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Upper Mantle Structure Beneath Mongolia with Surface Wave Tomography

(表面波トモグラフィーによるモンゴル地域の上部マントル構造)



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Abstract

Investigations of seismic structures of the upper mantle based on surface wave analysis can be useful for clarifying the regional tectonics and mantle dynamics, but such studies have still been limited in the Mongolian region. Although some local areas in Mongolia have been the targets of several earlier seismological studies, surface-wave tomography across the nationwide areas of Mongolia with the enhanced coverage of seismic stations has yet to be investigated.

Over the past decade, a number of seismic stations have been installed in Mongolia. Between 2011 and 2013, 69 temporary broadband seismic stations were installed in central-south Mongolia as a part of a collaborative project between China and Mongolia. From 2012 to 2016, several temporary broadband networks of IRIS-PASSCAL (Incorporated Research Institutions for Seismology – Program for the Array Seismic Studies of the Continental Lithosphere) were sporadically installed in western Mongolia. Since 2015, the nationwide regional broadband seismic network in Mongolia has also been in operation. These seismic networks have enabled us to carry out surface-wave dispersion measurements, which can be the basis for the construction of a 3-D S-wave model of the entire Mongolian region.

In this study, we employ the three-stage inversion technique of multimode surface waves. In the initial stage, we measured multimode phase speeds of Love and Rayleigh waves for a number of source-receiver pairs by using a fully nonlinear waveform fitting for individual seismograms based on the Neighborhood Algorithm. In the second stage, we retrieved multimode phase speed maps using the linearized inversions based on the method of surface-wave tomography incorporating approximate effects of finite frequency. In the third stage, localized multimode dispersion curves derived from the sets of multimode phase speed maps were used to construct local 1-D S-wave profiles, including radial anisotropy, which formed a final 3-D S-wave speed model beneath Eastern Eurasia including Mongolia.

The dataset utilized in this study consists of seismic waveforms from over 1655

teleseismic events ($M_w \geq 5.8$) that were recorded at permanent and temporary stations in and around Mongolia between 2009 and 2021. We reconstructed phase speed maps with $2.0^\circ \times 2.0^\circ$ grids (or greater size, depending on path coverage for each period) by covering the study region with many ray paths; e.g., about 70,000 paths for the fundamental-mode Rayleigh waves, 40,000 paths for the fundamental-mode Love waves at a period of 100 s, and about 5,000–40,000 paths for higher modes of both Rayleigh and Love waves. Broadband seismograms from the permanent and temporary seismic networks in Mongolia, as well as nearby stations of the Global Seismograph Network and temporary regional networks in Eurasia, were combined to retrieve a high-resolution S-wave model covering entire Mongolia. The new 3-D model gave us new insight into the large-scale tectonics of the Mongolian region, including the western margin of the Amurian plate.

The obtained 3-D S-wave model shows significant lateral variations of S-wave speeds in the uppermost mantle beneath Mongolia, indicating the structural differences between western and eastern Mongolia. In the radial anisotropy model, SH waves tend to be faster than SV waves in most areas of the Mongolian upper mantle above 100 km depth, except for the northeast of the Altay Mountains. The low-velocity anomalies are seen in western Mongolia beneath the Hangay Dome and the northern Altay Mountains. The low-velocity anomaly in the Altay Mountains persists to a depth of about 100 km, which may be linked to the northern extension of the collision tectonics between Asia and India. A large-scale low velocity beneath the northeast of the Hangay Dome persists to around 150 km depth, accompanied by significant radial anisotropy with slower SV velocity than SH in the top 100 km that may result from finely layered or laminated heterogeneity, possibly including partially molten layers. High-velocity anomalies corresponding to the lithosphere can be seen in the uppermost mantle of eastern Mongolia, reflecting the relatively stable tectonics in this region, in contrast to the tectonically active regions of central-western Mongolia. The differential motion between eastern and western Mongolia may be related to such regional variations of the upper mantle structure and dynamics beneath the Eurasian and Amurian plates.

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Chapter 1

Introduction

Seismic waves provide us with various information on the Earth's interior. Large-scale three-dimensional structures of the Earth's interior have been investigated through tomographic studies using seismic waves. In particular, seismic surface waves are the major tools for the detailed seismic imaging of the Earth's upper mantle, the main field where plate tectonics takes place. Unlike body waves, which propagate throughout the Earth's deep interior, including the lower mantle and core, surface waves mainly sample the Earth's crust and upper mantle, which provides valuable information on the lithosphere (or tectonic plates) and the asthenospheric mantle underneath. While the body-wave receiver functions can be used to detect the seismic discontinuities in the crust and upper mantle, surface waves are essential for reconstructing the three-dimensional anisotropic S-wave velocity model.

Many surface wave studies have been based on the analysis of fundamental mode, whose analysis is relatively straightforward. Still, the higher-mode surface waves are more sensitive to the deep structure of the Earth's mantle than the fundamental-mode surface waves. Thus, the use of higher modes can be of great help in improving the vertical resolution of surface-wave tomography models. In this study, we employ the method of multimode surface wave tomography to construct a radially anisotropic 3-D S-wave model in and around Mongolia as part of a large-scale tomography model of the Eurasian region. In this introductory chapter, we first briefly review the tectonic backgrounds and some earlier seismological studies in the Mongolian region. Then, the main purpose of this study and the structure of this thesis will be summarized.

1.1 Tectonic backgrounds in and around Mongolia

Mongolia is located inside the East Asian continent, bounded by the Siberian craton in the north, affected by the India-Asia collision zone in the south and probably by the far-field subduction zone of the Pacific plate in the east (Figure 1.1). The Central Asian Orogenic Belt (CAOB), a long-lasting accretionary orogen that spans much of central and eastern Asia, includes Mongolia. The CAOB, extending from North China Cratons and the northern margins of Tarim Basin to the northern Baikal Rift region, is one of the biggest Paleozoic orogens from the Mesoproterozoic to Paleozoic, with the accretion of ophiolites, oceanic islands, accretionary wedges, island arcs, and microcontinental fragments (Windley et al., 2007).

Due to the collision of the Indian and Eurasian plates as well as the subduction of the Pacific and Philippine Sea plates beneath Japan, East Asia has experienced extensive tectonic deformations during the Cenozoic (Wei et al., 2012). India has been moving steadily northward since 55 million years ago, raising the Himalayas and Tibetan Plateau, and controlling tectonics in western China and Mongolia (Molnar & Tapponnier, 1975; Hao et al., 2019). Nonetheless, the subductions of the Pacific and Philippine Sea plates have been significant contributors to the tectonic stress in Asia, producing island arcs, marginal seas, and continental rifting (Hall, 2002; Huang & Zhao, 2006; Yin, 2010; Hao et al., 2019).

The creation of continental rifts and active intraplate volcanoes in Northeast China has also been suggested by tomographic images that may relate the dynamic processes of the Western Pacific and East Asia with the subducted Pacific plate stagnated at the mantle transition zone beneath East China (Huang & Zhao, 2006). There are several intra-plate volcanoes in Mongolia, the Korean Peninsula, and China (Global Volcanism Program, 2023). Figure 1.1 displays the locations of Taryatu Chulutu (TC), Khanuy Gol (KG), Bus Obo (BO), Middle Gobi (MG), Dari Ovoo (DO), and Dariganga (DG) volcanic eruption sites from the Late Cenozoic. According to Hunt et al. (2012), Mongolia's volcanism comprises several alkali-basalt cones that have erupted since 30 million years ago. Even though the greatest lava flows occur in the eastern Dariganga, the majority of modern volcanism is located near the Hangay Dome and the south of

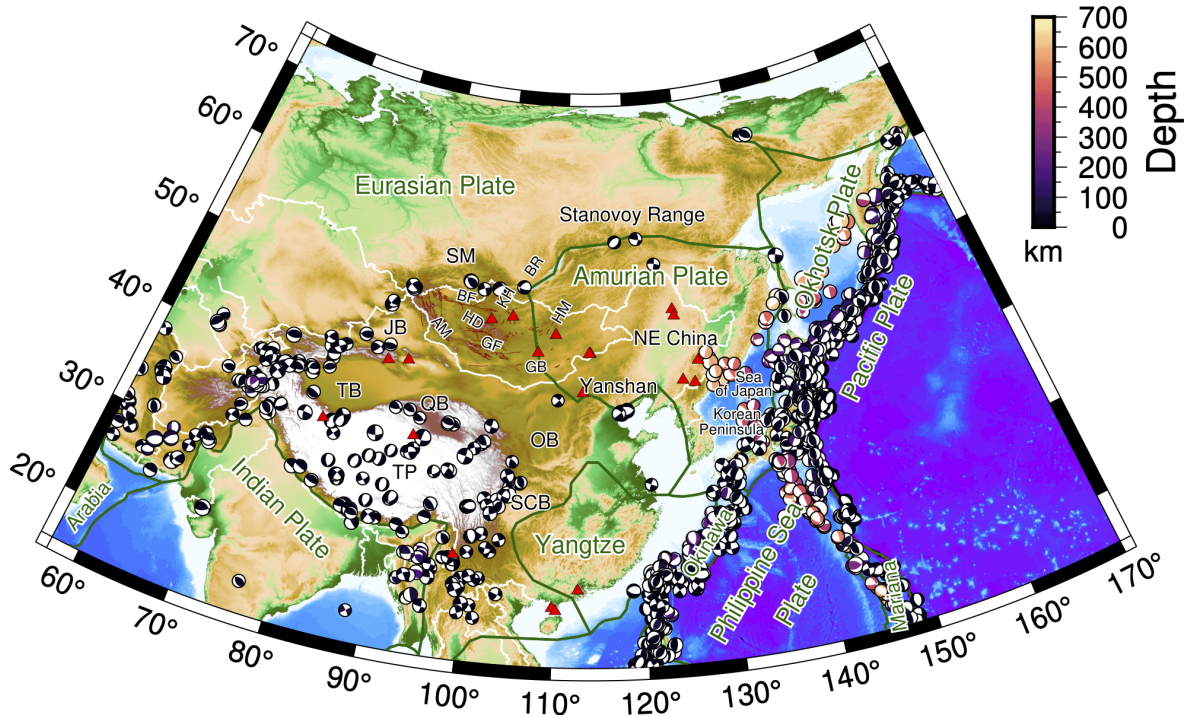


Figure 1.1: Topography and seismicity in the Eastern Eurasian region around Mongolia. Focal mechanisms ($M_w \geq 6.0$) are taken from the Global CMT catalog between 1976 and 2022 (Ekström et al., 2012). Dark green lines indicate plate boundaries by Bird (2003). Triangles represent Holocene volcanoes (Global Volcanism Program, 2023). Abbreviations from west to east: TB = Tarim Basin, JB = Junggar Basin, TP = Tibetan Plateau, QB = Qaidam Basin, AM = Altay Mountains, SM = Sayan Mountains, BF = Bulnay Fault, HD = Hangay Dome, GF = Gobi Altay Fault, KH = Khovsgol, BR = Baikal Rift, GB = Gobi Basin, SCB = Sichuan Basin, HM = Hentey Mountains, OB = Ordos Block.

the Baikal Rift.

A number of Precambrian micro-continents, island/continental arcs, accretionary complexes, and ophiolites make up the Mongolian region (Windley et al., 2007; Sengör et al., 1993; Badarch et al., 2002; Wilhem et al., 2012) (Figure 1.2). Generally, the geological structure of Mongolia is separated into the northern and southern domains by the Main Mongolian Lineament; i.e., an Early Paleozoic domain in the north and a Late Paleozoic domain in the south (Badarch et al., 2002; Wilhem et al., 2012). Cratonic pieces, Neoproterozoic ophiolites, Precambrian and Early Paleozoic metamorphic rocks, Early Paleozoic basins with clastic rock, island arcs, and related volcanoclastic deposits are the elements that make up the northern domain (Badarch et al., 2002). The history of the Paleo-Asian Ocean has been connected to Paleozoic island arcs, ophiolite

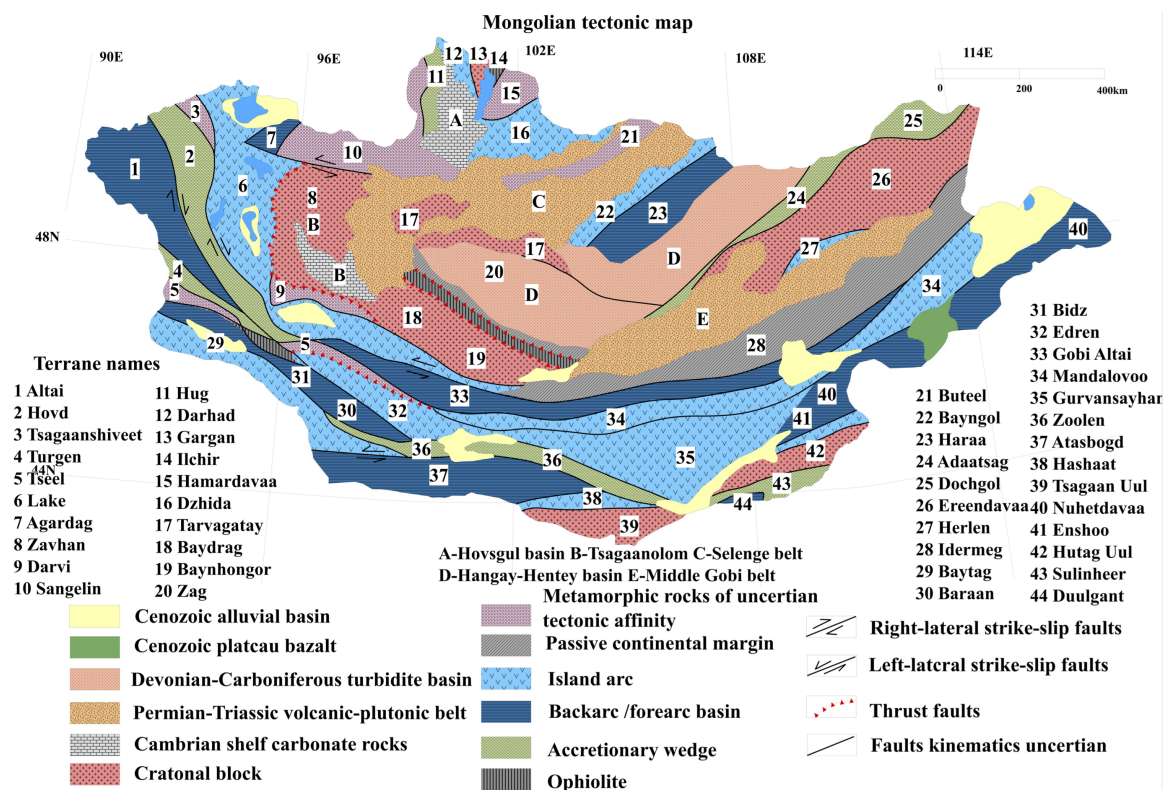


Figure 1.2: Tectonic map of Mongolia adapted from Badarch et al. (2002) with permission from Elsevier.

fragments, and late Carboniferous to Permian volcanic rocks found in the southern domain (Lamb & Badarch, 2001; Kozakov et al., 2007).

East Asia is indeed a very complex deformation zone that contains many micro-plates (Zonenshain & Savostin, 1981) and multiple potentially hard blocks (Molnar & Tapponnier, 1975). Precisely determining the geometry and border of sub-plates in these locations is challenging due to low seismicity and the absence of a distinct geographical limit, except for the Baikal rift zone and the Kuril-Japan trenches. The Mongolian region encompasses the enigmatic western boundary of the Amurian plate, whose border is not quite visible at the surface. Also, the uplifting of the Hangay Dome in Middle Mongolia is one of the major targets of tectonic studies in this research area. In the subsequent sections, two major tectonic features characterizing the large-scale Mongolian tectonics, the Amurian plate and the Hangay Dome, are reviewed in more detail, followed by the recent seismological studies in Mongolia.

1.2 Amurian Plate

Geological, seismological, and geodetic data have been used in various studies on East Asian plate tectonics to investigate the Amurian plate boundary. The Amurian plate is in northeastern Asia, covering northeast China, the Korean Peninsula, the Sea of Japan, the eastern shore of Lake Baikal in southeast Russia, and eastern Mongolia. Western Mongolia belongs to the Eurasian plate, whereas Eastern Mongolia belongs to the Amurian plate in the global plate model by Bird (2003) (Figure 1.1).

The Amurian plate, moving toward the southeast, was postulated as distinct from the Eurasian plate. The border between these two plates is thought to be confirmed by the Baikal Rift Zone. Zonenshain & Savostin (1981) investigated the rotation pole of the Eurasian and Amurian plates using focal mechanism solutions, and the angular rate estimated from the Baikal opening was about 1×10^{-7} degrees/year. Wei & Seno (1998) calculated the motion of the Amurian plate relative to the neighboring main plates using the slip vectors of the earthquakes along these boundaries as well as the NUVEL-1 data (DeMets et al., 1990) along the boundaries of North American-Pacific plates and Eurasian-North American plates. Based on a four-year continuous GPS study, Takahashi et al. (1999) proposed that the eastward velocity vectors recorded in

the western half of the Sea of Japan and south Sakhalin indicate the Amurian plate motion. The Amurian plate in the global plate model by Bird (2003) is predominantly grounded in the result of Wei & Seno (1998), as geodetic observations have limitations in precisely defining plate borders, unlike topography and seismicity.

According to the majority of current plate models, eastern Mongolia is considered to be a part of the Amurian plate, whereas western Mongolia is a part of the Eurasian plate. Nonetheless, a numerical plate motion model by Petit et al. (1996) and Petit & Fournier (2005) employed the Amurian plate model encompassing nearly the whole of Mongolia. They collected data from GPS and focal mechanisms in and around the Amurian plate to describe the characteristics of its borders as well as its relative speed in relation to neighboring plates. Then, using constant plate rheology, thermal state, crust and mantle thicknesses, Petit & Fournier (2005) used the finite-element code SHELLS to simulate the Amurian plate motion. Ashurkov et al. (2011) used five consecutive GPS observations on the geodynamic test site (Amurian region) between 2001 and 2007 to determine the velocity field of horizontal crustal movements. The rotation properties of the blocks were used to compute the Baikal rift opening rates. In the southern part of Lake Baikal, the velocity is 2.3 mm/yr, while at its northernmost point, it is 1.25 mm/yr. The velocity drastically drops further northeast along the Baikal rift system's terrestrial basins. Additionally, their computations have demonstrated that the Amurian Plate's eastern margin traverses a branch of the Tan-Lu fault system. Jin (2013) proposed the Amurian plate boundary using GPS data (1998-2005) from around 85 continuous and over 1000 analyzed campaign sites in East Asia (China, South Korea, and Japan), suggesting about 1.8 mm/year of extension between the South Korean and Amurian blocks, and about 3.0 mm/year of spreading at the Baikal Rift between the Eurasian and Amurian plates.

Based on the studies of active fault kinematics and the paleostress condition of the Earth's crust in three regions — the Burgut block, the Selenga block, and the Hangay-Hentey tectonic saddle — Sankov et al. (2022) postulated the western boundary of the Amurian plate as claimed by Zonenshain & Savostin (1981) and subsequently endorsed by other publications. They suggested that the western boundary of the Amurian plate inside the Mongolian territory represents the margin of the deformation zone covering

entire Western Mongolia. The eastern margin of the Amurian plate boundary is more consistent in many earlier studies, compared to other borders. Still, the existence of the Amurian plate and its geometrical border remain controversial due to the limited geophysical information.

