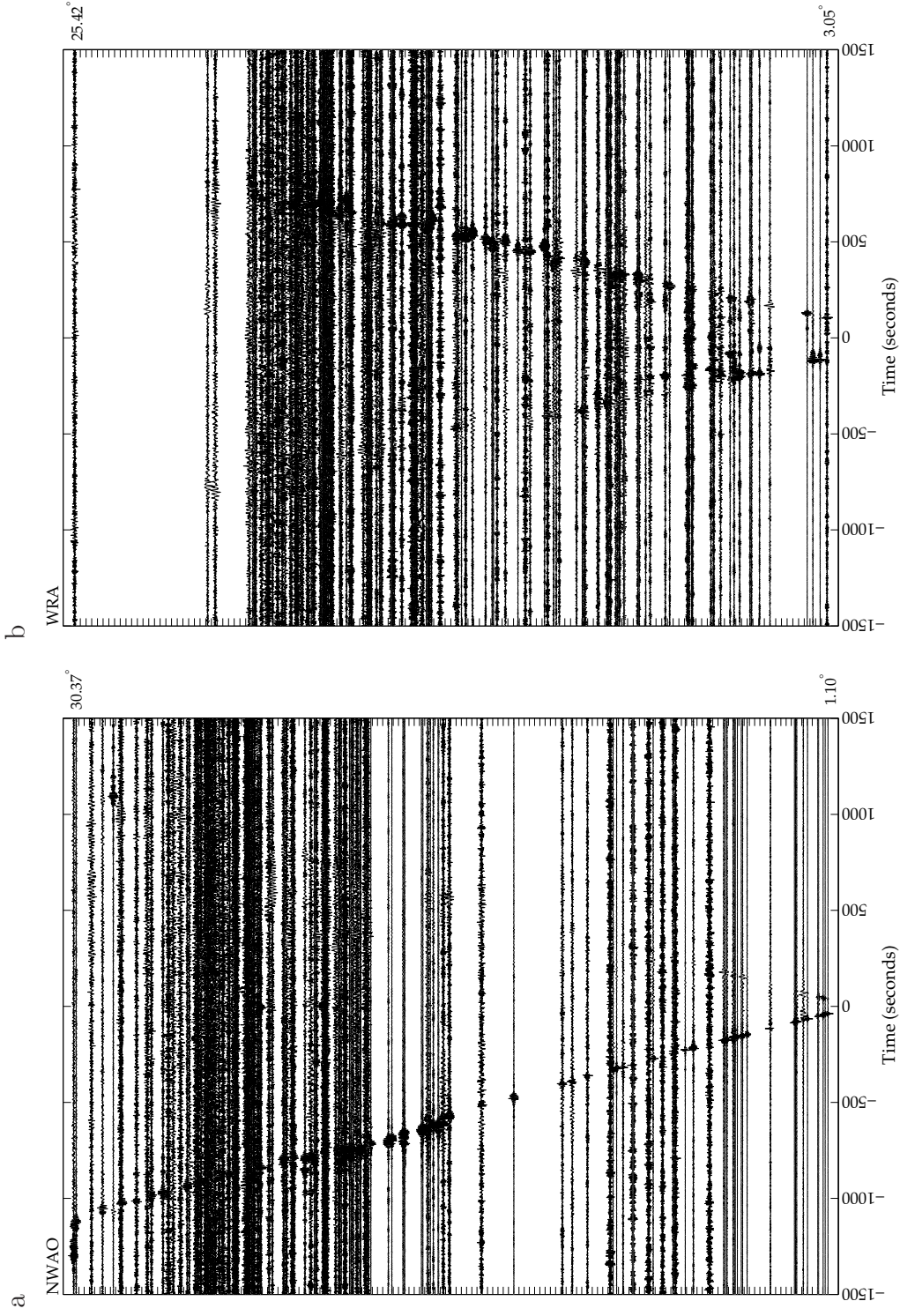


Figure 4.12: The extracted Green's functions of Rayleigh waves from cross-correlations between temporary and permanent seismic stations across Australia for three different seismic permanent stations with given minimum and maximum interstation distances. The increase in arrival time is evident with the increasing interstation distance. a) NWA0. b) WRA.

