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High Resolution Mapping of the Upper Mantle Using Interstation Phase and Amplitude Data of Surface Waves

(二観測点間の表面波位相及び振幅データを用いた
上部マントルの高分解能マッピング)



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the degree of Doctor of Philosophy

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Abstract

A new approach to the enhancement of regional-scale surface wave tomography incorporating inter-station phase and amplitude data is proposed, utilizing a high-density broadband seismic array. The use of observed amplitude anomalies for the reconstruction of tomographic models can be of great benefit for the further advancement of conventional techniques of surface wave tomography, most of which have been based primarily on the phase information of surface waves.

We have developed a method of fully non-linear waveform fitting between two stations located on a common great-circle path using the Neighborhood Algorithm as a global optimizer. By applying this technique to the Transportable Array (USArray) in the United States, we collect a large-number of inter-station phase speed data (about 130,000 for Rayleigh and 85,000 for Love waves) and amplitude ratio data (about 75,000 for Rayleigh waves) in a period range from 30 to 130 s for fundamental-mode surface waves. The inter-station distances of the measured dispersion data are mostly in a range between 300 and 800 km, which are nearly comparable to the wavelength of surface waves in the target periods of this study, and are much shorter than the epicentral distances between source and receiver for typical single-station dispersion measurements. Massive measurements of inter-station phase speeds and amplitudes of surface waves for such short paths can be of help in enhancing the horizontal resolution of a tomography model.

The phase information of surface waves, which reflects the average phase speed perturbation along a propagation path, has been the major source of information for the reconstruction of the lateral heterogeneity in the upper mantle. Amplitude anomalies of surface waves, on the other hand, have rarely been used in tomographic studies, since they are affected by a variety of uncertain factors such as seismic moment and source mechanism, local amplification at receiver, elastic focusing/defocusing due to lateral heterogeneity and anelastic attenuation. In the inter-station analysis, the effects of seismic source can be canceled out, so that we need to consider the other factors only. With an appropriate correction for anelastic attenuation, we can focus on the effects of the focusing/defocusing as well as the receiver amplification factors. Since the effects of elastic focusing/defocusing on amplitude anomalies depend on the second derivatives of phase speed perturbation across the ray path, amplitude data should be more sensitive to the shorter-wavelength structure than the conventional phase data. This indicates that the amplitude measurements of surface waves are effective in improving the lateral resolution of velocity models.

In this study, the measured inter-station phase speeds and amplitude ratios are inverted simultaneously for phase speed maps as well as local amplification factor at each receiver locations. The phase speed maps derived from both phase and amplitude data exhibits better recovery of the strength of velocity perturbations of small-scale heterogeneities than those from phase data only, with enhanced local-scale tectonic features characterized by strong velocity gradients. Furthermore, the spatial distribution of receiver amplification factor shows a clear correlation with the velocity structure, particularly in the longer period.

Isotropic and anisotropic 3-D S wave speed models in the United States are then constructed from the phase speed maps of fundamental-mode sur-

face waves. In the isotropic S wave model derived from both phase and amplitude data for Rayleigh waves, smaller-scale tectonic features with the size of about 200 km are emphasized clearly; e.g., Colorado Plateau in the western region and New Madrid Seismic Zone in the central region at the depth of around 100 km. These results indicate that inter-station amplitude ratios can be used as additional constraints on high-resolution mapping of the upper mantle structure. A preliminary radial anisotropy model derived from the simultaneous use of Rayleigh and Love waves has shown faster SH wave speed anomalies than SV above the depth of 100 km, particularly in tectonically active regions in the western and central U.S., suggesting the effects of current and former tectonic processes on anisotropic properties in the lithosphere.

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I would like to give special thanks to Malcolm Sambridge at Australian National University for the provision of his Neighborhood Algorithm code. I also thank the IRIS Data Management Center for the provision of seismic waveform data, particularly the stations of Transportable Array (TA) of USArray used in this study.

The development of the method for a nonlinear waveform fitting described in chapter 3 started as a part of the author's Master research (Hamada,

2014), which was primarily targeted for anelastic attenuation structure. The method has been extensively improved since then, and the full details of the latest version of the method is described in this thesis.

The basic concept of chapter 3 and 4 (i.e., the method of waveform fitting for inter-station phase and amplitude measurements, and their application to the joint inversion for phase speed maps) are based on the author's original paper, "Interstation phase speed and amplitude measurements of surface waves with nonlinear waveform fitting: application to USArray" by K. Hamada & K. Yoshizawa, *Geophysical Journal International*, 202, 35-54, 2015, which won a 2015 GJI Student Author Award. I would like to thank the editors of GJI for bestowing such a great honor on me.

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Chapter 1

Introduction

The seismic structure of the upper mantle reflects the dynamic processes, such as plate motion, earthquakes and volcanic activities, occurring inside of the Earth. Seismic surface waves can be of great help in mapping the lateral heterogeneity and anisotropy in the crust and upper mantle. A number of surface wave tomography models have been proposed in the past three decades, both on global and regional scales, to delineate the seismological images of the lithosphere and asthenosphere beneath continental and oceanic areas. In this introduction chapter, we briefly summarize observable data of surface waves that can be used in tomographic studies, and some technical backgrounds of surface wave tomography.

1.1 Surface wave tomography: Mapping the upper mantle

Studies on seismic tomography started in the 1970s for the reconstruction of 3-D velocity structures in the Earth utilizing travel-time anomalies of seismic body waves; e.g., Aki et al. (1977) retrieved local-scale lateral heterogeneity in several layers in the upper mantle beneath the NORSAR array in Norway, and Dziewonski et al. (1977) investigated the large-scale lateral heterogeneity in the lower mantle. Unlike seismic body waves whose energy propagates into

the deep interior of the Earth, seismic surface waves are radiated cylindrically from the source, and propagate along the Earth's surface, so that they have very good sensitivities to the shallow structures such as the crust and upper mantle.

Tomographic mapping of the upper mantle using surface waves and/or long-period waveforms have begun in the early 1980s (e.g., Nakanishi and Anderson, 1982; Woodhouse and Dziewonski, 1984). Owing to their good sensitivities to the crust and upper mantle, seismic surface waves have been the major sources of information for the reconstruction of global-scale 3-D shear velocity structures in the upper mantle (e.g., Trampert and Woodhouse, 1995; Ekström et al., 1997), including both azimuthal and radial anisotropy (e.g., Montagner and Tanimoto, 1991).

1.1.1 Phase data of surface waves

Most of the conventional surface wave studies have been based on the *single-station method* (Fig. 1.1a), in which path-average phase speeds are measured for a ray path between source and receiver. In general, the single-station method requires long epicentral distances more than 1000 km, in order to isolate the surface wave arrivals from the preceding arrivals of body waves. It is, therefore, generally applied to a large-scale tomographic mapping of global or large-scale regional structures (Fig. 1.2).

As an alternative approach to surface wave analysis for regional or local scale tomographic studies, the *two-station method* (Fig. 1.1b) has been widely used, in which the phase differences between seismograms at two stations located on a common great-circle path are measured (e.g., Dziewonski and Hales, 1972). Such a classical two-station analysis have been employed in many regional-scale studies to construct the upper mantle structure (e.g., Isse et al., 2006; Yoshizawa et al., 2010; Bakirci et al., 2012). There are several

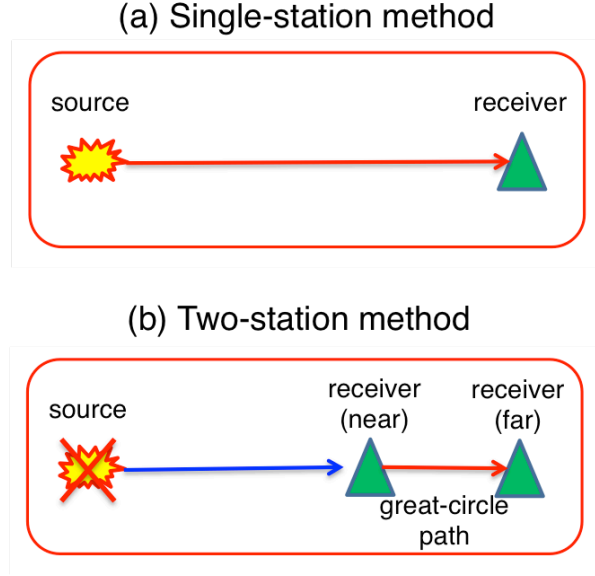


Figure 1.1: Schematic illustrations of geometrical configuration for (a) the single-station method and (b) the two-station method.

techniques to extract the inter-station phase delay information, such as the direct estimation of phase differences from unwrapped phase spectra (e.g., Yoshizawa et al., 2010; Bakirci et al., 2012), taking cross-correlation between two seismograms (e.g., Lebedev et al., 2006), and taking inter-station phase differences using the measured phases derived from single-station waveform analysis (Foster et al., 2014a).

There are two major advantages of the two-station method over the single-station method. First, while the single-station method requires well-known seismic source mechanism, no information of source mechanism is necessary in the two-station method, since source terms can be canceled out by taking the phase differences (and amplitude ratios) between two seismograms in the frequency domain. Second, inter-station measurements with path lengths shorter than 1000 km can be possible, which cannot readily be available in the single-station analysis and can be of great help in enhancing the horizontal

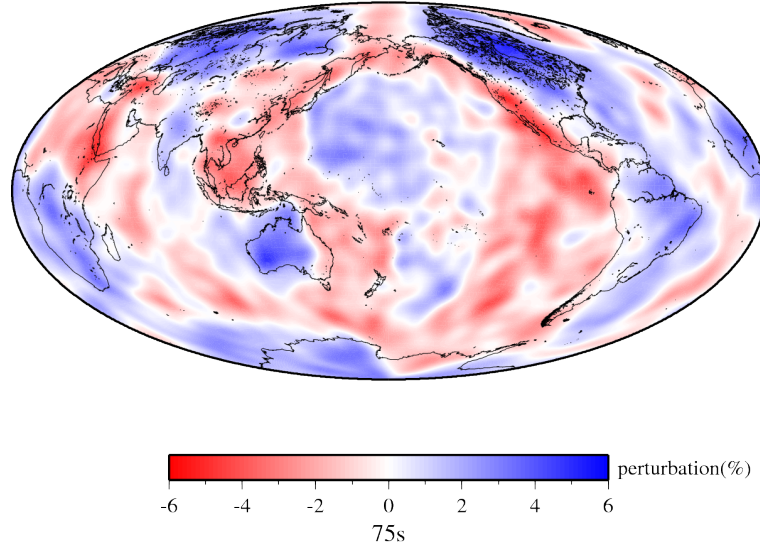


Figure 1.2: A global phase speed map of Rayleigh wave at 75 s (Ekström et al., 1997).

resolution of velocity models. There is, however, an intrinsic limitation in the inter-station analysis that the two stations have to be located on or near a common great-circle path, which tends to restrict the available number of inter-station dispersion measurements. Thus, the two-station method has been mainly used in surface wave mapping at regional or local scales for which a relatively dense broad-band seismic network can be available.

Recently, dense seismic arrays are deployed in many regions and countries; e.g., a transportable seismic array across the United States (USArray) deployed as a part of the Earthscope project, NECESSArray in the north-eastern China deployed by an international research group of Japan, China and United States, and the permanent broad-band seismic network (F-net) deployed and maintained by NIED (National Research Institute for Earth Science and Disaster Resilience) in Japan, which are of great help in activating the studies of surface wave mapping on regional scales with inter-station analysis.

Yoshizawa et al. (2010) have constructed a 3-D shear wave speed structure

of the upper mantle beneath the Japanese islands and surrounding regions from inter-station phase speed measurements of surface waves using several seismic networks including F-net, a temporary broad-band seismic network in Far-East Russia and permanent stations of global seismograph network in East Asia.

In North America, Foster et al. (2014a) have obtained the phase speed maps of Rayleigh and Love waves at periods from 25 to 100 s, based on the inter-station surface-wave phase estimation using USArray. They have extracted phase speeds between two stations by employing single-station measurements of surface-wave phases by Ekström (2011), and also estimated arrival angles of surface waves at each station using multi-station analysis for correcting the directions of surface wave propagation.

1.1.2 Amplitude data of surface waves

The investigations of the upper mantle structure using surface waves have been generally based on the measurements of phase delay, which represents the average phase speed perturbation along a ray path. To the contrary, the use of amplitude information of surface waves for tomographic studies has been rather limited. It has been well known that surface-wave amplitude data observed at a single station can be affected by many sources of uncertainties; e.g., the estimated seismic moment and source mechanisms, site responses at receiver locations, scattering and elastic focusing/defocusing during the propagation in laterally heterogeneous structure, and intrinsic attenuation caused by anelasticity in the Earth.

Nevertheless, with appropriate corrections as well as good knowledge on the velocity structures and seismic sources, amplitude data for long propagation paths have been used to construct the large-scale attenuation models (Q models), which represents the anelastic properties of the Earth materials.

Several global attenuation models of the upper mantle using the single-station measurements have shown the existence of large-scale lateral heterogeneity of anelastic attenuation in the upper mantle (e.g., Romanowicz, 1995; Selby and Woodhouse, 2000; Dalton and Ekström, 2006b; Dalton et al., 2008).

Surface wave amplitude can also be used to constrain the velocity structure, since they reflect the focusing/defocussing effects caused by lateral variations in the velocity structure as discussed in detail in the next section. There are only limited number of applications of surface wave amplitude anomalies for the reconstruction of velocity models using single-station amplitude measurements (Yomogida and Aki, 1987; Laske and Masters, 1996; Dalton and Ekström, 2006a). Amplitude ratios between two stations can also be used to extract the structural information in the Earth, but the use of inter-station amplitude data for tomographic study has been rather limited; e.g., for an attenuation study by Yang et al. (2004).

Another example of the use of amplitude data for the reconstruction of velocity models is an array-based analysis that represents the incoming wave-field as the sum of two interfering plane waves (Forsyth and Li, 2005). Yang and Forsyth (2006) have extended the method by incorporating the finite-frequency sensitivity kernels based on the Born approximation for regional-scale tomography.

1.2 Surface wave tomography with different observables

Majority of surface wave tomography have employed the information of phase data which can readily be measured using a single or two station analysis. We can also use the other types of surface-wave observables, such as arrival-angle anomalies (deviation of a propagation path from the great-circle) and amplitude anomalies. With an appropriate linearized relationship for these

observed data and the velocity structure, tomographic reconstruction can be done in a framework of surface-wave ray theory. In this section, we briefly summarize the linearized relationship between surface wave observables and phase speed perturbation, following the linearized relationship formulated by Woodhouse and Wong (1986) and Dahlen and Tromp (1998).

1.2.1 Linear relation for surface wave phases

The phase perturbation $\delta\psi$ of surface waves is related to the phase speed perturbation δc as follows,

$$\delta\psi = -k \int_0^\Delta \frac{\delta c(\theta, \phi)}{c_0} ds, \quad (1.1)$$

where k is the wavenumber, Δ the epicentral distance, ϕ the along-path coordinate, θ the path-perpendicular coordinate, c_0 the reference phase speed, and ds the ray path along the great circle. The observed phase data reflects the average phase speed along the ray path, and thus the lateral heterogeneity whose scale is much smaller than the path length and wavelength of surface waves is hardly reflected in the phase speed maps.

To overcome such issues, there have been attempts in the theoretical developments to incorporate the effects of finite-frequency of surface waves (e.g., Yoshizawa and Kennett, 2002a, 2004; Zhou et al., 2004, 2005; Fichtner et al., 2010), although such advanced analyses are based on the single station measurements with relatively long propagation paths over 1000 km. On the other hands, in the inter-station phase speed measurements using dense seismic arrays, the average length of inter-station paths can be shortened dramatically depending on the station density of the array, and thus the achievable lateral resolution of phase speed maps can be enhanced to some extent.

1.2.2 Linear relations for derivative-sensitive observables: arrival-angles and amplitudes

As alternative approaches for enhancing the resolution of velocity models, we can envisage the use of different types of surface wave observables, such as polarization (or arrival-angle) anomalies and amplitudes, which are sensitive to the lateral velocity gradient between the source and receiver.

The polarization or arrival-angle anomaly, ζ , which represents the deviation of the surface-wave propagation path from the corresponding great circle, can be represented by the path integral of the lateral gradients of phase speed (e.g., Woodhouse and Wong, 1986),

$$\zeta = - \int_0^\Delta \frac{\sin \phi}{\sin \Delta} \partial_\theta \frac{\delta c}{c_0} d\phi, \quad (1.2)$$

where ∂_θ represents the 1st derivative with respect to the path-perpendicular coordinate, and we omit the dependency of ζ and c on the angular frequency ω . There are only limited examples of the use of the arrival-angle information in the tomographic inversions (e.g., Laske and Masters, 1996; Yoshizawa et al., 1999). Yoshizawa et al. (1999) have shown that, from their analysis using synthetic and observed data for global-scale tomography, the polarization data exhibit better resolving power for smaller scale heterogeneities than the conventional phase data.

The original expression of amplitude anomaly caused by focusing/defocusing, A_F , for a source-receiver path has been derived by Woodhouse and Wong (1986), which is then modified slightly by Wang and Dahlen (1994) and Dahlen and Tromp (1998) as follows,

$$\begin{aligned} \ln A_F = & \frac{\delta c_R}{2c_0} + \frac{\delta c_S}{2c_0} \\ & + \frac{1}{2 \sin \Delta} \int_0^\Delta [\sin(\Delta - \phi) \sin \phi \partial_\theta^2 - \cos(\Delta - 2\phi)] \frac{\delta c}{c_0} d\phi, \end{aligned} \quad (1.3)$$

where δc_S and δc_R indicate the phase speed perturbation at the source and

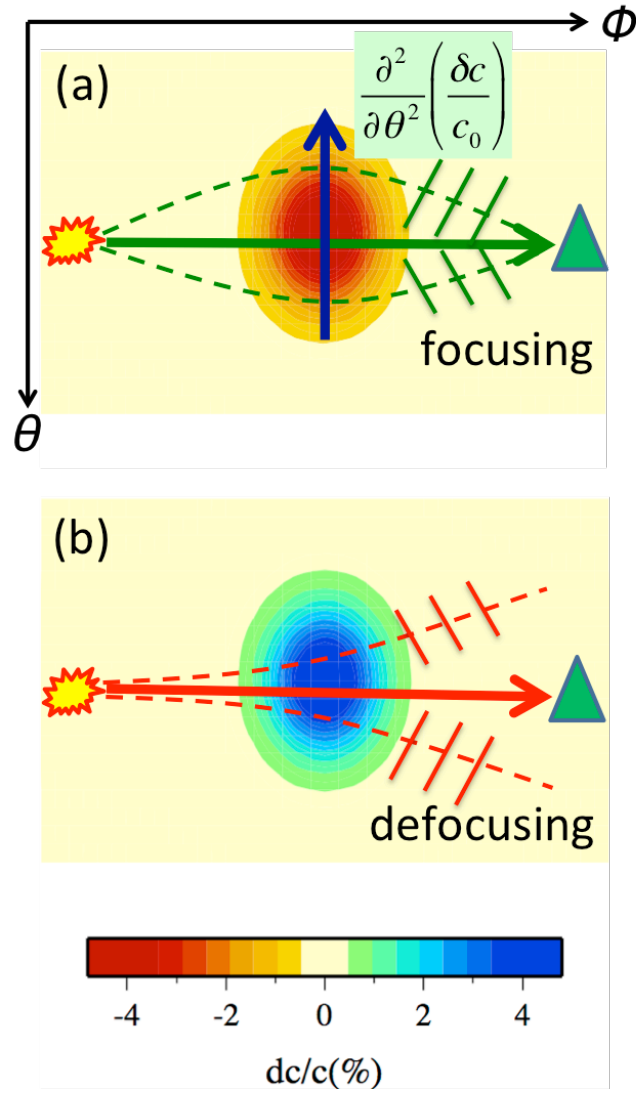


Figure 1.3: Schematic illustrations of wave propagation thorough (a) a small-scale low velocity area and (b) a high velocity area.

receiver, and ∂_θ^2 represents the second derivative with respect to the path-perpendicular coordinate (see Appendix A for more details). It is clear that amplitude anomalies of surface waves caused by elastic focusing/defocusing are sensitive to the second derivatives of phase speed perpendicular to the ray path.

Such effects deduced from (1.3) can be schematically illustrated in Fig. 1.3. The low velocity zone surrounded by relatively faster area produces large phase speed gradients on both sides of the central ray path. Thus, the observed amplitudes of surface waves, which propagate through such a low velocity zone will be affected by focusing (Fig. 1.3a). In the opposite case with fast anomaly surrounded by the relatively slower areas, the amplitude is affected by defocusing (Fig. 1.3b). The amplitude data are, therefore, more sensitive to smaller-scale heterogeneities than the conventional phase data, and suitably-corrected amplitude data can be useful for improving the lateral resolution of phase speed models (e.g., Laske and Masters, 1996; Yang and Forsyth, 2006).

Although the number of tomographic models incorporating amplitude data have been limited, there are several seismic velocity models derived from surface-wave amplitude based on the single-station analysis. Yomogida and Aki (1987) obtained phase speed maps of Rayleigh waves in the Pacific Ocean by inverting phase and amplitude data extracted from analogue seismograms based on their Gaussian beam method taking account of approximate effects of finite-frequency of surface waves. By using larger data set derived from digital seismograms, Laske and Masters (1996) have constructed global phase speed models using phase, polarization and amplitude data of surface waves. Their results suggested that the phase data in the frequency range of 3-15 mHz do not require the smaller scale structure corresponding to the spherical harmonic degrees greater than 8, but polarization and ampli-

tude data are more sensitive to such smaller-scale heterogeneity that cannot readily be retrieved well only by phase data. Dalton and Ekström (2006a) have recently used very large data sets of fundamental-mode Rayleigh wave amplitudes in the period range between 150 and 250 s to construct global phase speed maps. Their phase speed models derived from amplitude data showed significant correlation with those from phase data, indicating that the observed amplitude information contains meaningful information of the elastic properties in the upper mantle.

In the regional-scale mapping with the two or multiple station method, however, the majority of tomographic inversions have been based on the phase measurements, and the use of inter-station amplitude data for phase speed mapping has yet to be investigated. It is not straightforward to measure amplitude ratios of surface waves between two stations directly in the frequency domain, due to the effects of notches in amplitude spectra. As an example for stable phase and amplitude measurements at a single station, Trampert and Woodhouse (1995) introduced a waveform inversion method to extract path-specific phase speed and amplitude information using B-spline functions to represent the smoothed parameterization of phase dispersion and amplitude spectra as a function frequency. The application of such waveform fitting technique to the two-station method can be of help in stable and accurate measurements of inter-station phase speed and amplitude ratios, which can be used in the reconstruction of high-quality shear velocity models.

1.3 Outline of this thesis

In this study, we develop a fully non-linear waveform fitting approach for the surface wave phase speed and amplitude measurements between two stations on a common great-circle path, working with a global optimization method; the Neighbourhood Algorithm (NA) by Sambridge (1999). By employing the

direct matching of observed waveforms at two stations, we extract the inter-station dispersion curves for phase speeds and amplitude ratios (Hamada and Yoshizawa, 2015). The new method of two-station waveform matching is applied to the large number of station pairs of USArray in an automated manner, by employing quantitative evaluation criteria for the quality control of the estimated inter-station phase speed and amplitude.

We also derive phase speed maps for fundamental-mode Rayleigh and Love waves, and those from the joint inversions of the phase and amplitude data for the vertical-component Rayleigh waves, which exhibits the importance of amplitude information to better constrain the smaller-scale heterogeneity across the array (Hamada and Yoshizawa, 2015).

A new isotropic 3-D shear wave model in North America is then constructed using the phase speed maps of the fundamental-mode Rayleigh waves, which emphasizes the effects of inter-station amplitude data on the local-scale heterogeneity in the 3-D S wave velocity model. We also discuss station correction term that is derived from the inversions of the amplitude data, which represents local amplification factor beneath each station.

In this thesis, we start with the review of major tectonic features of North America and recent seismological studies of North American continent using USArray in chapter 2. In chapter 3, we explain the method of inter-station phase speed and amplitude measurements, and the practical procedure for inter-station waveform fitting as well as the quantitative evaluation of the estimated phase speeds and amplitude ratios. This method is applied to the actual data set of USArray and semi-automatic data selection process is demonstrated. In chapter 4, we make the detailed explanations of the inversion technique for the phase speed maps and amplification factors. The joint inversion of phase and amplitude data requires additional *a priori* parameter of relative weight on different types of data. We then show the maps of phase

speed and amplification factor in the United States. The final 3-D S wave velocity models (both isotropic and anisotropic models) are displayed and discussed in chapter 5, and we summarize this thesis in chapter 6, proposing some additional topics to be investigated in the future.

Chapter 2

Seismic structure of the North American continent

In this chapter, we focus on the upper mantle structure beneath North America, where the high-quality broadband seismic data derived from the high-density Transportable Array (USArray) have facilitated many new tomographic studies. In combination with such a dense seismic array, the new method developed in this thesis (i.e., inter-station measurements of phase and amplitude, and their application to the joint tomography using both phase and amplitude data) can be of help in the high-resolution mapping of the upper mantle beneath the North American continent. In this chapter, we introduce its tectonic features and review recent tomographic studies in this region.