1.3 Hangay Dome

The Hangay Dome is located in western Mongolia, characterized by a high topography, comparatively flat plateau, and widespread Quaternary basalts. The Hangay Dome area serves as a vital kinematic connection between the actively deforming Indo-Eurasian collision and the Cenozoic extensional deformation in the Baikal regions.

The N-S compression by the convergence of the Indian plate and Tibetan Plateau affects tectonically active western Mongolia, including the Hangay Dome, while the back-arc extension by the subduction of the Pacific plate might influence E-W oriented tectonics in relatively stable eastern Mongolia (Watson et al., 1987; Zheng et al., 2000; Meng et al., 2003).

GPS measurements in western Mongolia showed that northward shortening decreases from south (10 mm/year) to north (4 mm/year) in the Altay region, but central and eastern Mongolia moves eastward/southeastward at 4 mm/year (Calais et al., 2003). This transition area dominates the Hangay Dome region.

The Hangay topography is often regarded as a relatively recent formation that was elevated and formed within the past 20 – 30 million years (Walker et al., 2007), and the greatest active stage from 3 – 4 Ma until now (Barry & Kent, 1998). However, McDannell et al. (2018) indicated that this mountain environment is very old and has endured for almost 100 million years based on the geochemical and geological evidence. From geochemical and petrological investigations, Erdenesaihan et al. (2013) discussed the tectonic formation history of the Hangay region, which is considered an accretionary orogen formed by the closure of the Hangay-Hentey paleo-ocean during the early Paleozoic through early Mesozoic.

Two seismically active strike-slip faults, the Bulnay Fault and the Gobi-Altay Fault, which generated gigantic intracontinental earthquakes ($M \geq 8.0$ in 1905 and 1957, respectively), are situated respectively north and south of the Hangay Dome (Walker

et al., 2007). Distant influences of the India-Asia convergence, Pacific plate subduction, mantle plume activity, convective mantle flow, and magmatic underplating have all been implicated in the mountainous terrain of the Mongolian Plateau (Molnar & Tapponnier, 1975; Windley & Allen, 1993; Cunningham, 2001; Petit et al., 2002; Bayasgalan et al., 2005; Zorin et al., 2006; Petit et al., 2008). Thus, the Hangay Dome is one of Mongolia's most complex and enigmatic tectonic features that remains debated.

Previous geophysical studies present controversial models which can be categorized into two major hypotheses. One supposes high topography and intraplate volcanism are formed due to the deep-rooted mantle plume (Windley & Allen, 1993; Zorin et al., 2006; Wu & Bao, 2023). The regional heat flow measurements with 60–70 mW/m² in the Hentey Mountains and Hangay Dome (Khutorskoy & Yarmoluk, 1989), lithospheric thinning (Zorin et al., 1990), and late Cenozoic alkaline volcanism (Hunt et al., 2012) supported the mantle plume hypothesis. Also, Chen et al. (2015a) suggested a low-velocity zone could be caused by hot fertilized peridotite detached from the stagnant Pacific slab as it falls into the lower mantle (Zorin et al., 2006) based on a large-scale tomographic image in East Asia. Wu & Bao (2023) reported the thinning of the mantle transition zone beneath central Mongolia based on a receiver function study, indicating the evidence of intracontinental mantle upwelling from the lower to the upper mantle, which may cause the Cenozoic uplift and basaltic volcanism beneath the Hangay Dome.

Another hypothetical model is a lithospheric model that can be explained as the dispersed locations of the potential melting zones (Cunningham, 2001; Barry et al., 2003). This scenario is consistent with a strong negative gravity model (Petit et al., 2002), the result of magnetotelluric inversions (Comeau et al., 2018), thermo-mechanical numerical modeling (Comeau et al., 2021), and recent tomography models (Feng, 2021; Wu et al., 2021; Wang et al., 2022).

1.4 Tomographic studies in the Mongolian region

Mongolia and the neighboring regions have been investigated using various seismic tomography as part of the broader regional models covering eastern and central Asia (e.g., Tiberi et al., 2008; Bao et al., 2013; Chen et al., 2015b,a; He et al., 2016; Tao et al., 2018; Wu et al., 2021; Zhao et al., 2021; Witek et al., 2021; Feng, 2021; Huang & Zhao,

2022; Wang et al., 2022; Li et al., 2022; Wu & Bao, 2023). Radial anisotropy in eastern and central Asia including Mongolia has also been investigated through large-scale tomographic studies (e.g., Chen et al., 2015a,b; Tao et al., 2018; Witek et al., 2021), which generally indicate the positive radial anisotropy ($SH > SV$) at shallow depths, except for regions such as Tibet, Japan, and Northeast China, where negative radial anisotropy ($SH < SV$) can be observed.

Recent local-scale studies in western and central Mongolia, especially in the Hangay Dome, indicated a thinner lithosphere (60–80 km) than the surrounding areas, while the crust was found to be as thick as about 50–55 km (Petit et al., 2008; Feng, 2021; Zhao et al., 2021). Low-velocity anomalies were seen in the upper mantle at around 70–150 km depths (Tiberi et al., 2008; Chen et al., 2015a; Comeau et al., 2018; Wu et al., 2021; Feng, 2021; Wang et al., 2022; Huang & Zhao, 2022).

In contrast to western Mongolia, eastern Mongolia has been less investigated, mainly due to relatively poor seismic station coverage. A receiver function result of He et al. (2016) showed the crustal thickness varies between 38 and 46 km, and Rayleigh-wave phase speed maps by Yu et al. (2015) and Pan et al. (2015) exhibited high-velocity anomalies in Southern Gobi and Hangay-Hentey basin while low-velocity anomalies in Middle Gobi. Using travel-time tomography, Zhang et al. (2017) found slow P and S velocities under the Hentey Mountains and areas close to the comparatively low-lying Gobi Desert. Rayleigh-wave tomography by Wang et al. (2022) suggested that consistently faster upper mantle velocities may indicate the strong lithosphere beneath the East Gobi Desert.

1.5 Purpose of this study

Although various seismological studies have been done in and near Mongolia, the nationwide regional 3-D upper mantle image from surface waves beneath the whole areas of Mongolia, incorporating the enhanced station coverage, has yet to be obtained. This study aims to construct a new 3-D upper-mantle S-wave speed model, including radial anisotropy, beneath Eastern Eurasia around Mongolia using multimode surface waves with enhanced station coverage by employing a large number of permanent and temporary broadband seismic stations inside Mongolia. Such a 3-D model with improved

lateral and vertical resolutions will enable us to clarify the relationship between the present Mongolian tectonics and mantle dynamics.

In this study, we employ the three-stage inversion technique by Yoshizawa & Kennett (2004), which has been utilized to construct many regional-scale multimode surface wave tomography models in both continental and oceanic areas in the world (e.g., Yoshizawa, 2014; Isse et al., 2019; Tang et al., 2022). The 3-D distributions of SV and SH wave speeds are obtained from dispersion data of multimode Love and Rayleigh waves recorded at the permanent and temporary Mongolian networks and neighboring stations of Global Seismograph Network (GSN) and temporary networks in the Eurasian continent (e.g., IRIS-PASSCAL networks and NECESSArray). The new 3-D model exhibits clear lateral structural variations between the Eurasian and Amurian plates, indicating new seismological evidence of the enigmatic western boundary of the Amurian plate.

Chapter 2 of this thesis explains our dataset of broadband seismic waveforms and multimode phase speed measurements based on the method of nonlinear waveform fitting (Yoshizawa & Kennett, 2002b; Yoshizawa & Ekström, 2010). Using a large data set of path-specific multimode phase speeds derived from the dense seismic networks, we construct high-resolution phase speed maps in Chapter 3 for the fundamental and up to the fourth higher modes of Rayleigh and Love waves in the target region. In Chapter 4, we create radially anisotropic 3-D S wave speed models in the Eastern Eurasian region around Mongolia using these multimode phase speed maps. The structural features and their tectonic and geodynamic interpretations are discussed in Chapter 5. In Chapter 6, we summarize this thesis and provide some recommendations for further research.

Chapter 2

Data and Method of Multimode Phase Speed Measurements

The number of permanent and temporary seismic stations installed in Mongolia has increased recently, which enables us to investigate the detailed seismic structure in the Mongolian upper mantle. In this section, we will review the current status of seismic networks in Mongolia, which are used to extract the multimode phase speeds for global seismic events. The method of data analysis and results of dispersion measurements are also summarized in this chapter.

2.1 Seismic observations in Mongolia

Since the first digital seismic station was installed in 1994, Mongolia has gradually increased the number of digital short-period seismic stations. To enhance the monitoring capability of seismic activity and expand seismological research in the Mongolian region, many short-period seismometers have been replaced by high-sensitivity broadband seismometers from 2013 to 2015. Seismic stations in Mongolia have increased yearly, and currently, 28 permanent broadband seismic stations (Figure 2.1 (b) blue triangle) are operated in the country, along with many mini-arrays and five digital regional three-component short-period seismometers. These sixteen broadband stations were outfitted with Gralp Inc.'s CMG-3ESPC (60 s and 120 s) broadband seismometers, which have a sensitivity of 2000 V/m/s in the Ulaanbaatar area. A data logger CMG-DM24S3EAM (Gralp Inc.) is connected to each sensor. All stations continuously record seismic signals at a 50 Hz sampling frequency.

From August 2011 till July 2013, 69 temporary seismic networks of broadband sta-

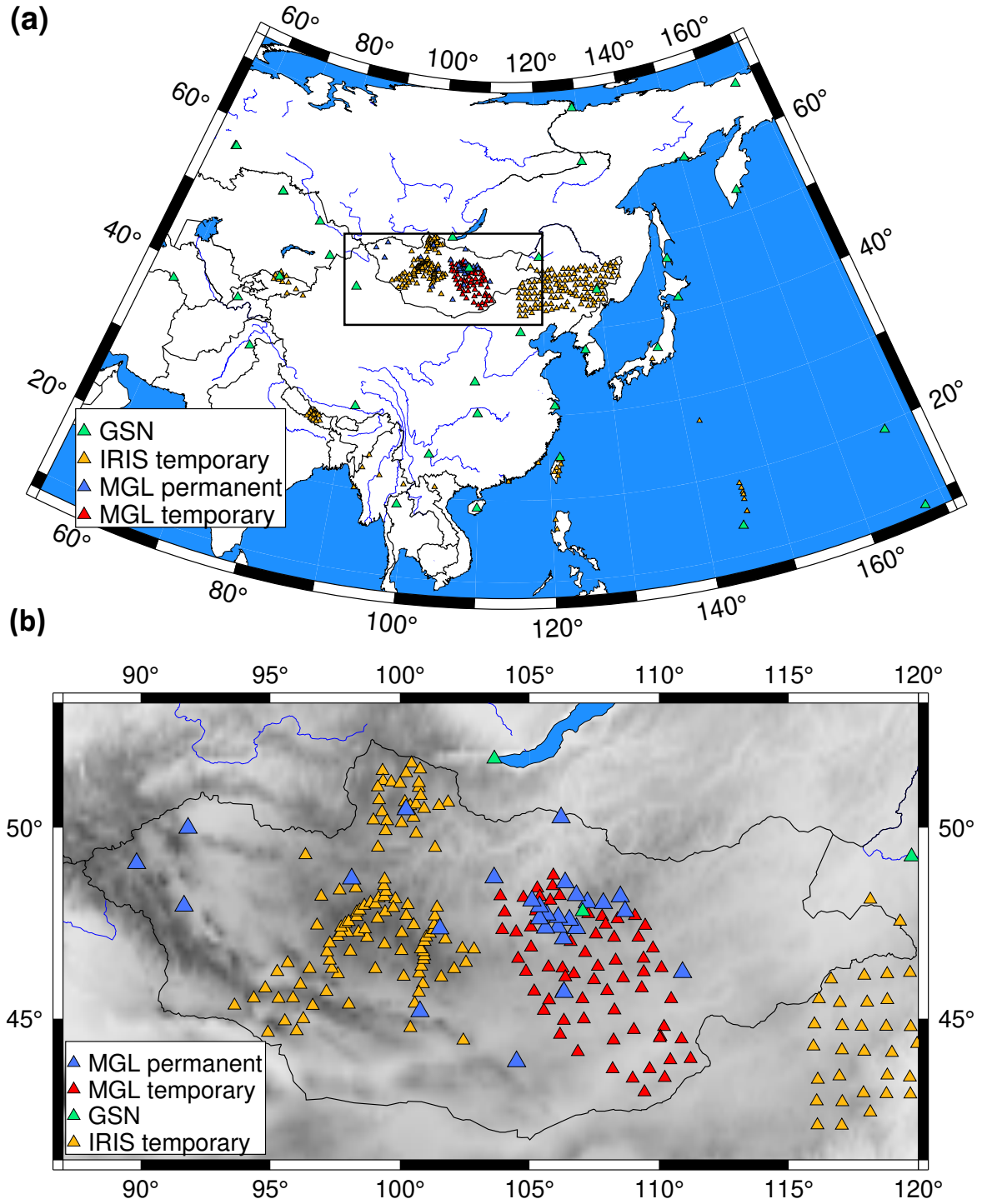


Figure 2.1: Seismic stations used in this study. (a) All stations, (b) stations in Mongolia.

Stations	Operating time														Area
	2009.8	2010	2011.8	2012.6	2013.7	2014		2015	2016.3	2017	2018	2019	2020	2021	
						7	8								
GSN network															57°E–171°E, 15°N–75°N
PASSCAL NECESSArray															115°E–134°E, 41°N–49°N
PASSCAL HD*.XL															95°E–104°E, 44°N–50°N
PASSCAL AT*.XL															93°E–97°E, 44°N–47°N
PASSCAL HV*.XL															98°E–102°E, 49°N–52°N
Mongolian permanent station															88°E–122°E, 44°N–52°N
Mongolian temporary (CM network)															102°E–113°E, 42°N–50°N

Table 2.1: Operation period of the seismic networks used in this study.

tions were installed in the central-southeast Mongolian region (103.5°E–111.5°E, 43°N–49°N) as a part of the China-Mongolian cooperation project. CMG-3ESP seismometers are used at these sites (Figure 2.1 (b) red triangles). Also, through the Central Mongolian Seismic Experiments (Meltzer et al., 2019), temporary broadband seismic arrays using the IRIS-PASSCAL instruments were installed in western Mongolia, e.g., in the Gobi-Altay area (14 stations) and Khovsgol area (26 stations) from August 2014 to March 2016, as well as in the Hangay Dome region (72 stations) from June 2012 to July 2014 (Figure 2.1 (b) yellow triangles). These temporary broadband seismic stations used the STS-2 seismometers and Quanterra Q330 digitizers with a 100 Hz sampling frequency.

Seismic waveform data from the regional seismic stations are invaluable for facilitating seismological research and seismic hazard assessments in Mongolia and understanding the current tectonics and mantle dynamics in the Mongolian region. By incorporating local data from temporary and permanent broadband seismic stations installed and operated inside Mongolia for about a decade, the detailed seismic velocity structure of the whole region is expected to be unraveled.

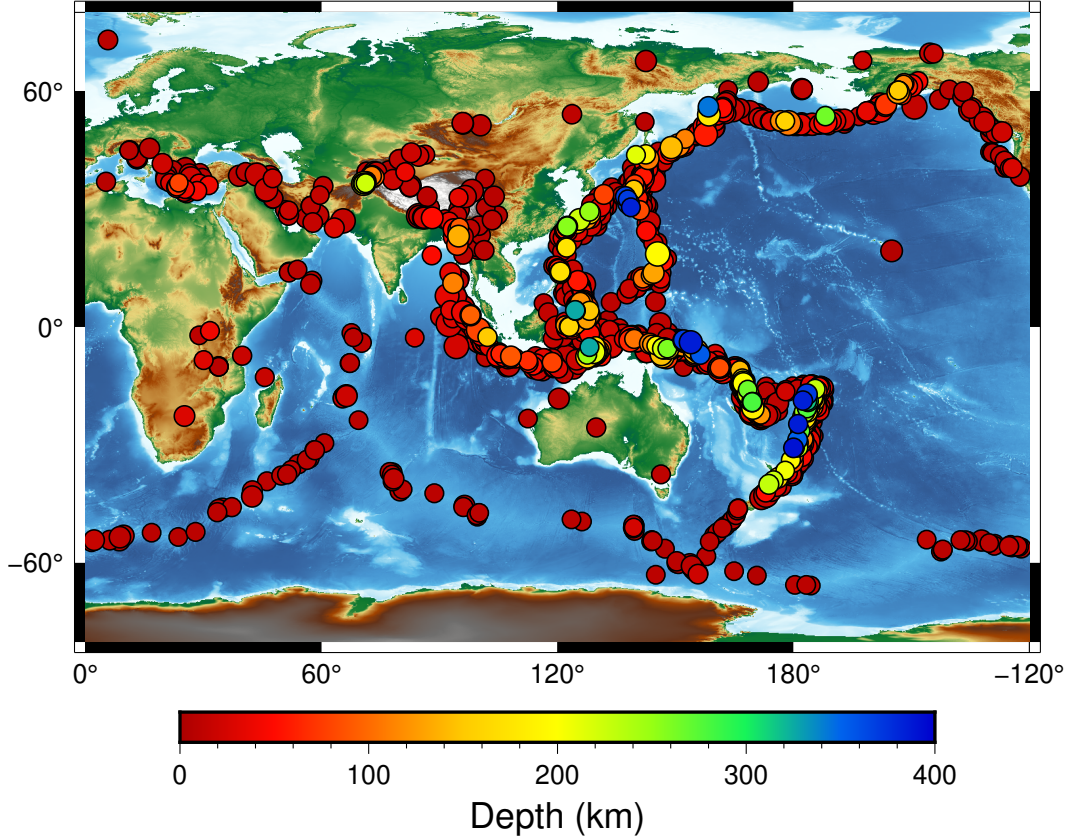


Figure 2.2: Seismic events with moment magnitude 5.8 or greater from 2009 to 2021 used for the multimode phase speed measurements.

2.2 Seismic stations and events for dispersion measurements

We employed three-component broadband seismograms of the permanent and temporary Mongolian networks, Global Seismograph Network (GSN), Central Mongolian Seismic Experiments (Meltzer et al., 2019) (PASSCAL network), and temporary networks in Northeast China (NECESSArray). In total, 425 broadband seismic stations (Figure 2.1) are used for the dispersion measurements. Table 2.1 summarizes the operation period of the seismic networks used in this study. We have analyzed 1655 teleseismic events with a moment magnitude 5.8 or greater, located at an epicentral distance longer than 1000 km (Figure 2.2). At first, all original waveforms are deconvolved with corresponding instrument responses. Then, the horizontal components are rotated to the radial and transverse components. Using the vertical and transverse components independently, path-average multimode phase speeds of Rayleigh and Love waves are measured for each event-station path across the target area using

the single-station waveform fitting method described in the next subsection.