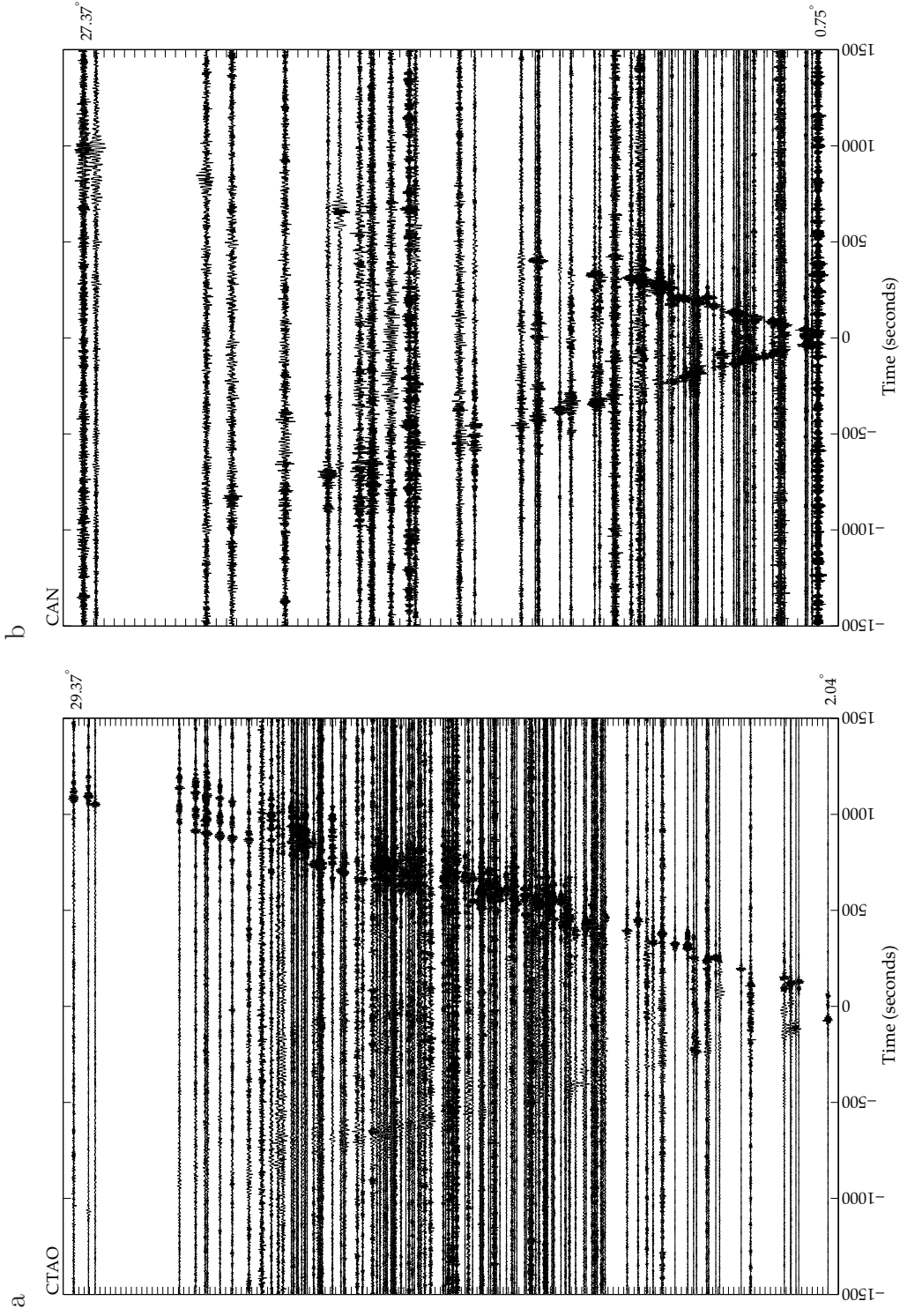


Figure 4.13: The extracted Green's functions of Rayleigh waves from cross-correlations between temporary and permanent seismic stations across Australia for three different seismic permanent stations with given minimum and maximum interstation distances. The increase in arrival time is evident with the increasing interstation distance. a) CTAO. b) CAN.

The number of raypaths after filtering varied slightly with the different frequencies due to the occasional failure of the filtering process on the selected waveform. These signals were removed so that the number of raypaths changed. The table 4.8.2 shows the total number of raypaths for the selected frequencies which were used in this study to create the tomographic image.

Frequency (Hz)	Total
0.08	946
0.12	1166
0.15	1166
0.20	1158

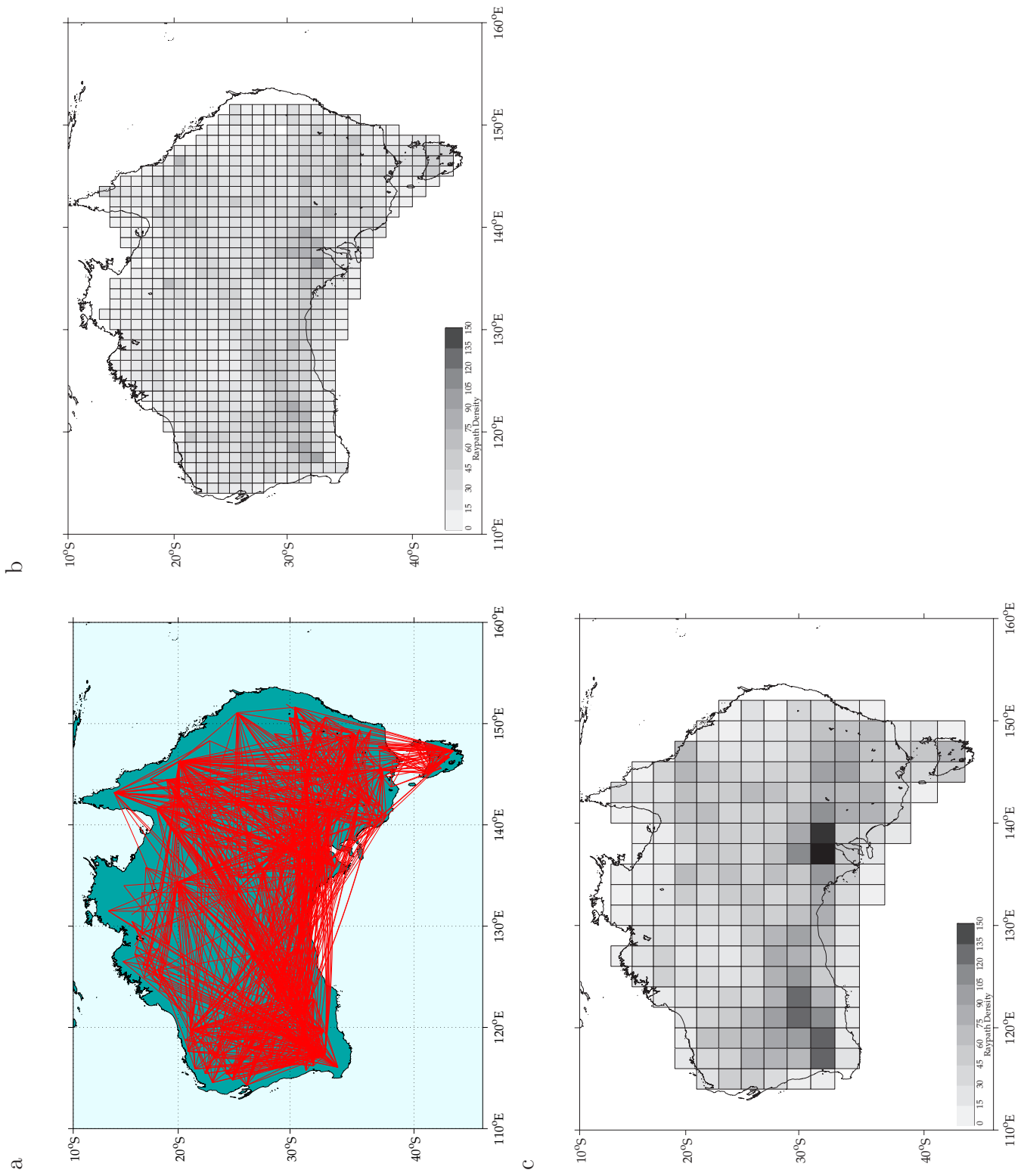
Table 4.3: The number of raypaths which were used in the tomographic inversion respect to the frequency. For some of the frequencies, particular waveform failed to give reliable result so that it was excluded.

The map of raypaths used in this study and an accompanying raypath density diagrams for two different cell sizes are given in figure 4.8.2 for the frequency 0.12 Hz which correspond to maximum amount of raypaths used in this study. Although the general coverage is sufficient, there are some gaps in the mid-west of Australia and some coastal areas where deployments have not been taken place yet. In the ray density diagrams, the effect of permanent stations as tie points is clearly observed with dense raypaths. The path distribution is quite different from that employed in the surface wave tomography studies of the Australian Continent (see, e.g., Fishwick et al. (2005)). The Green's function estimates are confined to the continent, rather than from regional earthquakes into Australian Stations. In particular this means that there is less sampling of northern Australia, but all paths are continental.

With the increasing number of broadband stations in Indonesia, and new or upgraded stations in northern Australia it should in future be feasible to attempt Green's function estimation extending to Indonesia. However, at short periods the contrasts in continental and oceanic structures may prove problematic for interpretation.

Figure 4.8.2: a) The map of connecting raypaths between two stations for 0.12 Hz. b) Path density diagram from the data coverage for cell size

of $1^\circ \times 1^\circ$. c) Path density diagram from the data coverage for cell size of $2^\circ \times 2^\circ$.