2.1 Tectonic features in North America

The North American continent encompasses a variety of complex structural features, including regions with east-west extension, active volcanoes, and stable cratons. Fig. 2.1 displays a topography map with tectonic boundaries in North America. The major tectonic features of the U.S. are divided by the Rocky Mountains; a tectonically active western region including the subduction of the Juan de Fuca plate as well as the Yellowstone hotspot, and

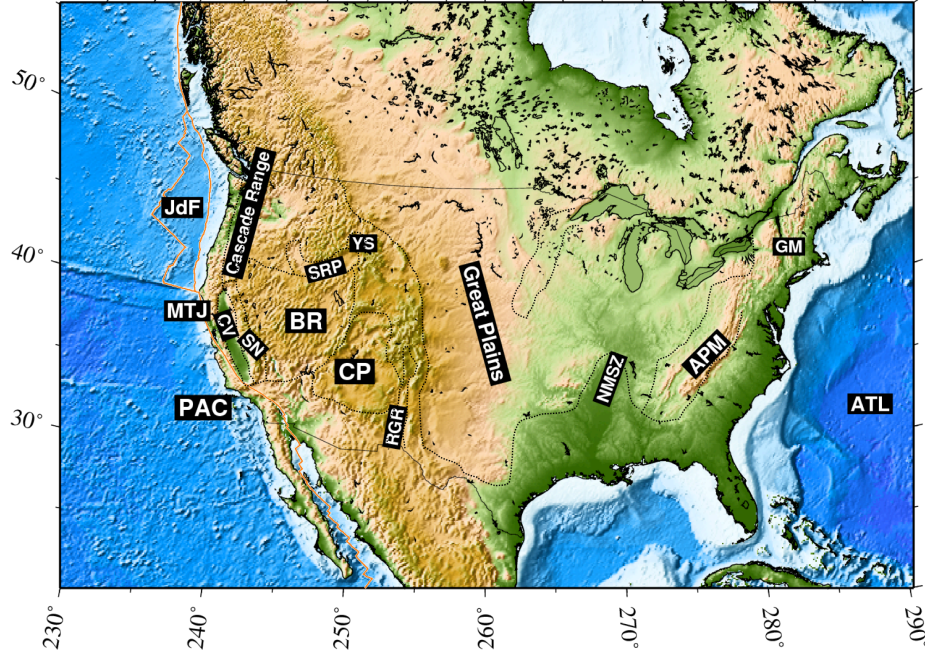


Figure 2.1: Topography map of North America. Boundaries of tectonic provinces (dotted line) are reproduced from Obrebski et al. (2011) and Sleep (2005), and tectonic boundaries (orange lines) are taken from Bird (2003). YS: Yellowstone Caldera, SRP: Snake River Plain, RGR: Rio Grande Rift, CP: Colorado Plateau, BR: Basin and Range, SN: Sierra Nevada, GV: Great Valley, JdF: Juan de Fuca Plate, MTJ: Mendocino Triple Junction, PAC: Pacific Plate, NMSZ: New Madrid Seismic Zone, APM: Appalachian Mountains, GM: Great Meteor hotspot track and ATL: Atlantic Ocean.

central and eastern regions with stable cratons (e.g., van der Lee and Nolet, 1997).

Fig. 2.3 shows a map of surface heat flow in the United States (Blackwell et al., 2011). The western U.S. is characterized by high surface heat flow and thinner crust (e.g., Mooney et al., 1998). The Yellowstone caldera is the highest heat flow area in the North American continent, and a large-scale volcanic system has formed a hot spot track in the Snake River Plain since 16 Ma (Smith and Braile, 1994). The extension and subduction has started

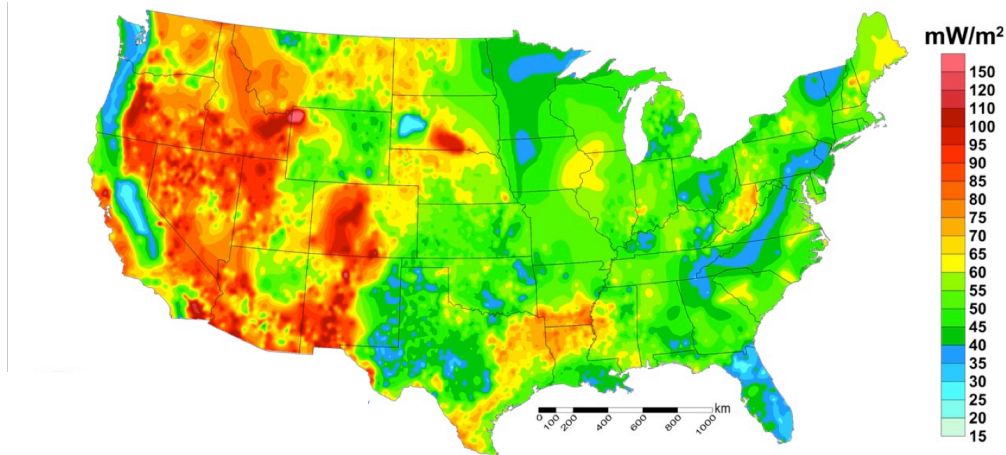


Figure 2.2: Heat flow map of North America from Blackwell et al. (2011).

in the Basin and Range since 40 Ma (e.g., Atwater, 1989), and the extension in the Rio Grande Rift has begun since 30 Ma (e.g., Morgan et al., 1986). Colorado Plateau is known as a stable cratonic area, and the temperature in the shallow lithosphere is lower (400-700 °C) than its surrounding areas (Goes and van der Lee, 2002).

To the contrary, the eastern U.S. is characterized by low surface heat flow (e.g., Blackwell et al., 2011) and thick crust (e.g., Mooney et al., 1998). The thickness of lithosphere in the eastern cratonic region reaches 200-250 km, and the temperatures at 50 km depth is estimated to be about 500 °C (e.g., Artemieva and Mooney, 2001).

2.2 Seismic imaging with USArray

The shear wave structure of North America, which encompasses complex tectonic features (Fig. 2.1), has been studied by a variety of surface wave analyses. The single-station method has been primarily used to constrain its large-scale S-wave structure (e.g., van der Lee and Nolet, 1997; Godey et al., 2003; Marone et al., 2007; Nettles and Dziewonski, 2008), employing permanent global networks and local stations in the U.S. The deployment

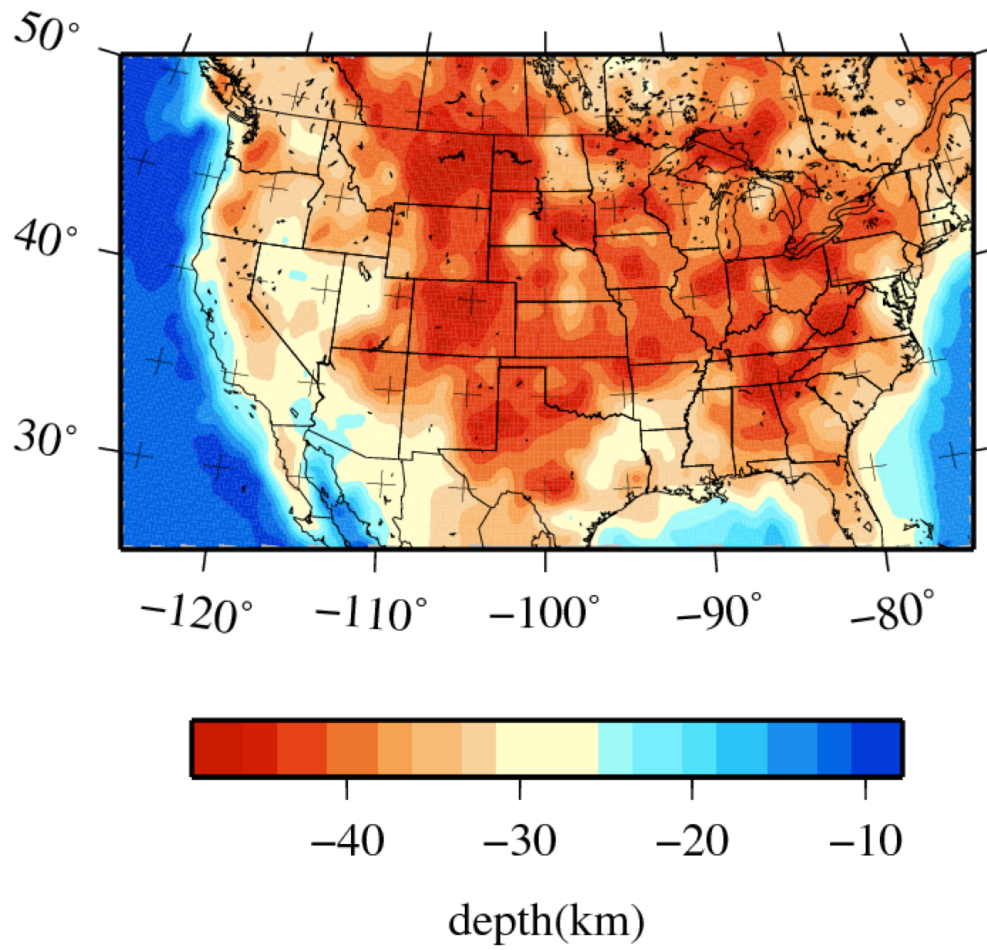


Figure 2.3: Moho depth map of the North America from CRUST1.0 (Laske et al., 2013).

of the Transportable Array (USArray) in the last decade provides us with high-quality broadband seismic waveform data, which has facilitated a variety of tomographic studies in this region. In this section, we review recent tomographic studies of surface waves, body waves and other seismological analyses with the USArray.

2.2.1 Surface wave tomography with single-station methods

There are many tomographic studies for large-scale structures in North America, using USArray as well as permanent stations in this area, based on the single-station method. Schaeffer and Lebedev (2014) applied an automated multimode waveform inversion of USArray data for the reconstruction of a 3-D S wave model, and found that the lithosphere beneath the cratonic boundary at depth closely follows the Rocky Mountain Front at the surface. Yoshizawa and Ekström (2010) employed a fully non-linear waveform inversion for measuring multimode phase speeds from a single seismogram, and applied it to North American stations, including USArray, for constructing the multi-mode phase speed maps of both Rayleigh and Love waves (Fig. 2.4).

Studies on local-scale tomography with USArray data alone have also been investigated. For example, Pollitz and Snoke (2010) employed a single taper method, which identifies frequency-dependent Rayleigh-wave dispersions, and derived a 3-D shear velocity structure in the western U.S., suggesting that low velocity mantle underlays active rift zones as well as high velocity mantle associated with the subducting Juan de Fuca plate. In the 3-D S wave models of southern central U.S. by Pollitz and Mooney (2014), the New Madrid Seismic Zone is identified as a slightly low velocity region in the upper mantle, while ancient rifts are generally associated with low crustal velocity due to the existence of thick sedimentary sequences.

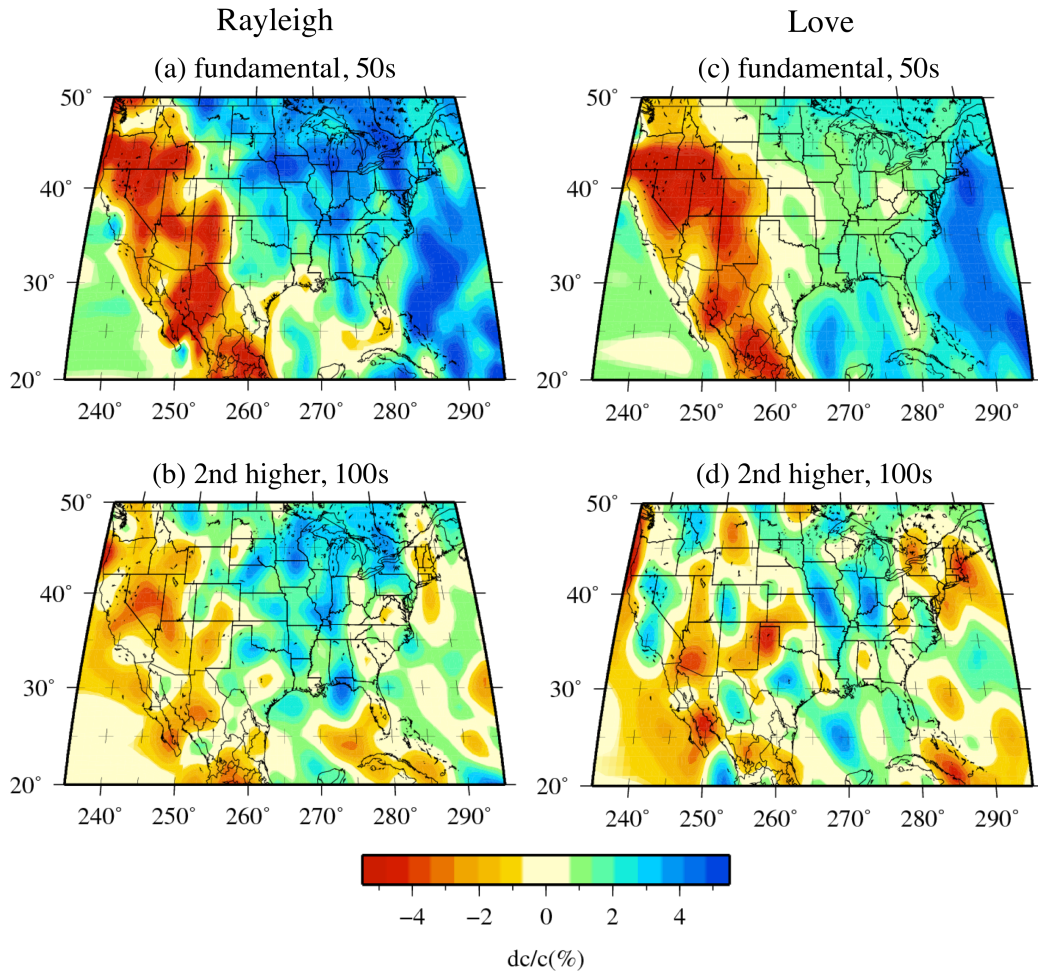


Figure 2.4: Phase speed maps (Yoshizawa and Ekström, 2010) of (a) fundamental mode Rayleigh wave at 50 s, (b) second higher mode Rayleigh wave at 100 s, (c) fundamental mode Love wave at 50 s, and (d) second higher mode Love wave at 100 s .

2.2.2 Two- or multiple-station approaches for surface waves and ambient noises

New two or multiple station techniques for constraining seismic structure have been also developed for making the efficient use of the USArray data. Foster et al. (2014a) extracted inter-station phase speeds based on the single-station measurements of the surface wave phases collected by Ekström (2011). They reconstructed phase speed maps of fundamental-mode Rayleigh and Love waves (Fig. 2.5), incorporating the arrival-angle corrections with the mini-array method.

Lin et al. (2009) developed a new method of surface wave tomography, applying the eikonal equation to the phase delay calculated from seismic ambient noise, by tracking phase front across an array. The eikonal tomography suggests that the gradient of phase speed is useful to estimate the local phase speed and the direction of wave propagation. The eikonal tomography has been improved by incorporating amplitude measurements with the Helmholtz equation (Lin and Ritzwoller, 2011). The resolution of the Helmholtz tomography, which takes account of the effects of wave interference, waveform healing and backward scattering, seems better than the conventional models.

Ambient noises have been also used in tomographic studies for the crustal structures beneath North America. Ekström (2014) obtained phase speed maps for Rayleigh and Love waves at short periods (5-40 s), with a noise-correlation technique. The results are highly consistent with the previous crustal models, with the conspicuous slow anomalies associated with basins and rifts in the western U.S.

Lawrence and Prieto (2011) constructed attenuation models in the western U.S. from ambient noise, evaluating spatial coherency with an elastic geometrical spreading term and a distance dependent decay (an attenuation

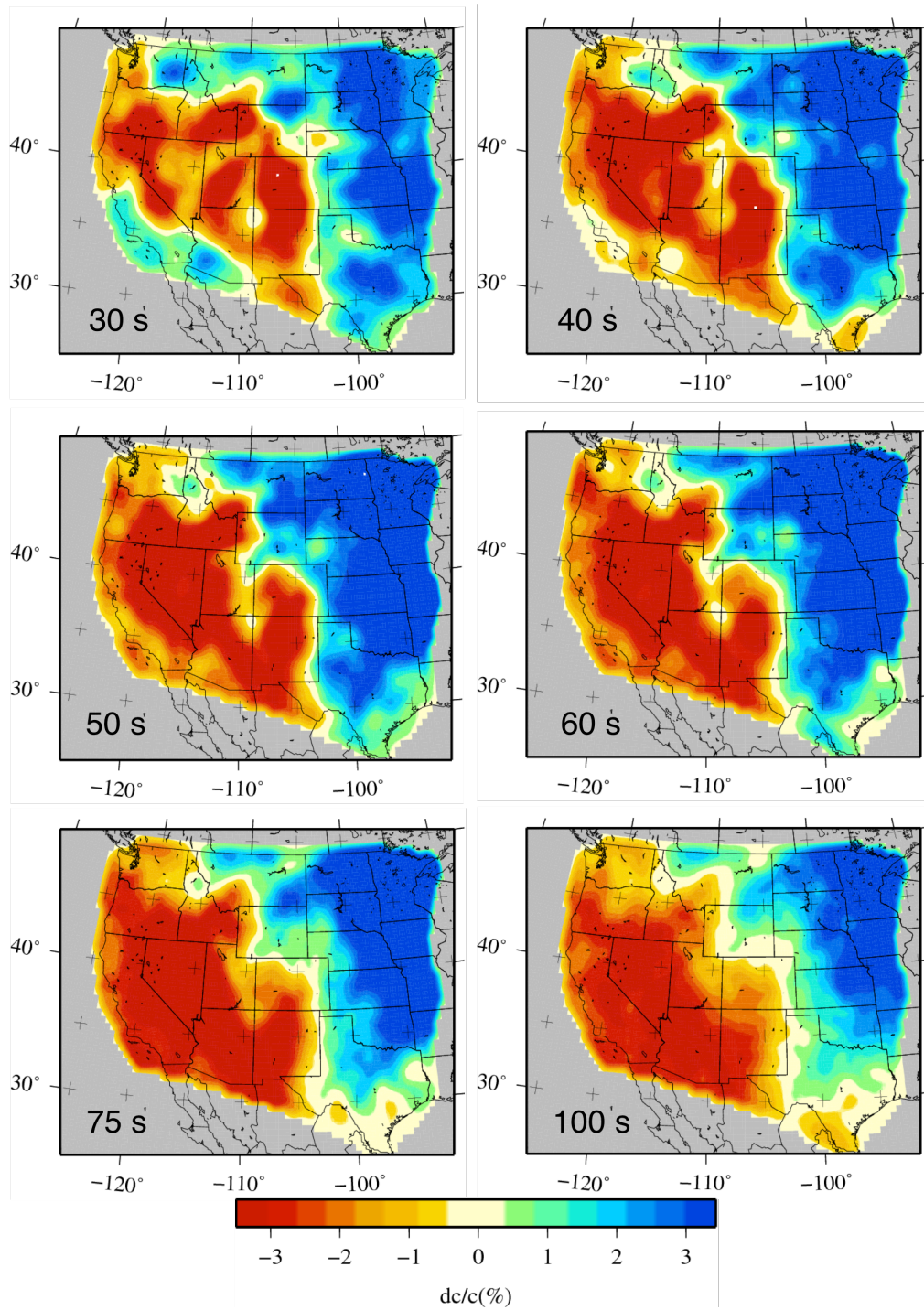


Figure 2.5: Rayleigh wave phase speed maps in the western U.S. from two-station phase measurements corrected for arrival angle at 30, 40, 50, 60, 75, and 100 s, reproduced from Foster et al. (2014a).

coefficient). The attenuation coefficient maps at periods of 8-32 s show the high attenuation anomalies in both Yellowstone and the west coast of the U.S.

2.2.3 Joint analysis with body-wave travel times

There are also many tomographic studies using body-wave travel times, which are far more sensitive to deep structures than surface waves. Burdick et al. (2010) measured P wave travel times using USArray, and derived high-resolution P wave velocity models in the western region. Joint tomography of body waves (S-wave travel times) and surface waves (phase speeds) was done by Obrebski et al. (2011) to enhance both lateral and vertical resolutions in the upper mantle. The results of their joint inversion showed an improved resolution above the depth of 200 km, owing to the incorporation of surface waves, compared with the velocity models derived from body waves only.

Chapter 3

Inter-station phase speed and amplitude measurements of surface waves with non-linear waveform fitting

3.1 Introduction

In this chapter, we describe the measurement technique of surface wave phase speeds and amplitude ratios between two stations on a common great-circle path, based on a fully nonlinear inter-station waveform fitting method with a global optimization using the Neighbourhood Algorithm (NA) (Sambridge, 1999). Although the basic concept on the prototype of this author's original method was described in Hamada (2014), we have largely modified and improved the practical approach of this inter-station waveform fitting afterward (Hamada and Yoshizawa, 2015), and thus the process of the latest technique is fully explained in this chapter.

In a recent study of two-station method using USArray (Foster et al., 2014a), inter-station phase information was derived from phase difference between two stations based on the single-station phase measurements of Ekström et al. (1997) for each station. We, instead, employ the direct comparison of waveforms at two stations to extract the inter-station phase speeds and amplitude ratios simultaneously. The basic idea of this method of non-linear waveform fitting is modeled on earlier works of Yoshizawa and Kennett

(2002b) and Yoshizawa and Ekström (2010), in which a single station seismogram is inverted with NA to extract path-specific multi-mode phase speeds for source-receiver pairs. This method of two-station waveform matching is applied to a large number of station pairs of USArray in an automated manner, by employing quantitative evaluation criteria for the quality control of the estimated inter-station phase speed and amplitude. In this chapter, we summarize the method of the inter-station phase and amplitude measurements for the fundamental-mode surface waves using the fully non-linear waveform fitting, and their application to the observed data of a dense broadband seismic array.

3.2 Method of phase speed and amplitude measurements

3.2.1 Formulation for inter-station waveform fitting

We define the seismogram of a nearer station in the frequency domain as $u^{near}(\omega)$, and that of the farther station as $u^{far}(\omega)$, which can be represented as,

$$u^{far}(\omega) = A^{far}(\omega) \exp[i\phi^{far}(\omega)], \quad (3.1)$$

$$u^{near}(\omega) = A^{near}(\omega) \exp[i\phi^{near}(\omega)], \quad (3.2)$$

where ϕ^{far} and ϕ^{near} are the phase terms, and A^{far} and A^{near} are the amplitude terms, respectively. A^{far} and A^{near} are corrected by geometrical spreading factors $\sqrt{\sin \Delta^{far}}$ and $\sqrt{\sin \Delta^{near}}$, where Δ^{far} and Δ^{near} are their epicentral distances. Let us find the best estimate of a transfer function $T(\omega)$ between u^{far} and u^{near} , which eventually gives us the inter-station phase and amplitude ratios as functions of frequency. Using the transfer function defined as $T(\omega) = D(\omega) \exp[i\delta\phi(\omega)]$, where D is the amplitude ratio and $\delta\phi$ the phase differences between the two stations, we can express the perturbed

seismogram u^{pert} as follows,

$$\begin{aligned} u^{pert}(\omega) &= T(\omega)u^{near}(\omega) \\ &= A^{near}(\omega)D(\omega)\exp[i\{\phi^{near}(\omega) + \delta\phi(\omega)\}]. \end{aligned} \quad (3.3)$$

The amplitude ratio D and phase difference $\delta\phi$ can be explicitly given as follows,

$$D = \frac{A^{far}}{A^{near}} = D_0 + \delta D, \quad (3.4)$$

$$\delta\phi = -\frac{\omega d\Delta}{c_0 + \delta c}, \quad (3.5)$$

where $d\Delta$ is the distance between two stations. $D_0 (= \exp[-\omega d\Delta Q_0^{-1}/2U_0])$ is the reference value of amplitude ratio. U_0 , c_0 and Q_0^{-1} are group speed, phase speed and inverse quality factor for a reference model. In this study, a modified version of Preliminary Reference Earth Model (PREM) of Dziewon-ski and Anderson (1981), in which the 220 km discontinuity is smoothed out, is used as the reference model.

For the stable estimation of phase dispersion curves and amplitude ratios as a function frequency by fitting two seismograms, u^{far} and u^{pert} , the fractional perturbations of path-specific phase speed δc and amplitude ratio δD are expanded in a set of B-spline functions $f_i(\omega)$,

$$\delta c = \sum_{i=1}^N b_i f_i(\omega), \quad (3.6)$$

$$\delta D = \sum_{i=1}^N d_i f_i(\omega), \quad (3.7)$$

where N is the number of parameters and B-spline functions, b_i and d_i are the coefficients for the i -th B-spline that represent the model parameters to be estimated through a nonlinear waveform fitting.

An example of a set of B-splines as a function of period is shown in Fig. 3.1. In this study, after several trials, we use 9 B-spline functions (i.e.,

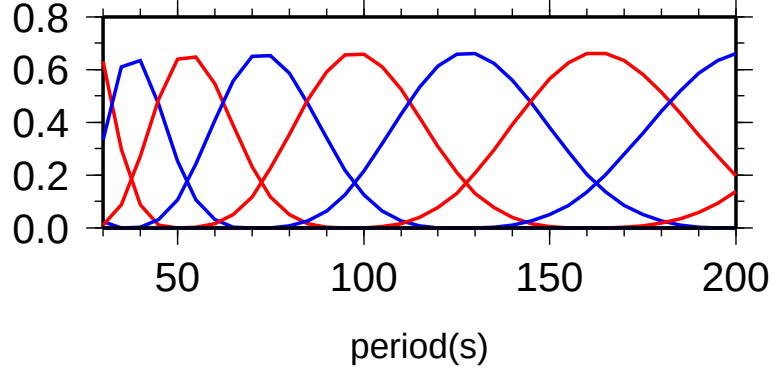


Figure 3.1: An example of a set of B-spline functions in a period range between 30 and 200 s.

$N = 9$) to represent the smoothly varying perturbation of dispersion curves in the period range from 30 to 200 seconds. In the NA process, we set the parameter search range such that ± 0.6 km/s for phase speed coefficients b_i and ± 0.8 for amplitude ratios coefficients d_i . Details of our non-linear waveform fitting will be explained in section 3.2.2.