Table 2.2: Time-window setting and multiple band-pass filters employed to fit synthetic and observed waveforms, following Yoshizawa & Ekström (2010). The begin and end times of each window are specified by group speeds and/or the theoretical travel times of S or SS wave (T_S) for the ak135 model (Kennett et al., 1995).

window	Rayleigh Wave		Love Wave	
	group speed (km/s)	freq. range (mHz)	group speed (km/s)	freq. range (mHz)
Win1	2.9 – 4.8	5 – 10	$3.3 - T_S$	5 – 10
		7 – 12		7 – 12
Win2	3.2 – 4.5	10 – 18	$3.5 - T_S$	10 – 20
		15 – 25		15 – 30
Win3	$4.3 - T_S$	20 – 33	$4.3 - T_S$	20 – 33
		25 – 50		25 – 50
Win4	$2.9 - T_S$	5 – 12		
		10 – 20		
		15 – 30		

2.3 Multimode phase speed measurements

Using global seismic events and many seismic stations in and around the Mongolian region, we measure the multimode phase speeds of Rayleigh and Love waves. This is the initial step for the three-stage surface wave tomography (Kennett & Yoshizawa, 2002; Yoshizawa & Kennett, 2004). To extract path-average multimode phase speeds for each event-station pair, we performed fully nonlinear waveform inversion for path-specific 1-D S-wave profiles by fitting observed and synthetic waveforms based on global optimization approach (Yoshizawa & Kennett, 2002b; Yoshizawa & Ekström, 2010). The details of the automated measurement technique are explained by Yoshizawa & Ekström (2010). Appendix A.1 concisely explains this procedure with some fundamental formulations. Here, we summarize the essence of the measurement technique briefly with examples of actual waveform fitting results for Rayleigh and Love waves (Figures 2.3 and 2.4).

Table 2.2 summarizes the multiple time-window settings and band-pass filter ranges employed for the waveform fitting. Figure 2.3 displays an example of the Rayleigh-wave case for an event near Mindanao, Philippines (Ms 7.0, 80 km depth) observed at the

U05M station in Mongolia close to the capital city with an epicentral distance of about 5100 km. Figure 2.4 displays another example of the Love-wave case for an event near the Kuril Islands (M_s 7.2, 110 km depth) observed at the HD01 station in Mongolia with an epicentral distance of about 3600 km.

For each event-station path, more than 5000 path-specific 1-D S wave profiles are sampled using the Neighborhood Algorithms by Sambridge (1999), which is used as a global optimizer. The surface-wave WKBJ method (e.g., Dahlen & Tromp, 1998) is used to compute synthetic waveforms for all models by employing the focal mechanism from the global CMT catalog (Ekström et al., 2012). Multiple time windows are selected automatically for the fundamental and higher modes based on the predefined group-speed ranges (Table 2.2). Waveform misfits between the synthetic and observed seismograms are assessed in these time windows with multiple band-pass filters in the frequency range from 5 to 30 mHz for the fundamental mode and up to 50 mHz for the higher modes (Figures 2.3 (d) and 2.4 (d), Table 2.2). Then, the best 1-D profile, which provides the minimum waveform misfit, is selected (the red solid line in Figures 2.3 (c) and 2.4 (c)) to compute the path-average multimode phase speeds for each path (Figures 2.3 (f) and 2.4 (f)). To evaluate the quality of measured phase speeds, we employ the frequency-dependent reliability parameter (Figures 2.3 (e) and 2.4 (e)) computed from the relative power of each mode and relative waveform fit as defined by Yoshizawa & Ekström (2010). In this study, we employed the same empirically determined constraints and thresholds as Yoshizawa & Ekström (2010) (Table A.1) for selecting reliable dispersion measurements and outlier removal. The reliable phase speeds used in the subsequent tomographic inversions are chosen by considering multiple quantitative information, including the waveform misfit values representing the entire waveform matching, the reliability parameter, and the relative radiation amplitude of each mode at the seismic source as summarized in Table A.1.

In general, the higher modes are overlapped in a seismogram due to similar group speeds of multiple modes over a broad frequency range, making it challenging to extract the individual higher-mode phase speeds without mode separation (e.g., van Heijst & Woodhouse, 1997; Matsuzawa & Yoshizawa, 2019; Brown et al., 2022). However, the full waveform fitting procedure based on fully nonlinear inversion with global optimiza-

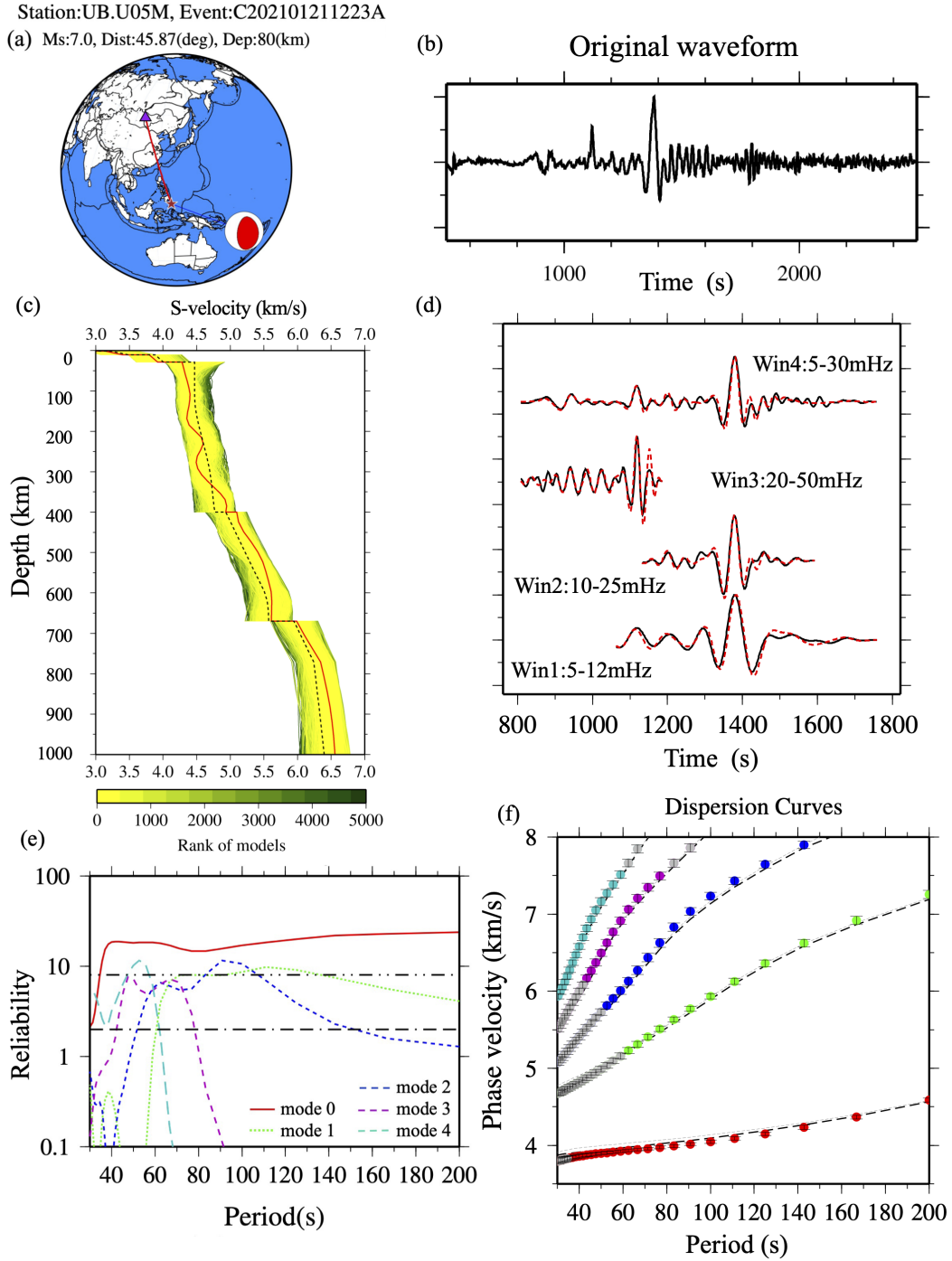


Figure 2.3: An example of the nonlinear waveform inversion for multimode Rayleigh waves in the vertical component at station U05M of the Mongolian seismic network. (a) Locations of the earthquake (red star) with Ms 7.0 at 80 km depth on January 21, 2021, and the U05M station (purple triangle) in Mongolia. (b) An original observed seismogram at U05M. (continue)

Figure 2.3: (continued.) (c) Path-specific 1-D S-wave profiles for the path in (a). The 5050 models evaluated during the inversion are sorted in order of decreasing misfit and shown with colors from green (greater misfit) to yellow (lower misfit). The red line is the best 1-D profile with the lowest waveform misfit. (d) Best-fit waveforms between the observed (solid black lines) and synthetic (red dashed lines) seismograms for 4-time windows defined in Table 2.2. (e) The reliability parameters for the dispersion curves as a function of the period for each mode. In this example, the threshold values for the reliability parameters are set to 8.0 for the fundamental mode and 2.0 for the higher modes. (f) The estimated dispersion curves of the fundamental and up to 4th higher modes are calculated from the best-fit 1-D profile. The colored dots with error bars represent the selected frequencies of dispersion data whose reliability parameter is higher than the threshold value shown in (e), while gray dots with error bars indicate less reliable measurements, which are excluded in subsequent phase speed mapping. The error bar for each measurement is estimated from the standard deviation of the best 1500 models.

tion used in this study allows us to compute the multimode phase speeds using the path-specific 1-D profile that provides the best waveform fit between the synthetic and observed seismograms, including the higher-mode time window. Although we do not consider the mode separation during the waveform fitting process, we must carefully consider the practically available frequency range for each mode, because the relative amplitudes of overlapped higher modes depend significantly on the depth of seismic sources. For the quantitative evaluation of the extracted multimode phase speeds, we employ the reliability parameters (Figures 2.3 (e) and 2.4 (e)), which reflect the quality of individual measurements. Since the reliability parameter reflects the effects of the relative excitation power of each mode at the seismic source, the frequency ranges with similar group speeds, which result in severe overlap of different modes in an observed seismogram, tend to exhibit lower reliability and be automatically discarded. This fact is well reflected in Figures 2.3 (e,f) and 2.4 (e,f)) and 2.5.

2.4 Data sets of multimode phase speeds

Our phase speed dataset derived from the waveform fitting consists of about 70,000 paths for the Rayleigh-wave fundamental mode, around 40,000 paths for the Love-wave fundamental mode at a period of 100 s, and about 5,000 – 40,000 paths for higher modes of both Rayleigh and Love waves (Figure 2.5 a, b). Figure 2.5 (c, d) shows the

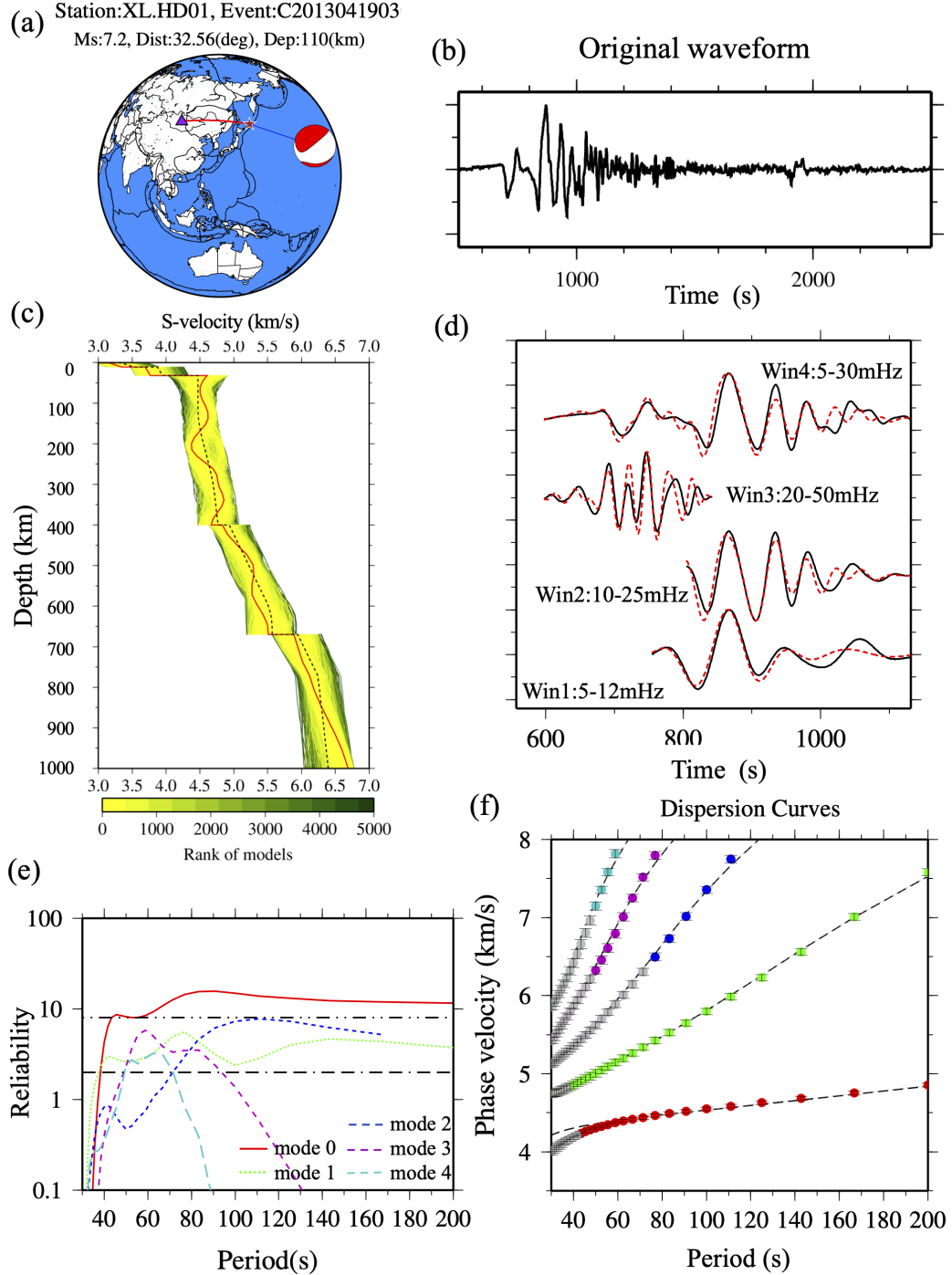


Figure 2.4: Same as Figure 2.3, but for Love waves in the transverse component at the HD01 station in Mongolia, for an event near the Kuril Islands with Ms 7.2 at 110 km depth on April 19, 2013.

average phase speeds with corresponding standard deviations from 30 to 200 s, which are computed from all measurements for each mode and discrete frequency. Reference phase speed dispersion curves for the PREM model (Dziewonski & Anderson, 1981) are also plotted as dashed black lines in Figure 2.5 (c, d). We could collect reliable phase speeds for many paths with reasonable levels of standard deviations around the average phase speeds.

The ray path coverage and ray density maps for the fundamental and higher modes at the selected periods of Rayleigh and Love waves are displayed in Figures 2.6 – 2.8. The entire study region is covered well with the dense ray paths owing to many broadband seismic stations deployed in and around Mongolia. We could obtain high ray-path density with good azimuthal coverage, particularly in East Asia and Mongolia. However, the ray density in the western marginal region of Mongolia is somewhat limited compared to central and eastern Mongolia. As described in the subsequent section, we reconstructed phase speed maps for each mode and frequency, for which we could gather more than 5000 paths. These multimode phase speed maps are employed to compute a 3-D shear wave model of the continental lithosphere and asthenosphere beneath Eastern Eurasia around Mongolia.

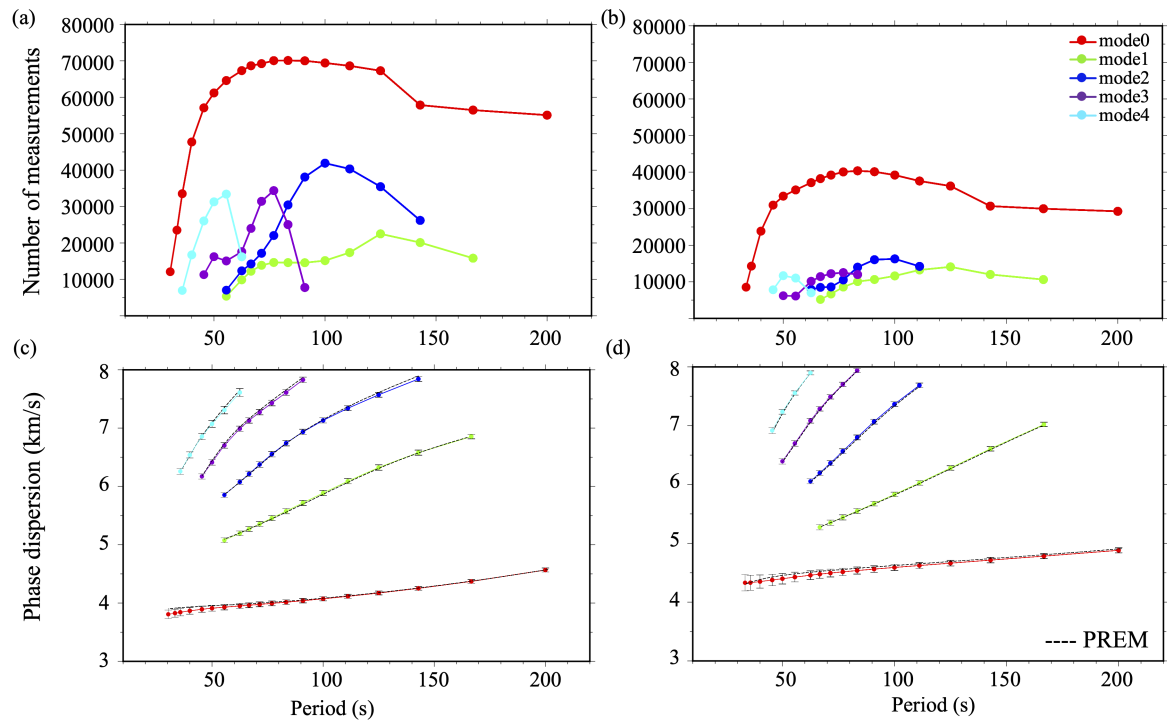


Figure 2.5: Numbers of measured multimode phase speeds for each period: (a) Rayleigh wave, and (b) Love wave. The average phase speeds from all the measurements for each period for (c) Rayleigh wave and (d) Love wave with corresponding standard deviations. Different colors represent the dispersion curves of different modes (fundamental mode and up to the 4-th higher modes).