4.8.3 Resolution Tests

To illustrate the recovery success of the tomographic results, a synthetic test was carried out initially. The checkerboard test consists of specifying horizontal/vertical low and high velocity perturbations and then trying to recover it with the current ray path coverage. This approach gives a chance for assessing the success of the recovery for a hypothetical model which has sharp velocity perturbations from the actual data raypaths.

With the current raypath coverage, two tests were conducted for horizontal resolution for $2^\circ \times 2^\circ$ and $1^\circ \times 1^\circ$ with a maximum velocity perturbation of 1 km/sec respectively and it was inverted with FMM and subspace inversion. A flat rate of Gaussian noise is added for each of the raypath with a deviation of 0.8.

Although, checkerboard tests gives an important measure of information for assessing the reliability of the tomographic inversion, some care has to be taken for assuming that the recovery success of the synthetic structure would be the indicator for the recovery of the real structure. It has been showed by L  v  que et al. (1993) that the simple checkerboard test only depicts the resolution for a certain anomaly, since it is possible to recover small scale structure using certain path coverage that may not able to resolve larger scale anomalies.

From the checkerboard results, it can be inferred that the general recovery is good for both of the cell sizes. But at the edges and mid-west Australia, lack of recovery of some patches and horizontal smearing are evident due to the gaps of the instrumentation in these parts and nonoverlapping time periods of the previous temporary deployments. As a result, the current coverage of the raypaths offers a reasonable imaging horizontal gradients of the Australian crust.

4.8.4 Tomographic Inversion Parameters and Models

The tomographic inversion has some dependencies on the user chosen parameters of the objective function given at eq.(4.79) related to the inverse problem which can affect the solution i.e.: the seismic image. In order to show to effects of these, a number of experiments with the inversion scheme were done before determining the final image for geological interpretation.

The trade-off between the data and the satisfying the regularization con-

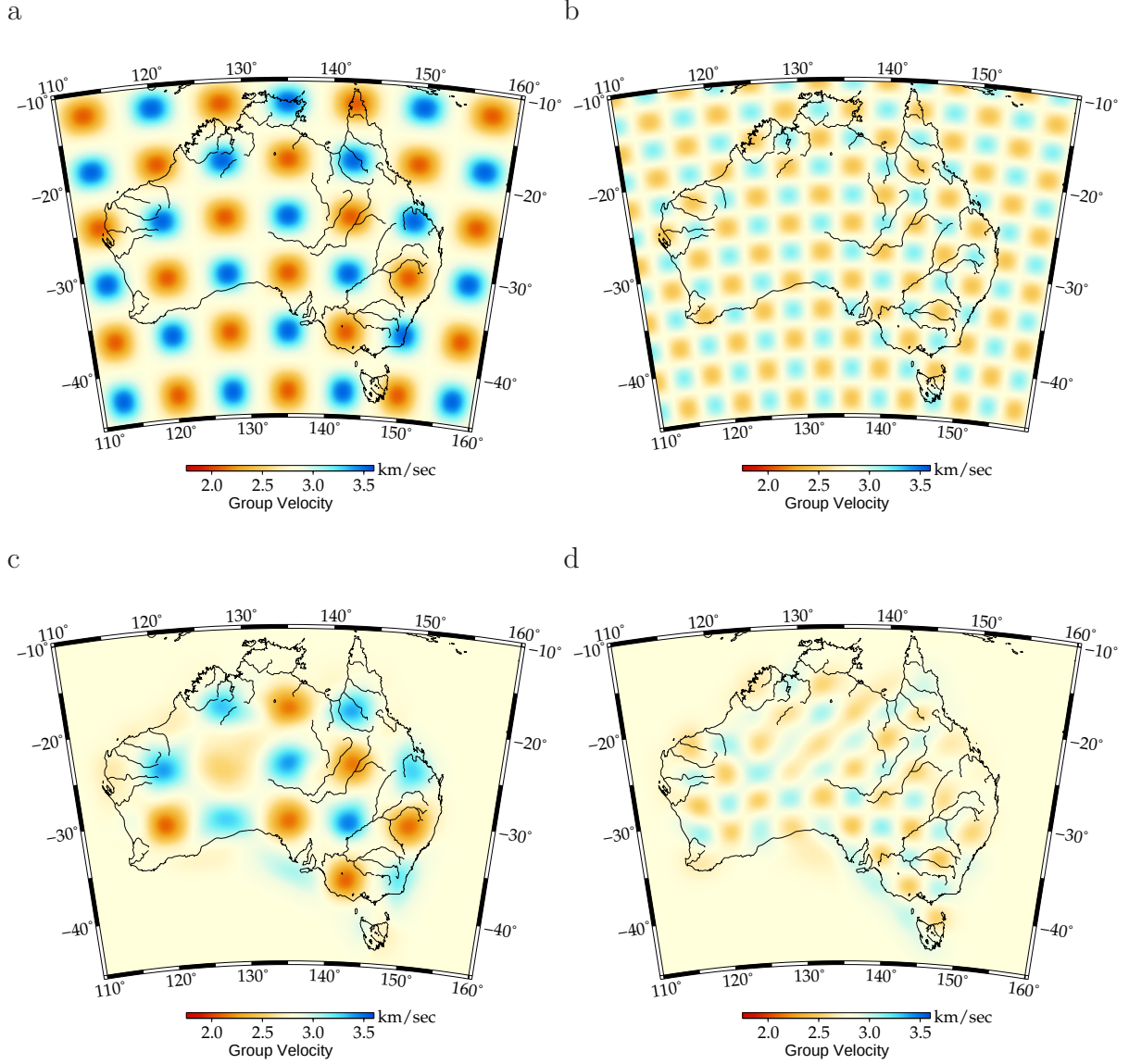


Figure 4.14: Checkerboard tests: a) Given velocity model for $2^\circ \times 2^\circ$ with a maximum velocity perturbation of 1 km/sec. b) Given velocity model for $1^\circ \times 1^\circ$ with a maximum velocity perturbation of 1 km/sec. c) Estimated velocity model for $2^\circ \times 2^\circ$ with the maximum raypath coverage. d) Estimated velocity model for $1^\circ \times 1^\circ$ with the maximum raypath coverage.

straints were explored by inverting with the highest raypath coverage with different sets of damping ϵ and smoothing η values. Although the vast given range of parameters, the final images do not show major differences shown in figure 4.15, which shows the success of parameterization.