The waveform fitting process of this study takes two steps. First, we perturb only the phase term of u^{near} to fit the seismogram at a farther station. This will enable us to determine an inter-station dispersion curve of phase speeds. The measurement reliability (see section 3.2.3) for phase speeds is evaluated by the fitted waveforms whose amplitudes are normalized. In the second step, we compute the perturbed waveforms by modulating the amplitude term to further fit u^{pert} to u^{far} , and evaluate the reliability for the estimated amplitude ratios using the total waveform fit. The reliability parameter employed in this study will be fully described in section 3.2.3. The procedure of our non-linear waveform fitting is summarized visually in Fig. 3.2.

3.2.2 Global optimization with Neighbourhood Algorithm

In this study, we employ the Neighborhood Algorithm (NA) (Sambridge, 1999) as a global optimizer, which explores a model parameter space to find suitable models with smaller misfit. The model parameter space is divided into Voronoi cells based on randomly generated models at each iteration, in which the parameter search is made to find smaller misfit models. NA is implemented in a simple way as follows,

- (a) n_s sets of initial models are generated in the model parameter space.
- (b) Compute the misfit function for the most recently generated set of n_s models and extract the n_r models with the lowest misfit from the all generated models.
- (c) Perform a uniform random walk in the Voronoi cell for each of the n_r chosen models to generate n_s new models.
- (d) Repeat (b) and (c).

In this study, 50 initial models are generated at first, and then the model parameter search is iterated 300 times with $n_s = 10$ and $n_r = 5$, so that 3050 models are explored in total for each pair of seismograms.

To evaluate the misfit between the two waveforms $u^{far}(t)$ and $u^{pert}(t) = T(t) * u^{near}(t)$ in the time domain (where the asterisk represents convolution), we use the misfit function defined by,

$$\Phi = \sum_{i=1}^{n_f} \{F_i u^{far}(t) - F_i u^{pert}(t)\}^2, \quad (3.8)$$

where F_i is i -th bandpass filter ($i = 1, \dots, n_f$), and n_f is the number of bandpass filters with different frequency ranges. We employ 6 bandpass filters to achieve the best waveform fit in a wide frequency range (Table 3.1). Prior

Table 3.1: A summary of time window for the fundamental-mode Rayleigh and Love waves used in this study. $\hat{U}$ is the reference group speed for Rayleigh/Love waves evaluated at the middle of each frequency band, calculated from the global group speed model (Larson and Ekström, 2001) for each path.

Frequency range(mHz)	Min. (km/s)	Max. (km/s)
5-10	$\hat{U} - 0.42$	$\hat{U} + 0.52$
7-12	$\hat{U} - 0.40$	$\hat{U} + 0.50$
10-18	$\hat{U} - 0.36$	$\hat{U} + 0.46$
15-25	$\hat{U} - 0.32$	$\hat{U} + 0.42$
20-30	$\hat{U} - 0.27$	$\hat{U} + 0.37$
25-40	$\hat{U} - 0.24$	$\hat{U} + 0.34$

to the band-pass filtering, time windows are automatically determined by taking account of the path-average group speeds for each path and frequency, which are estimated from global group speed maps by Larson and Ekström (2001). As a result, the fundamental-mode Love and Rayleigh waves are properly included in the chosen window. The time window and frequency ranges for bandpass filters used in this study are summarized in Table 3.1.

One of the critical issues in the inter-station phase measurement is $2n\pi$ ambiguity in the phase term (where n is an arbitrary integer), which may results in a phase cycle skip. Our smoothed parameterization of dispersion curves using the B-spline functions as well as the waveform matching with multiple band-pass filters including two long-period filters over 100 s suppresses the occurrence of a phase skip during the waveform fitting. In addition, a parameter search range for phase speed perturbations is limited within ± 0.6 km/s from the reference dispersion curve for PREM in our global optimization method, much less than the expected velocity perturbations caused by a $\pm 2\pi$ phase shift in the long period. Thus, our measurements are unlikely to be severely affected by the phase cycle skips, except for the period range

shorter than 50 s with inter-station distances longer than 2000 km, though the number of such data are very limited.

An example of Rayleigh waves before and after the nonlinear fitting process is displayed in Fig. 3.2. We can see that the observed waveforms at the far station can be matched fairly well in a wide frequency range. Fig. 3.2 also shows an example of a set of 3050 dispersion curves for phase speed and amplitude ratio as a function of period, with the nonlinear NA waveform fitting. Yellow lines in the background indicate the dispersion curves with smaller misfit, while the best-fit dispersion curve is shown by the green dots with error bars that are estimated from the ensemble of all the dispersion curves. Note that such rough estimates of measurement errors depend on the convergence rate in the global optimization process, and it does not necessarily reflect meaningful errors in the dispersion measurements. We, therefore, consider a reliability parameter that quantifies how well the perturbed waveform is fitted with the reference waveform in the next section.

3.2.3 Reliability of dispersion measurements

In this study, we use a reliability parameter for waveform fitting (e.g., van Heijst and Woodhouse, 1997; Yoshizawa and Kennett, 2002b; Yoshizawa and Ekström, 2010) to estimate the frequency-dependent accuracy of dispersion measurements. We define a residual seismogram $u^{res}(t) \equiv u^{far}(t) - u^{pert}(t)$, and calculate the corresponding spectrograms $S^{res}(\omega, t)$ and $S^{far}(\omega, t)$ in the frequency-time domain. For the quantitative evaluation of the waveform fit between the synthetic waveform u^{pert} and the observed waveform u^{far} , we introduce a waveform fit parameter $f(\omega, t)$ as follows,

$$f(\omega, t) = \exp \left[-\frac{S^{res}(\omega, t)}{S^{far}(\omega, t)} \right]. \quad (3.9)$$

Integrating $f(\omega, t)$ by time at a fixed frequency with a normalization by its time window T_w , the reliability parameter $r(\omega)$ can be estimated as a function

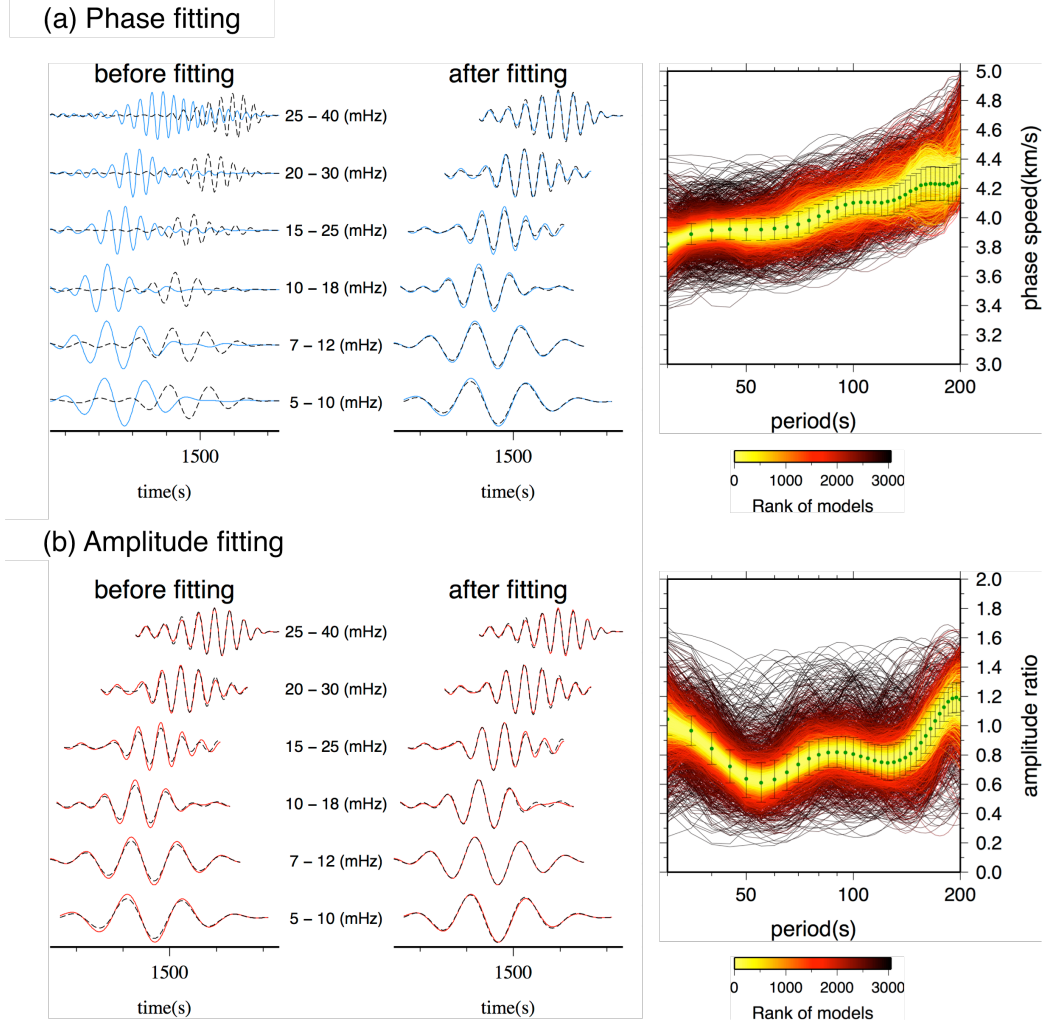


Figure 3.2: Examples of waveform fitting for the fundamental-mode Rayleigh waves, showing six time windows with different frequency ranges, for a station pair L20A (u^{near}) and P13A (u^{far}) from a seismic event in Samoa at 10 km depth on December 13, 2007. (a) Before and after the waveform fitting by perturbing the phase speed term only (no amplitude fitting), and the resultant phase speed dispersion curves (on the right), derived from the phase-term fitting through the NA. All the 3050 models exploited by the NA are ranked in order of the smaller misfit, and are plotted with colors varying from dark brown (larger misfit) to yellow (smaller misfit). The green dots represents the best-fit dispersion measurements with error bars calculate from the ensemble of all dispersion curves. (b) Before and after the waveform fitting by perturbing the amplitude term, and estimated amplitude ratios as a function of period.

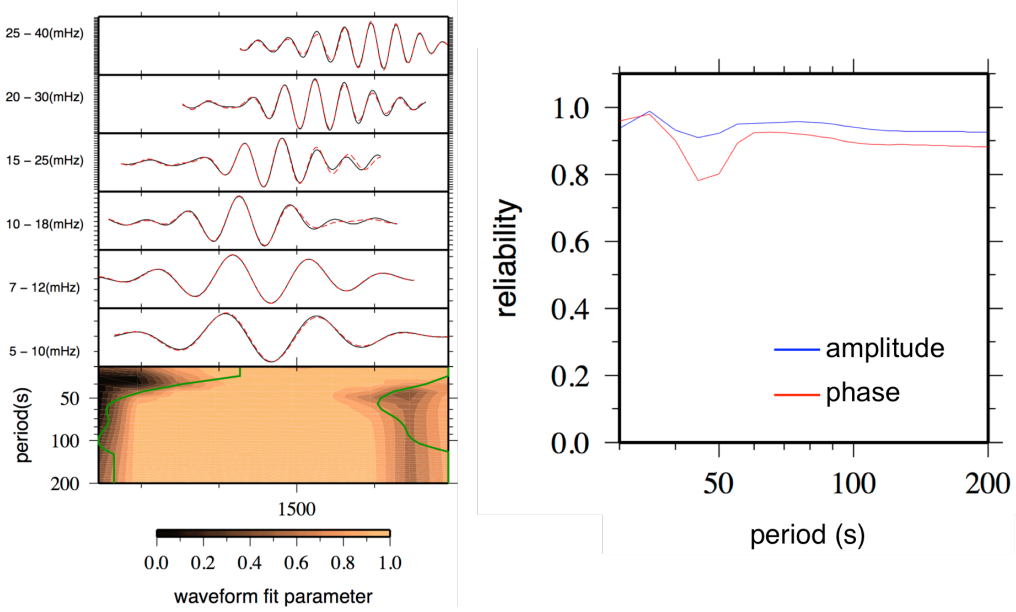


Figure 3.3: (Left): (Top) A perturbed waveform (red dashed line) and an observed waveform for the farther station from the source (black line) used in Fig. 3.2, (bottom) the waveform fit parameter $f(\omega, t)$ in the frequency-time domain. Green lines represent the start and end time of the time window used for the waveform fitting. (Right): Reliability parameters $r(\omega)$ as a function of period, calculated from the matched waveforms with phase fitting (red line) and with amplitude fitting (blue line).

of ω ,

$$r(\omega) = \frac{1}{T_w} \int_0^{T_w} f(\omega, t) dt. \quad (3.10)$$

Fig. 3.3 displays examples of the parameters $f(\omega, t)$ and $r(\omega)$ with fitted waveforms at several frequency ranges. The reliability parameter $r(\omega)$ varies from 0 (unreliable, mismatched waveforms) to 1 (reliable, perfect waveform fit).

Although the reliability parameter is basically an independent parameter from a measurement error, it can also be affected by the similar error sources in the measurements, such as the effects of off-great-circle propagation and the interferences of non-plane waves and/or higher modes, which may dete-

riorate the quality of waveform fitting under the assumption of our analysis; i.e., a single plane-wave propagation along the great-circle path. We use the reliability parameter as an automatic data selection criterion and a weighting factor in the subsequent process of phase speed mapping in chapter 4.

3.3 Application to USArray

3.3.1 Events and stations

The inter-station waveform fitting technique described in the previous section is applied to the fundamental-mode Rayleigh and Love waves observed at the Transportable Array (TA) of USArray and the Southern California Seismic Network (CI) stations. We use seismic events occurred from January 2007 to December 2014 with moment magnitudes 6.0 or greater. In this study, we eliminate the events occurred at the depth below 100 km, in which the influence of interference of higher modes upon fundamental mode seismograms tends to be larger. Fig. 3.4 shows the distributions of the events and stations used in this study. The USArray stations has migrated gradually from west to east, and it has covered the whole United States by the end of 2014.

3.3.2 Waveform processing

Prior to the waveform fitting, the instrument response of each seismogram is deconvolved, and then all the seismograms are convolved with the response function of STS-1 seismometer. Surface waves in the horizontal components (particularly, Love waves in this study) are likely to be affected by the misorientation of seismometers (e.g., Laske and Masters, 1996; Yoshizawa et al., 1999). We therefore use the database of polarization anomalies estimated by Göran Ekström, which is provided through the web site of Lamont-Doherty Earth Observatory of Columbia University (<http://www.ldeo.columbia.edu/~ekstrom/Projects/USARRAY/QC.2010/index.html>), to correct the plausi-

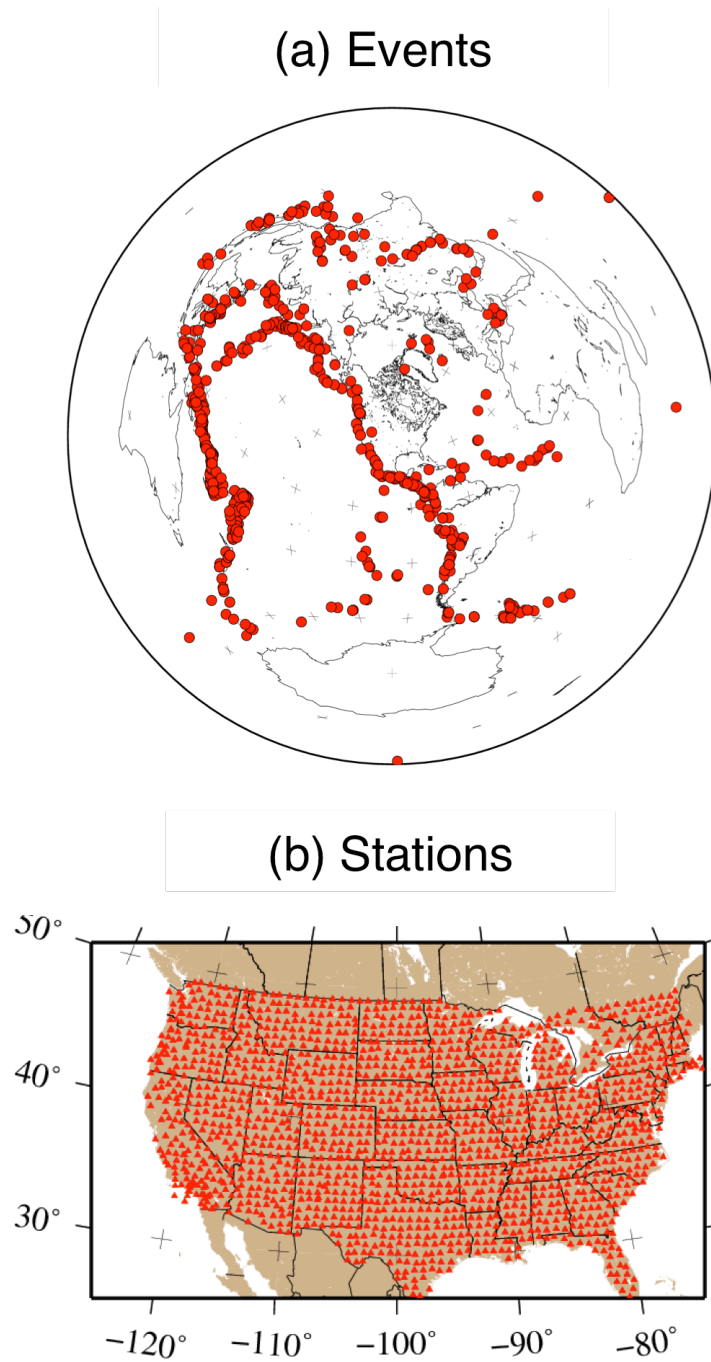


Figure 3.4: Distributions of (a) seismic events and (b) seismic stations used in this study.

ble mis-orientation of horizontal components at each station, when we rotate the horizontal components to the great-circle directions. It should be noted that, as of November 2016, the database for the mis-orientation of USArray stations are available only for the observation period from 2006 to 2010, which are utilized in this study. The horizontal components of USArray stations in other periods are simply rotated to the great-circle directions.

We also evaluate the radiation pattern of each event referring the Global CMT solutions (Ekström et al., 2005) to minimize large uncertainties in measurements in the nodal direction of surface wave radiation. In this study, we eliminate the stations located close to the nodal direction, in which the radiated amplitude is less than 50 % of the maximum surface-wave radiation.

In the inter-station waveform fitting method, two stations must be located nearly on a common great-circle path. The geometry of this system is shown in Fig. 3.5, where α is the difference in azimuths from a source to the two stations, and β is the difference in back-azimuths. In this study, we select station pairs, which satisfy that α is less than 0.5° and β is less than 2.5° . The waveform fitting is then applied, and the phase speeds as well as amplitude ratios between the two stations are estimated. The reliability parameters defined in section 3.2.3 are used as a guide to determine the frequency range with satisfactory accuracy for each dispersion curve.

With several empirical experiments, we use the threshold (or minimum) values of the reliability parameter r_c for the automatic selection of high-quality phase speed measurements as a function of period T , which is given as, $r_c(T) = r_0 \times \bar{r}(T)/\bar{r}(T_0)$, where T_0 is the reference period, and $\bar{r}$ is the average reliability value for each period. In this study, we used a reference period of $T_0 = 80$ s, and $\bar{r}(T_0) = 0.7$ for both Rayleigh and Love waves. The threshold values of the reliability parameter for the amplitude ratios is fixed to 0.9 at all the period ranges.

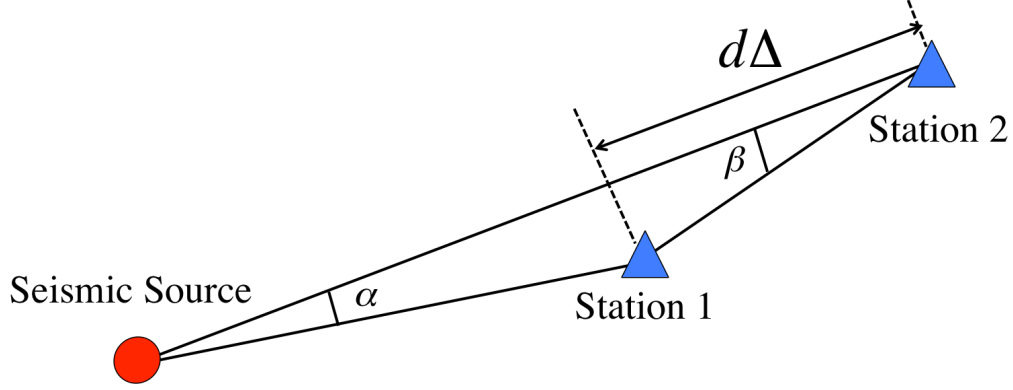


Figure 3.5: A geometrical configuration of two stations and a seismic source.

3.3.3 Data sets for phase speed and amplitude ratio

Tables 3.2 and 3.3 display the numbers of measurements, station pairs, standard deviations, and the threshold of reliability parameters as a function of period for Rayleigh and Love waves, respectively. We gather more than 90,000 station pairs for the Rayleigh waves and over 70,000 for Love waves in the majority of our target period range for the phase measurements. The number of amplitude data for the vertical component Rayleigh waves are collected for about 60,000 station pairs. The phase and amplitude measurements for Love waves are more affected by noise than Rayleigh waves, mainly due to lower signal-to-noise ratios in the horizontal components and higher-mode interference with the fundamental mode of Love waves. In addition, the number of amplitude measurements for Love waves are very limited, and we use the amplitude measurements only for Rayleigh waves in the phase speed mapping of chapter 4.

Table 3.2: The numbers of measurements, station pairs, standard deviations σ_c and σ_A for the measured phase speed and amplitude ratios, and the threshold of reliability parameters r_c and r_A for phase speed and amplitude ratios, as a function of period of Rayleigh waves.

Period (s)	Rayleigh wave							
	Measurements		Station pairs		Standard deviation		Threshold	
	Phase	Amplitude	Phase	Amplitude	σ_c	σ_A	r_c	r_A
30	114,747	52,554	79,002	42,482	0.323	0.461	0.647	0.900
40	132,684	75,715	90,382	59,131	0.290	0.476	0.661	0.900
50	140,220	72,420	94,180	56,462	0.254	0.488	0.647	0.900
60	140,877	79,167	94,631	61,380	0.222	0.495	0.668	0.900
70	135,245	81,049	91,671	62,803	0.250	0.533	0.686	0.900
80	129,356	80,529	88,369	62,490	0.203	0.523	0.700	0.900
100	120,129	79,540	83,032	61,823	0.247	0.572	0.715	0.900
130	103,453	81,021	73,091	66,527	0.280	0.725	0.735	0.900

Table 3.3: Same as Table 3.2, but for the phase speed data of Love waves.

Love wave				
Period (s)	Measurements	Station pairs	σ_c	Threshold: r_c
30	78,482	63,513	0.132	0.734
40	86,852	70,609	0.129	0.793
50	85,879	70,488	0.150	0.751
60	89,352	73,399	0.166	0.730
70	90,733	75,163	0.203	0.711
80	90,143	75,786	0.218	0.700
100	81,968	72,150	0.257	0.693
130	62,508	61,023	0.308	0.712

Fig. 3.6 displays the path density maps and the histogram of the numbers of measurements as a function of inter-station distance for Rayleigh-wave phase and amplitude and Love-wave phase at 80 s. This figure indicates that we collect a large-number of inter-station data for short distances less than 1000 km. Such short-path information cannot readily be extracted from the conventional single-station analysis for surface waves, and can be of great help in enhancing the horizontal resolution of phase speed models in North America.

Fig. 3.7 displays two examples of dispersion curves of both phase speed and amplitude ratios of Rayleigh waves for pairs of D13A-G05A stations and FUR-MPP stations. The phase speed measurements for a longer path (Fig. 3.7b, inter-station distance of 655 km) exhibit fairly consistent values in a wider period range up to 150 s with increasing deviations in a long period, while those for a shorter path (Fig. 3.7c inter-station distance of 319 km) are more scattered in the period longer than 100 s, for which the wavelength of Rayleigh waves becomes much longer than the inter-station distance. The amplitude ratios between two stations are more scattered than phase speeds, which results in lower variance reductions for the amplitude data in the joint inversions discussed later in section 4.2. We assume a plane wave propagation

for the waveform fitting process, but this assumption may be violated if the effects of non-plane waves or off-great-circle signals are non-negligible (e.g., Pedersen, 2006). Some outlying measurements, which have passed the quality control requirements, may contaminate our large data set extracted from the automated measurements. However, the majority of the measurements well reflect the average local features for each path, and the number of outliers is rather limited compared to the total number of our data set. The measured phase speeds in a long periods tend to be unstable, particularly in the case that the wavelength of surface waves becomes much longer than the inter-station distances, resulting in large measurement errors. Since the majority of our measured data sets comprises inter-station distances shorter than 600 km, we avoid to use the phase speeds in the period longer than 130 s in the subsequent discussion of this study.