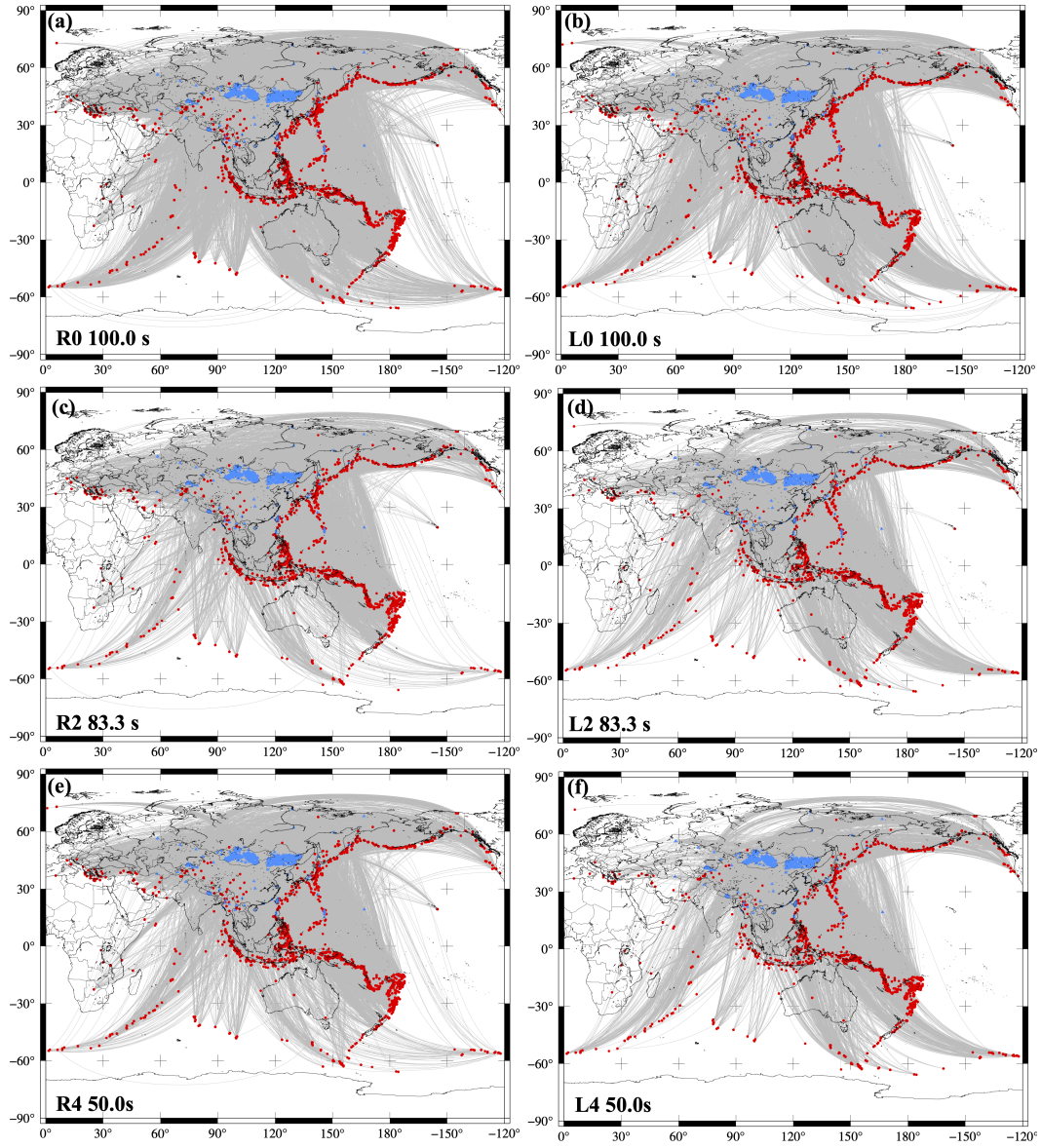


Figure 2.6: The ray-path coverage of the selected periods of the fundamental and higher modes of Rayleigh and Love waves.

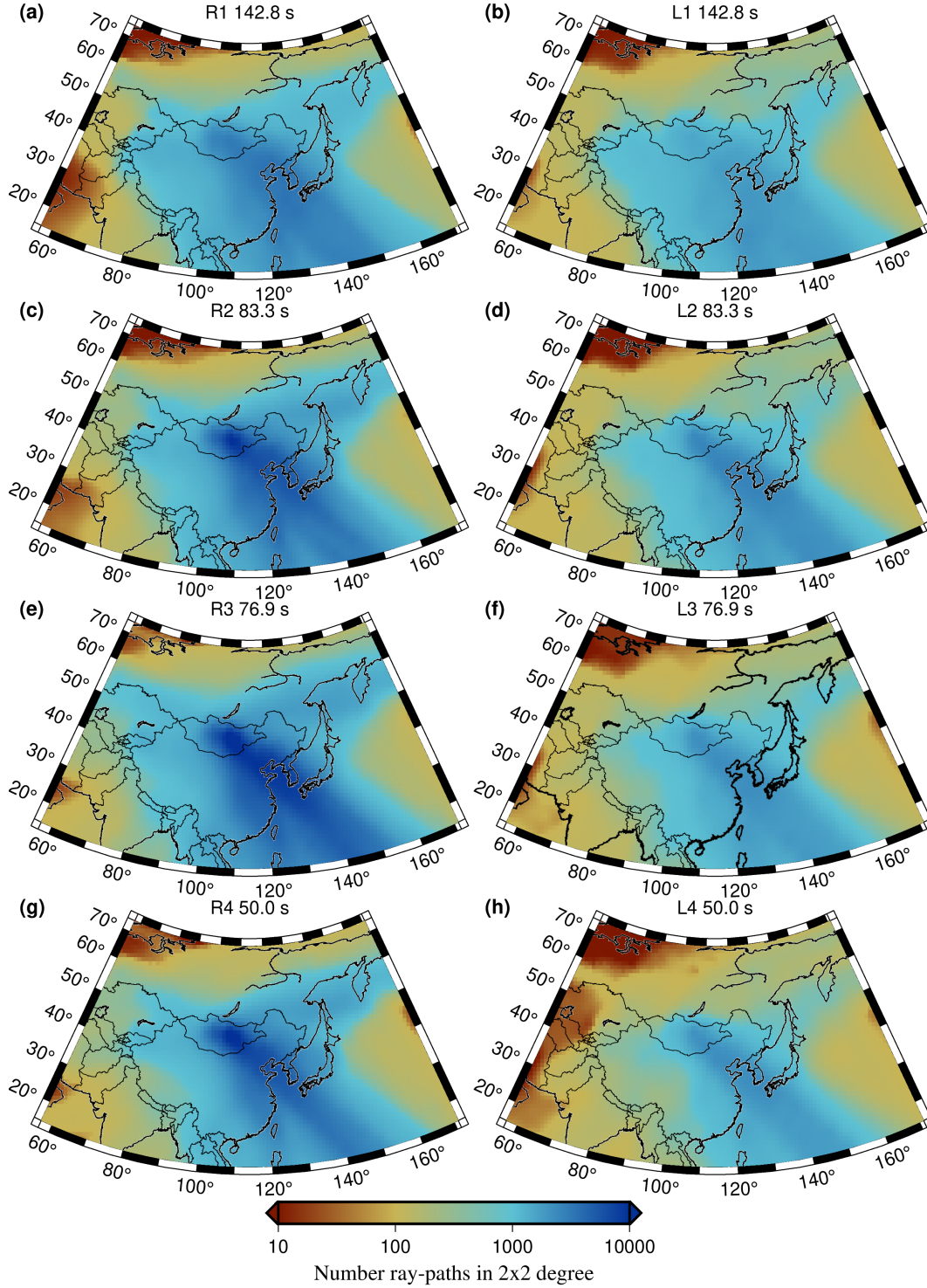


Figure 2.7: Ray-path density maps for the fundamental-mode Rayleigh and Love wave at 100 s. The numbers of ray paths are counted in $2^\circ \times 2^\circ$ grids.

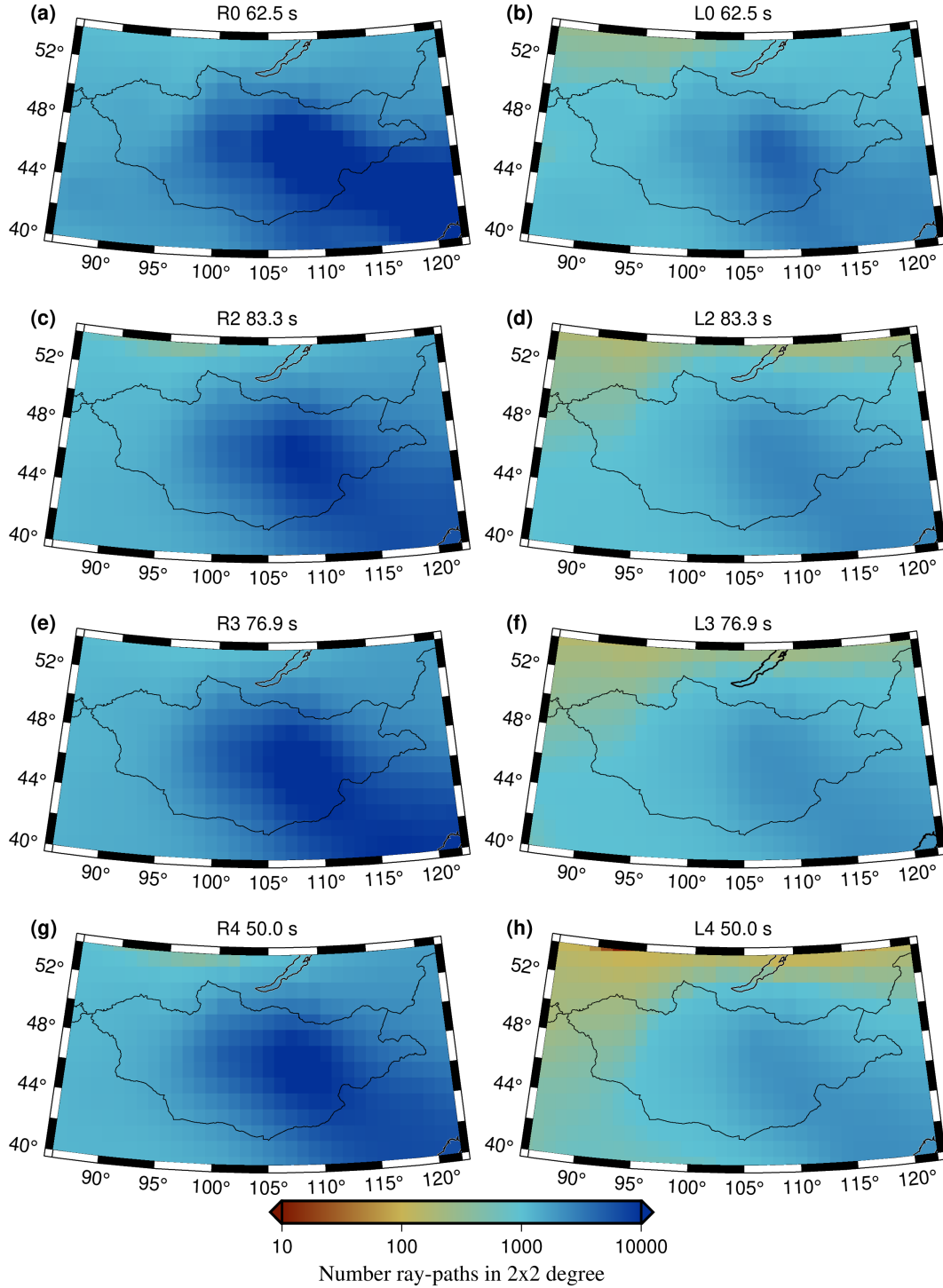


Figure 2.8: The numbers of ray-paths in $2^\circ \times 2^\circ$ grids for the selected periods of the fundamental and higher modes of Rayleigh and Love waves in the Mongolian region.

Chapter 3

Multimode Phase Speed Mapping

The measured multimode phase speeds from the previous chapter are utilized to retrieve 2-D phase speed maps for each mode and frequency. This process corresponds to the second stage of the three-stage surface wave tomography. In this chapter, we summarize the method and results of multimode phase speed maps, which provide intermediary structural information for the construction of the final 3-D S-wave model in the next chapter. Some results of the experiments to check the spatial resolution and model uncertainty are also discussed. We have also performed additional model recovery tests in Appendix B.2 to confirm the validity of the tomographic inversions.

3.1 Inversions for 2-D phase speed maps

We used linearized tomographic inversion considering approximate influences of finite-frequency for surface waves (Yoshizawa & Kennett, 2002a). At first, phase speed maps were obtained from the inversions using the great-circle path. These initial phase speed maps were then updated, incorporating the effects of ray-path bending due to lateral heterogeneity and the influence zone around the propagation path that allows us to account for the approximate finite-frequency effects of surface waves. The procedure of phase speed mapping has been explained by Yoshizawa & Kennett (2004) in detail. Appendix A.2 also briefly describes the methodology with formulations. Phase speed distribution is parameterized using spherical B-spline functions at rectangular grids. The grid spacing of this study depends on the ray-path coverage for each mode and frequency; i.e., 2.0° for the fundamental mode with more than 30,000 paths, 3.0° for more than 15,000 paths, 4.0° for more than 8,000 paths, and 5.0° for all modes with

more than 5000 paths. We also considered additional criteria for the minimum grid size for each mode and frequency so that the grid size must not be smaller than half of the wavelength. We solve the linear inverse problem with a damped least-squares method based on the LSQR algorithm (Paige & Saunders, 1982), using the reliability parameters for individual phase speed measurements as a weight on each path.

3.2 Resolution test

We performed checkerboard tests to clarify the lateral resolution of our tomography model. The input checkerboard models include velocity fluctuations of around $\pm 5\%$ from the reference velocity. We employed various cell sizes to investigate the spatial resolution of phase speed maps.

We show the results of checkerboard tests for the fundamental-mode Rayleigh wave at 100 s and fourth higher-mode Rayleigh at 50 s in Figure 3.1, and those for Love wave in Figure 3.2. Figure 3.3 shows the results of checkerboard tests for the fundamental mode at 100 s and the fourth higher mode at 50 s of Rayleigh and Love, focusing on the Mongolian region. The retrieved checkerboard patterns of these figures suggest that our data set achieves a reasonably good lateral resolution of about 3.0° or greater for Rayleigh-wave phase speed maps beneath entire Mongolia owing to many crossing paths in each cell. The checkerboard results indicate that the southern and central parts of the study regions are particularly well resolved by denser ray coverage, which allows us to discuss even smaller-scale heterogeneity of about 3.0° . In contrast, the rest of the parts, including the western edge of Mongolia, show somewhat smeared patterns due to limited numbers of crossing paths for smaller-scale checkerboards. Since the numbers of available ray paths are smaller, the lateral resolution of Love waves is somewhat limited compared to that of Rayleigh waves, although the large-scale features of about 4° or greater can be mostly recovered well in the target region. Additional model recovery tests for phase speed maps are also summarized in Appendix B.2.

3.3 Jackknife resampling test

We performed jackknife resampling tests to estimate standard errors in our multimode phase speed maps following Panning & Romanowicz (2006) and Isse et al. (2019).

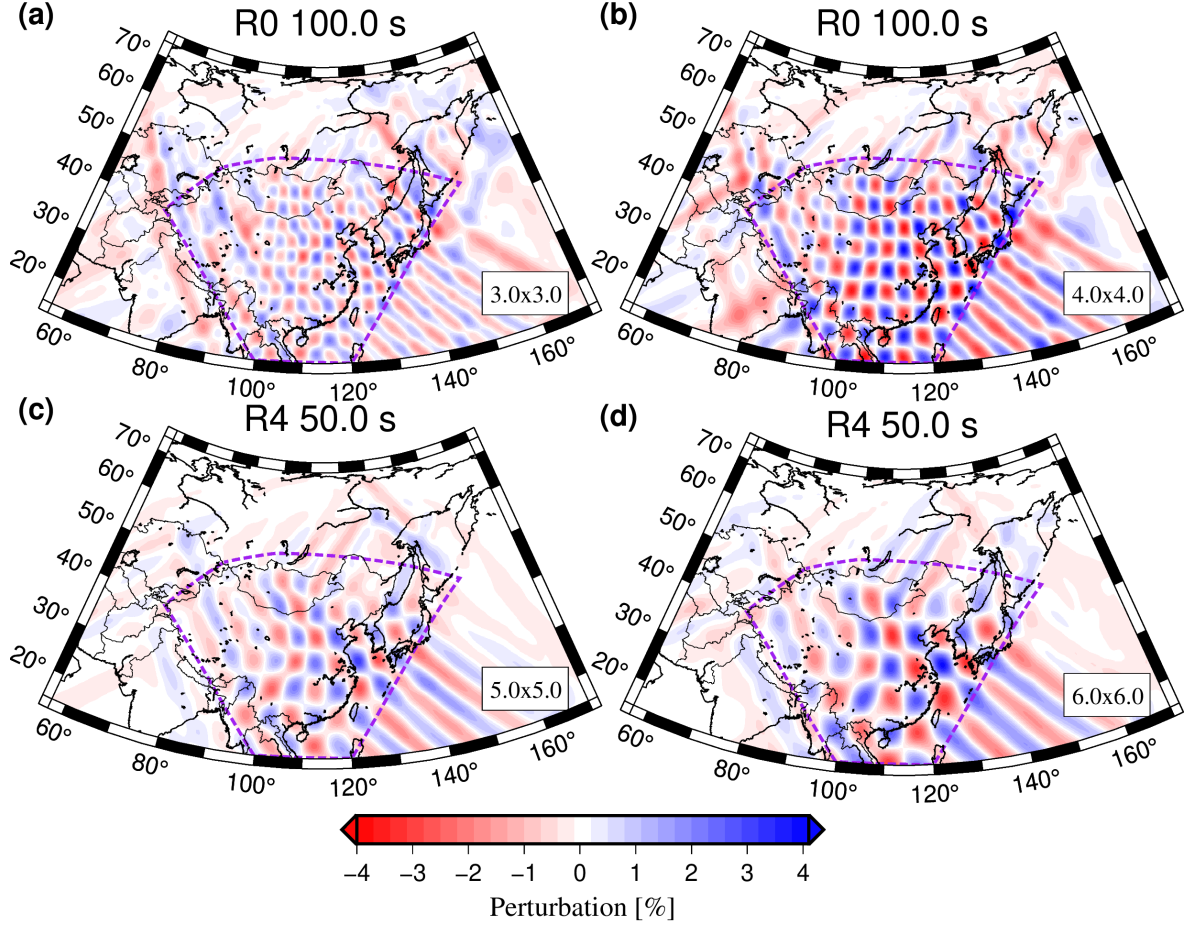


Figure 3.1: Results of checkerboard resolution tests for the fundamental and fourth higher modes of Rayleigh wave with $3.0^\circ - 6.0^\circ$ cells. Fundamental-mode Rayleigh (R0) wave at 100.0 s with (a) $3^\circ \times 3^\circ$ and (b) $4^\circ \times 4^\circ$ cell patterns, 4th higher-mode Rayleigh (R4) wave at 50.0 s (c) $5^\circ \times 5^\circ$ and (d) $6^\circ \times 6^\circ$ cell patterns. Purple dashed line indicates the well-resolved area.

Jackknife resampling was made by dividing the original phase speed data set into 10 subsets by random selection without overlap. Then, we performed 10 inversions by removing one subset to obtain 10 phase speed maps. The standard error at each location is estimated from the inverted models.