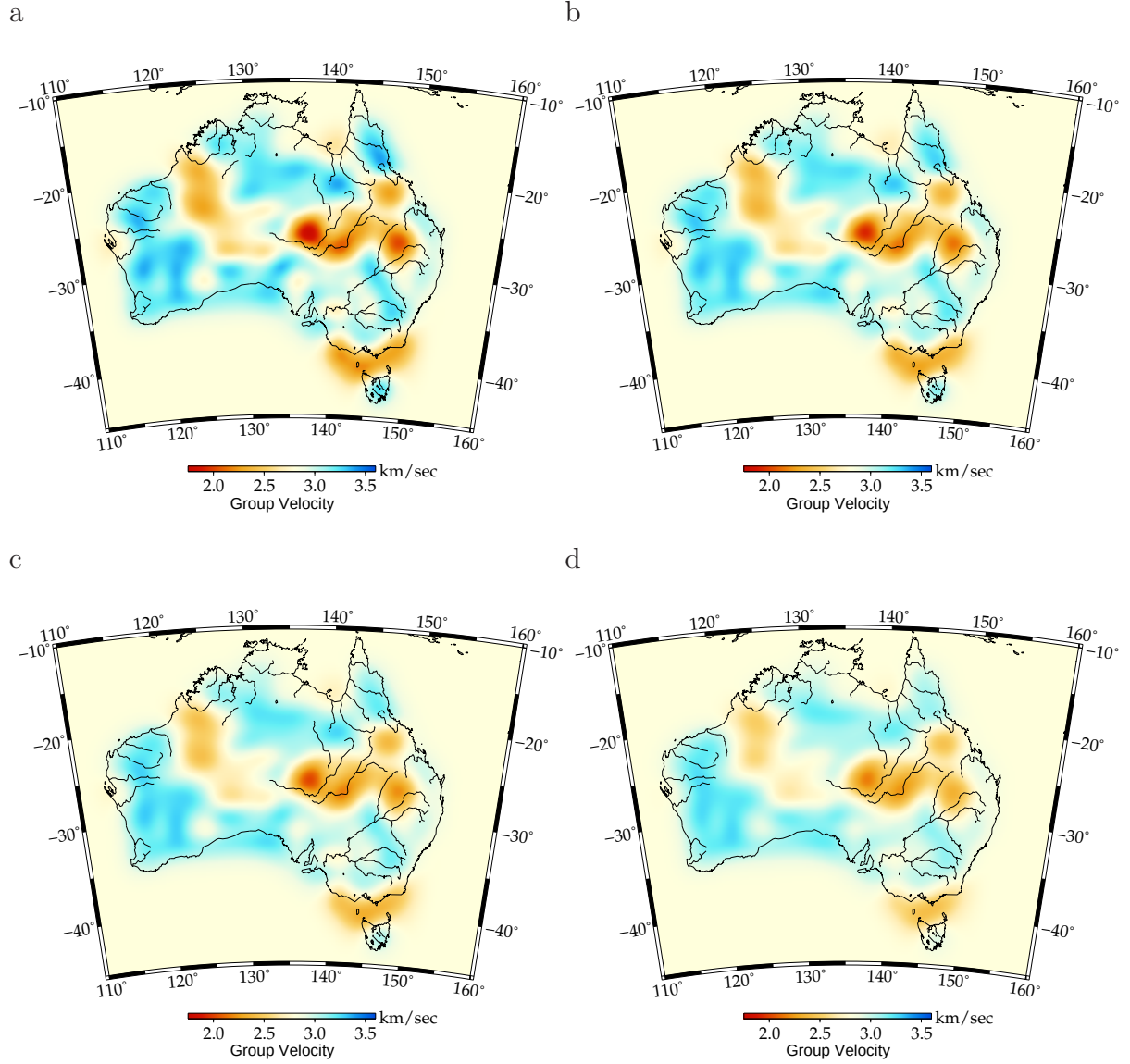


Figure 4.15: The tomographic images for different damping ϵ and smoothing η values. a) $\epsilon = 1000$, $\eta = 1000$. b) $\epsilon = 20000$, $\eta = 50000$. c) $\epsilon = 1000$, $\eta = 1000$. d) $\epsilon = 50000$, $\eta = 20000$.

The great circle paths between the different stations are plotted in figure 4.8.2a. The fully nonlinear inversion with iterative calculation of raypaths shows that the effective paths between stations can deviate very significantly from the great circle. Figure 4.16 shows the raypath distribution for the final model given in Figure 4.19b. The paths try to avoid the zones of the low velocities, because we allow the for smoothing via regularization effective imaging can be achieved.

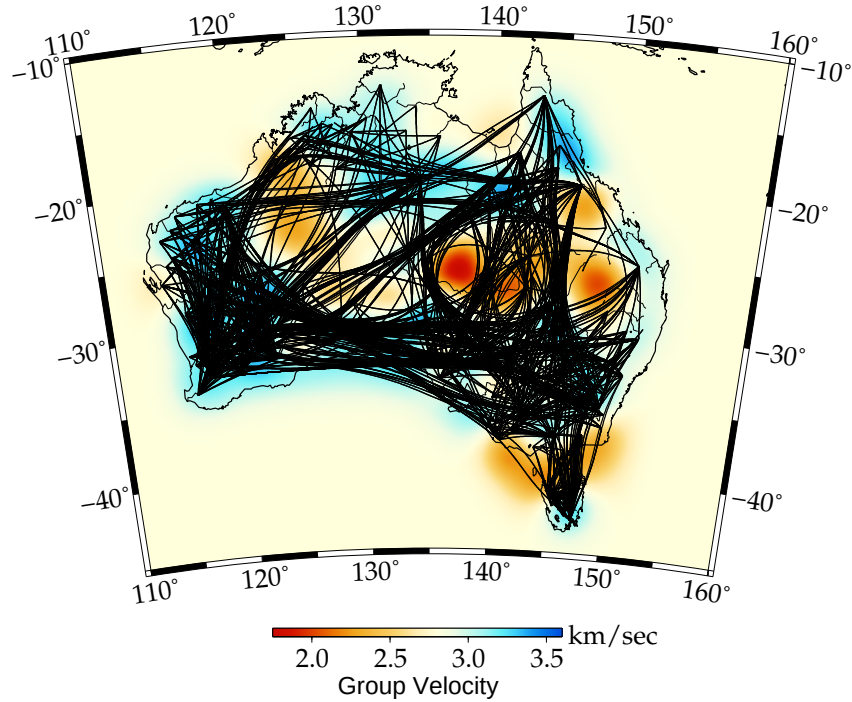


Figure 4.16: Deviation of great circle paths between two stations after fully nonlinear inversion.

The results of the group velocity tomography of the surface waves from the ambient noise are given in figure 4.17 and figure 4.18.