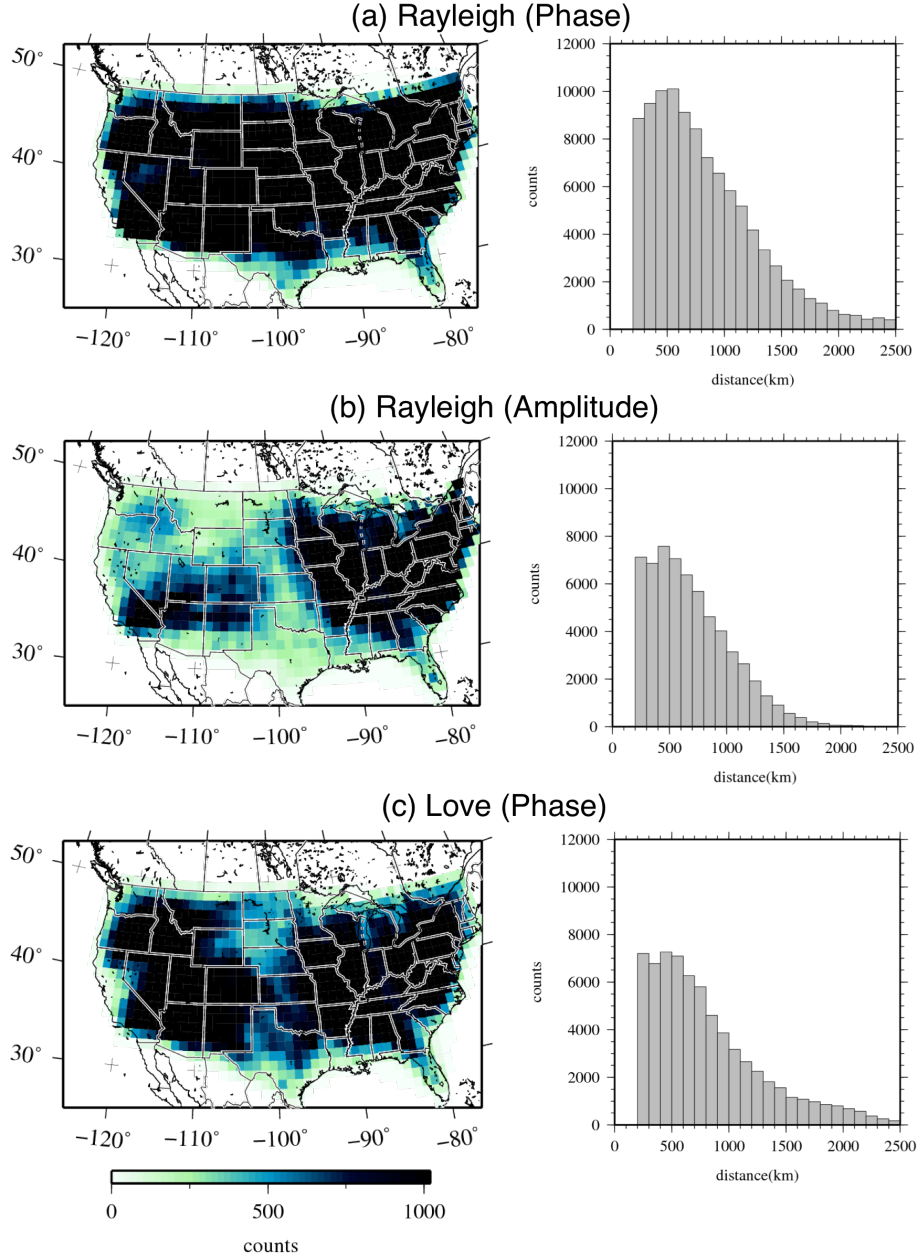


Figure 3.6: (Left) The numbers of ray paths in $1^\circ \times 1^\circ$ cells for (a) phase speed measurements and (b) amplitude ratio measurements of Rayleigh waves at 80 s, and (c) the phase speed measurements of Love waves at 80s. (Right) Numbers of measurements for each data set as a function of inter-station distances.

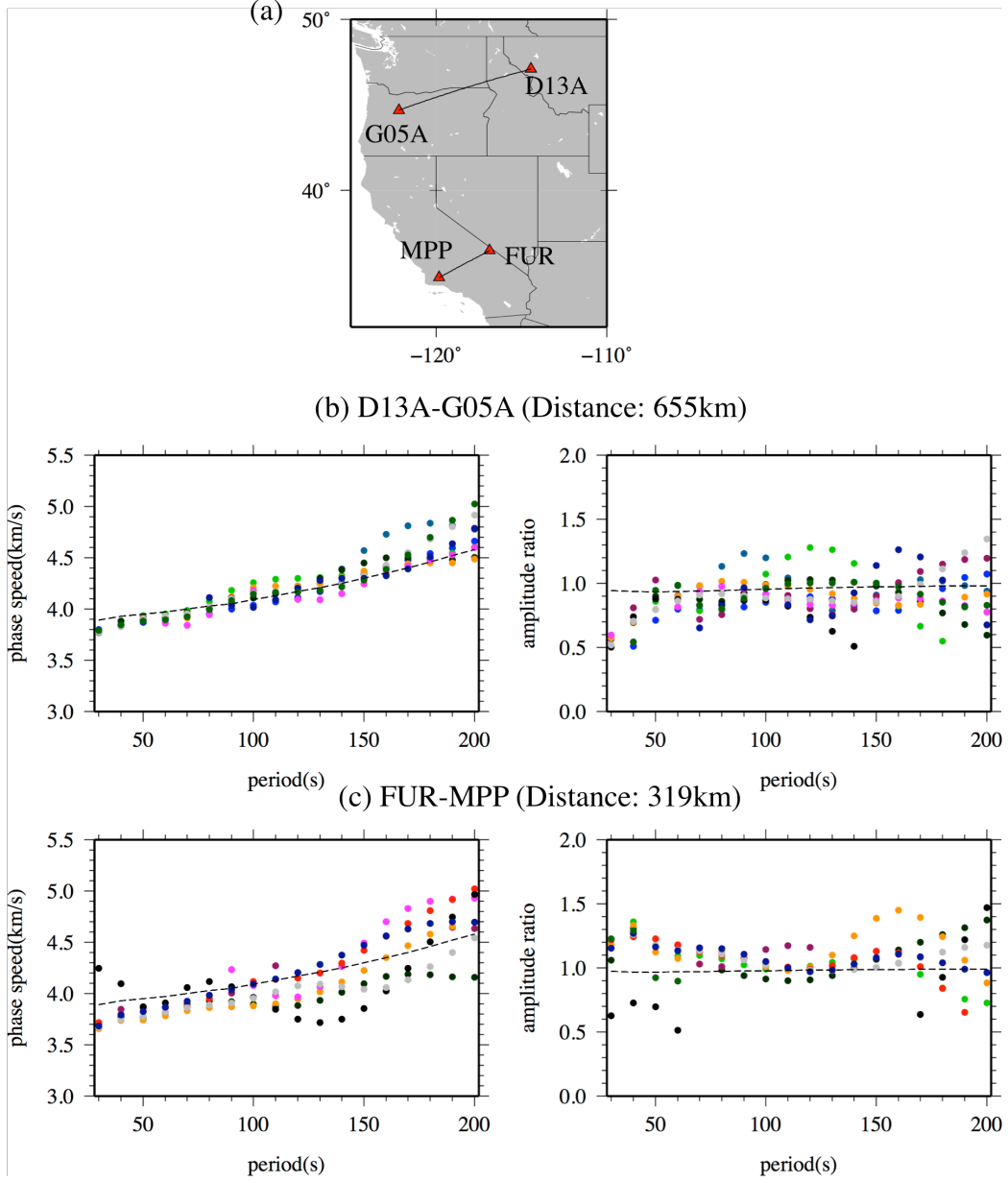


Figure 3.7: Examples of observed dispersion curves for Rayleigh-wave phase speeds and amplitude ratios. (a) Inter-station paths between D13A and G05A, and between FUR and MPP stations. (b), (c) Observed dispersion curves for the two inter-station paths. Different colors of plots represent the measurements for different seismic sources (12 events for D13A-G05A, and 13 events for FUR-MPP), and dashed lines indicate the reference values calculated for PREM with a smoothed 220 km discontinuity.

Chapter 4

Mapping phase speed models

4.1 Introduction

The measured inter-station phase speed and amplitude ratios, described in chapter 3, are inverted simultaneously for phase speed distributions and receiver amplification terms. The method of tomographic inversion scheme used in this study is based on the surface-wave tomography technique developed by Yoshizawa and Kennett (2004) for phase speed mapping. We have modified this method so that two different types of data (phase and amplitude) can be inverted simultaneously for a phase speed map and station amplification terms. In this chapter, we briefly summarize the method to map phase speeds with the observed phase and amplitude data, and display the maps of phase speeds and local amplification factors in the U.S.

4.2 Formulation of inversion for phase speed maps

4.2.1 Phase speed mapping using phase data

The observed path-average phase speeds between two stations can be expressed as follows,

$$\frac{\delta c^{obs}}{c_0} = \frac{1}{d\Delta} \int_{g.c.} \frac{\delta c(s)}{c_0} ds, \quad (4.1)$$

where $\delta c^{obs} \equiv c^{obs} - c_0$, c^{obs} is the measured average phase speed along an inter-station path, and c_0 the reference phase speed. In this study, we use the average phase speed of all the measurements for c_0 . $d\Delta$ is the inter-station distance and s represents the distance along each ray path. On the assumption that surface wave propagates along the great-circle path, phase speed maps for each period can be constructed by using the equation (4.1).

The equation (4.1) can be expressed in a generalized form, $\mathbf{d} = \mathbf{G}\mathbf{m}$, where $\mathbf{d}$ is the data vector that comprises the observed phase speed perturbation, $d_i = (\delta c^{obs}/c_0)_i (i = 1, 2, \dots, M)$, and M is the total number of observed data. $\mathbf{G}$ is the kernel matrix and $\mathbf{m}$ is the model-parameter vector $m_j (j = 1, 2, \dots, N)$, where N is the total number of model parameters.

A local phase speed perturbation on the earth at latitude θ and longitude ϕ can be expanded in spherical B-spline functions $B_j(\theta, \phi)$,

$$\frac{\delta c(\theta, \phi)}{c_0} = \sum_{j=1}^N m_j B_j(\theta, \phi), \quad (4.2)$$

where the model parameter m_j is the coefficient for the j -th basis function. In this study, we use the constant grid interval of $1.0^\circ \times 1.0^\circ$ when we invert for phase speed maps. The elements of the kernel matrix $\mathbf{G}$ for i -th measurement and j -th model parameter can be written as follows by using the spherical B-spline functions,

$$G_{ij} = \frac{1}{d\Delta} \int_0^{d\Delta} B_j ds. \quad (4.3)$$

The inverse problem can be solved by minimizing the following objective function,

$$\Phi(\mathbf{m}) = |\mathbf{d}' - \mathbf{G}'\mathbf{m}|^2 + \lambda^2 |\mathbf{m}|^2, \quad (4.4)$$

where the components of $\mathbf{d}'$ and $\mathbf{G}'$ are $d'_i = (w_i/\sigma_c)d_i$ and $G'_{ij} = (w_i/\sigma_c)G_{ij}$, respectively, σ_c is the standard deviation of phase speed data for each period, and w_i is the weighting factor for i -th measurement for which we use

the reliability parameter in each measurement (i.e., $w_i = r(\omega)$). λ is an *a priori* damping parameter that controls the trade-off between the data misfit $|\mathbf{d}' - \mathbf{G}'\mathbf{m}|^2$ and the model norm $|\mathbf{m}|^2$. The linear inverse problem is solved iteratively using the LSQR algorithm (Paige and Saunders, 1982).

4.2.2 Phase speed mapping using inter-station amplitude ratios

Observed amplitudes of surface waves include a variety of effects. The amplitude A at a single station, after the corrections for the geometrical spreading and the instrument response, can be represented as

$$A(\omega) = A_S(\omega)A_R(\omega)A_F(\omega)A_Q(\omega), \quad (4.5)$$

where A_S represents the source term including radiation pattern and source excitation, A_R the receiver term including site amplification effects, A_F the elastic focusing/defocusing term, and A_Q the anelastic attenuation term during propagation.

Now, we consider the amplitudes for both the farther and nearer stations to the source as follows, omitting the frequency dependency,

$$A^{far} = A_S^{far} A_R^{far} A_F^{far} A_Q^{far}, \quad (4.6)$$

$$A^{near} = A_S^{near} A_R^{near} A_F^{near} A_Q^{near}. \quad (4.7)$$

Considering that the source terms can be canceled out in our two-station measurements along the great-circle paths (i.e., $A_S^{far} = A_S^{near}$), dividing (4.6) by (4.7) leads to the following relation,

$$\frac{A^{far}}{A^{near}} = \frac{A_R^{far}}{A_R^{near}} \frac{A_F^{far}}{A_F^{near}} \frac{A_Q^{far}}{A_Q^{near}}. \quad (4.8)$$

Taking the logarithm of both terms of (4.8) yields,

$$\ln \frac{A^{far}}{A^{near}} = \ln A_R^{far} - \ln A_R^{near} + \ln \frac{A_F^{far}}{A_F^{near}} + \ln \frac{A_Q^{far}}{A_Q^{near}}. \quad (4.9)$$

The third term on the right-hand side represents the amplitude anomaly between the two stations caused by the effects of elastic focusing/defocusing due to the lateral heterogeneity of velocity structure, as given by (1.3) for a single-station method. The corresponding amplitude anomaly for the case of inter-station amplitude ratios can be derived from the difference between the two stations as follows:

$$\begin{aligned} \ln \left(\frac{A_F^{far}}{A_F^{near}} \right) &= \frac{\delta c_f}{2c_0} - \frac{\delta c_n}{2c_0} \\ &+ \frac{1}{2 \sin(\Delta_n + d\Delta)} \int_0^{d\Delta} [\sin(d\Delta - \phi') \sin(\Delta_n + \phi') \partial_\theta^2 \\ &- \cos(d\Delta - \Delta_n - 2\phi')] \frac{\delta c}{c_0}(\omega) d\phi', \end{aligned} \quad (4.10)$$

where Δ_n is the epicentral distance of the nearer station, $d\Delta$ is the inter-station distance and $\phi' = \phi - \Delta_n$. We approximate that $\Delta_n \sim \Delta_n + d\Delta$ on the assumption that $\Delta_n \gg d\Delta$. For the details of derivation of (4.10), see Appendix A.

The final term in (4.9) represents the anelastic attenuation between the two stations. The majority of our inter-station amplitude ratios are measured for short distances around 500 km, which encompasses only a couple of phase cycles of intermediate/long period surface waves. The effects of lateral variations of anelastic attenuation on the amplitude anomaly for such short paths are comparable or much less than the measurement errors of amplitude ratios. Thus, in this analysis, we simply evaluate this attenuation factor using the reference values (Q_r) from the global attenuation models of Dalton and Ekström (2006b) as follows,

$$\ln \frac{A_Q^{far}}{A_Q^{near}} = -\frac{\omega d\Delta}{2U_r Q_r}, \quad (4.11)$$

where U_r is the reference group velocity derived from the global group velocity models of surface waves (Larson and Ekström, 2001).

The equation (4.9) can then be written as,

$$A'(\omega) = \ln A_R^{far}(\omega) - \ln A_R^{near}(\omega) + A'_F(\omega), \quad (4.12)$$

where $A' = \ln(A^{far}/A^{near}) - \ln(A_Q^{far}/A_Q^{near})$, which represents the observed amplitude ratio corrected with a reference attenuation (Note that the geometrical spreading and instrument responses are corrected beforehand). A' is used as input amplitude data for the joint inversion with phase data. $A'_F = \ln(A_F^{far}/A_F^{near})$ represents the amplitude anomaly caused by elastic focusing/defocusing due to the lateral variations of phase speeds corresponding to (4.10).

Using (4.1) and (4.12), the measured inter-station phase speed and amplitude ratios are inverted simultaneously for the receiver amplification term $\ln A_R$ (or the station correction term) for each station as well as phase speed distribution $\delta c/c_0$, expanded in the spherical B-spline functions, as given in (4.2).

We use the weighting scheme for equalizing different data sets of phase and amplitude, in the similar way to Julia et al. (2000),

$$E = \frac{p}{\sigma_c^2 N_c} \sum_{i=1}^{N_c} w_i^2 \left(s_i - \sum_{j=1}^M G_{ij} m_j \right)^2 + \frac{1-p}{\sigma_A^2 N_A} \sum_{i=1}^{N_A} w_i^2 \left(a_i - \sum_{j=1}^M F_{ij} m_j \right)^2, \quad (4.13)$$

where s_i and a_i are the observed phase speed perturbations and logarithm of amplitude ratios, N_c and N_A are the numbers of phase and amplitude data, and p is the factor to determine the relative weight between the phase and amplitude data that varies from 0 to 1. F_{ij} represents the equation (4.10) with spherical B-spline functions. σ_c and σ_A are the standard deviations of phase and amplitude ratios for each period respectively (summarized in Table 3.2 and 3.3), and w_i is the weight factor (or the reliability parameter explained in section 3.2.3) for each measurement.

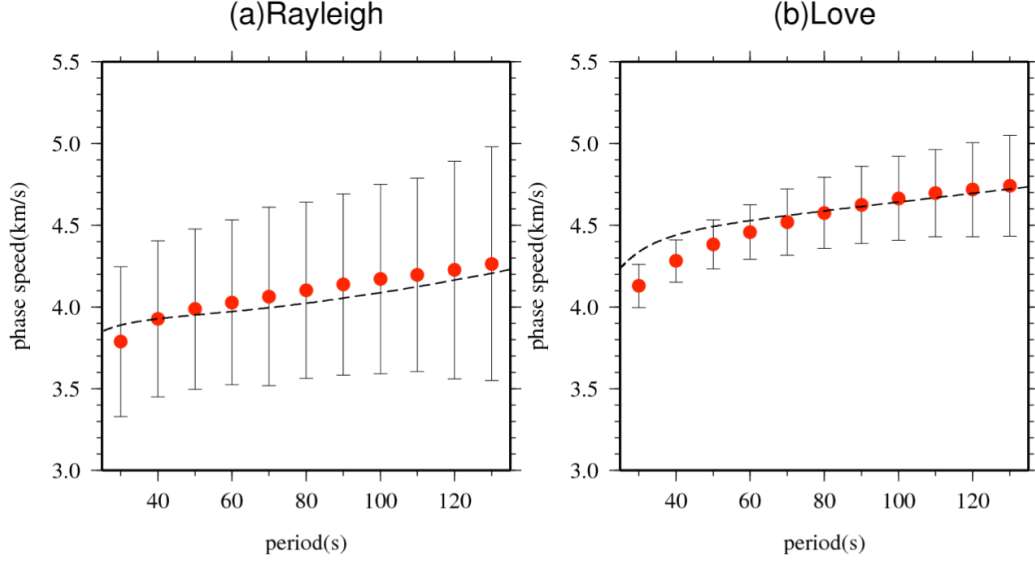


Figure 4.1: Average phase speeds (red dots) and standard errors for (a) Rayleigh waves and (b) Love waves calculated for our final data set used for phase speed mapping. Dashed lines indicate the dispersion curves for PREM.

4.3 Phase speed maps and local amplifications

4.3.1 Phase speed maps of North America

The average phase speeds estimated from our total data set are displayed in Fig. 4.1, which is slightly slower than the reference phase speed of PREM in the period range less than 40 s for Rayleigh waves and 80 s for Love waves. The standard deviations of Rayleigh waves are larger than those of Love waves, suggesting larger lateral variations of phase speeds for Rayleigh waves between the western and eastern areas than those for Love waves.

We perform checkerboard resolution tests with several cell sizes for the evaluation of the achievable horizontal resolution with our current data set. Some examples are displayed in Fig. 4.2 for Rayleigh-wave phase speed maps at 80 s using phase data only, with cell sizes of 2° , 3° and 4° . The input checkerboard patterns contain the maximum perturbations of $\pm 4.8\%$ from the reference phase speed. In the map with the cell size of 2.0° , the strength of

velocity perturbation in the eastern area of North America is better recovered than that in the western area. This is mainly because a number of crossing paths have become available in the eastern area, owing to many seismic events in the western coastal areas of North and Central America with ideal conditions for inter-station analysis.

Figs. 4.4 and 4.5 display phase speed maps for the fundamental-mode Rayleigh and Love waves with phase data only in the period range between 30 and 130 seconds. These maps are plotted as perturbations from the average phase speed at each period shown in Fig. 4.1. The variance reductions achieved by the present inversions, with respect to the average phase speeds of our measurements (Fig. 4.1), are about 13-75% for Rayleigh waves and 12-50% for Love waves in the period range between 30 and 130 s (Fig. 4.3). The variance reductions of our model are smaller than those of other phase speed models with single-station analyses (e.g., Yoshizawa and Ekström, 2010) or the two-station analysis based on single-station measurements by Foster et al. (2014a) in the same region. This implies the interference of non-plane waves, higher modes and off-great-circle propagation may not be negligible in our inter-station measurements, resulting in large scatters in the dispersion measurements for the similar paths (e.g., Fig. 3.7). Nevertheless, the resultant models are consistent with the earlier North American models on a large scale.

Fig. 4.6 shows the results of checkerboard resolution tests for the phase speed maps derived from joint inversions of phase and amplitude data of vertical-component Rayleigh waves. The patterns of checkerboards are smeared to some extent as we weight more on the amplitude data, because the number of paths for amplitude is limited compared to phase data (see Table 3.2). However, the retrieved strength of velocity perturbation in the checkerboard models is clearly improved as the weight on amplitude becomes large (or, as

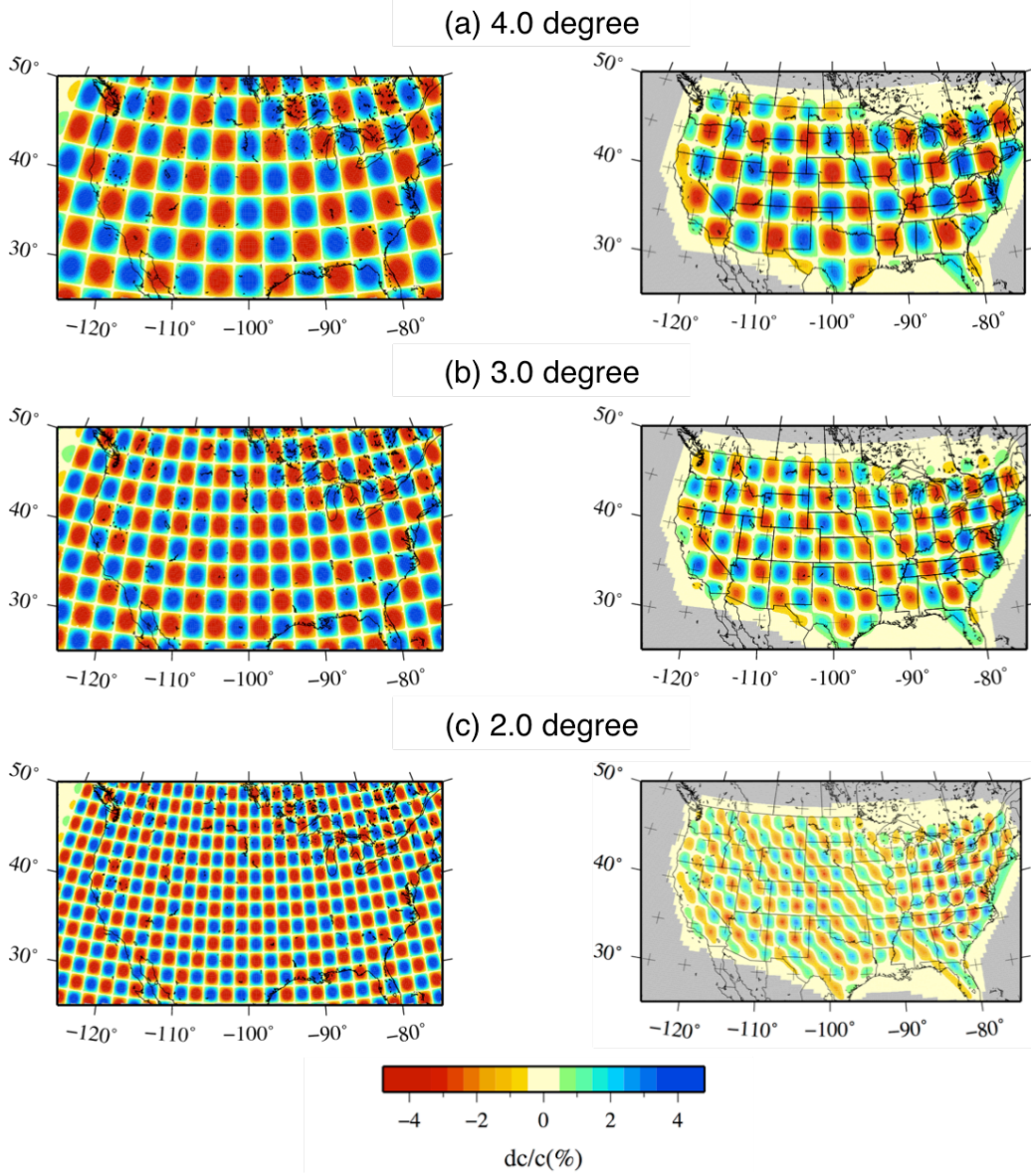


Figure 4.2: Examples of checkerboard resolution tests for Rayleigh-wave phase speed maps at 80 s with cell sizes of (a) 4.0°, (b) 3.0° and (c) 2.0°. Left panels are input checkerboard patterns, and right retrieved models.