Figure 3.4 shows the results of the jackknife resampling test for the fundamental-mode and 2nd higher-mode of Rayleigh and Love waves at 100 s. The spatial average of estimated errors for the fundamental-mode Rayleigh wave (average phase speed: $c_{ave}^{R0} = 4.11$ km/s) and Love wave ($c_{ave}^{L0} = 4.61$ km/s) in the study area are about 0.03 km/s, while those for the 2nd higher-mode Rayleigh wave ($c_{ave}^{R2} = 7.18$ km/s) and Love wave ($c_{ave}^{L2} = 7.38$ km/s) are around 0.05 and 0.04 km/s, respectively. The larger errors are

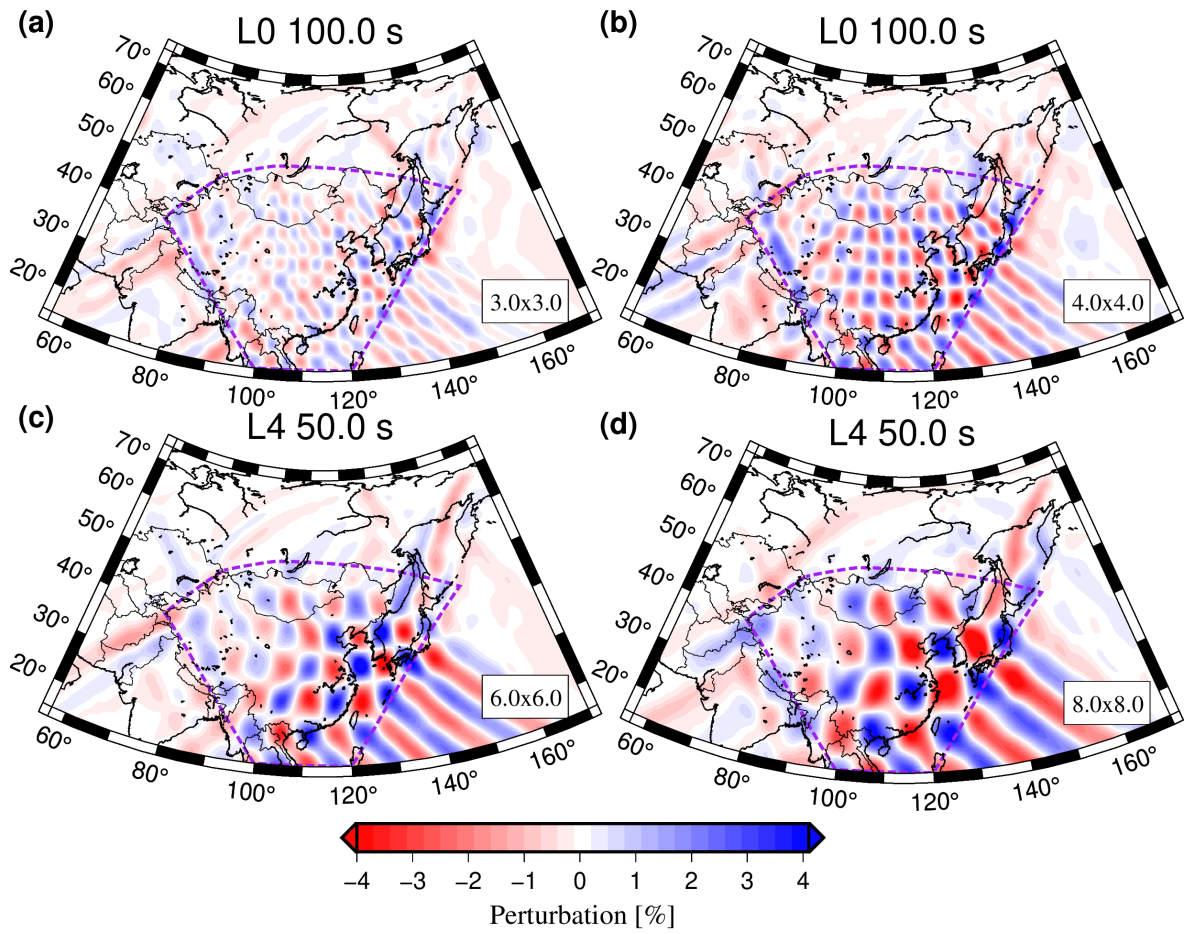


Figure 3.2: Same as Figure 3.1, but for Love waves (Fundamental-mode Love (L0) and 4th higher-mode Love (L4)), with greater cell patterns of (c) $6^\circ \times 6^\circ$ and (d) $8^\circ \times 8^\circ$ for L4 at 50.0 s.

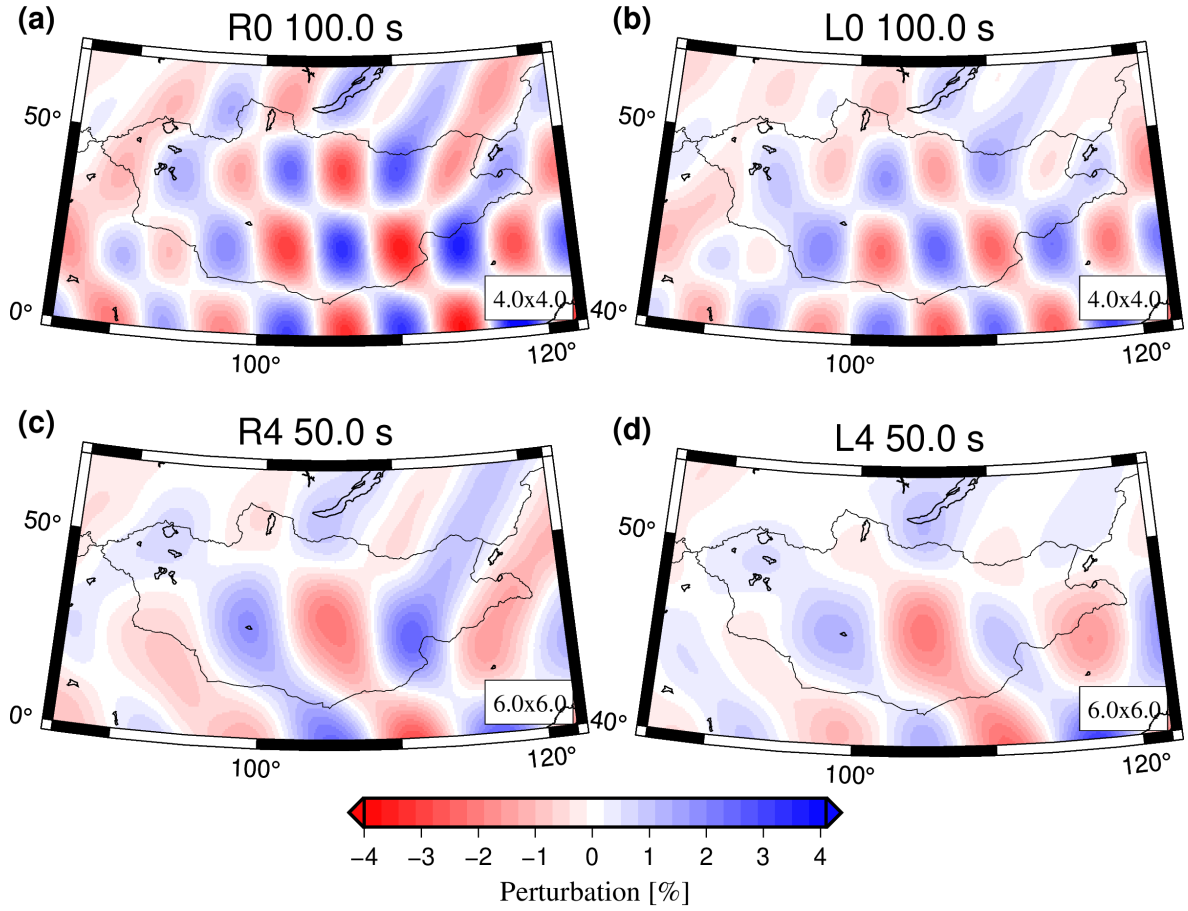


Figure 3.3: Results of checkerboard resolution tests focusing on the Mongolian region for the fundamental and fourth higher modes of Rayleigh wave and Love with $4.0^\circ - 6.0^\circ$ cells. These results are the same as Figures 3.1 and 3.2, but the Mongolian region is magnified in these plots. Fundamental-mode Rayleigh (R0) wave and Love (L0) at 100.0 s with (a) $4^\circ \times 4^\circ$ and (b) $4^\circ \times 4^\circ$ cell patterns, 4th higher-mode Rayleigh (R4) wave and Love (L4) at 50.0s (c) $6^\circ \times 6^\circ$ and (d) $6^\circ \times 6^\circ$ cell patterns.

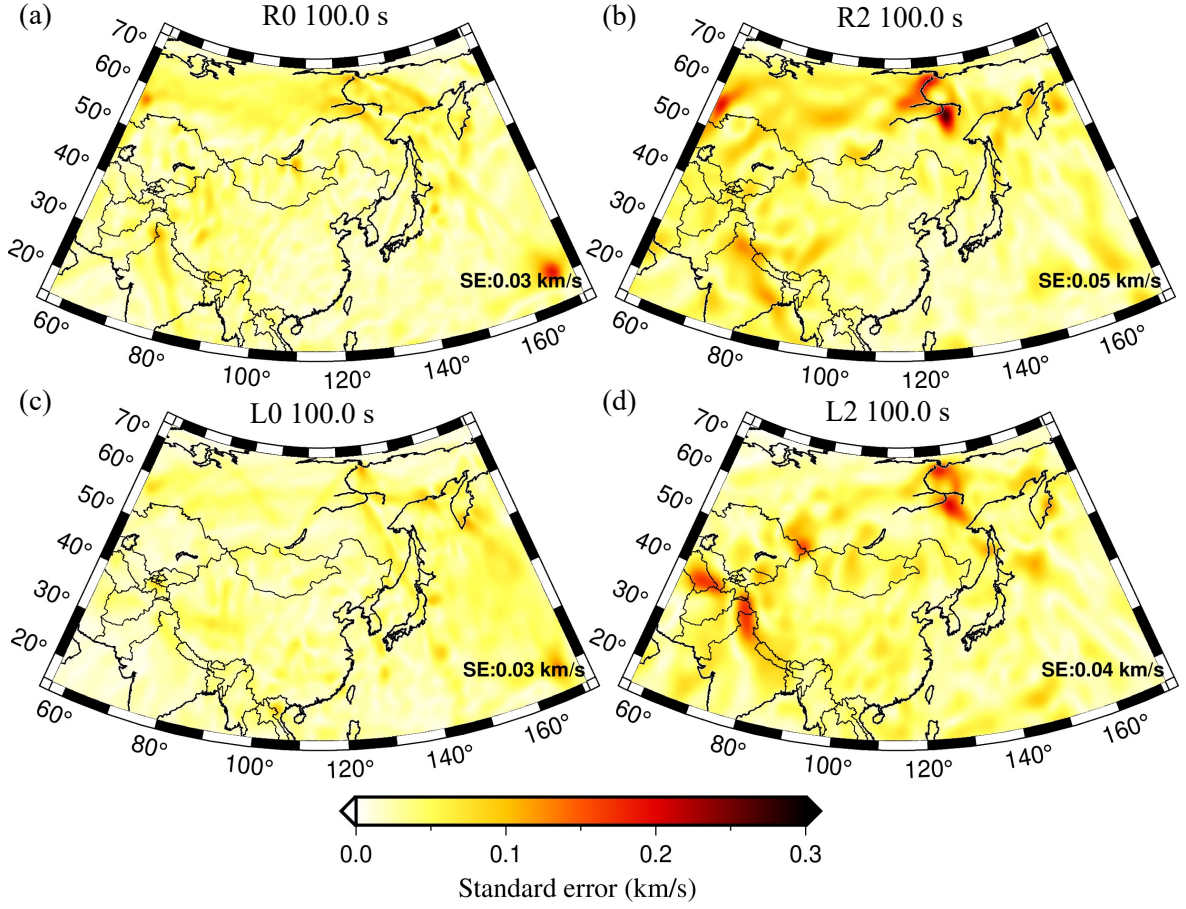


Figure 3.4: Results of jackknife resampling tests for (a) the fundamental-mode Rayleigh wave, (b) the second higher-mode Rayleigh wave, (c) the fundamental-mode Love wave, and (d) the second higher-mode Love wave at 100 s.

seen in some limited areas such as Tiksi and Yakutsk of Russia, where we do not have sufficient ray paths with limited lateral resolution. The resultant error distributions suggest that the estimated errors in Mongolia are sufficiently small.

3.4 Variance reduction for phase speed maps

Here, we summarize the achieved variance reductions (VRs) for all phase speed maps of Rayleigh and Love waves as a period function in Figure 3.5. VRs are computed as, $[1 - |\mathbf{d} - \mathbf{G}\mathbf{m}|^2 / |\mathbf{d}|^2] \times 100$ (%), where $\mathbf{d}$ is the data vector, $\mathbf{G}$ is the kernel matrix, and $\mathbf{m}$ is the model parameter vector, as given in Appendix A.2. VRs reach about 65 – 90 percent for shorter periods than 100 s and decrease in the longer period range (25 – 65 percent) for the fundamental-mode Rayleigh and Love waves, while VRs are about

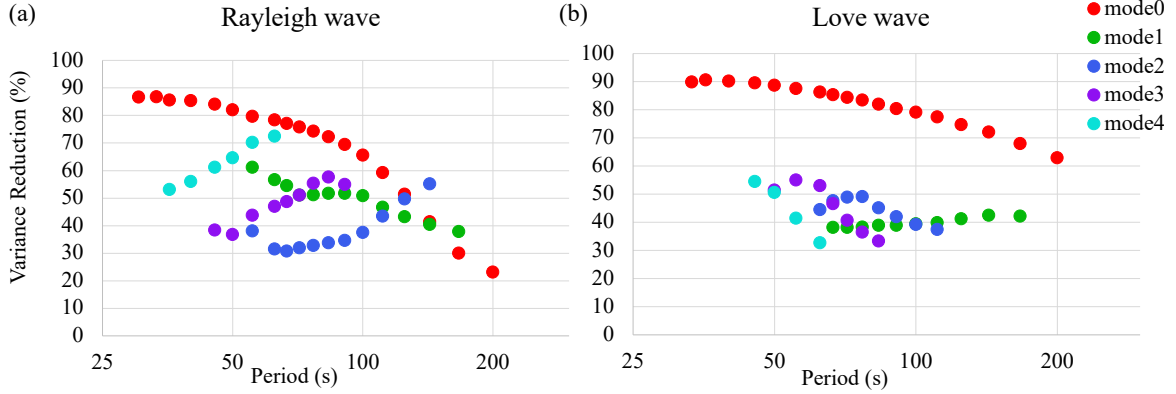


Figure 3.5: Variance reductions for all multimode phase speed maps as a function of period. (a) Rayleigh wave and (b) Love wave.

30 – 80 percent for higher modes. In general, VRs for higher modes tend to be smaller than those for fundamental modes, though it depends on the modes and periods.

3.5 Multimode phase speed maps

The phase speed maps are constructed in the period range of 30 – 200 s for Rayleigh, and 33 – 200 s for Love waves, depending on modes. The period ranges for all modes are summarized in Table 3.1. To reduce the influence of outlying dispersion measurements with large residuals relative to the initial phase speed maps derived from preliminary tomographic inversions, we eliminated outliers whose absolute residuals were greater than 2σ , where the standard deviation σ is derived from all the measured phase speeds for each mode and period, based on the approach by Bao et al. (2013, 2015). Note that all the results of model-checking processes in this thesis (e.g., resolution tests, jackknife tests, etc.) are done with such updated phase speed data sets.

Figures 3.6 and 3.7 show examples of phase speed maps for the fundamental-mode and higher-modes of Rayleigh and Love waves in the Eurasian region and the Mongolian region, respectively. Additional examples of multimode phase speed maps for Rayleigh and Love waves are displayed in Appendix B.3. The average phase speed C_0 shown in each map is used as a reference for the representation of fractional phase speed perturbations. The reference phase speed is derived from the average of path-specific phase speeds within the whole area of this study.

The large-scale features of the fundamental-mode maps show fast anomalies in the

Table 3.1: The period ranges of the multimode phase speed maps used to construct 3-D S wave models of this study

Mode	Rayleigh wave	Love wave
Fundamental	30.3 – 200.0 s	33.3 – 200.0 s
1st-higher	55.5 – 166.7 s	66.7 – 166.7 s
2nd-higher	55.5 – 142.9 s	62.5 – 111.1 s
3rd-higher	45.5 – 90.9 s	50.0 – 83.3 s
4th-higher	35.7 – 62.5 s	45.5 – 66.5 s

Siberian Craton, Tarim Basin, Sichuan Basin, and Ordos area and slow anomalies in Tibet, Okhotsk, and western Mongolia. The overall patterns of the phase speed distribution of the fundamental-mode Rayleigh and Love waves are consistent with the global model GDM52 (Ekström, 2011) as shown in Figure 3.8. While the large-scale structural features are consistent with the global model, the lateral resolution of our model is much higher than that of the global model, particularly in our target region, owing to the use of many regional seismic networks in and around Mongolia. These phase speed maps will be used in the next section to obtain the 3-D S-wave speed model.

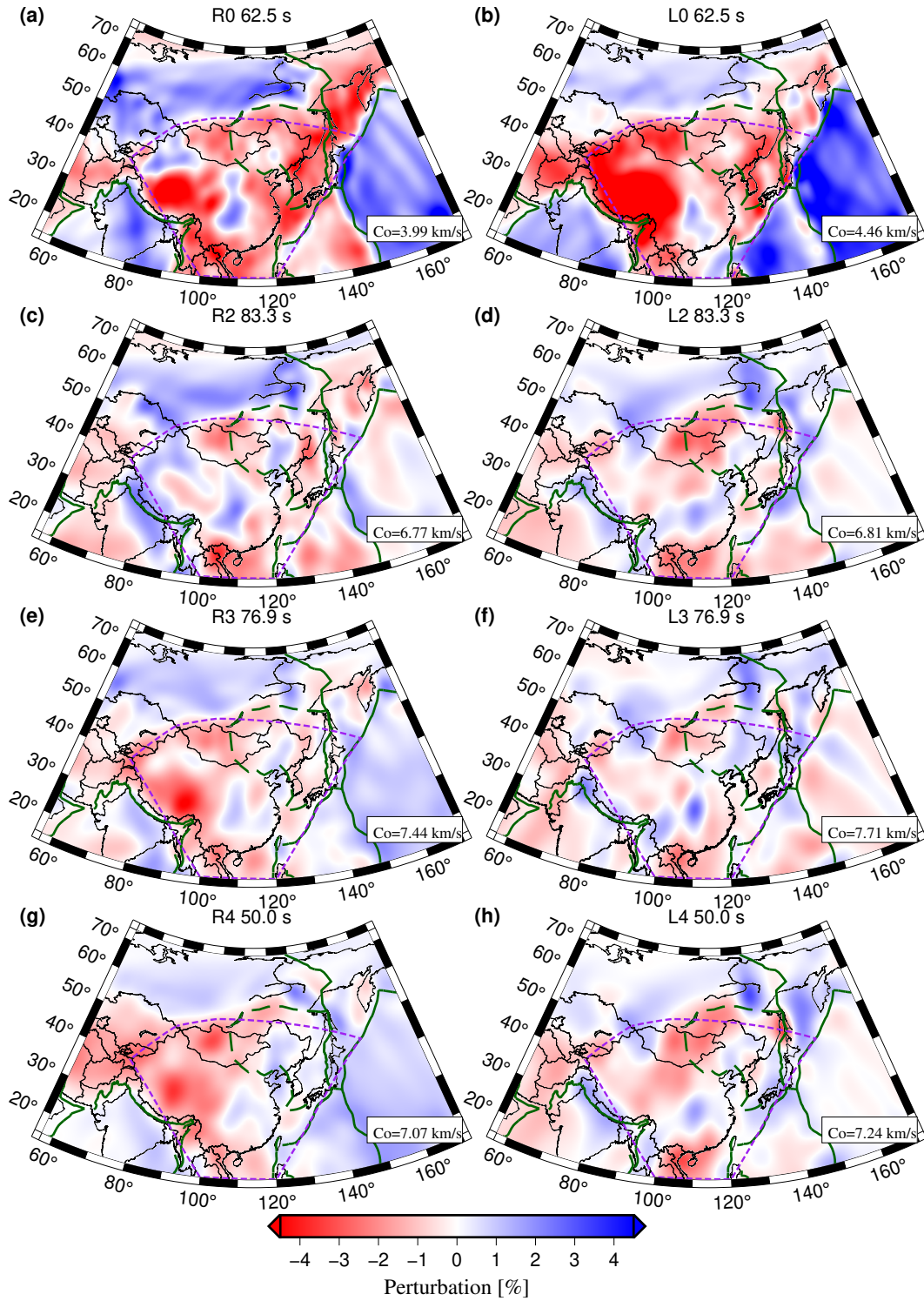


Figure 3.6: Phase speed maps for the fundamental and higher modes of Rayleigh (R) and Love (L) waves, with the numbers representing the mode number and period. Purple dashed line indicates the well-resolved area estimated from the checkerboard tests.