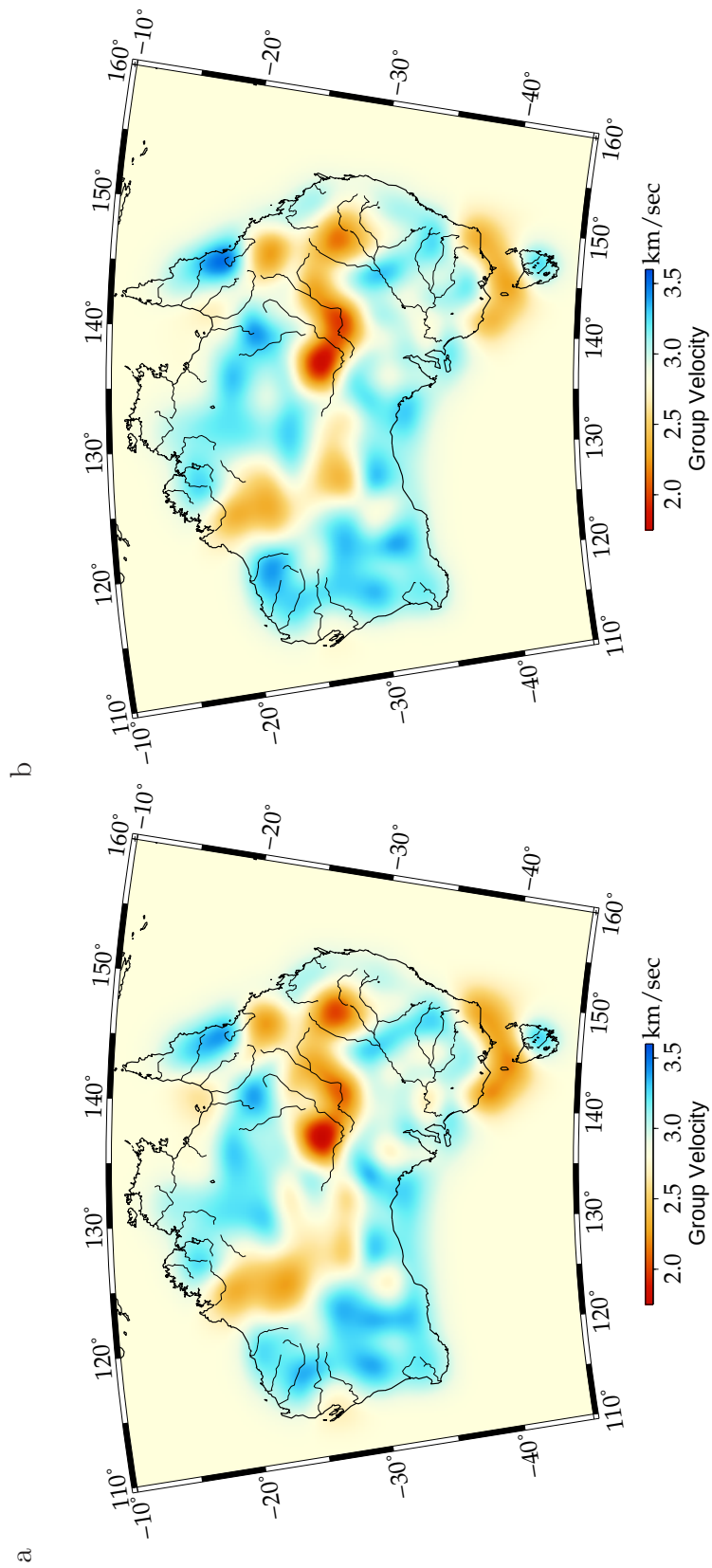


Figure 4.17: Results of group velocity tomography from ambient noise for different frequency ranges. a) 0.2 Hz. b) 0.15 Hz.

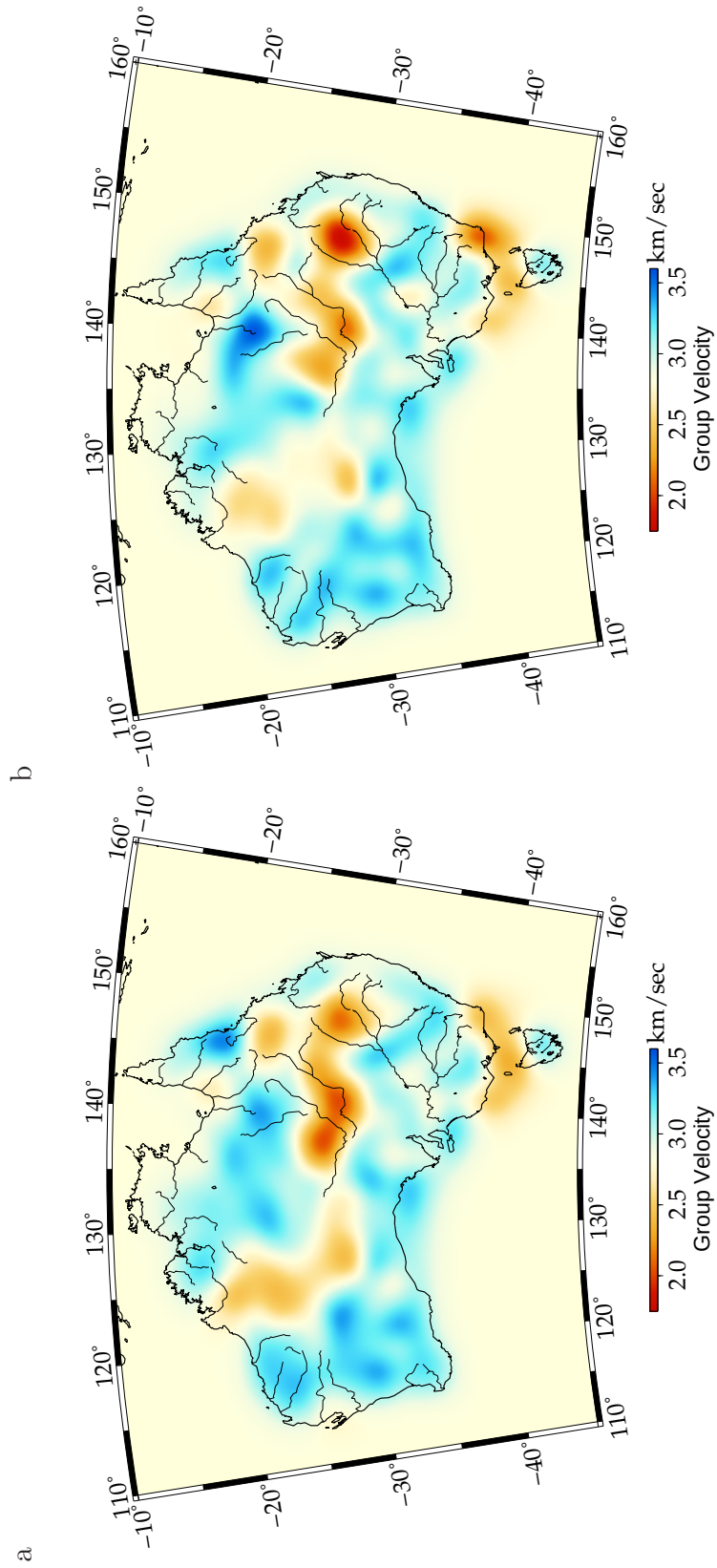


Figure 4.18: Results of group velocity tomography from ambient noise for different frequency ranges. a) 0.12 Hz. b) 0.08 Hz.

4.9 Discussion

There is a good spatial correspondence between features of the group velocity anomaly map and aspects of the major geological provinces of Australia shown in figure 4.19. The estimated group velocities varies from 1.8 km/sec at the east to 3.6 km/sec to the west with an average velocity of 3.2 km/sec. The lowered wavespeeds have a strong correlation with the thick sedimentary cover and petroleum deposits. The boundary between the Precambrian rocks of western and central Australia and Phanerozoic belts in the east have signature on the images as low group wavespeeds.