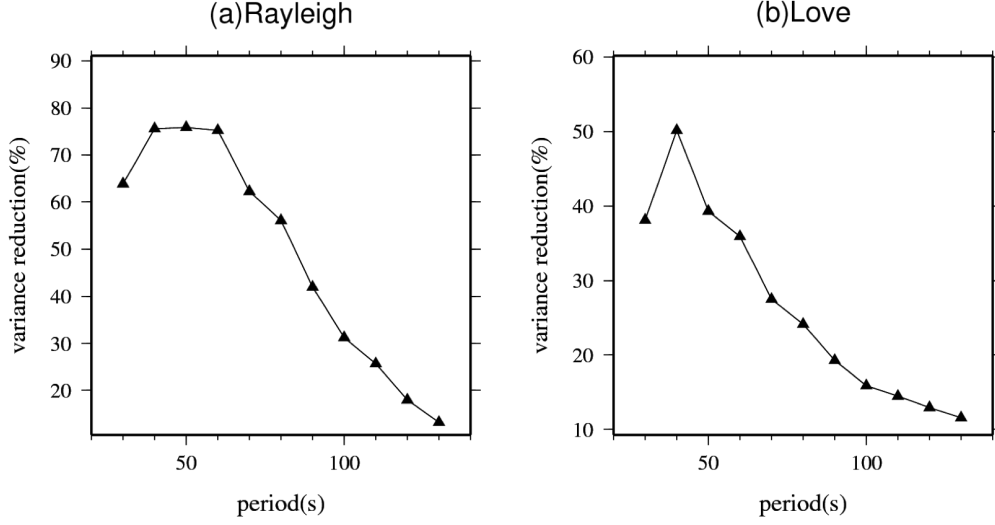


Figure 4.3: Variance reductions by inversions for phase speed maps of (a) Rayleigh waves and (b) Love waves with phase data.

p becomes small).

The average strengths of retrieved phase speed perturbations with respect to the input model are 51% for phase data only ($p = 1.0$), 75% ($p = 0.8$), and 94% for amplitude data only ($p = 0.0$). This represents better resolving power of amplitude data that are sensitive to the second derivatives of velocity perturbation, resulting in the better recovery of the strength of velocity perturbation for small-scale heterogeneity than phase data alone. The achievable horizontal resolution may be conversely lowered to some extent, due to the limited number of amplitude measurements. The variance reductions for the amplitude and phase terms, and the total data sets as a function of the data weight factor p are summarized in Fig. 4.7, where the total variance reduction is calculated from E in the equation (4.13). In this study, we choose the map for $p = 0.8$ as the best compromised model derived from the joint inversions of phase and amplitude data. In Fig 4.8, we display the resultant phase speed maps obtained from phase data only and those from

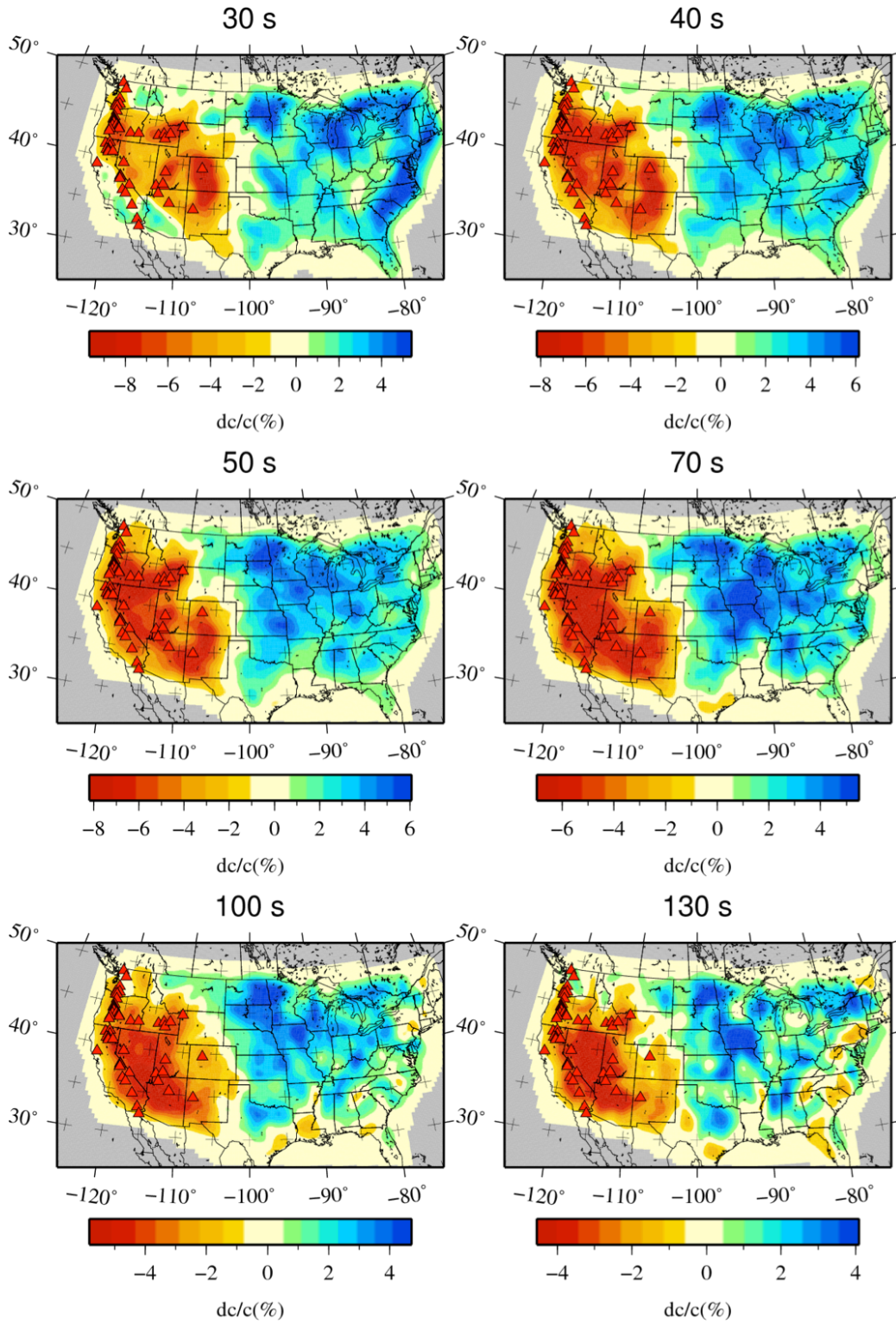


Figure 4.4: Rayleigh wave phase speed maps derived from phase data at periods from 30 to 130 s. Red triangles indicate the distribution of volcanoes (Siebert and Simkin, 2002).

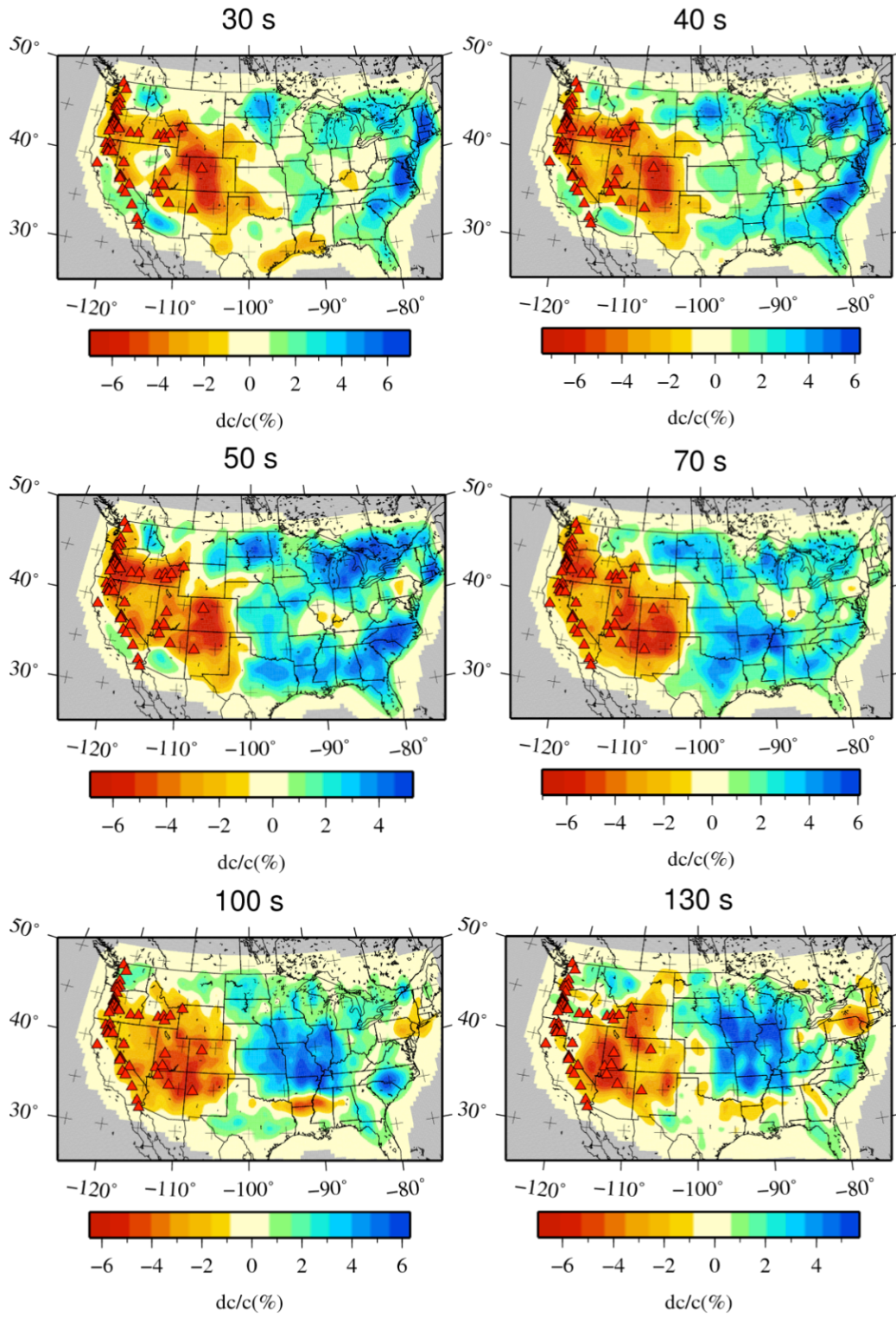


Figure 4.5: Same as Fig.4.4, but for Love waves.

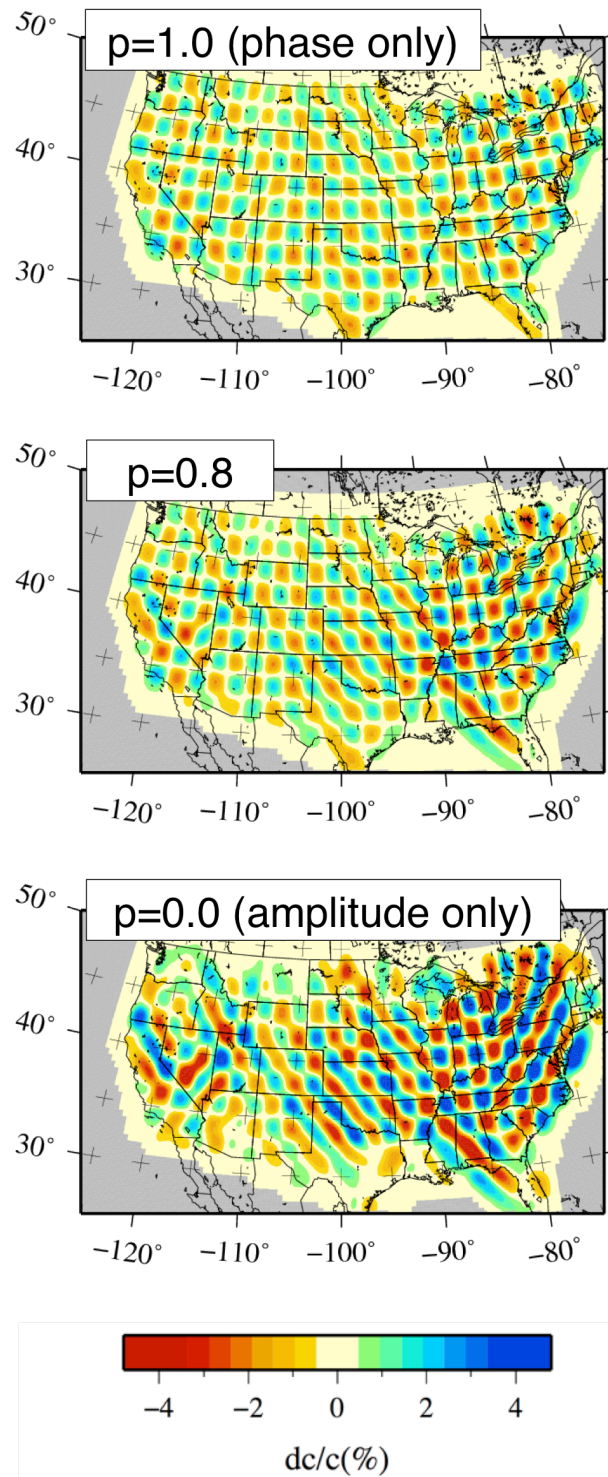


Figure 4.6: Checkerboard resolution tests with joint inversion of phase and amplitude data at 80 seconds, for a varying weighting factor p on phase data.

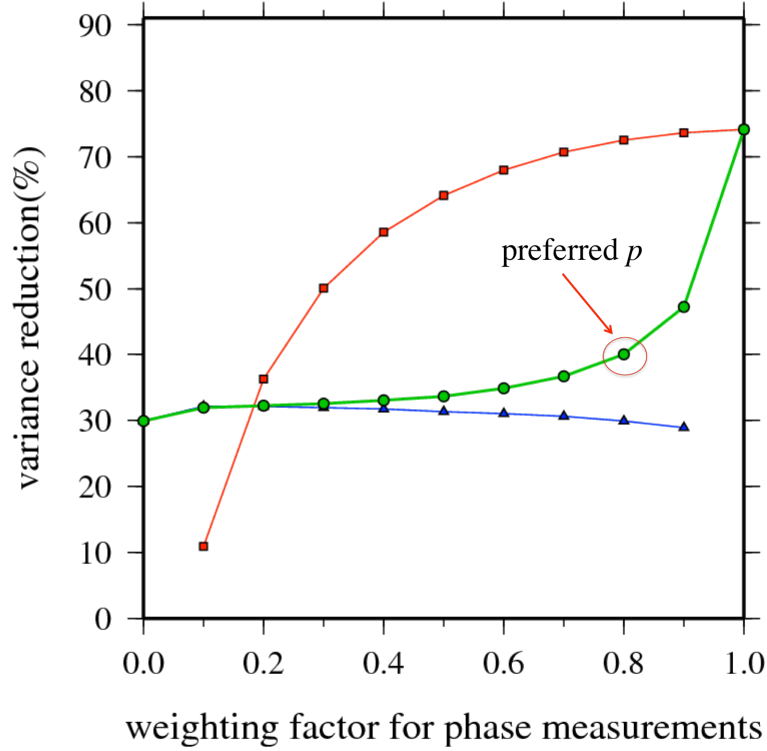


Figure 4.7: Variance reductions of amplitude (blue), phase (red) and total (green) as a function of the data weight factor p , from the joint inversions of Rayleigh-wave phase and amplitude data at 40 s.

the joint use of phase and amplitude data.

4.3.2 Local amplification at receiver locations

Observed interstation amplitude data of surface waves is inverted for the maps of phase speed and station correction term simultaneously. The station correction term represents the local amplification factor beneath the stations, while the phase speed maps derived from amplitude data reflect the effect of elastic focusing/defocusing. Fig. 4.9 shows the results of resolution tests for receiver amplification term derived from the joint inversions of phase and amplitude data of Rayleigh waves. The strength of amplification is better recovered by the inversion using amplitude data only ($p = 0.0$), while the

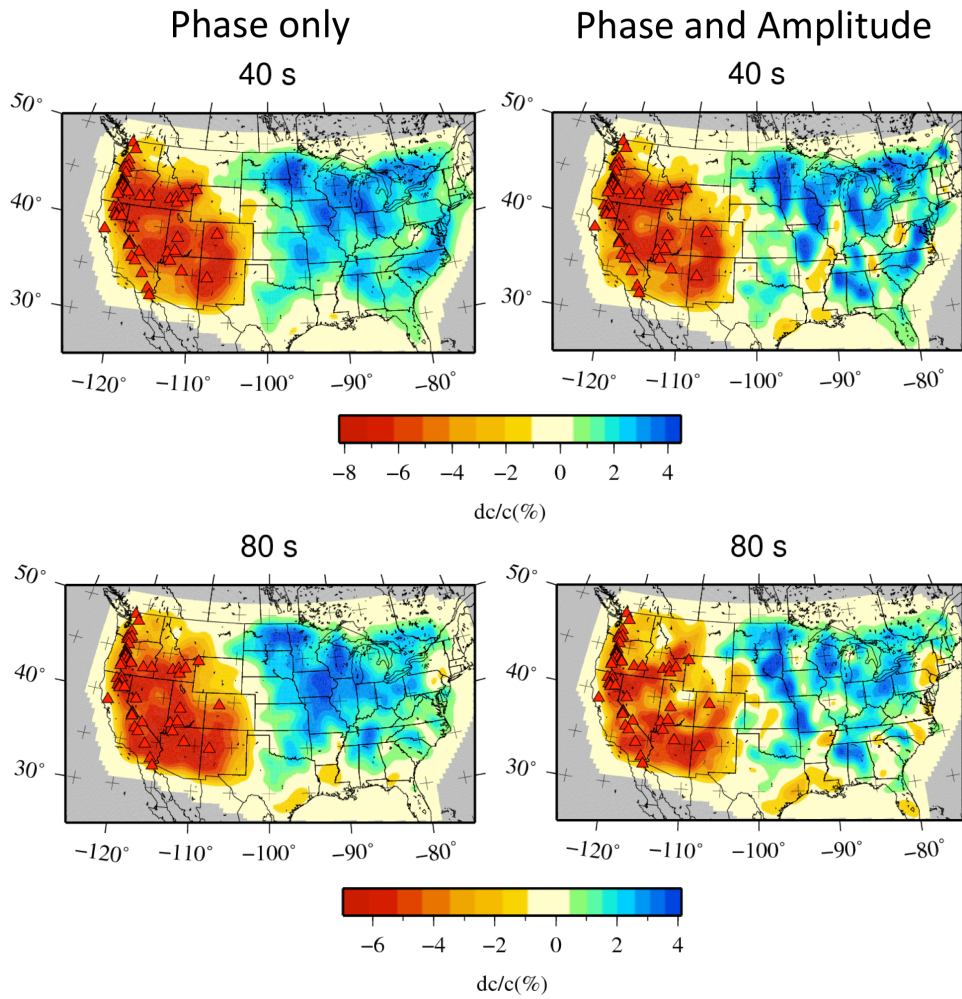


Figure 4.8: Phase speed maps derived from phase data only (left) and from joint inversions of Rayleigh-wave phase and amplitude data (right) at 40 and 80 seconds.

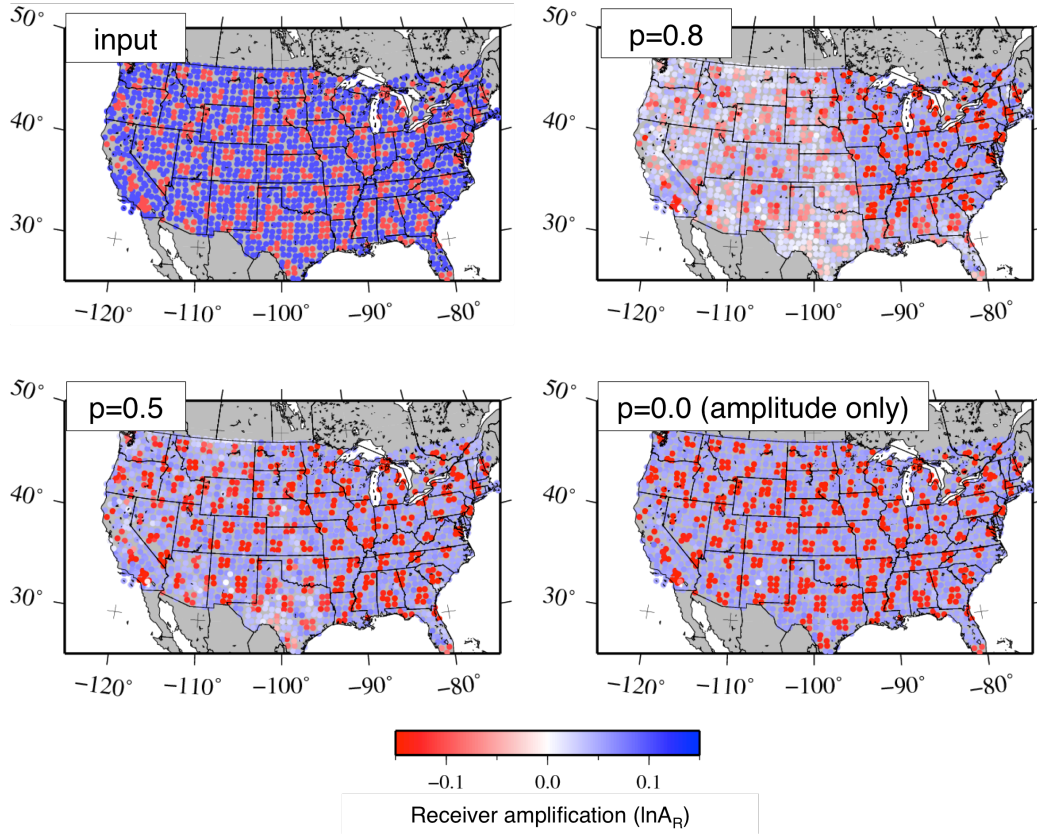


Figure 4.9: Resolution tests of receiver amplification term with joint inversion of phase and amplitude data at 80 seconds, for a varying weighting factor p on phase data.

resolution in the western and central U.S. tends to be limited as we weight more on phase data, which may be affected by the regional variation of path density for amplitude data across the U.S.

The spatial variation of the station correction term between 30 and 120 seconds is displayed in Fig. 4.10. The results represent local amplification of observed amplitude beneath the station, which shows a clear correlation with the velocity structure in the period longer than 50 seconds. For example, significant amplification can be seen in the Snake River Plain and Rio Grande Rift, and deamplification in the central cratonic region.

The major features of our station correction maps are similar to the local

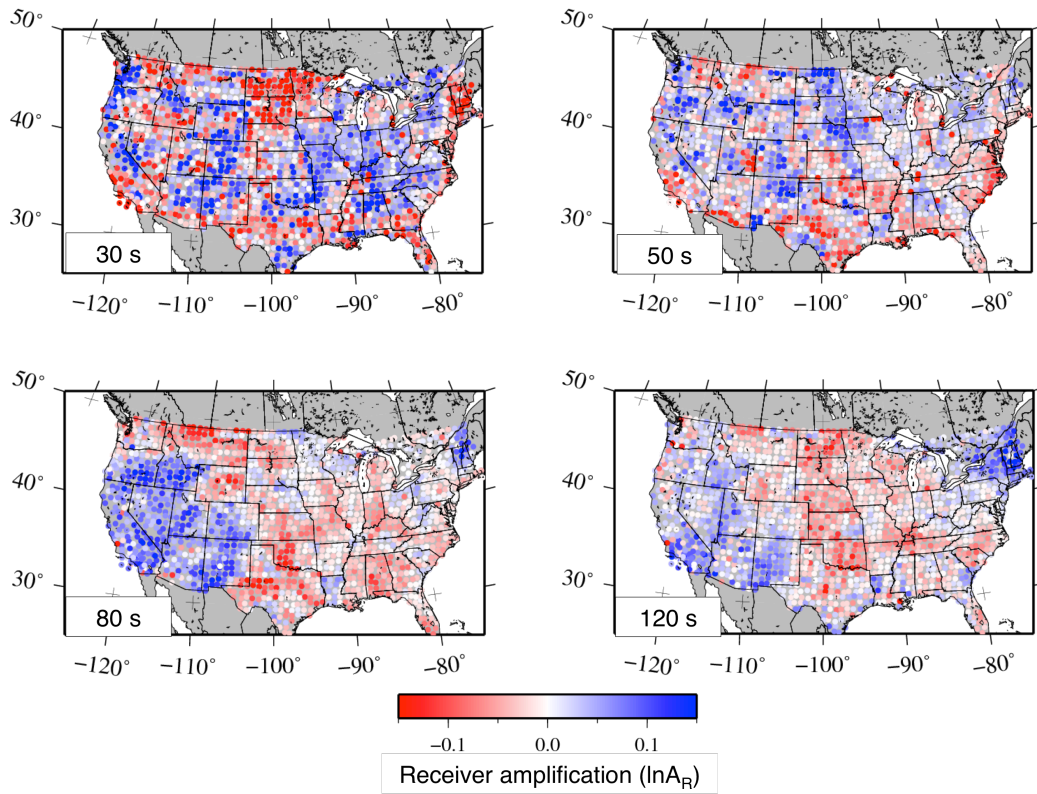


Figure 4.10: Spatial distributions of local amplification term at receiver locations at 30, 50, 80 and 120 seconds.

amplifications derived by Lin et al. (2012), in which the amplitude variations as well as the effect of focusing has been discussed using fundamental-mode Rayleigh waves. The distributions of local amplification term in the longer periods are also highly consistent with those derived by Eddy and Ekström (2014), which has measured amplitude ratios at adjacent stations and isolate the effects of local structure from those of the source and propagation. Thus, our results indicate that the effects of the focusing are isolated well from the local amplification through the joint inversions of the observed amplitude and phase data.

Chapter 5

3-D models in North American continent

5.1 Introduction

We construct 3-D S wave velocity models from the phase velocity maps of the fundamental-mode Rayleigh and Love waves obtained in chapter 4, employing the method of phase speed inversion for a shear-wave speed model following Yoshizawa and Kennett (2004) and Yoshizawa (2014). Local dispersion curves derived from the set of phase velocity maps are used to constrain local 1-D shear wave velocity models that form a 3-D shear wave model in the end. In this chapter, we summarize the method of constructing the 3-D S wave model, and display and discuss the results of 3-D shear wave speed structure in North America.