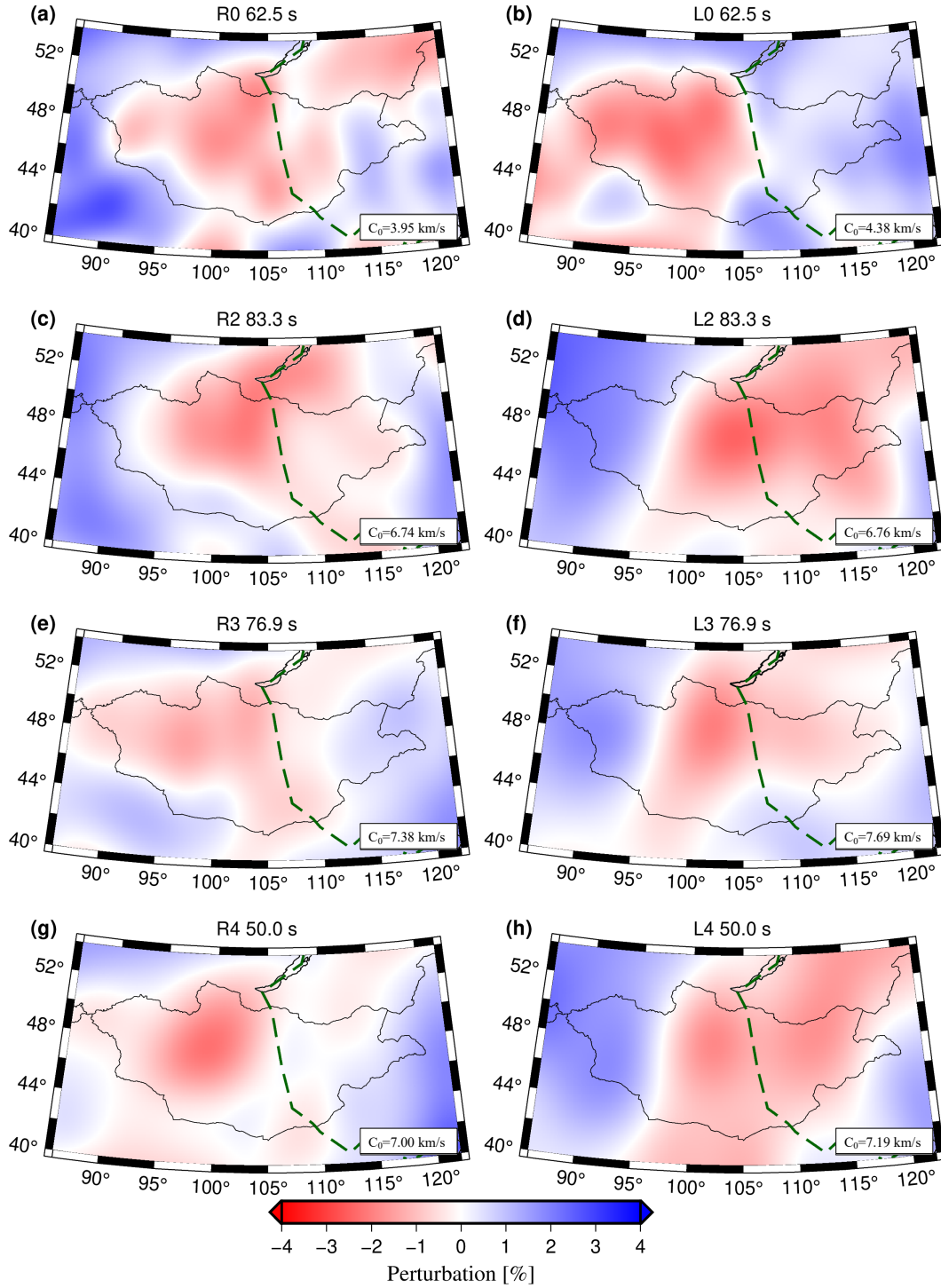


Figure 3.7: Phase speed maps for the fundamental and higher modes of Rayleigh (R) and Love (L) waves focusing on the Mongolian region, with the numbers representing the mode number and period.

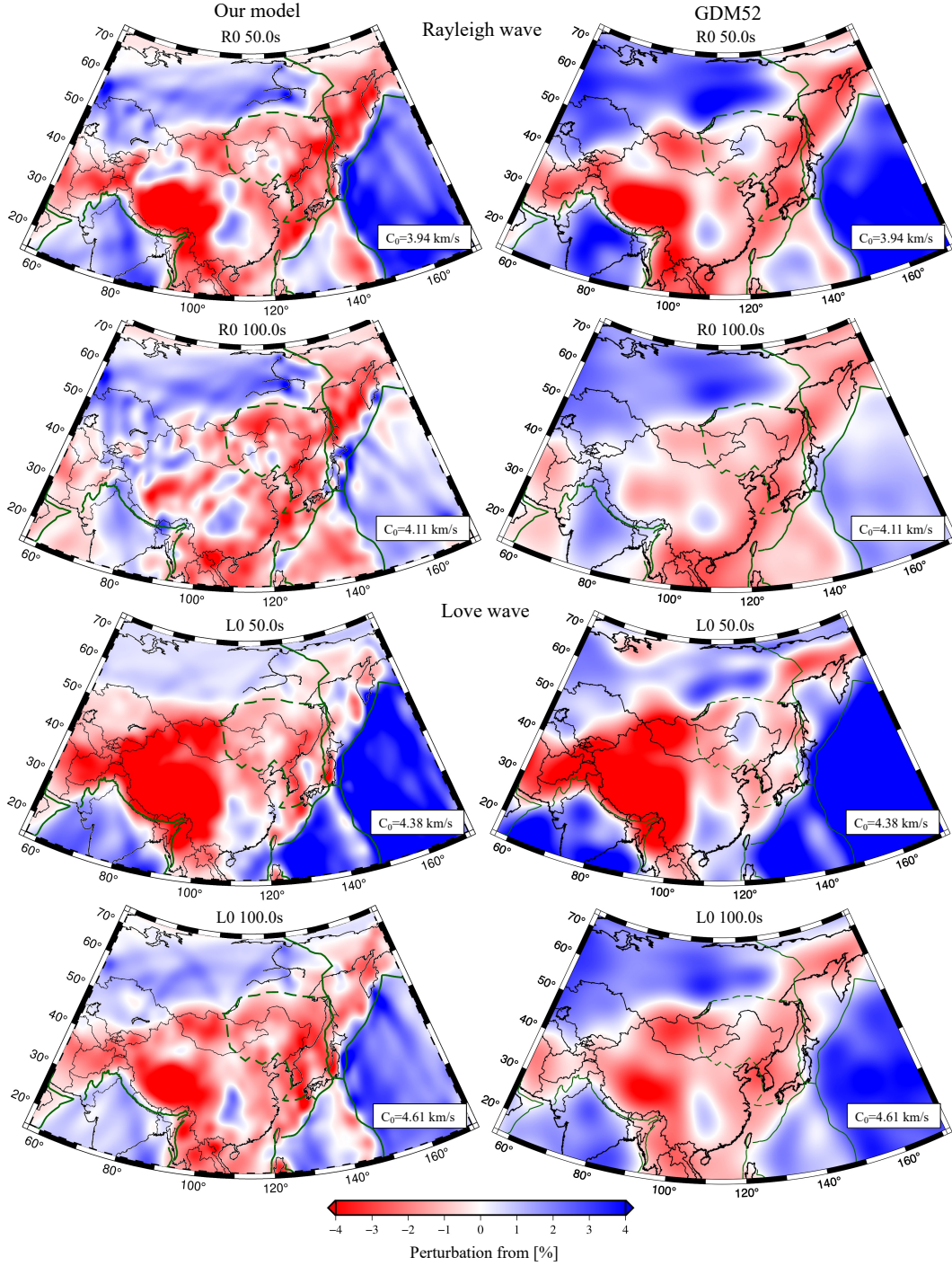


Figure 3.8: Comparison of phase speed maps of this study (left) with those of the GDM52 model (right) for the fundamental modes of Love (L) and Rayleigh (R) waves at 50 s and 100 s.

Chapter 4

Radially Anisotropic 3-D S-wave Model

The multimode phase speed maps described in the previous chapter are inverted for the 3-D S-wave model, which is the final step of the three-stage surface wave tomography. In this chapter, we briefly review the employed method of inversions for localized 1-D velocity profiles, including numerical experiments for crustal corrections and vertical resolution tests. Then, the final 3-D S-wave model on a broader scale in Eastern Eurasia, including Mongolia, is displayed, which will be further discussed in detail in the next chapter.

4.1 Inversions for radially anisotropic 1-D S wave models

To build the 3-D radially anisotropic S-wave model using the multimode phase speed maps, we repeatedly perform inversions for local 1-D S-wave models using the local multimode phase dispersion curves derived from the phase speed maps in the previous chapter. We employed the generalized nonlinear least-squares inverse scheme of Taran-tola & Valette (1982), following Yoshizawa (2014). The detailed procedure has been described by Yoshizawa (2014), and some technical aspects of the method are briefly summarized in Appendix A.3. Here we briefly summarize the essence of the method.

We consider a set of 6 independent model parameters $(\rho, \alpha_H, \alpha_V, \beta_H, \beta_V, \eta)$, where ρ is density, α_H and α_V represent horizontally and vertically polarized P wave speeds, β_H and β_V represent horizontally and vertically polarized S wave speeds, η is a dimensionless anisotropic parameter. P wave speeds and density are considered via scaling relations for S wave speeds (e.g., Gung et al., 2003; Panning & Romanowicz, 2006), following Montagner & Anderson (1989). We invert local multimode Rayleigh and

Love wave phase speeds simultaneously for local SV (β_V) and SH (β_H) models, which is repeatedly performed for all locations in the target areas at 1.0 degree grid interval, to construct the final anisotropic 3-D S wave model. The employed period ranges for each mode are summarized in Table 3.1.

To control the amplitude of shear wave speed perturbations and the smoothness of the vertical velocity model during the inversions, we employ the model covariance matrix comprising two *a priori* parameters; standard deviations σ and correlation length L between the vertical grids (e.g., Nishimura & Forsyth, 1989; Yoshizawa & Kennett, 2004; Yoshizawa, 2014). We used the following *a priori* parameters, i.e., $\sigma = 0.045$ km/s from Moho to 670 km depth, $L = 10$ km from Moho to 400 km depth and $L = 15$ km from 400 to 670 km depth. Beneath 670 km depth, σ decreases linearly to 0.025 km/s, and L increases linearly to 30 km at 1500 km depth. The velocity in the crust can be perturbed with $\sigma = 0.025$ km/s and $L = 5$ km. Thus, the crust and uppermost mantle structures change rapidly while the deeper structure changes gradually.

Figure 4.1 displays an example of the radially anisotropic local 1-D S-wave model (SH and SV wave speeds) derived from the inversions of local multimode dispersion curves at 46°N and 99°E. Based on the retrieved SH (β_H) and SV (β_V) models, we compute their Voigt average as the isotropic S wave model, $\beta_{iso} = \sqrt{(2/3)\beta_V^2 + (1/3)\beta_H^2}$ (e.g., Panning & Romanowicz, 2006) and the radial anisotropy parameter $\xi = (\beta_H/\beta_V)^2$. Measured dispersion curves derived from the multimode phase speed maps in the previous section can reasonably fit well in all modes over the wide period ranges.

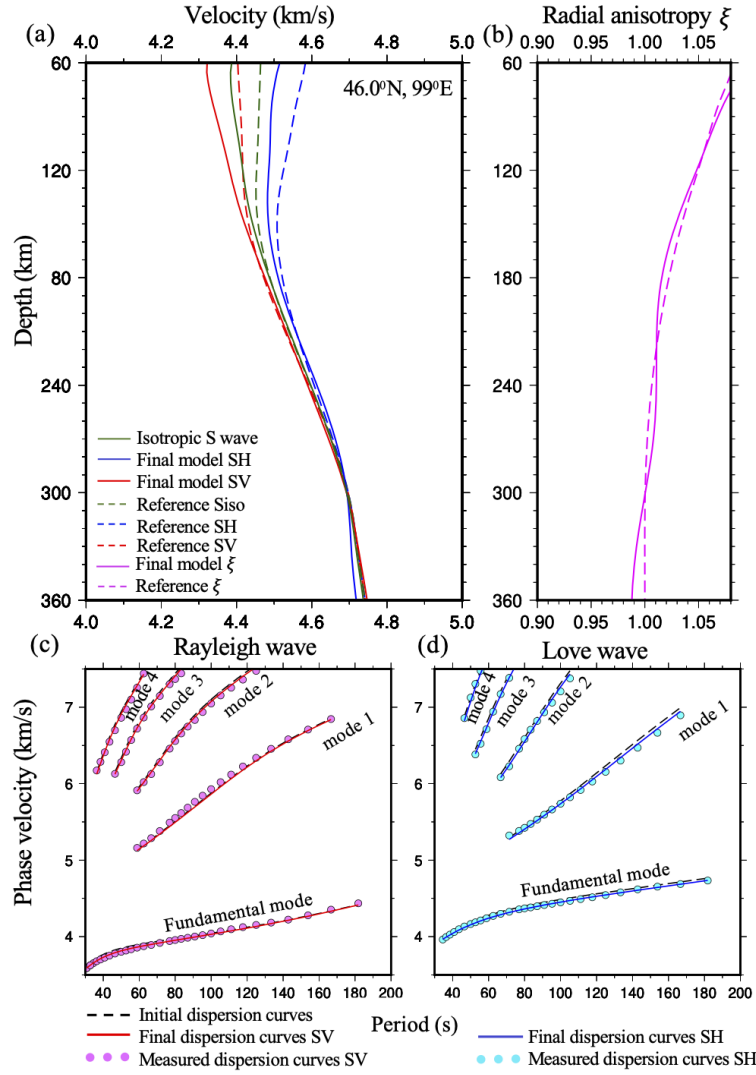


Figure 4.1: (a) An example of the 1-D shear wave model at a location of 46°N and 99°E in Mongolia, and (b) Radial anisotropy. The reference SV and SH velocities as well as radial anisotropy (dashed lines) are taken from the modified PREM with a smoothed 220 km discontinuity. (c, d) The results of dispersion curve fitting of synthetic phase speeds computed from the inverted SV and SH velocity models in (a) and the measured local phase speeds for (c) Rayleigh wave and (d) Love wave.

4.2 Effects of the crustal corrections

Since we employ surface-wave phase speeds in the period range longer than 30 s, our phase speed data have limited vertical sensitivities to the shallow crustal structure (see, e.g., Figure A.1). Therefore, we need a proper local crustal model in the target region to make corrections for the crust in the inversions for local 1-D velocity profiles. We performed experiments to examine the effects of the crustal corrections on the shear wave speed model using two 3-D crustal models and without the crustal correction (with fixed crust to PREM). Figures 4.2 and 4.3 show the comparisons of the retrieved shear wave models with crustal corrections using 3SMAC (Nataf & Ricard, 1996) and CRUST1.0 (Laske et al., 2013) and without the crustal correction using a modified PREM (Dziewonski & Anderson, 1981) with a smoothed 220 km discontinuity. The inverted 1-D SH and SV models using the three crust models are compared in Figure 4.2. Note that in these figures (and all equivalent figures in this thesis), the depth is measured from the average Earth's radius (6371 km), although the local topography is considered during the inversions.

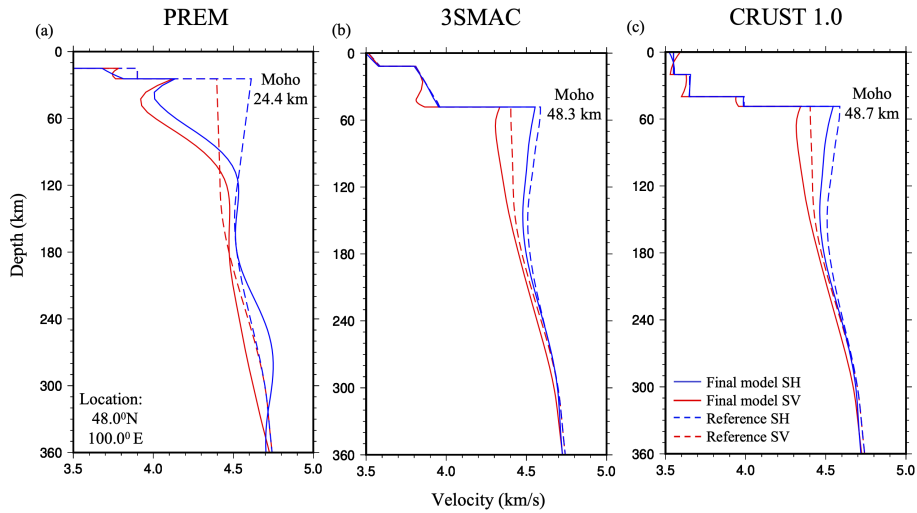


Figure 4.2: Examples of SH and SV wave models derived from the different crust models (PREM, 3SMAC, and CRUST1.0). The depth in the vertical axis is measured from the average Earth's radius and the local topography is omitted in the diagram.

The two different crust models (3SMAC and CRUST1.0) provide us with similar results of 1-D SV and SH models in the upper mantle despite the slight differences in the crustal thickness and velocity, while the use of PREM's crust (which does not

represent the thick Mongolian crust) results in unstable S velocities in the upper mantle, indicating the importance of the use of a proper crust model.

The retrieved 3-D isotropic S wave speed model (the Voigt average of SH and SV models) in the Mongolian region is compared in Figure 4.3. According to local studies, the crustal thickness in western Mongolia has been reported to be around 50–55 km (Petit et al., 2008; Feng, 2021; Zhao et al., 2021), while that in eastern Mongolia around 38–46 km (He et al., 2016). The CRUST 1.0 model with a 1-degree grid is likely to better represent the local crustal structure in Mongolia compared to the 3SMAC model with a 2-degree grid. Thus, in this study, we employed the local reference models combining the CRUST1.0 model for the crust and the modified PREM (with smoothed 220 km discontinuity) for the mantle.