Due to the sensitivity of the surface waves on the structure employed in this study, we have used two different but closely linked mechanisms to interpret the anomalies of the tomographic images for different depths. From the modeling of the Rayleigh wave derivatives (figure 4.7 and figure 4.8), the high frequency part of the surface waves are confined in the first 3 km, the effect of the sediments will be more dominant on the propagation which also correlates with some of the major low velocity anomalies of the images. In figure 4.21, the depth of the surface sediment cover and its related shallow depth tomographic image is given. There is a general agreement on the presence of the sediments and for some of the low velocity anomalies. For longer periods, the sampling area will be in midcrustal depths (10-15 km), where seismic velocity increases. However, the low velocity anomalies present on the longer periods can be described with the heat flow processes. Relatively heated crust will result in lowered seismic velocities. In figure 4.22, the map of the crustal temperatures at 5 km depth estimated from the boreholes and the tomographic image for longer period (12.5 sec) are given. Some of the features in the tomographic images which do not directly correlate with the surface geology and surface sediment information, can be explained by incorporating the effect of the heat on the seismic velocity.

Western Australia is the oldest and most stable part of the continent. The Archaean cratons have the fastest group velocity structure which also complies with the previous findings of surface wave studies, e.g., Fishwick et al. (2005); Simons et al. (1999); Zielhuis & van der Hilst (1996) for larger depths. However, the tomographic images of this study do not show a clear separation on the boundaries of Pilbara Craton in the north and Yilgarn in the south where the extracted group velocities are higher than the other

parts of the continent (figure 4.19a[2] and figure 4.19a[3]). Due to the limited path coverage at the western margin of the continent, structural information has not been extracted.

The central Australia has some striking features. Between western Australia and North Australian Craton, a low velocity region with velocities around 2.5 km/sec indicating the existence of thick sedimentary features, located. The two branch structure of the low velocity anomaly correlates with the location of the Amadeus Basin in the north and Officer Basin in the south (figure 4.19a[4] and figure 4.21a[B&C]). The Musgrave Block is located between the two basins and shows a signature on the tomographic images. In the north, the location of Kimberley Block is well recovered from east (figure 4.19a[1] and figure 4.21a[A]). Although Kimberley Block is covered with ancient sediments with little basement exposed, Graham et al. (1999) suggested the existence of an Archaean lithospheric-mantle keel from the analysis of isotopic data. This idea is supported with the recovered fast velocities from the tomographic images for longer periods. In addition to this, a high velocity anomaly for this region was also observed in the shallow depth images of Fishwick et al. (2005). Towards to east, Mt. Isa Block exists, which comprises largest Proterozoic areas in the continent (Betts et al., 2006). Also from the receiver function part of this study and Clitheroe et al. (2000), the greatest crustal thicknesses (> 40 km) have been found in this block. The seismic velocity change is significantly high in the tomographic images (figure 4.19a[5]). In the east of Mt. Isa Block, Georgetown Inlier located. This Proterozoic unit has fast group velocities around 3.2 km/sec (figure 4.19a[6]). Also Simons et al. (1999) concluded that part of a rather thin high wave speed exists with a pronounced low velocity zone underneath this block.

Towards the east Australia, the most important feature is the hypothetical *Tasman Line*. Tasman Line was originally proposed by Hill (1951), which separates Precambrian west and central Australia from Phanerozoic east. As more studies were carried out, various propositions were made to define Tasman Line. In figure 4.20, the proposed location of Tasman Lines from various studies are compiled and given from the work of Kennett et al. (2004). Direen & Crawford (2003) compiled various geophysical evidence and definitions of Tasman Line. Fishwick et al. (2005) and Kennett et al. (2004) showed the shear-wave velocity structure for lithosphere for Australia. In these stud-

ies, tomographic images show a significant contrast between east, centre and west of the continent for larger depths. However, the structure presents complex picture than the others at 75 km depth. The shorter period tomographic images show multiple short velocity zones with velocities lower than 2.0 km/sec, which also coincide with given sediment thickness map (figure 4.21a[D]). However, the orientation of the anomalies do not present a single well defined boundary between the Precambrian and Phanerozoic Australia in this period range. In the eastern tip of the continent, there exists another low velocity region. However, one should be careful for attributing this to a structure, where the raypath distribution is limited in this area.

For longer periods, some of the low velocity anomalies show strong correlation with the estimated crustal temperatures at depth even though surface sediments do not extend to these depths (figure 4.22b[I,II,III,IV]). For example, there is no significant sediment cover on the Bass Strait. However, the estimated low velocities correlates with crustal temperature anomaly map (figure 4.22b[IV]).

Mitchell et al. (1998) used attenuation tomography with L_g coda waves for delineating crustal structure of the Australian continent from the permanent stations across Australia. They concluded that the obtained crustal Q_0 values are lower than the expected values for other stable regions of the Earth. The crustal structure of Australia from their work is given at figure 4.23. If we compare the anomaly around Tasman Line with the images of this study, e.g., figure 4.18b, the presence of a broader agreement can be inferred.