5.2 Formulation of inversion for 3-D S wave model

5.2.1 Inversion for isotropic 3-D S wave model

The phase speed maps obtained in chapter 4 are used to construct a 3-D S wave model. The linear relationship between a phase speed perturbation and a local 1-D structure can be expressed as follows (e.g., Takeuchi and Saito,

1972; Dahlen and Tromp, 1998),

$$\frac{\delta c(\omega)}{c} = \int_0^R \left\{ K_\rho(\omega, r) \frac{\delta \rho(r)}{\rho} + K_\alpha(\omega, r) \frac{\delta \alpha(r)}{\alpha} + K_\beta(\omega, r) \frac{\delta \beta(r)}{\beta} \right\} dr, \quad (5.1)$$

where R is the Earth's radius, δc is a phase speed perturbation, $\delta \rho$, $\delta \alpha$ and $\delta \beta$ are the perturbations of density, P-wave speed and S-wave speed, respectively. K_ρ , K_α and K_β are sensitivity kernels to represent the partial derivatives of phase speed with respect to each of the model parameter. The effects of parameters of the density and P-wave speed are smaller than those of S-wave speed in the period range between 30 and 130 s. Thus, we take the density and P-wave speed parameters into account through the scaling relationship to S-wave parameters as used in Yoshizawa (2014). We, therefore, consider only shear wave speed perturbations as independent parameters for inversions.

We use the iterative least squares inversion scheme of Tarantola and Valette (1982) to invert a dispersion curve for a local 1-D shear wave speed model. Here we consider a non-linear problem, in which the observed data vector $\mathbf{d}$ is represented by a function of model parameter vector $\mathbf{p}$ as follows:

$$\mathbf{d} = \mathbf{g}(\mathbf{p}), \quad (5.2)$$

where $\mathbf{d}$ consists of a set of phase speed perturbations $\delta c(\omega)$, and $\mathbf{p}$ consists of a set of 1-D shear wave perturbations $\delta \beta(r)$.

A model parameter vector at the $(k + 1)$ -th iteration for an underdetermined problem can be represented as,

$$\mathbf{p}_{k+1} = \mathbf{p}_0 + \mathbf{C}_{pp} \mathbf{G}_k^T (\mathbf{C}_{dd} + \mathbf{G}_k \mathbf{C}_{pp} \mathbf{G}_k^T)^{-1} [\mathbf{d} - \mathbf{g}(\mathbf{p}_k) + \mathbf{G}_k (\mathbf{p}_k - \mathbf{p}_0)], \quad (5.3)$$

where $\mathbf{p}_k$ is the model vector at the k -th iteration, $\mathbf{p}_0$ is the reference model vector, $\mathbf{C}_{dd}$ is the *a priori* data covariance matrix, $\mathbf{C}_{pp}$ is the *a priori* model covariance matrix, and $\mathbf{G}_k$ is the kernel matrix that comprised of the partial derivatives of data with respect to the model parameter (Tarantola and Valette, 1982).

The *a priori* model covariance matrix is assumed to be represented by a Gaussian distribution as follows,

$$\mathbf{C}_{pp}(r_i, r_j) = \sigma(r_i)\sigma(r_j) \exp \left\{ -\frac{1}{2} \frac{(r_i - r_j)^2}{L^2} \right\}, \quad (5.4)$$

where r_i is the depth of the i -th model parameter, $\sigma(r_i)$ is the standard deviation for the i -th model parameter, which controls the amplitude of variations for the model parameter, and L is the correlation length between the model parameters r_i and r_j , which determines the smoothness of model variations as a function of depth.

In this inversion for isotropic S wave speed models from Rayleigh-wave phase speed maps, we set the standard deviation σ to be 0.025 km/s and the correlation length L to be 3 km above the Moho, and σ to be 0.05 km/s and L to be 10 km below the Moho. The reference 1-D model is constructed by combining the crustal structure from the 3SMAC model (Nataf and Ricard, 1996), and the mantle structure from isotropic PREM (Dziewonski and Anderson, 1981).

5.2.2 Inversion for a radial anisotropy model

In addition to an isotropic S wave model, we invert for a radially anisotropic S wave model, using phase data of both Rayleigh and Love waves. In the inversion for radial anisotropy model, we may use two sets of 6 independent model parameters (e.g., Takeuchi and Saito, 1972), (A) $[\rho, \alpha_H, \phi, \beta_V, \xi, \eta]$ or (B) $[\rho, \alpha_H, \alpha_V, \beta_V, \beta_H, \eta]$, where α_H and α_V are PH and PV wave speeds, which are horizontally and vertically polarized P wave speeds, β_H and β_V are SH and SV wave speeds, which are horizontally and vertically polarized S wave speeds, η is the anisotropic parameter that controls the incident-angle dependency of wave speeds, $\phi = (\alpha_V/\alpha_H)^2$ and $\xi = (\beta_H/\beta_V)^2$. In the parameterization (B) with SV and SH wave speeds, the sensitivity kernels for Love waves exhibit better sensitivities to SH wave speeds and are less

affected by SV wave speeds (Yoshizawa, 2014), so that we adopt the model parameter set (B) in this study.

In this anisotropic inversions, we set a standard deviation σ to be 0.02 km/s and the correlation length L to be 5 km above the Moho, and σ to be 0.06 km/s and L to be 10 km below the Moho for the equation (5.4). We use a modified anisotropic PREM model (with a smoothed 220 km discontinuity), as a reference model of our inversion with the corrected crustal structure (3SMAC) as in the case of isotropic inversions.

5.3 Isotropic 3-D S wave model of North America

In Fig. 5.1, we compare two types of the 3-D S wave speed models in the whole U.S. derived from the different weight parameters between phase and amplitude data. Both S wave speed models reflect large-scale tectonic features well, for example, slow anomalies in the tectonically active western U.S., while fast anomalies in the eastern cratonic region. The 3-D shear wave models derived from both phase and amplitude data emphasize smaller-scale anomalies characterized by strong lateral velocity gradients.

Zoom-in maps of S wave speed and vertical cross-sections in the western and central U.S. are displayed in Figs. 5.2 and 5.3. We show 2-D S wave maps with absolute velocities, while vertical cross-sections are displayed as perturbations from the average velocity of the whole U.S. A significant difference between the two types of models can be seen in areas where the typical scale of velocity anomaly is about 200 km accompanied by the large velocity contrast with the surrounding areas, such as Colorado Plateau (CP) in Fig. 5.2 and New Madrid Seismic Zone (NMSZ) in Fig. 5.3. These results indicate that inter-station amplitude ratios extracted from dense seismic array enable us to reconstruct shorter-wavelength elastic properties, owing to its sensitivities to the second derivatives of phase speeds across each path.

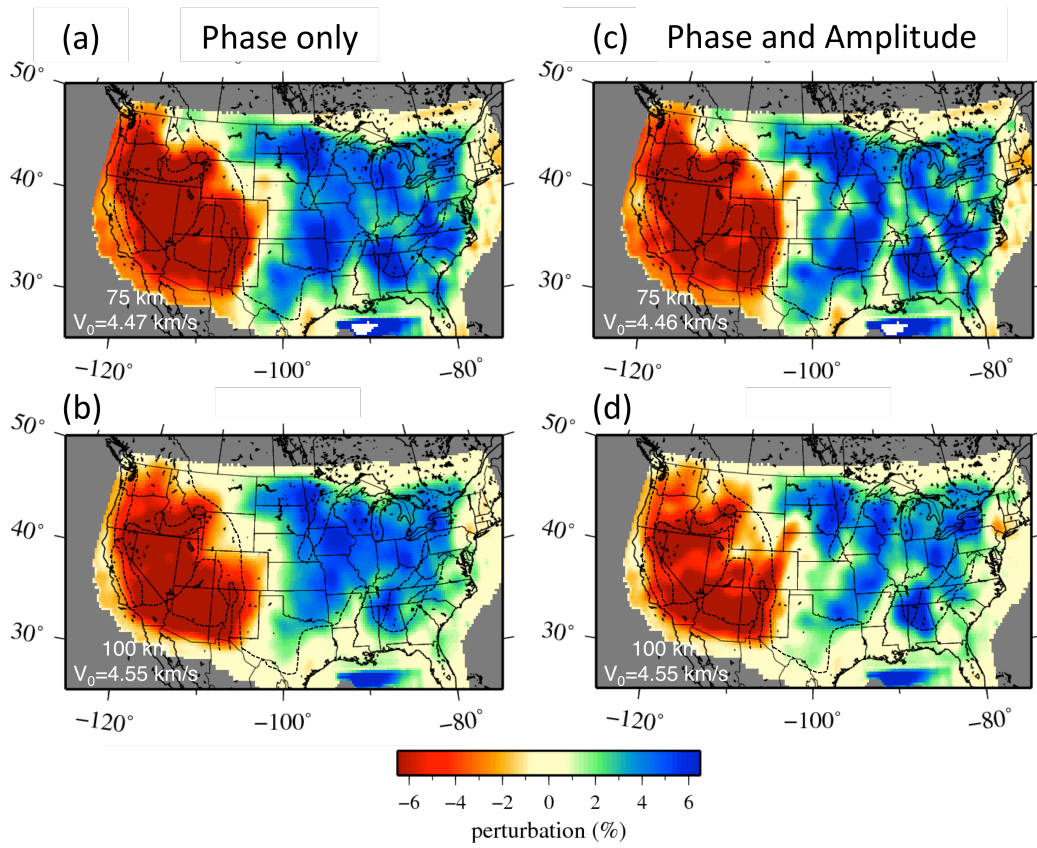


Figure 5.1: 3-D S wave velocity model at the depth of 75 (a,c) and 100 km (b,d) derived from (a,b) phase data only and (c,d) phase and amplitude data.

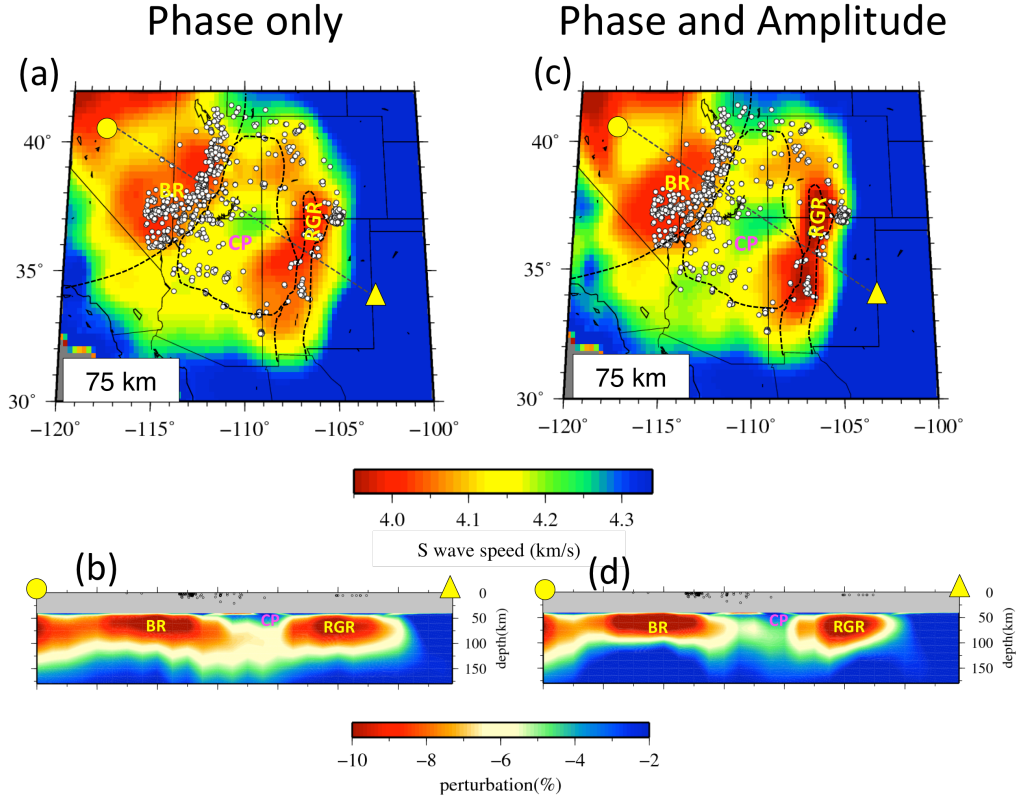


Figure 5.2: (a,c): 3-D S wave speed model in the western U.S. at the depth of 75 km derived from phase data only and phase and amplitude data. White circles indicate the locations of earthquakes with magnitude greater than 2.5, taken from the catalog by the United States Geological Survey in the period between 2000 and 2015. (b,d): Depth cross-section along the lines shown in the top of Fig. 5.2, which passes from Basin and Range (BR) to Rio Grande Rift (RGR), including the area of Colorado Plateau (CP).

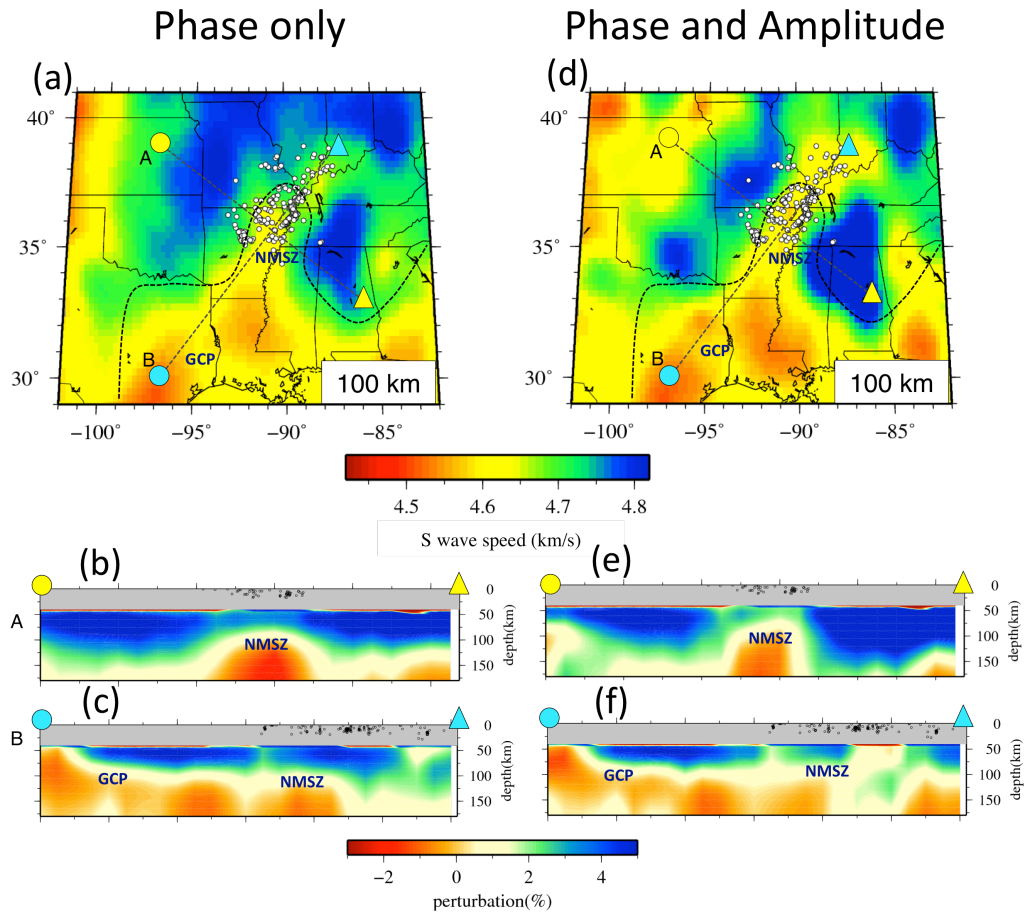


Figure 5.3: (a,d) Same as Fig. 5.2, but in the central U.S. including the area of New Madrid Seismic Zone (NMSZ) at the depth of 100 km. (b,c,e,f) Two vertical cross-sections passing through the NMSZ from midcontinent region (A), and from Gulf Coastal Plain (GCP) (B).

Our isotropic 3-D shear wave speed model (Figs. 5.1, 5.2, 5.3) with phase and amplitude data of surface waves emphasizes smaller-scale structures characterized by strong lateral velocity gradients, which have not been so clear in the conventional tomographic studies derived from phase data only.

While significant slow anomalies are seen in the eastern edge of Basin and Range and Rio Grande Rift at depth shallower than 100 km in both models, Colorado Plateau can be identified clearly as fast anomaly spots compared with the surrounding areas in the joint model (Fig. 5.2). Smaller-scale tectonic features are also mapped clearly in the central U.S., that is, slow anomalies can be seen in New Madrid Seismic Zone at around the depth of 100 km, as shown in Fig. 5.3.

The joint model derived from phase and amplitude data show clear anomalies with a horizontal scale around 200 km, comparable to the wavelength of Rayleigh waves at 50 s, although such anomalies are blurred in the model derived from phase data only. These results indicate that inter-station amplitude data are useful to reconstruct short-wavelength structures, owing to their sensitivities to the second derivative of phase speed across the ray path. In the next section, we will discuss in detail the prominent features of small-scale heterogeneities in the new 3-D model derived from the joint inversions of phase and amplitude data.

5.4 Comparisons with tectonic features in North America

The North American continent encompasses a variety of complex structural features, including a tectonically active region in the west and stable cratons in the east, divided by Rocky Mountains. We discuss in detail the local-scale tectonic features in the United States, retrieved by our joint inversion models utilizing amplitude data.

5.4.1 Southwestern U.S.

Fig. 5.2 exhibits a significantly slow anomaly localized in the Basin and Range province and Rio Grande Rift. In the Basin and Range province, seismic activities dominated by normal faulting are concentrated along its eastern and western boundaries (e.g., Pancha et al., 2006), suggesting that large extensional strains have deformed this region. The Rio Grande Rift has also undergone lithospheric extension since the middle to late Cenozoic deformation (e.g., Baldrige et al., 1991). The slow anomalies in this region is consistent with the high heat flow in the crust and uppermost mantle (e.g., Goes and van der Lee, 2002).

To the contrary, Colorado Plateau is characterized by a faster patch encompassed with slow anomalies (Fig. 5.2). This feature become much clearer in the shear wave model derived from phase and amplitude data, owing to the well-constrained local velocity gradient. A recent tomographic study of body waves in this region (Sine et al., 2008) has shown that Proterozoic mantle lithosphere reaches down to the depth of 100 km or deeper beneath the Colorado Plateau, which is well reflected in our S wave models with faster anomalies than its surroundings.

In Fig. 5.2, the central Basin and Range shows a relationally fast anomaly compared with the northern and southern areas of Basin and Range. The investigation of azimuthal anisotropy of split SKS arrivals indicates that the rotational flow may exist around the edges of narrow subducting slabs in this area (Zandt and Humphreys, 2008). Another study, which has constructed P-wave tomographic model, has shown that the fast anomaly is seen beneath the central Basin and Range down to the depth of 100 km, suggesting that a lithospheric drip has been generated (West et al., 2009). Our S wave model derived from phase and amplitude data shows a strong correlation with the tectonic feature in this area, which has not been clearly mapped in other

tomographic models derived from surface wave phase information.

5.4.2 Central U.S.

In Fig. 5.3, a significant slow anomaly is mapped right beneath the New Madrid Seismic Zone (NMSZ), where three large historic seismic events occurred in 1811-1812. Regional tomographic studies in the NMSZ have shown that the slow anomaly exists at deeper depth below 100 km (e.g., Bedle and van der Lee, 2006; Pollitz and Mooney, 2014). The descent of the ancient Farallon slab beneath the central U.S. is expected to drive localized viscous flow and stress directly below the NMSZ (Forte et al., 2007), and thus the low velocity in the NMSZ extends deeper than other typical rift zones in the central U.S., where slow anomalies are seen in the crust and shallow mantle (Pollitz and Mooney, 2014). Our model (Fig. 5.3 e and f) indicates slow anomalies in the depth below 70 km, and the lithosphere here is much thinner than the surrounding cratonic lithosphere.

Slow anomalies also appear beneath Gulf Coastal Plain (Fig. 5.3 d and f) in the upper mantle. A recent tomographic study in this region has shown that V_p is slightly faster while V_s is slower beneath the Gulf Coast, which suggests the existence of a layer of partial melt that marks the top of a shear zone or the Lithosphere-Asthenosphere Boundary (Evanzia et al., 2014). In the cross section in Fig. 5.3 (e and f), we can see the thin continental lithosphere with the thickness of about 75-90 km, which is lain by slow velocity asthenosphere in the Gulf Coast areas.

5.5 Radial anisotropy in the upper mantle beneath North America

We display the radially anisotropic S wave speed model; SV wave speed β_V (Fig. 5.4), SH wave speed β_H in Fig. 5.5 and radial anisotropy parameter $\xi =$

$(\beta_H/\beta_V)^2$ (Fig. 5.6). In this study, we have not used amplitude measurements for Love waves in the phase speed mapping, due to higher noise levels in the horizontal component, and higher-mode interference with the fundamental mode. Thus, the radial anisotropy model is derived from phase data only for both Rayleigh and Love waves, to ensure comparable resolution between SV and SH wave speed maps.

In Fig. 5.6, we can see large-scale anomalies with faster SH wave speed in the western and central active regions above the depth of 100 km, such as Snake River Plain, Basin and Range, Rio Grande Rift, and New Madrid Seismic Zone, which may indicate the effects of the current or former lateral extension underneath these regions. On the other hand, the radial anisotropy ξ becomes less than 1.0 below the depth of 75 km in the eastern region. Our model is highly consistent with the 3-D anisotropy models derived by Yuan et al. (2011, 2014), with strong anisotropy with $\xi < 1$ in this area below 60 km depth.

The quality of measurements for Love-wave phase speed is remarkably lower than that for Rayleigh waves, due to the worse signal-to-noise ratios in the horizontal components, and the higher-mode interference with the fundamental mode, which will be discussed in chapter 6, so that suitably-corrected phase speed data for the fundamental-mode Love waves are necessary for enhancing the resolution and accuracy of radial anisotropy model.

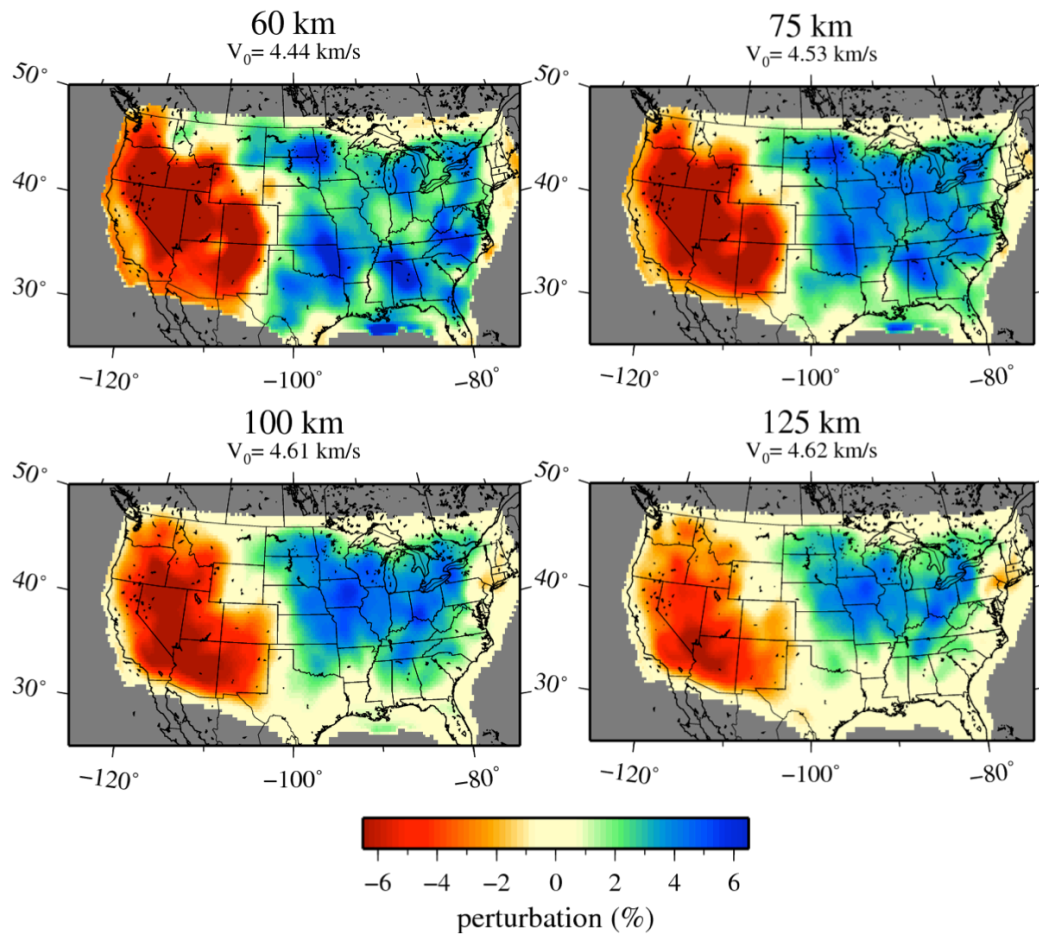


Figure 5.4: SV wave speed models at 60, 75 100 and 125 km depth, derived from the anisotropic inversion using Rayleigh and Love waves.