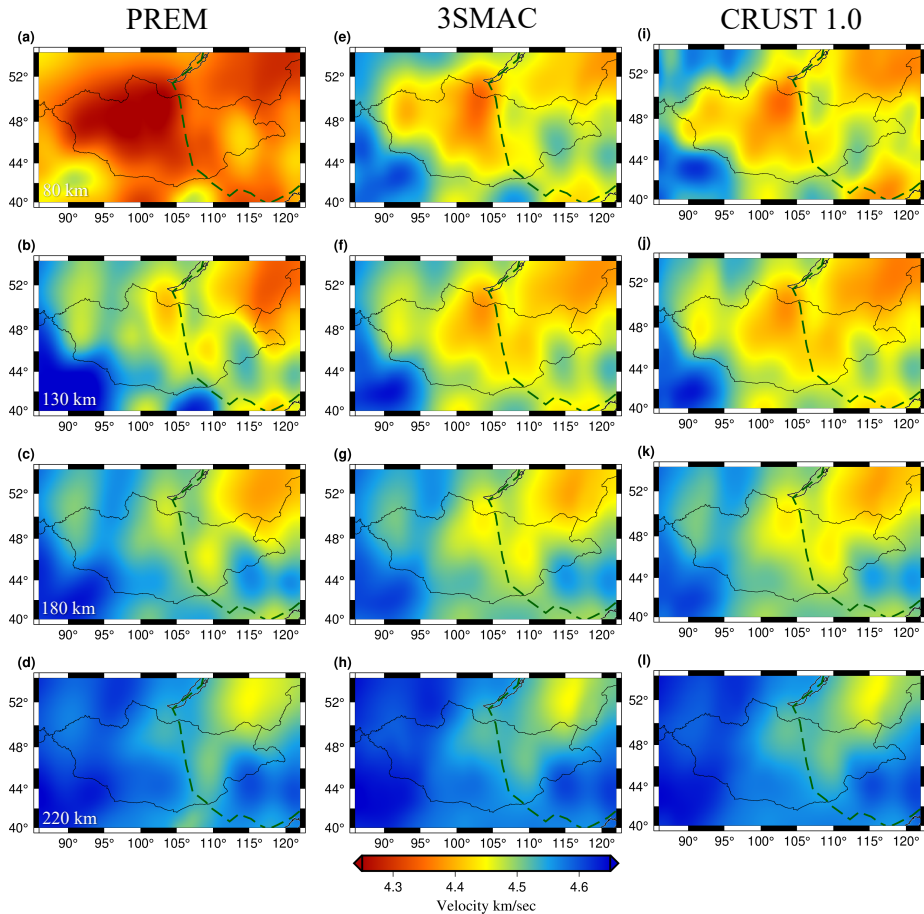


Figure 4.3: Comparison of isotropic S wave speed model beneath Mongolia with crustal corrections from PREM (left), 3SMAC (middle), and CRUST1.0 (right).

4.3 Vertical resolution tests

To confirm the depth resolution of the 3-D model, we performed vertical resolution tests using synthesized S-wave models with spike-like anomalies in the upper mantle. To generate synthetic dispersion curves for multimode Love and Rayleigh waves including higher-modes up to the 4-th mode, we create five synthetic models with a 5 % faster anomaly relative to the reference model (modified PREM with a smoothed 220 km discontinuity) in a limited depth range around 80, 100, 150, 200, and 300 km. Then, using the same parameter sets for the real data inversions as explained in the earlier section, radially anisotropic S-wave speed profiles are retrieved through the linearized inversions. The results of the vertical resolution tests are summarized in Figure 4.4, which indicate that we can achieve sufficient vertical resolution down to 300 km depth with the current dispersion data sets, despite the vertical smearing effects from the long-period surface waves.

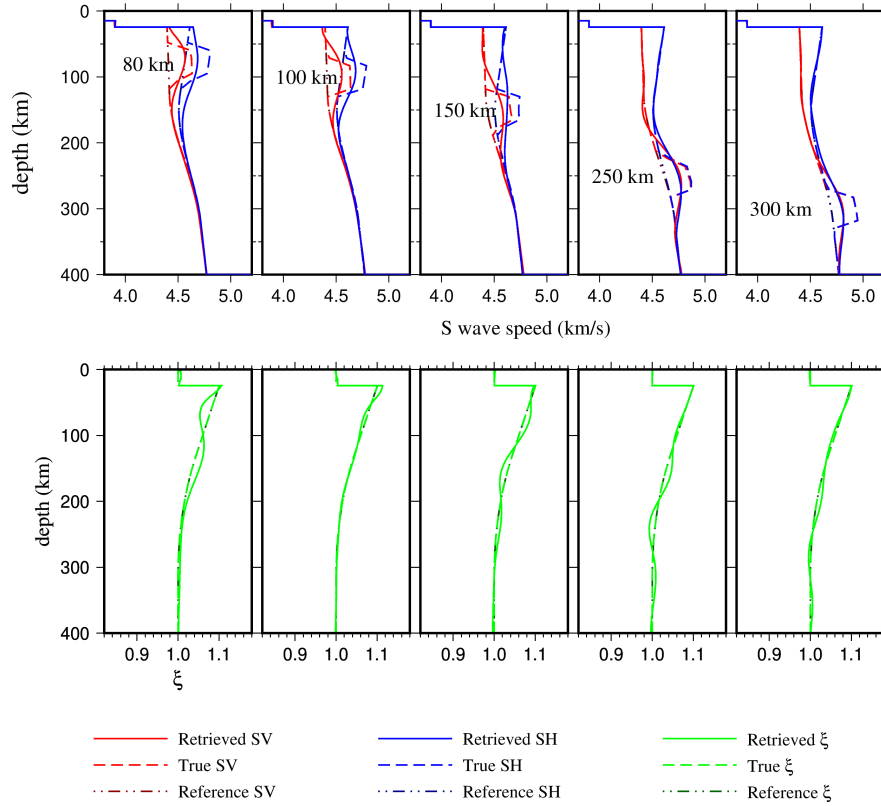


Figure 4.4: Examples of depth resolution tests for radially anisotropic shear wave models. (Top) SV and SH wave speed models, and (bottom) radial anisotropy profiles.

4.4 3-D S-wave model in Eastern Eurasia and Mongolian regions

4.4.1 Regionally-averaged 1-D S-wave profiles

The radially anisotropic local 1-D S-wave models at every 1.0-degree grid across the target region are merged to form the final 3-D SV and SH wave models. Figure 4.5 displays regionally-averaged 1-D S-velocity profiles for (a) the Eurasian region and (b) the entire Mongolian region, in comparison with the modified PREM. Some of these 1-D profiles are used as a reference model for plotting the perturbation maps and cross

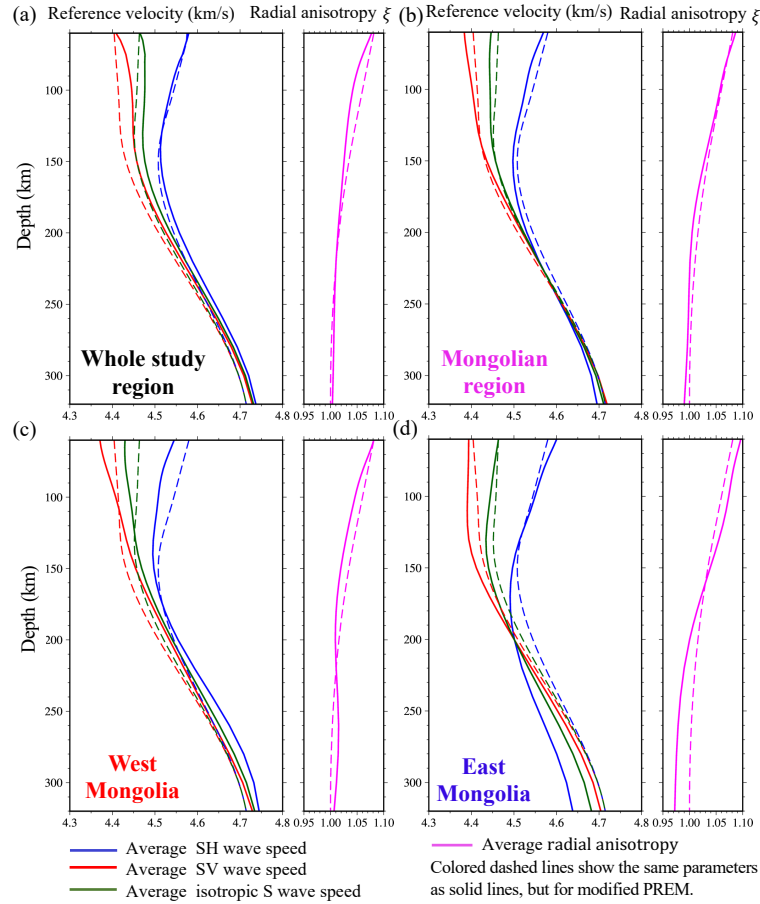


Figure 4.5: The average 1-D velocity profiles of SV (red), SH (blue), isotropic S wave speeds (green) and radially anisotropy ξ (magenta) for (a) the whole region of this study (Eastern Eurasia), (b) the entire Mongolian region, (c) western Mongolia, and (d) eastern Mongolia. Colored dashed lines represent the same parameter for the modified anisotropic PREM with a smoothed 220 km boundary. The areas used for computing the regional average in (b, c, and d) are shown in Figure 4.6.

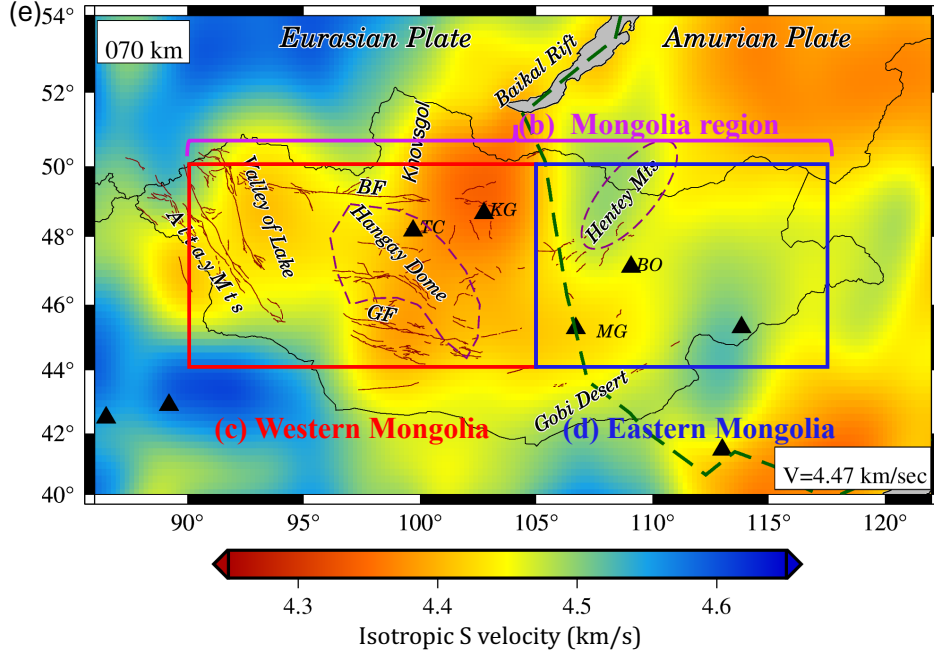


Figure 4.6: S-velocity model in Mongolia indicating the western and eastern areas of Mongolia used to compute the averaged 1-D profiles in each region in Figure 4.5 (c, d).

sections of the 3-D S-wave models displayed in this section and the next chapter.

The average 1-D profiles of the SV and SH wave speeds below 60 km in the broader Eurasian region are somewhat faster than the modified anisotropic PREM model (Figure 4.5 a), while those in the Mongolian region tend to exhibit slower shear wave speeds (Figure 4.5 b). The average radial anisotropy in the Mongolian region is similar to the PREM model.

We also display the averaged 1-D profiles for the western and eastern Mongolian regions in Figure 4.5 (c) and (d), respectively, indicating the regional variations of the seismic structure across Mongolia, which will be discussed in the next chapter in detail based on the 3-D S-wave model. Figure 4.6 shows regions used to compute the locally averaged 1-D profiles in western and eastern Mongolia.

4.4.2 3-D S-wave model in broader areas of Eurasia

Figure 4.7 shows the final 3-D SH and SV wave speed models, and Figure 4.8 shows isotropic S wave speed model (β_{iso}) and radial anisotropy models ξ in the whole region of this study. For the 3-D model plots, we apply the 2-D Gaussian filter with a standard

deviation of 0.67° to smooth our 3-D velocity model laterally.

Here we briefly overview the large-scale features of the retrieved 3-D model at various depth slices shown in Figure 4.7 (SH and SV wave speed models) and in Figure 4.8 (isotropic S wave speeds and radially anisotropy ξ). Some selected vertical cross-sections (E-W and N-S directions) of isotropic S wave speeds and radial anisotropy in Eastern Eurasia are summarized in Figure 4.9. In Appendix C, we also show a gallery of broader-scale cross-sections across Eastern Eurasia (Figures C.1 – C.5).

In the 3-D model at 80 – 100 km depth, we find relatively low velocities in western Mongolia, Tibet, and in the East and South China Seas, Yellow Sea, Ryukyu Trench, Sea of Japan, and Sea of Okhotsk, while other areas in the Siberian craton, Sichuan basin, Ordos block, and Tarim basin exhibit higher velocities. At a depth of 150 km, the low-velocity anomalies in western Mongolia have weakened, while the eastern marginal seas and Tibet still show slower anomalies, reflecting the higher temperature in the mantle wedge region in the subduction zone. The fast S-wave anomalies in the Tarim, Ordos, and Sichuan basins are apparent in both SH and SV wave models down to 150 km depth. Earlier surface wave studies agree well with the overall pattern of shear wave speed anomalies in the upper mantle in this region (Ritzwoller & Levshin, 1998; Chen et al., 2015b; Bao et al., 2015; Tao et al., 2018; Witek et al., 2021).

In the radial anisotropy maps at 80 – 100 km depth, most regions exhibit positive radial anisotropy ($\xi > 1$ or $\beta_H > \beta_V$), except for some regions such as a part of Tibet, and under Japan. Below 150 km depth, the strength of radial anisotropy tends to be weaker than shallower depth in most of the region.

Figure 4.9 summarize vertical cross-sections for locations indicated in Figure 4.8 (g). In profiles A–A', C–C', D–D', and E–E', notable fast velocity anomalies are seen beneath the Siberian Craton, Sichuan Basin, Tarim Basin, Ordos Block, and Pacific Plate. Significant low-velocity anomalies can be seen in profiles C–C', D–D', and F–F', beneath the Tibet and Yellow Sea.

In the next chapter, we will further discuss characteristic seismological features in the Mongolian region inferred from the anisotropic 3-D S-wave model.

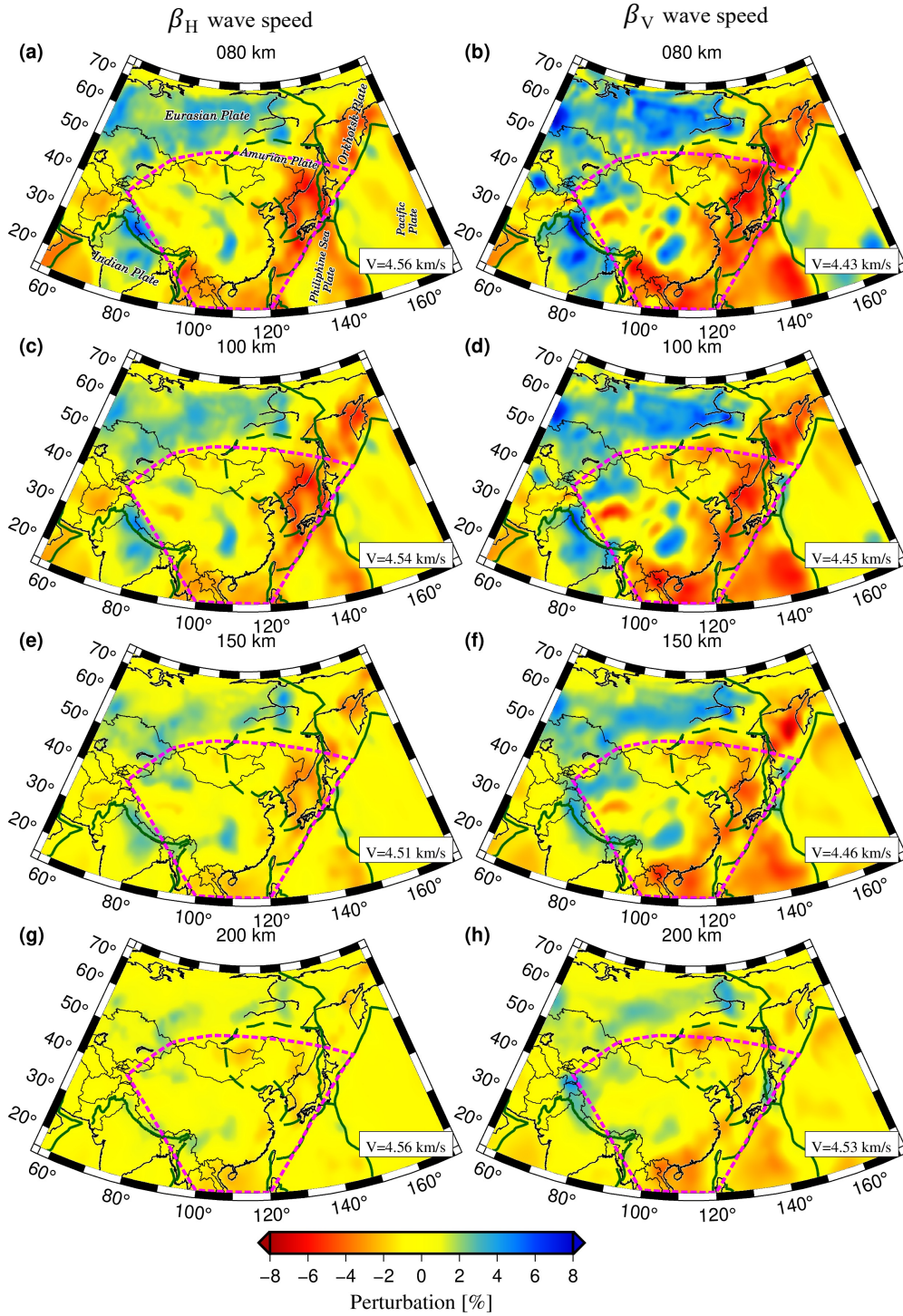


Figure 4.7: SH and SV wave speed models for whole areas of this study from 80 to 200 km depth. Reference velocities are indicated in each map. Plate boundaries (green lines) are taken from Bird (2003). Magenta dashed line indicates the well-resolved area estimated from the checkerboard tests.