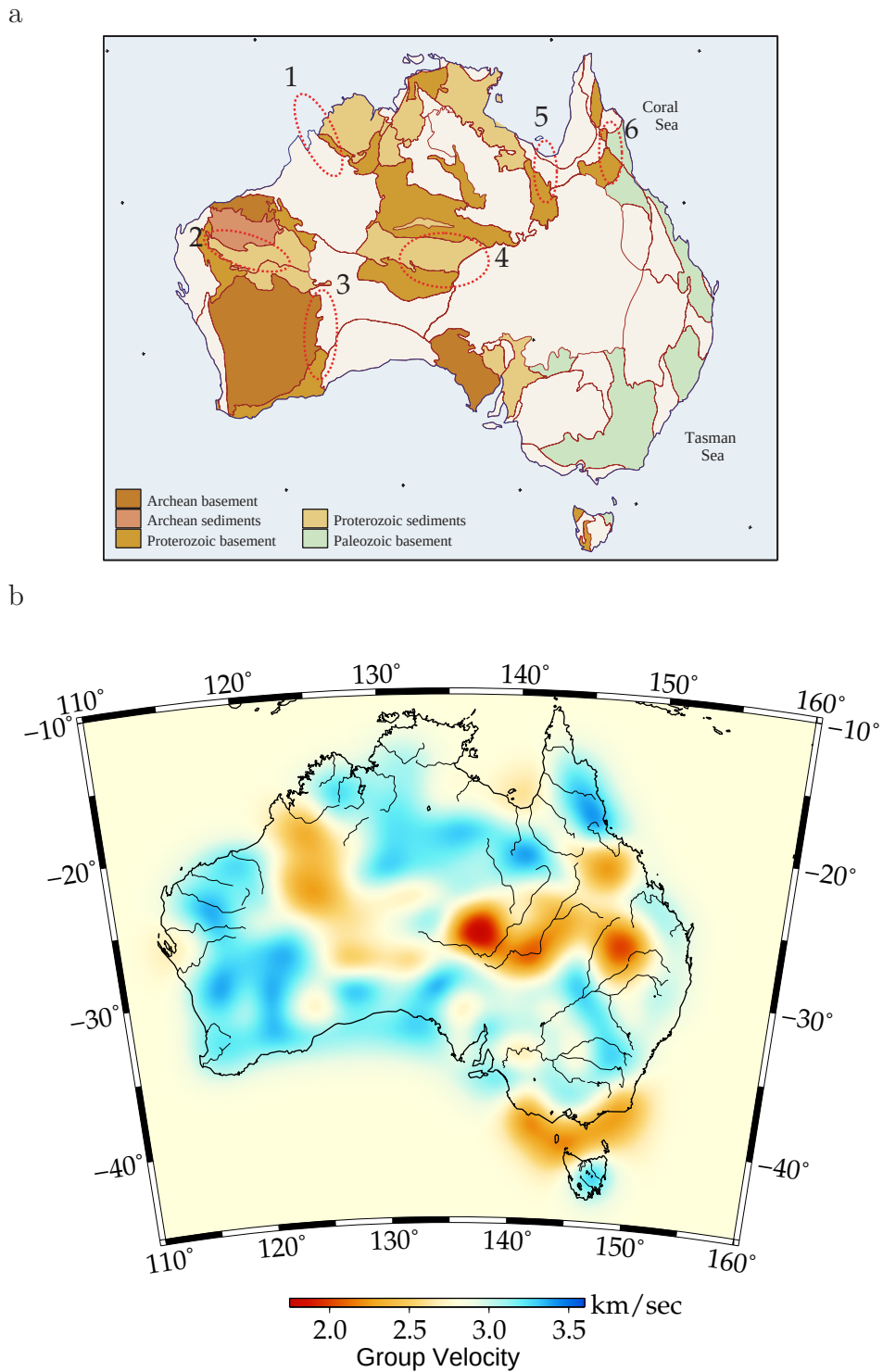


Figure 4.19: a) The surface geology of Australia is given with the matching anomalies (red ellipses). b) The group velocity image for 5 sec.

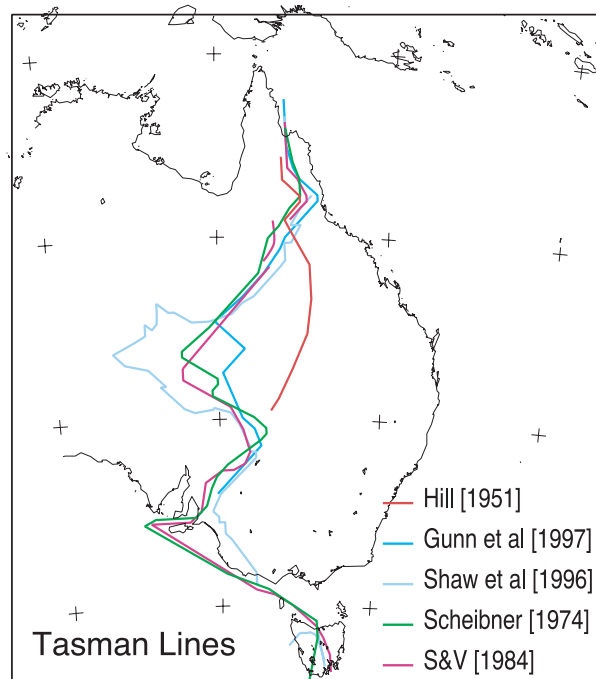
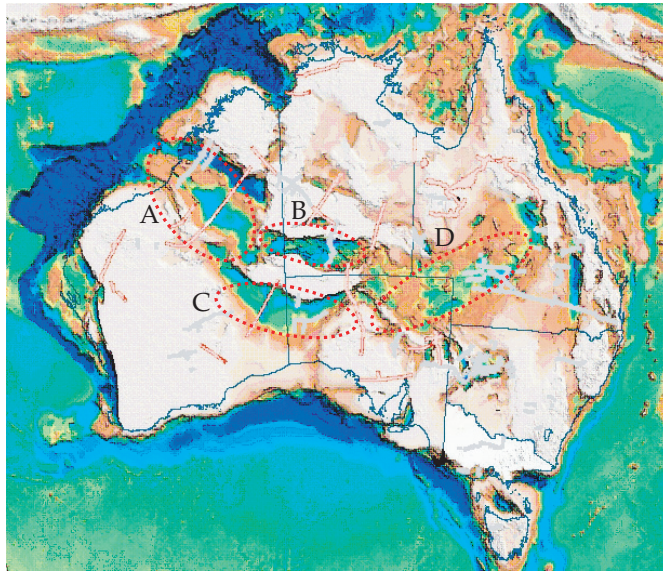


Figure 4.20: Locations of variants of the Tasman Line from Kennett et al. (2004).

a



b

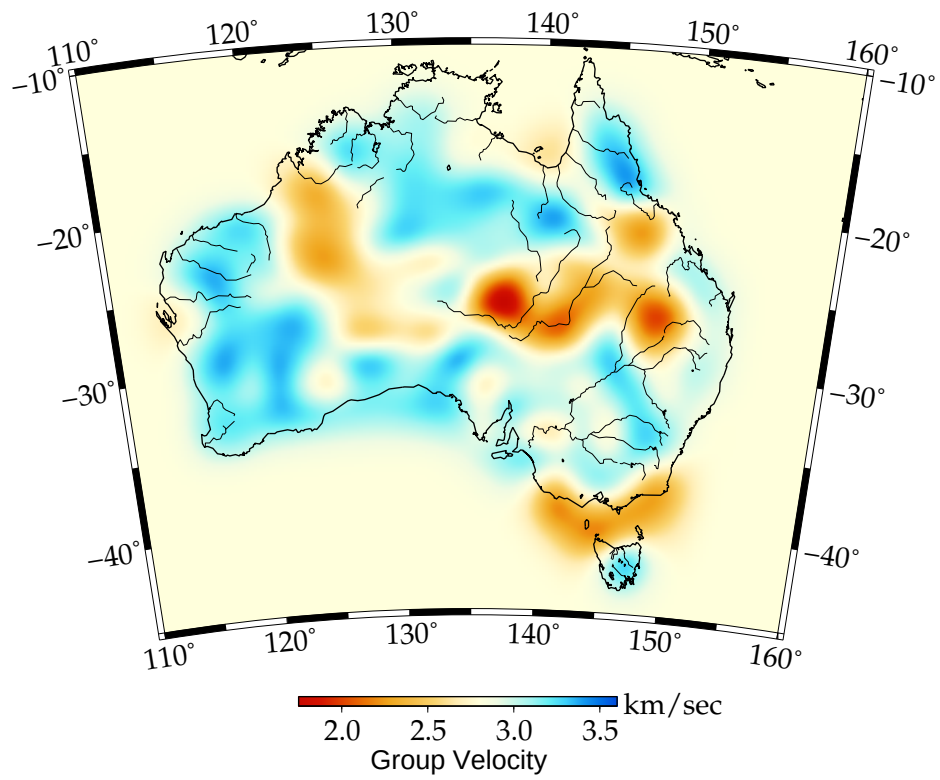
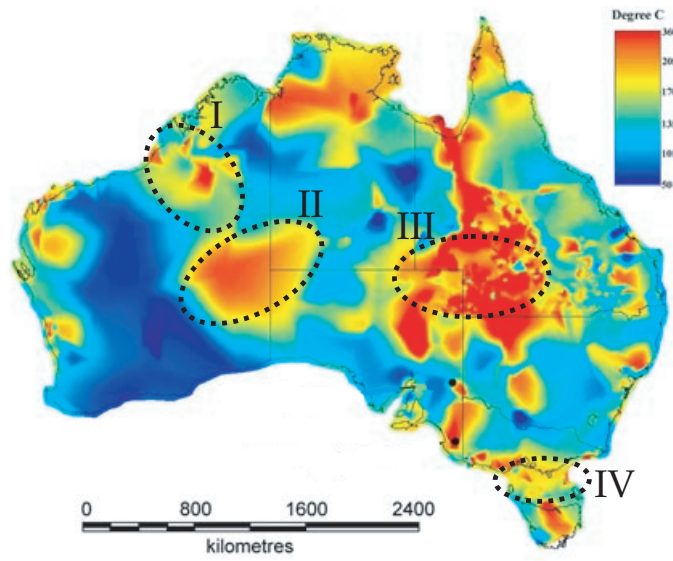