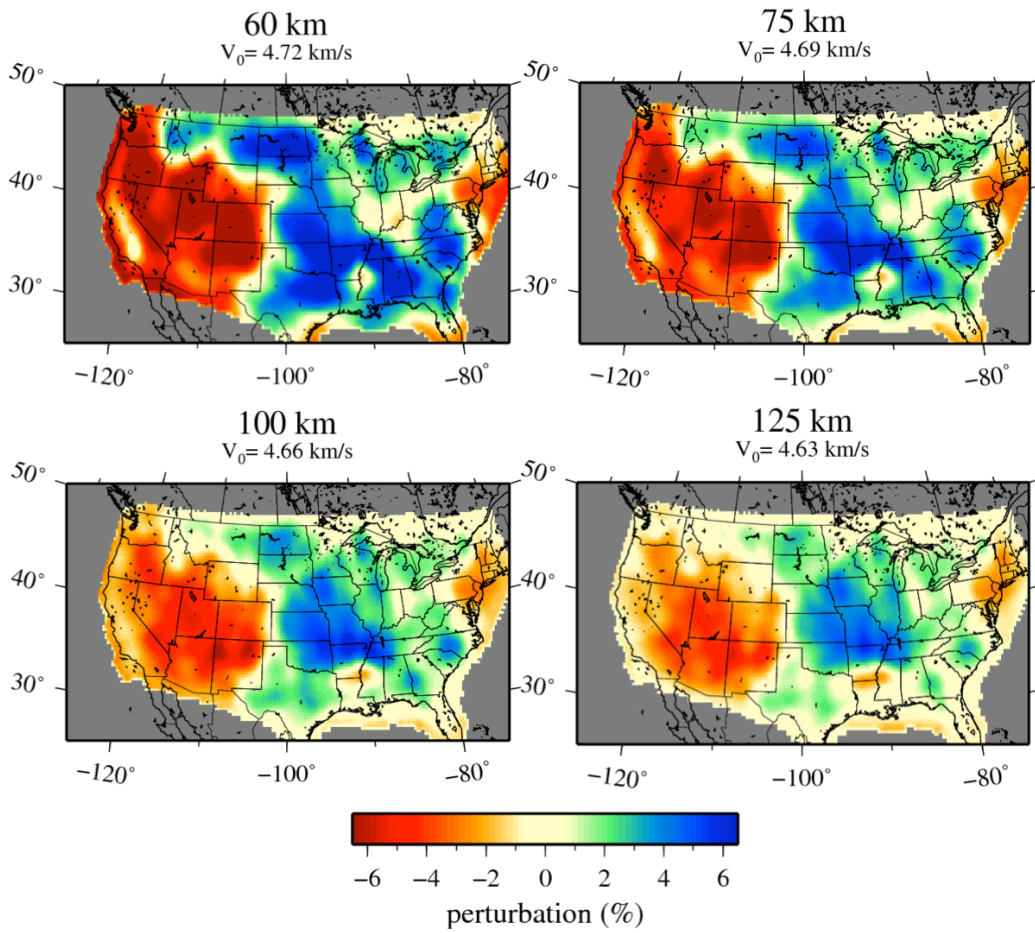


Figure 5.5: Same as Fig. 5.4, but for SH wave speed models.

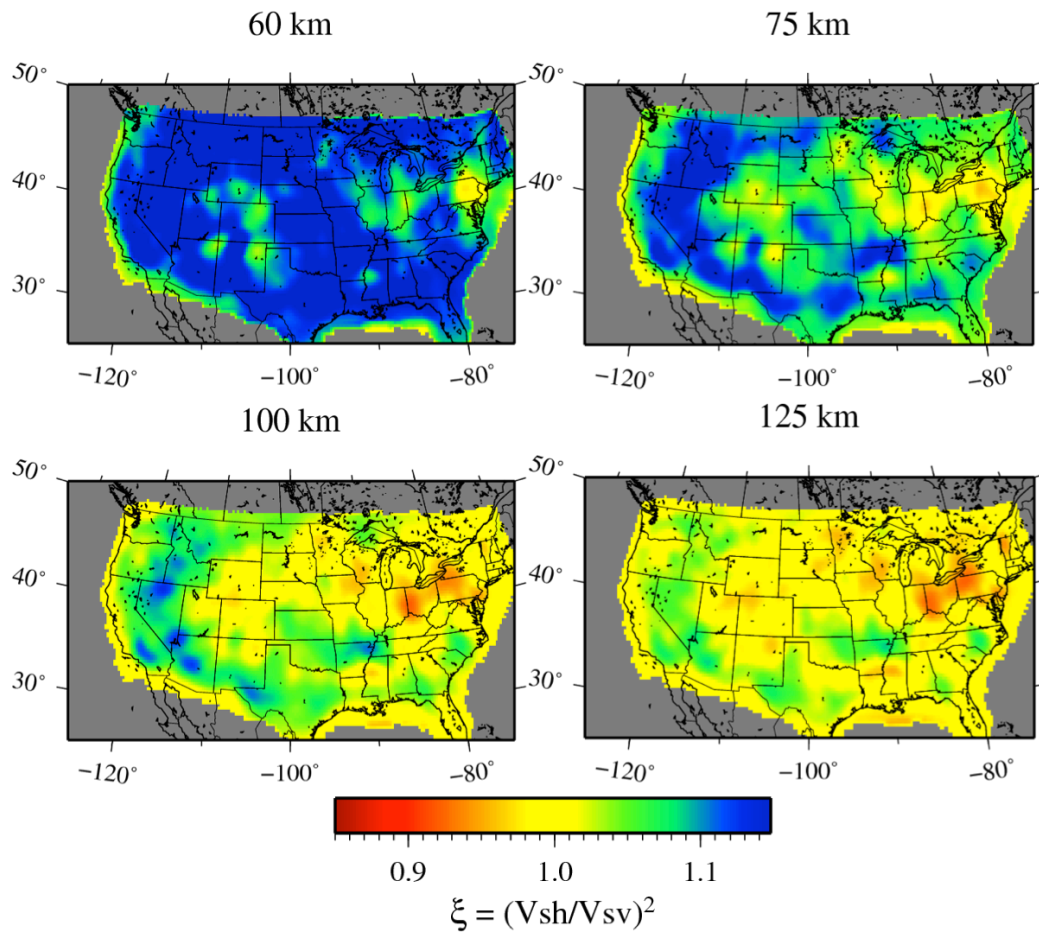


Figure 5.6: Same as Fig. 5.4, but for radial anisotropy parameter $\xi = (\beta_H/\beta_V)^2$.

Chapter 6

Conclusions and future prospects

6.1 Summary of the thesis

In this thesis, we have developed new techniques of inter-station phase speed and amplitude measurements, and the joint inversion of phase and amplitude data for reconstructing phase speed maps and station amplification factors. The methods are applied to the dense broadband seismic array, USArray, deployed across the United States in the past decade, and a new high-resolution 3-D S wave speed model of North American upper mantle has been obtained.

In chapter 3, we have explained the fully non-linear waveform fitting technique to measure inter-station phase speeds and amplitude ratios simultaneously, using a global optimization method. Model parameters of the phase speed and amplitude ratio are explored by fitting two observed waveforms located on a common great-circle path by perturbing both phase and amplitude of the fundamental-mode surface waves. We have introduced the reliability parameter that quantifies how well the waveforms at two stations can be fitted in a time-frequency domain, which can be used as a data selection criterion for the subsequent phase speed mapping. This method of waveform fitting has been applied to observed seismograms of USArray, and a large-number of inter-station phase speed and amplitude data are collected in a period range from 30 and 130 seconds. The typical inter-station distances

for the majority of measured dispersion data are much less than 1000 km, which is much shorter than the average path-length of conventional single-station measurements between source and receiver. Thus, the inter-station data can be helpful for improving the horizontal resolution of tomography models.

In chapter 4, the measured inter-station phase speed and amplitude ratios are inverted simultaneously for the phase speed maps as well as local amplification factor at each receiver location. The checkerboard resolution tests suggest that the phase speed maps derived from both phase and amplitude data tend to exhibit better recovery of the strength of phase speed perturbations, particularly for the smaller-scale heterogeneity. The results indicate that the amplitude anomalies caused by elastic focusing/defocusing, which are sensitive to the second derivatives of phase speed across the ray path, can be of help in retrieving the actual strength of shorter-wavelength heterogeneity. The distribution of local amplification factor tends to be correlated well with the velocity structure in the upper mantle; e.g., higher amplification anomalies in Snake River Plain and Rio Grande Rift. The results are consistent with another recent study of local amplification factor (e.g., Eddy and Ekström, 2014), particularly in the longer periods, so that we can expect that, in our joint inversion of phase and amplitude, the effects of local amplification have been isolated well from those of focusing.

In chapter 5, a new isotropic 3-D S wave speed model of North American continent has been obtained from the phase speed maps using phase and amplitude data for the vertical component Rayleigh waves. With the aid of amplitude data, our isotropic 3-D S wave model tends to emphasize local-scale tectonic features associated with conspicuous lateral velocity gradients; e.g., fast anomaly in the Colorado Plateau surrounded by slow anomalies, and slow anomaly in the New Madrid Seismic Zone encompassed by fast

anomalies. Such local tectonic characters, whose scale-length of heterogeneity is typically around 200 km, can hardly be identified in the conventional surface wave tomography, and thus the inter-station amplitude ratio data can be of great help in reconstructing shorter-wavelength structures of the upper mantle. We have also constructed a preliminary radial anisotropy model from simultaneous inversion of the phase speed maps of Love and Rayleigh waves derived from phase data only. The radial anisotropy model shows faster SH wave speed anomaly than SV in the western and central active regions, while the model exhibits faster SV wave speed anomaly than SH in the eastern region below the depth of 75 km.

6.2 Prospects for future studies

6.2.1 Hybrid surface-wave mapping: single-station, inter-station and inter-array measurements

In this study, we have used the seismograms of the USArray and the Southern California Seismic Network stations only, so that we can achieve path coverage as uniform as possible, and also keep the uniform data quality for stable measurements of surface-wave phase and amplitude. This restriction of the employed seismic stations were successful in retrieving high-resolution model of the United States, but, at the same time, the target region of the model is restricted to the mainland of the U.S. In the North American region outside of the U.S., seismic stations are located sparsely compared with US-Array, and thus the single-station method is required for the reconstruction of a large-scale 3-D S wave model encompassing wide areas of North American continent, which is inevitable for unveiling the seismic structure of the cratonic lithosphere and asthenosphere.

To enhance the vertical resolution of 3-D S wave speed models, the higher-mode surface waves, which are more sensitive to deep structures than

the fundamental-mode surface waves (Fig. 6.1), are necessary for the tomographic reconstruction. Yoshizawa and Ekström (2010) have employed fully non-linear waveform inversion for multimode phase speed measurements with the single-station method, which has been applied to North American stations for the phase speed maps of multimode Rayleigh and Love waves. The combination of the inter-station phase and amplitude measurements described in this thesis and the single-station measurements of multimode surface waves will provide us with the new combined 3-D S wave model for the full range of North American continent with enhanced vertical resolution throughout the upper mantle.

Our method of inter-station phase and amplitude measurements can also be applied to other dense seismic arrays all around the world. As an example, the inter-station method developed in this thesis is being applied to the dense broadband seismic arrays in northeast China (NECESSArray) and in Japan (F-net) for updating the existing upper mantle model in the Japanese Islands and its surrounding regions (Yoshizawa et al., 2010). With the combination of two arrays in the continent and in the island arc, we can enhance the inter-station path coverage across the Sea of Japan. Such inter-array analysis of surface wave phase and amplitude is expected to enhance the horizontal resolution of velocity model in the marginal sea.

6.2.2 Higher-mode interference on Love waves

While the majority of surface-wave models on regional scales have been based on Rayleigh wave observations, the number of Love wave models tend to be limited mainly due to a variety of uncertainties in the horizontal component seismograms, including higher noise levels, misalignment of seismometers, arrival-angle anomalies, and the interference of higher modes, which makes it difficult to measure phase dispersion of the fundamental-mode Love waves

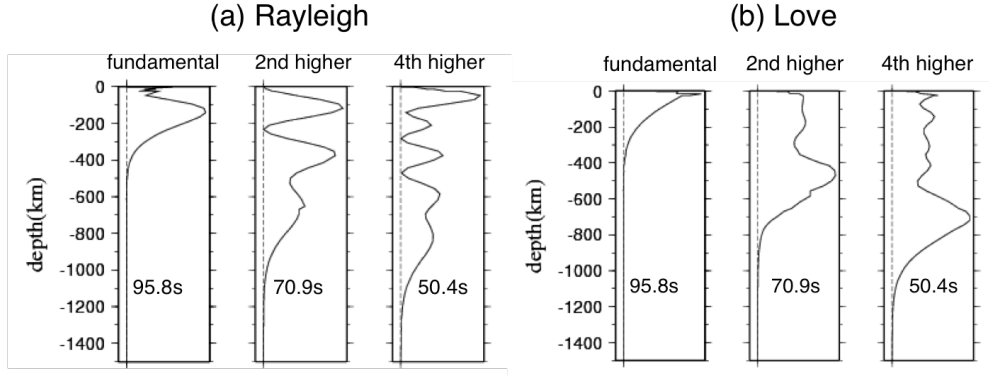


Figure 6.1: Vertical sensitivity kernels of (a) Rayleigh and (b) Love waves calculated from PREM (Dziewonski and Anderson, 1981). All sensitive kernels are normalized so that the maximum amplitude of K_β for each mode becomes unity.

accurately (e.g., Yoshizawa and Ekström, 2010; Foster et al., 2014a).

One of the intrinsic difficulties of Love wave analysis is attributed to the fact that the group speeds of fundamental-mode Love waves are similar to those of higher-modes, particularly at the periods from 50 to 100 s (Fig. 6.2). In other words, the wave packets of the fundamental mode and higher mode Love waves are overlapped in an observed seismogram. Foster et al. (2014b) have examined the effect of higher-mode interference on phase measurements of fundamental-mode Love waves for single-station and two-station methods. They have shown that the contamination effects of higher-mode interference on a single-station measurements are about 1 % of the measured phase speeds, while those on two-station measurements are up to 10 %.

There are several attempts to isolate fundamental-mode phase speeds from higher modes. van Heijst and Woodhouse (1997) have employed a mode-stripping technique to measure phase speeds at relatively long epicentral distances greater than 30 degrees, which have been used in the global-scale mapping of the multi-mode phase speed maps (van Heijst and Woodhouse, 1999). Luo et al. (2015) have recently applied linear radon transform

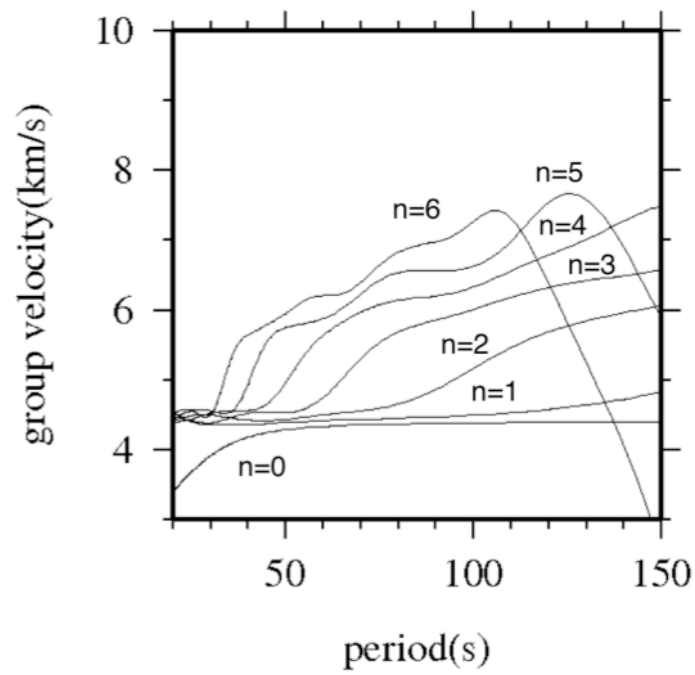


Figure 6.2: Group speed dispersion curves for Love waves of fundamental mode ($n=0$) and 1st-6th higher modes calculated from PREM.

(LRT) to synthetic and real seismograms of USArray to demonstrate the utility of LRT in separating the fundamental-mode Love waves from higher modes. They have suggested that the contamination effects by higher modes reach up to 10 % for two-station measurements of Love waves, which is consistent with the independent estimation by Foster et al. (2014b). Such efforts for improving the measurement quality of the inter-station Love wave phase speeds through the modal separation are important for the accurate reconstruction of radial anisotropy in the upper mantle, which provides us with precious information on mantle dynamics.

6.2.3 Joint analysis of surface waves and receiver functions

S receiver functions, which represent wave conversions from S to P at interfaces underneath a seismic station, are useful to identify the boundaries in the upper mantle, while P receiver functions (conversion from P to S) reflect the discontinuity at shallower depth (Yuan et al., 2006). Kind et al. (2015) have recently used S receiver functions of USArray stations, and constructed the structure of the upper mantle between the Moho and the 410 km discontinuity in the western and central U.S. The mid-lithospheric discontinuity (MLD), whose cause is largely unknown, has been identified in the cratonic region at the depth of about 100 km as well as the lithosphere-asthenosphere boundary (LAB) has been found at 200 km depth.

Julia et al. (2000) have developed the method of joint inversion of receiver function and surface wave dispersion for the fundamental mode. Bodin et al. (2012) have employed the joint inversion method of receiver function and the fundamental mode surface waves, based on a Bayesian approach for obtaining a posterior probability distribution of the velocity model. They have shown that, using synthetic experiments, 1-D S wave velocity and the discontinuity are resolved well with the joint inversion, while they become rather obscure in

the independent inversions using surface wave dispersion or receiver functions only. Thus, 3-D S wave models derived from the joint inversion of surface waves and S receiver function are expected to improve the accuracy of the retrieved upper mantle structure, including the plausible interfaces, such as the MLD and LAB underneath the continental and oceanic lithosphere.

Appendix A

Linear equation for amplitude anomaly due to focusing/defocusing

A.1 Single-station amplitude anomaly due to focusing/defocusing

Surface wave amplitude can be affected by focusing and defocusing of surface-wave rays due to lateral heterogeneity in the earth, which can be mathematically represented by a linear equation based on the first-order perturbation theory (e.g., Woodhouse and Wong, 1986; Wang and Dahlen, 1994).

As discussed in chapter 1, Woodhouse and Wong (1986) first derived a linear relation between the amplitude anomaly caused by elastic focusing and the phase speed perturbation along the propagation path from the source to receiver, by considering the first-order perturbation of the geometrical spreading factor,

$$\ln A_F = \frac{1}{2 \sin \Delta} \int_0^\Delta \sin(\Delta - \phi) [\sin \phi \partial_\theta^2 - \cos \phi \partial_\phi] \frac{\delta c}{c_0}(\omega) d\phi. \quad (\text{A.1})$$

The same equation has also been obtained by Romanowicz (1987) in a framework of asymptotic normal mode formulation. The equation (A.1) have been utilized in some earlier studies; e.g., phase velocity mapping incorporating amplitude anomaly (e.g. Laske and Masters, 1996), and focusing corrections for mapping attenuation models (e.g. Selby and Woodhouse, 2000).

A slightly modified version of a linear relation for focusing/defocusing of surface wave amplitude for a source-receiver path have later been proposed by Wang and Dahlen (1994) and Dahlen and Tromp (1998) based on a formulation in a framework of surface-wave WKB theory,

$$\begin{aligned} \ln A'_F = & \frac{\delta c_R}{2c_0} + \frac{\delta c_S}{2c_0} + \frac{1}{2 \sin \Delta} \int_0^\Delta [\sin(\Delta - \phi) \sin \phi \partial_\theta^2 \\ & - \cos(\Delta - 2\phi)] \frac{\delta c}{c_0}(\omega) d\phi, \end{aligned} \quad (\text{A.2})$$

where δc_R and δc_S are the phase speed perturbations at the receiver and source locations, respectively. Dalton and Ekström (2006a,b) have employed this equation for mapping phase speed models and Q structure using the amplitude anomaly measured at a single station.

The only difference between equations (A.1) and (A.2) is that an additional correction term for the fractional phase speed perturbation at the receiver location, $\delta c_R/2c_0$, is included in (A.2). It should be noted that the second term in the righthand side of (A.2) (i.e., $\delta c_S/2c_0$) as well as the second term of the integrand (i.e., $\cos(\Delta - 2\phi)$) can be derived from the integration of the term involving an along-path gradient $\partial_\phi(\delta c/c)$ in (A.1) by part. Thus, we can recast the relation between (A.1) and (A.2) as, $\ln A'_F = \delta c_R/2c_0 + \ln A_F$. As argued in Wang and Dahlen (1994) and Dahlen and Tromp (1998), $\ln A'_F$ in equation (A.2) satisfies the source-receiver reciprocity.

A.2 Approximated representation for inter-station amplitude anomaly due to focusing/defocusing

Now we consider the inter-station amplitude anomaly based on these two equations. By taking the difference between the amplitude anomalies for the farther station, A_F^{far} , and that for the nearer station, A_F^{near} , using (A.1),

inter-station amplitude ratios can be represented as follows,

$$\begin{aligned} \ln \left(\frac{A_F^{far}}{A_F^{near}} \right) &= \frac{1}{2 \sin (\Delta_n + d\Delta)} \int_0^{\Delta_n + d\Delta} \sin(\Delta_n + d\Delta - \phi) \\ &\quad \times [\sin \phi \partial_\theta^2 - \cos \phi \partial_\phi] \frac{\delta c}{c_0}(\omega) d\phi \\ &\quad - \frac{1}{2 \sin \Delta_n} \int_0^{\Delta_n} \sin(\Delta_n - \phi) \\ &\quad \times [\sin \phi \partial_\theta^2 - \cos \phi \partial_\phi] \frac{\delta c}{c_0}(\omega) d\phi, \quad (\text{A.3}) \end{aligned}$$

where Δ_n is the epicentral distance from the source to the nearer station, and $d\Delta$ is the inter-station distance.

With the approximation such that $\Delta_n \sim \Delta_n + d\Delta$ under the assumption of $\Delta_n \gg d\Delta$, (A.3) can be simplified as,

$$\begin{aligned} \ln \left(\frac{A_F^{far}}{A_F^{near}} \right) &\simeq \frac{1}{2 \sin (\Delta_n + d\Delta)} \int_{\Delta_n}^{\Delta_n + d\Delta} \sin(\Delta_n + d\Delta - \phi) \\ &\quad \times [\sin \phi \partial_\theta^2 - \cos \phi \partial_\phi] \frac{\delta c}{c_0}(\omega) d\phi. \quad (\text{A.4}) \end{aligned}$$

Finally, by changing the independent variable, $\phi' = \phi - \Delta_n$, (A.4) can be expressed as follows,

$$\begin{aligned} \ln \left(\frac{A_F^{far}}{A_F^{near}} \right) &\simeq \frac{1}{2 \sin (\Delta_n + d\Delta)} \int_0^{d\Delta} \sin(d\Delta - \phi') [\sin(\Delta_n + \phi') \partial_\theta^2 \\ &\quad - \cos(\Delta_n + \phi') \partial_\phi] \frac{\delta c}{c_0}(\omega) d\phi'. \quad (\text{A.5}) \end{aligned}$$

In a similar way, the inter-station amplitude anomaly based on (A.2) by Wang and Dahlen (1994) and Dahlen and Tromp (1998) can be given as follows,

$$\begin{aligned} \ln \left(\frac{A_F'^{far}}{A_F'^{near}} \right) &\simeq \frac{\delta c_f}{2c_0} - \frac{\delta c_n}{2c_0} \\ &\quad + \frac{1}{2 \sin (\Delta_n + d\Delta)} \int_0^{d\Delta} [\sin(d\Delta - \phi') \sin(\Delta_n + \phi') \partial_\theta^2 \\ &\quad - \cos(d\Delta - \Delta_n - 2\phi')] \frac{\delta c}{c_0}(\omega) d\phi', \quad (\text{A.6}) \end{aligned}$$

where δc_f and δc_n are the local phase speed perturbations at the farther and nearer stations.

(A.5) represents the inter-station focusing term based on the formulation given by Woodhouse and Wong (1986), while (A.6) represents that given by Dahlen and Tromp (1998). The essential differences between these two are the correction terms of phase speed perturbations at the farther and nearer stations in (A.6). To check the effects of such slight differences on the estimated phase speed models and the station correction terms discussed in chapter 4, we applied both equations to our joint inversions using both phase and amplitude data.

Fig. A.1 displays phase speed maps and receiver amplification terms derived from the two types of the equation of amplitude anomaly caused by focusing between two stations. As we can see in these maps, both phase speed models and station amplification terms exhibits nearly the same patterns of heterogeneity, and the differences between them are rather minor. These results indicate that both formulations (A.5) and (A.6) can be acceptable to be used in the phase speed mapping with inter-station amplitude anomaly.

The typical values of phase speed perturbation $\delta c/2c_0$ of our model are less than ± 0.02 , that is, less than one tenth of the value of local amplifications. Thus, there are little visible differences between these amplification maps. Still, it should be noted that the amplification term in Fig A.1(b) reflects solely the effect of receiver amplification, while that in Fig A.1(a) is likely to reflect the mixed information with local phase speed perturbation. In this study, we prefer to use the approximate linear relation (A.6) based on the formulation of amplitude anomaly by Dahlen and Tromp (1998), so that the results of receiver amplification terms may be comparable with those of some recent studies (e.g. Lin et al., 2012; Eddy and Ekström, 2014).