Figure 4.21: a) The sediment thickness distribution for Australia is given with the matching anomalies (red ellipses). b) The group velocity image for 5 sec.

a



b

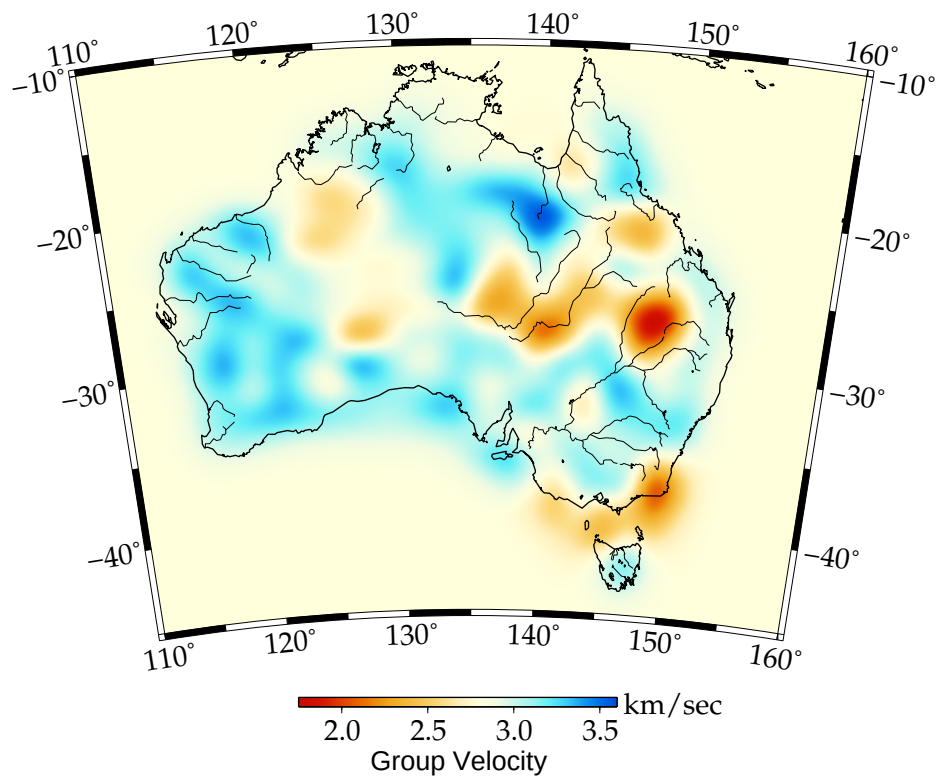


Figure 4.22: a) The estimated crustal temperature at 5 km depth of Australia is given with the matching anomalies (black ellipses) (Wyborn et al., 1994). b) The group velocity image for 12.5 sec.

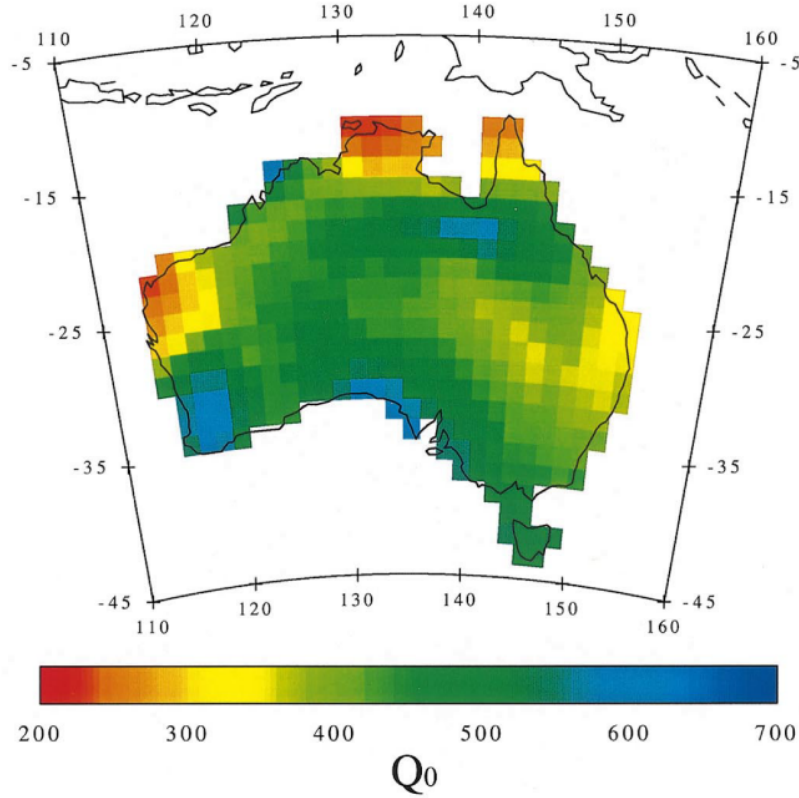


Figure 4.23: The crustal structure of Australia from attenuation tomography (Mitchell et al., 1998). The notable feature is the low Q values around Tasman Line.

In conclusion, the group wavespeed tomography of the ambient noise imaged most of the major tectonic blocks of the continent in crustal scale, which was not done in the past with the earthquake data due to the insufficient ray-path coverage. The thick sedimentary cover which is present on most of the places of the continent has a profound effect on the images. It is also interesting to note that, for most of the locations of the low velocity anomalies, a known petroleum deposit exist, e.g., Bass Strait production area. The Tasman Line in the east does not constitute a simple transition from low to high group velocities unlike in the lithosphere and upper mantle.