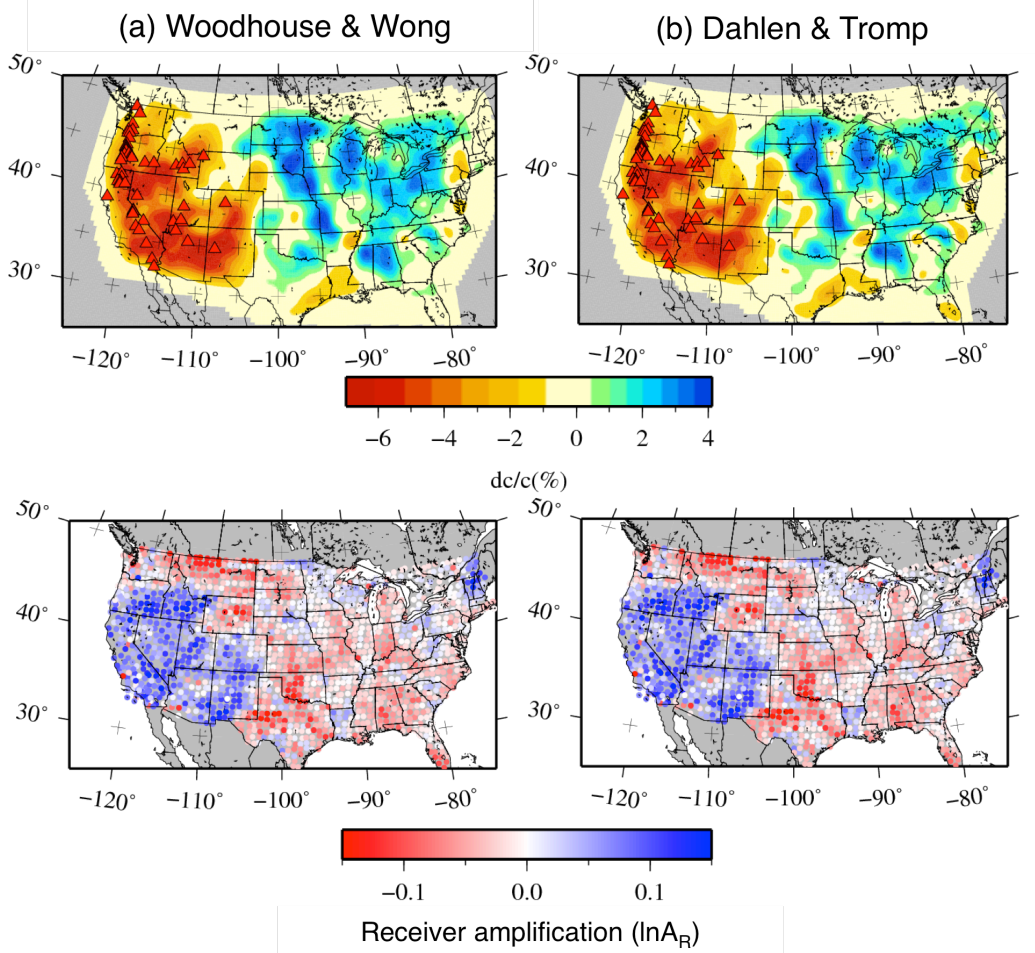


Figure A.1: Rayleigh-wave phase speed maps derived from joint inversions of phase and amplitude data (top), and receiver amplification terms (bottom) at 80 s, based on the linear equation of amplitude anomaly by (a) Woodhouse and Wong (1986), and (b) Dahlen and Tromp (1998).

Appendix B

Phase speed maps gallery

In this appendix, we display Rayleigh-wave phase speed maps derived from joint inversions of phase and amplitude data for the weighting factor $p = 0.8, 0.4$, and 0.0 , respectively.

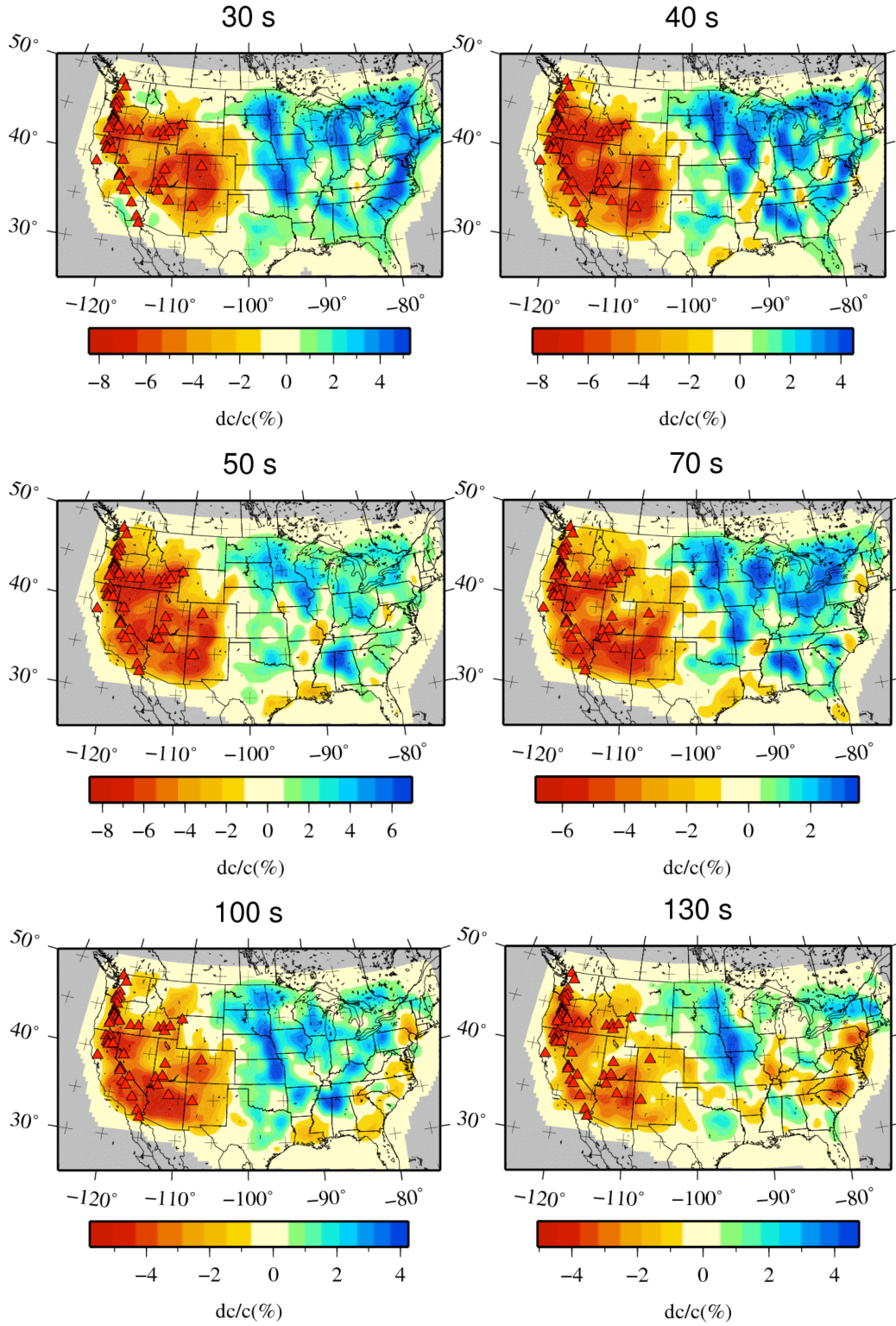
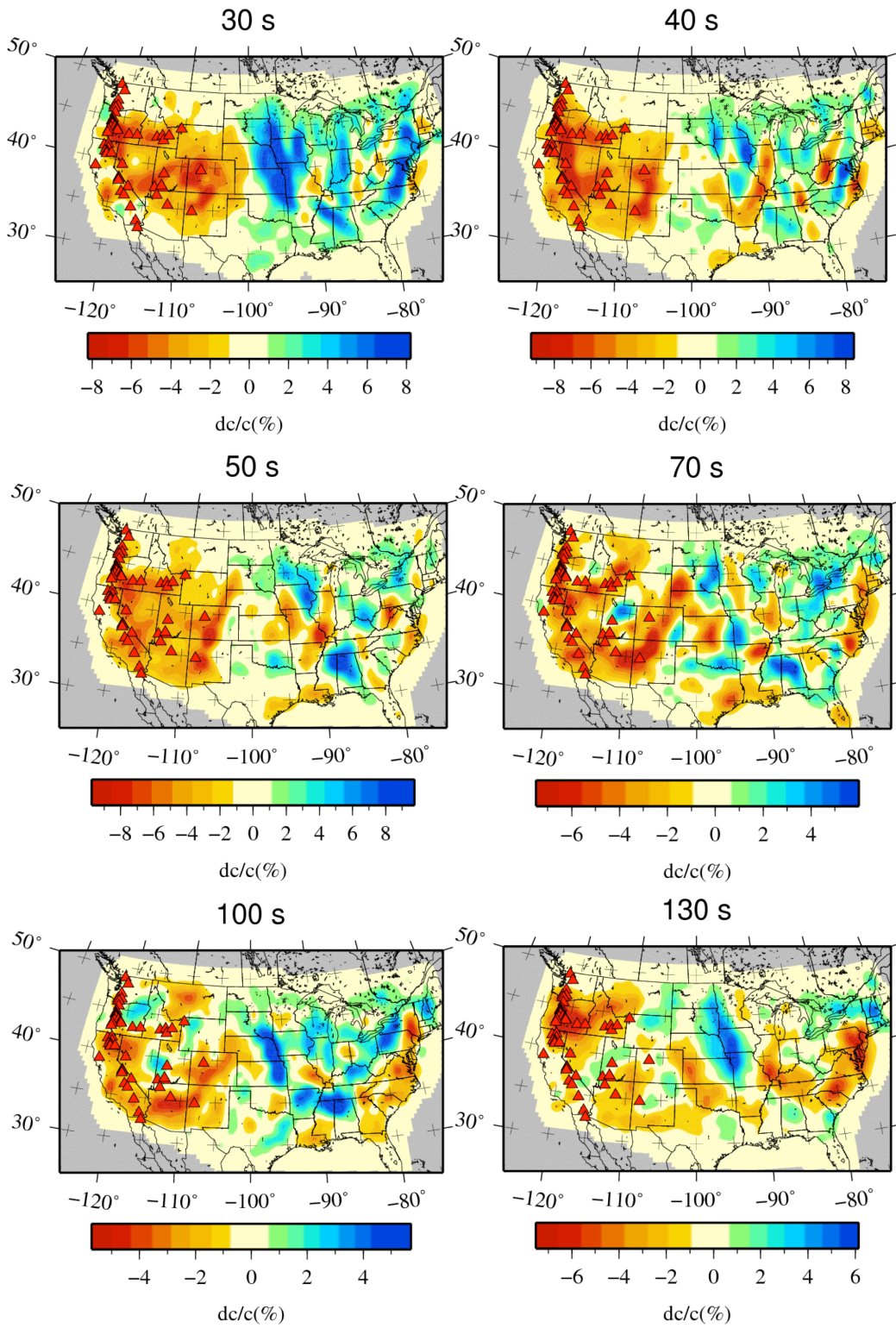
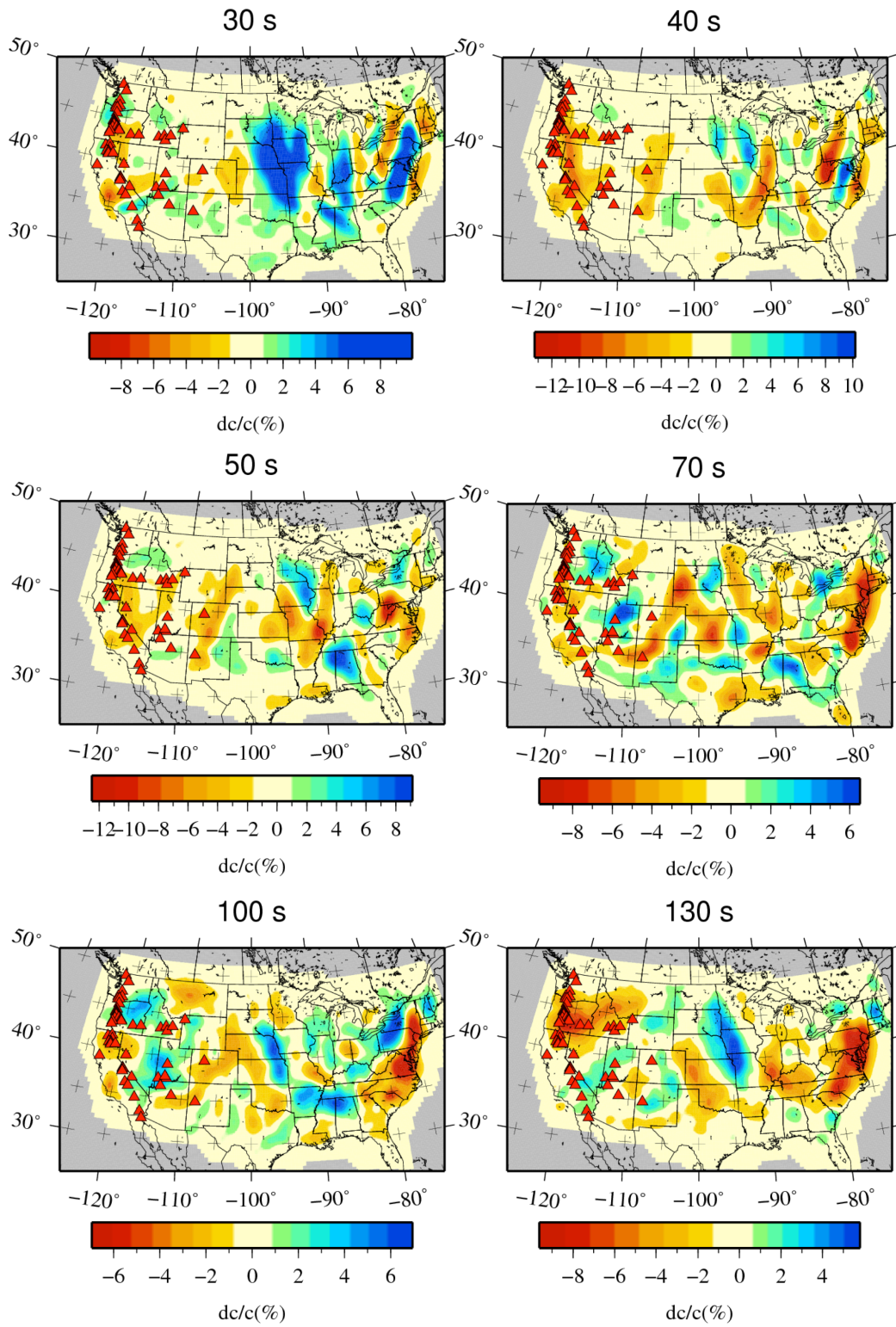


Figure B.1: Rayleigh-wave phase speed maps from joint inversions for $p = 0.8$. We used these maps for constructing 3-D S wave models in this study.

Figure B.2: Same as Fig. B.1, but for $p = 0.4$.

Figure B.3: Same as Fig. B.1, but for $p = 0.0$ (amplitude only).

Appendix C

3-D shear wave speed maps gallery

The isotropic 3-D S wave speed model and corresponding cross sections are displayed in this appendix.

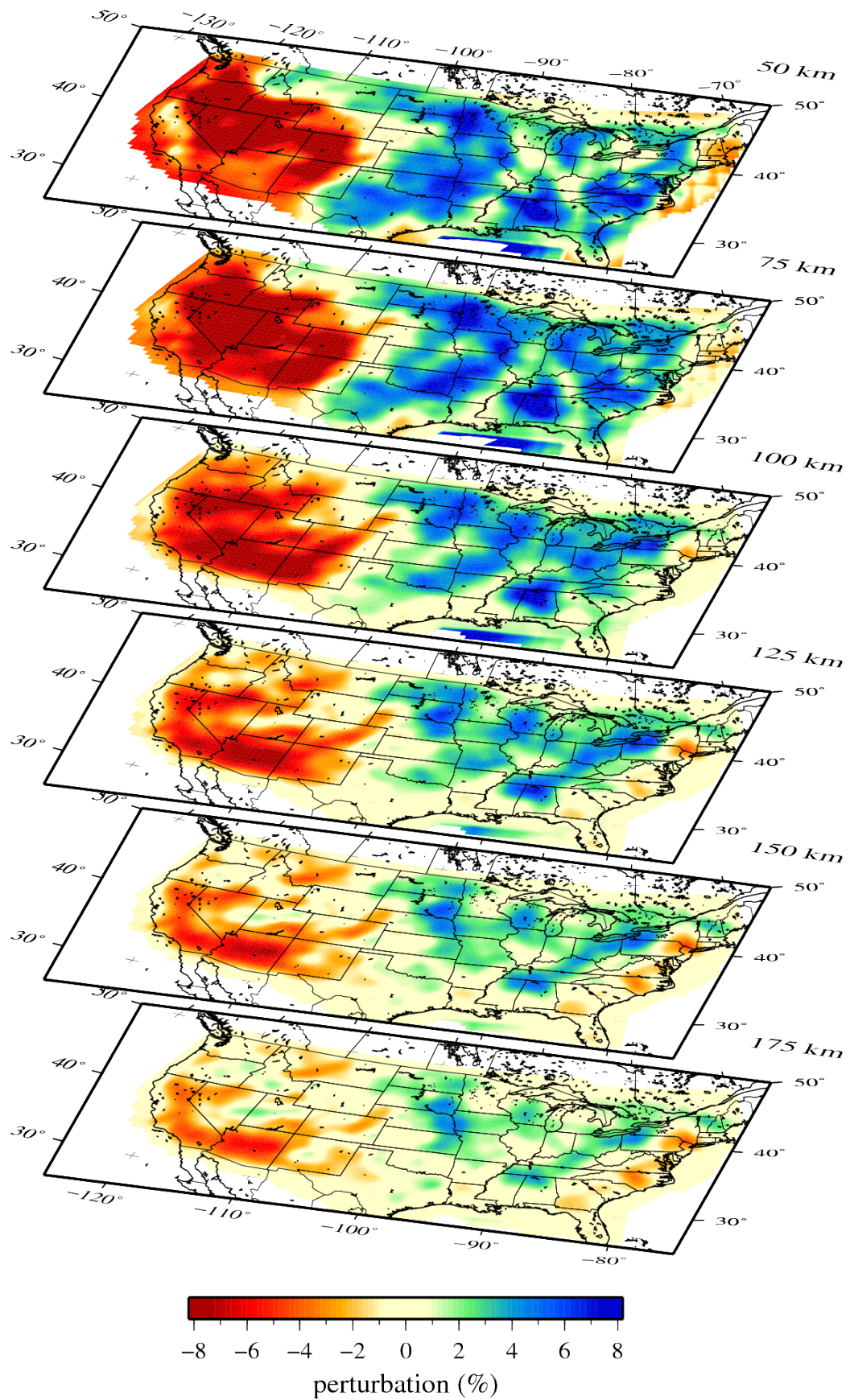


Figure C.1: Isotropic S wave speed model derived from both phase and amplitude data.

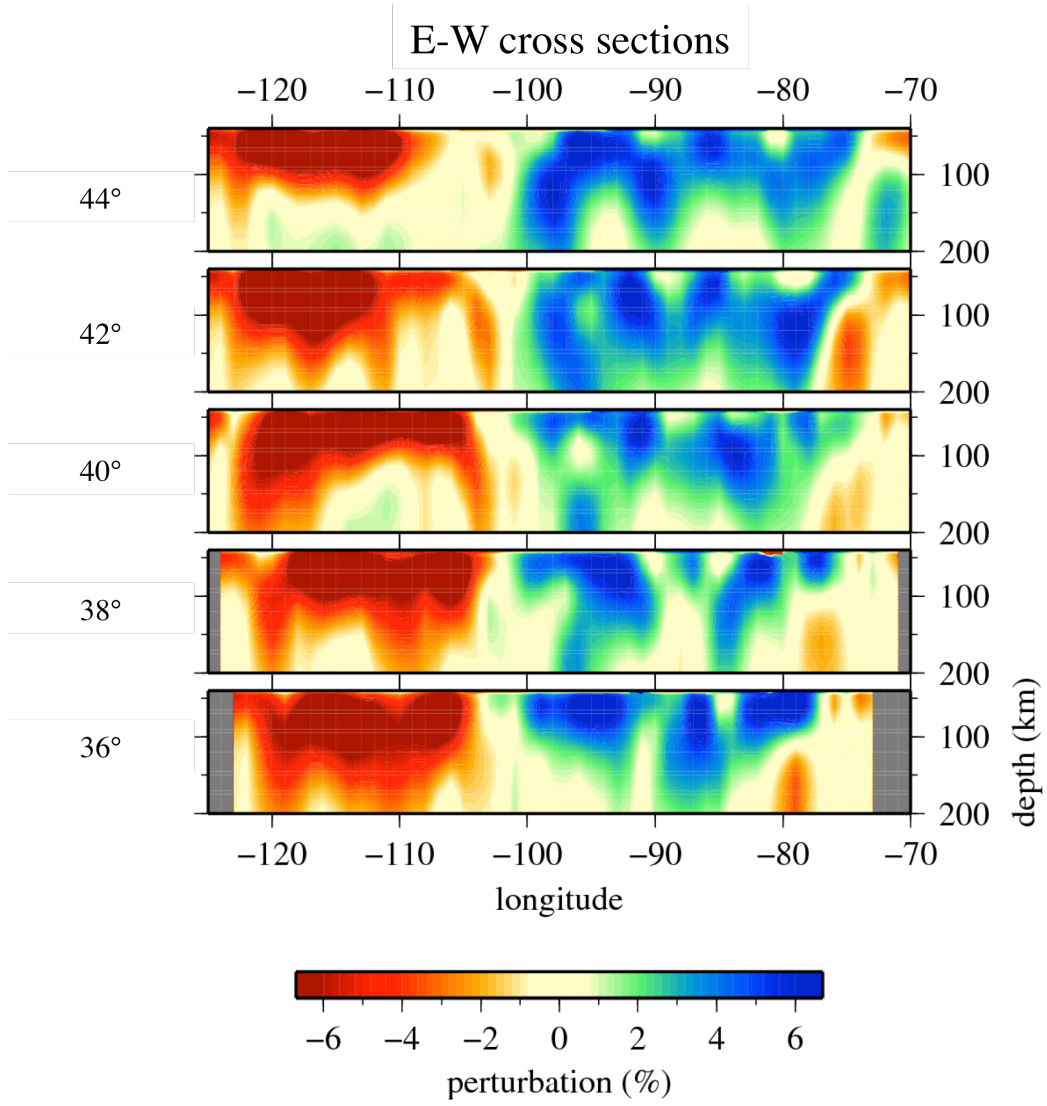


Figure C.2: East-west cross sections of S wave speed model in Fig. C.1.

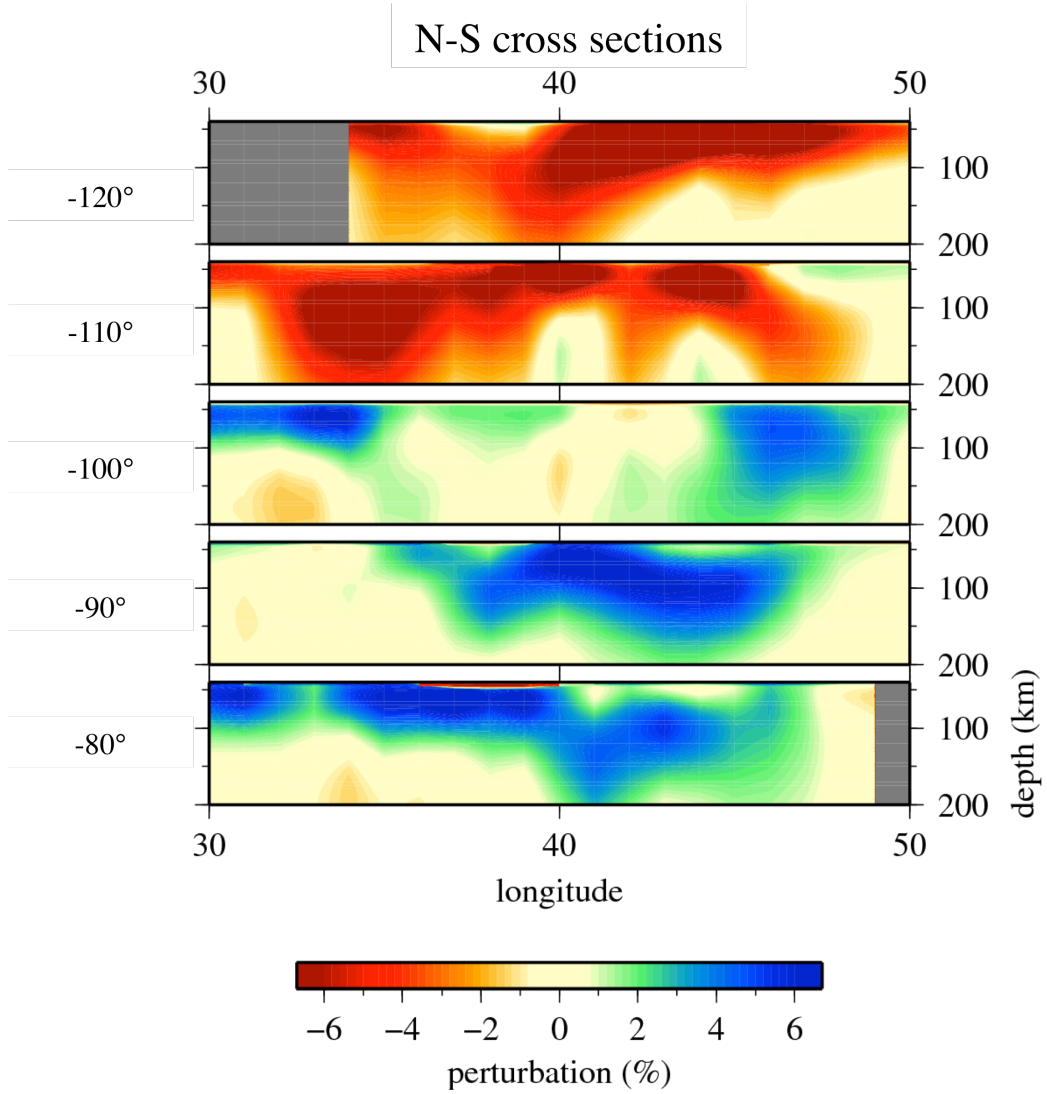


Figure C.3: North-south cross sections of S wave speed model in Fig. C.1.

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