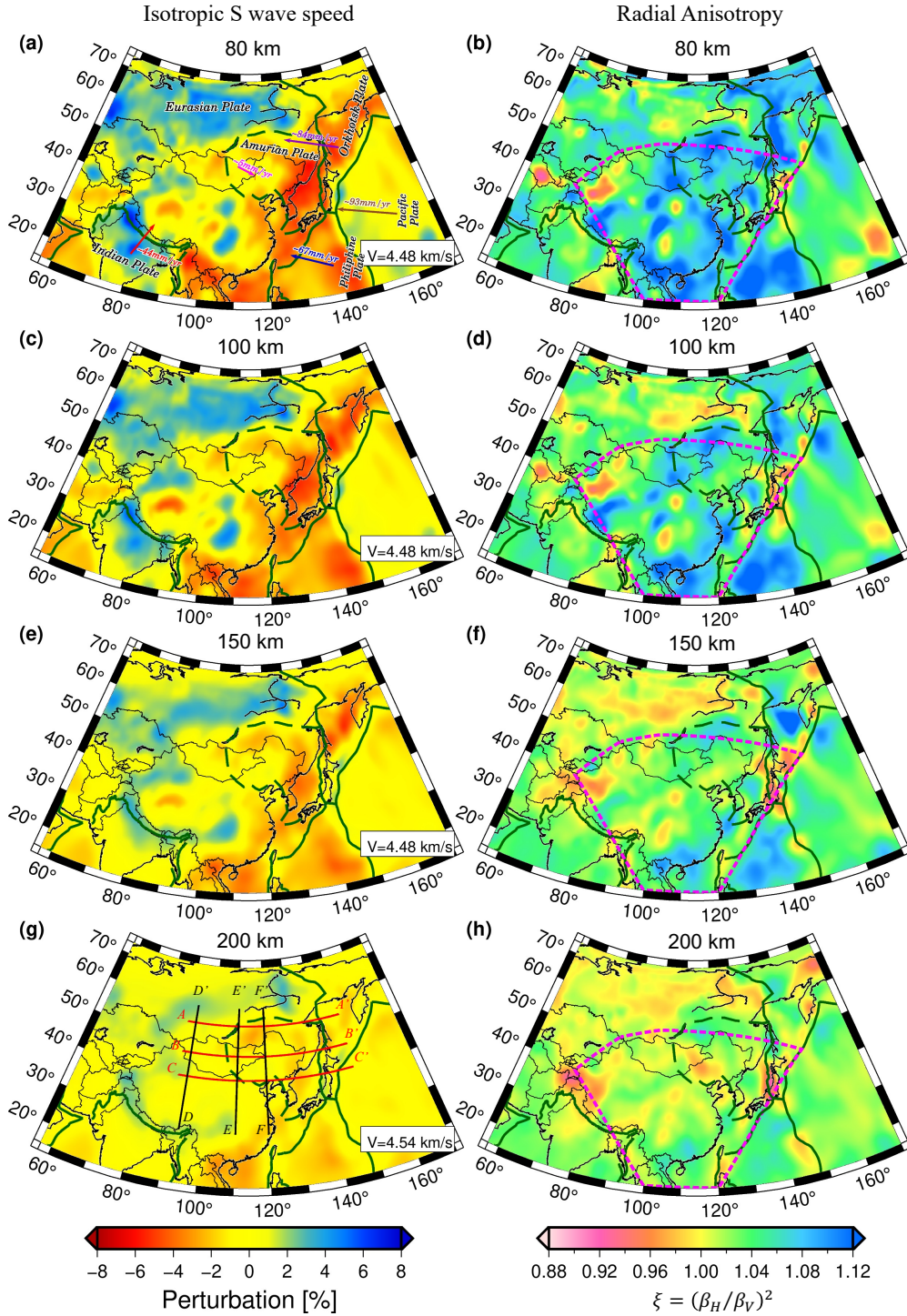


Figure 4.8: Same as Figure 4.7, but for isotropic S wave speed model (β_{iso} , left) and radial anisotropy model (ξ , right). The arrows in (a) indicate relative plate motions based on the MORVEL plate model (DeMets et al., 2010). Locations of cross sections in Figures 4.9 are shown in (g).

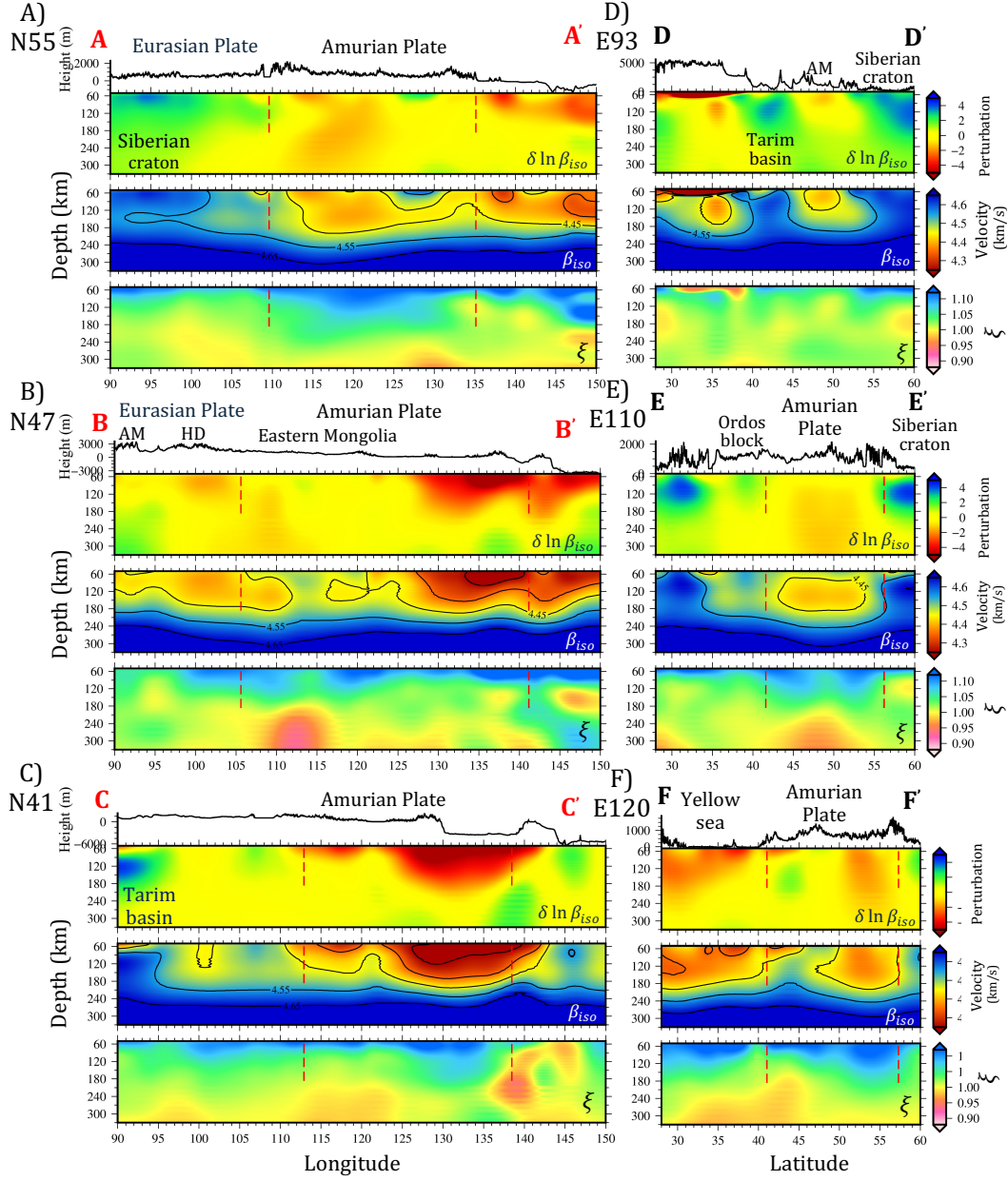


Figure 4.9: Cross-sections for fractional perturbation of β_{iso} (top), absolute isotropic S velocity β_{iso} (middle), and radial anisotropy ξ (bottom) along the profiles (A) A-A', (B) B-B', (C) C-C', (D) D-D', (E) E-E', and (F) F-F' shown in Figure 4.8 (g). The reference isotropic S wave speed profile (a modified PREM) used for the perturbation plot is shown as a dashed green line in Figure 4.5(a). The surface topography on top is from ETOPO1 (Amante & Eakins, 2009). Major tectonic provinces are labeled on topography. The vertical red dashed lines show the vertical extension of the Amurian plate boundary by Bird (2003).

Chapter 5

Discussion of Mongolian Upper Mantle Structure

We constructed a new 3-D S-wave model including radial anisotropy beneath Mongolia and its surrounding region using a number of local seismic stations in this region. The lateral structural changes revealed by our new 3-D S-wave model beneath the Mongolian region suggest some intriguing geographical relations that reflect Mongolian tectonics and mantle dynamics. Here we focus on the S-wave structure in the upper mantle from 60 to 250 km depth, which has been well resolved by our surface wave dispersion data, and discuss some prominent features of the 3-D model in relation to the tectonic and dynamics processes in the Mongolian region.

5.1 Entire Mongolia

Figure 5.1 displays various depth slices of the original SH and SV wave speed models, and Figure 5.2 displays the isotropic S wave speed model (β_{iso}) and radial anisotropy (ξ) beneath Mongolia. Cross sections in Figure 5.3 depict some significant characteristics in our model in the depth range from 60 to 250 km. Many more figures are displayed in Appendix C.2, with a gallery of cross-sections focusing on the Mongolian region (Figures C.6 – C.10).

In the shallower depth at 75 – 100 km beneath Mongolia, the isotropic S-wave speeds (Figure 5.2) are generally slower in the tectonically active western area and become faster in the relatively stable eastern area. The fast anomaly in eastern Mongolia indicates the lithosphere with a thickness of around 70–100 km, which disappears in the west of the expected Amurian plate border by Bird (2003) (Figure 5.3 (A)). Slower

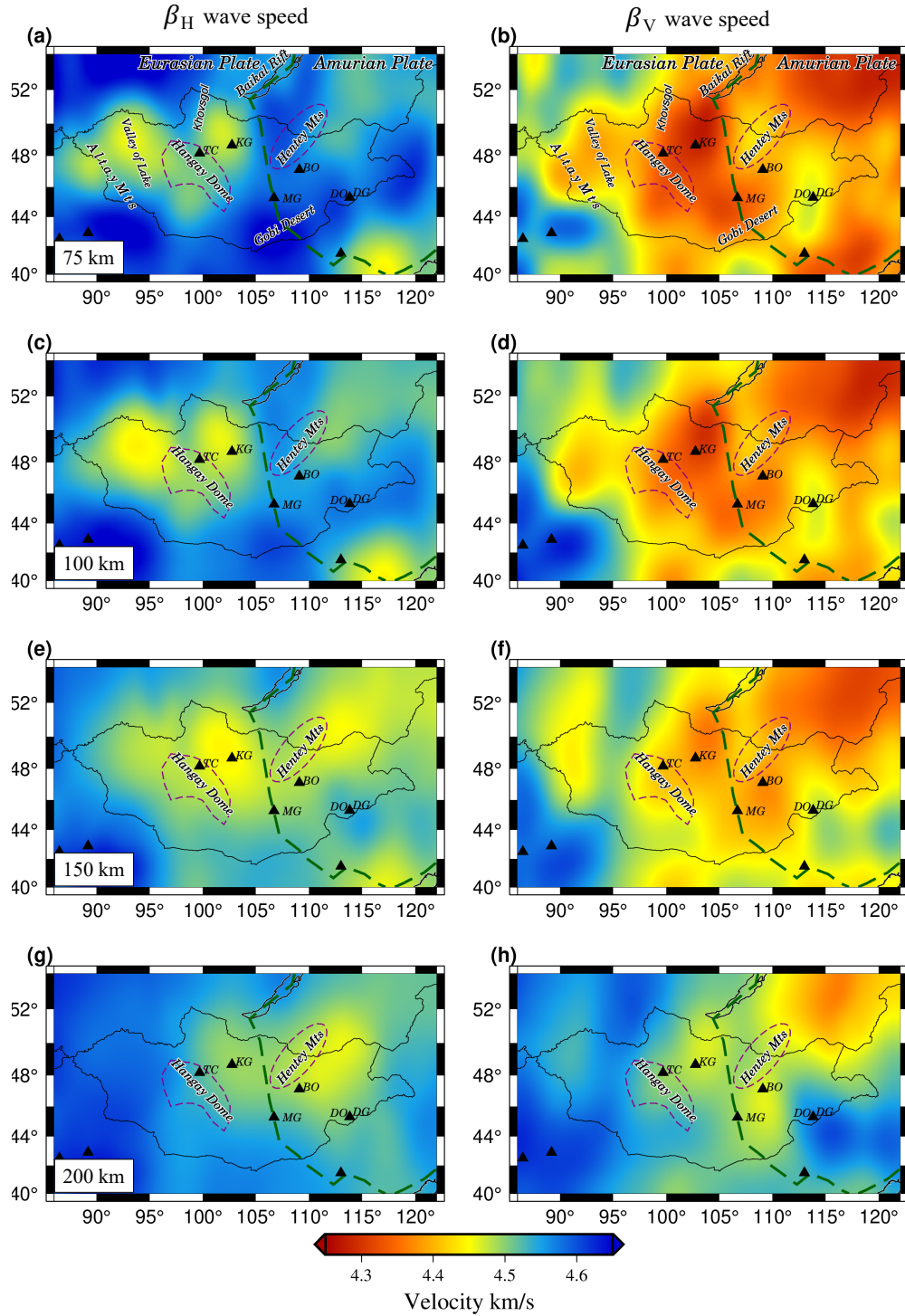


Figure 5.1: SH and SV wave speed models beneath Mongolia from 75 to 200 km depth. Green dashed line is the Amurian plate boundary by Bird (2003). Black triangles denote the Holocene volcanoes; Taryatu Chulutu (TC), Khanuy Gol (KG), Middle Gobi (MG), Bus Obo (BO), Dari Ovoo (DO), and Dariganga (DG).

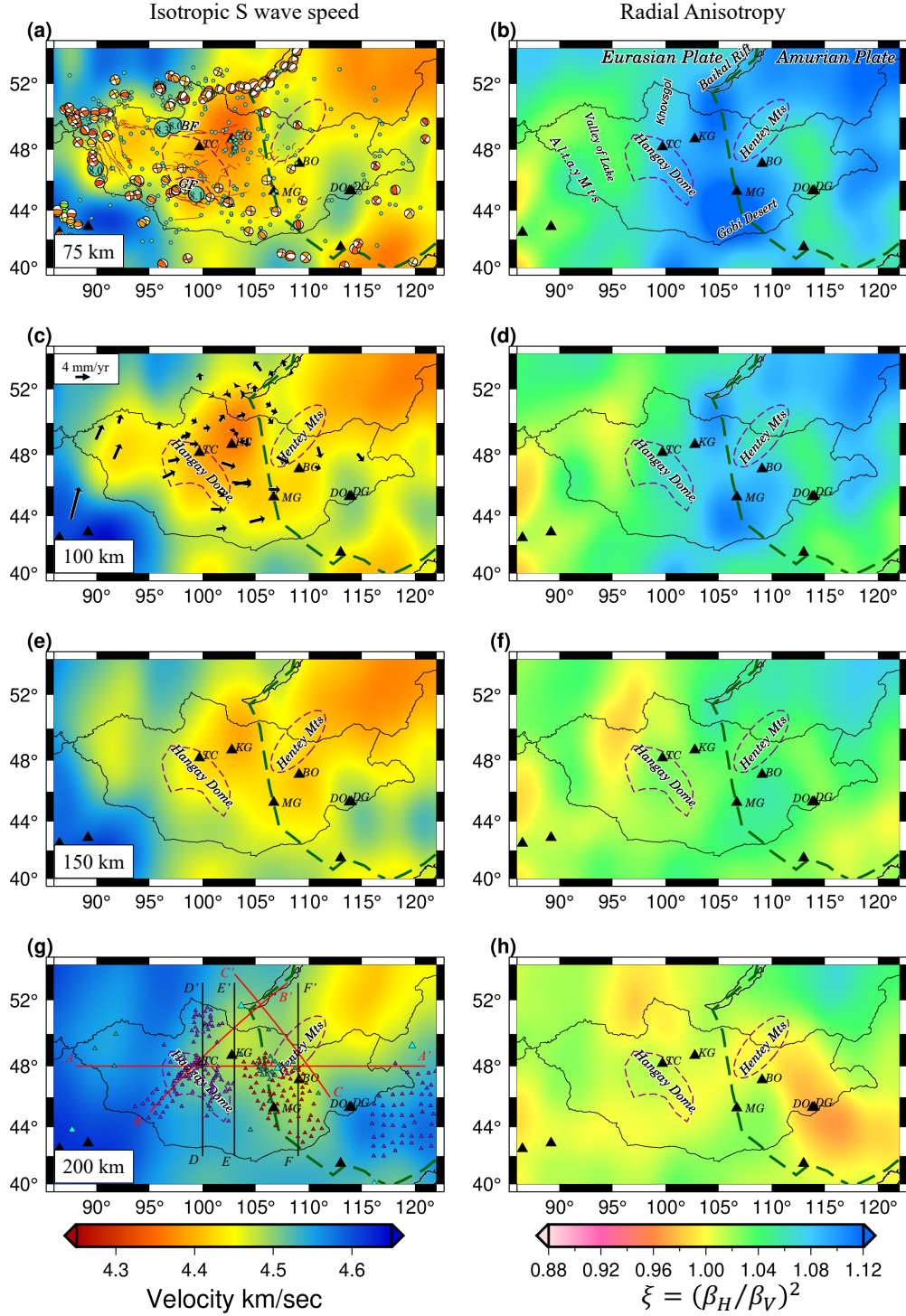


Figure 5.2: Same as Figure 5.1, but for isotropic S wave speed (β_{iso}) and radial anisotropy (ξ) model. Green circles in (a) indicate seismicity from the event catalog by IAG (Institute of Astronomy and Geophysics of Mongolia) from January 1964 to May 2023 with $M_l \geq 3.5$. Some historical large earthquakes ($M_w \geq 8.0$) are labeled. (continue.)

Figure 5.2: (continued.) Focal mechanisms greater than Mw 4.0 taken from the Global CMT catalog between January 1976 and May 2023 (Ekström et al., 2012) are shown in (a). Black arrows in and around Mongolia in (c) denote the rates of crustal motion relative to Eurasia (Calais et al., 2003). Black and red lines in (g) indicate the locations of cross sections in Figure 5.3. Small colored triangles in (g) indicate the employed seismic stations.

velocities in western and central Mongolia can be linked to tectonics under the Hangay Dome and Altay Mountains, which are also seen in earlier tomography models (e.g., Ritzwoller & Levshin, 1998; Chen et al., 2015a; Tao et al., 2018; Witek et al., 2021; Li et al., 2022).

In western Mongolia, slow velocity anomalies are seen in the southeastern area of the Khovsgol Rift Zone (Khanuy Gol volcanic field), Baikal Rift, eastern Hangay Dome, the northern edge of the Altay Mountains, and Valley of the Lake. On the contrary, in the eastern regions at shallower depths from 70 to 100 km, fast anomalies are observed in the Gobi Desert, Hentey Mountains, and Dariganga. The fast anomalies around the Hentey Mountains and the Gobi Desert have weakened at a greater depth of around 150 km (Figures 5.1 and 5.2).

The radial anisotropy model exhibits a faster SH wave speed (β_H) than SV (β_V) in nearly the entire Mongolian lithosphere above 150 km depth, except for the southern part of the western Sayan Mountains. The Altay and Sayan Mountains have experienced strong convergence as a result of the northward motion of the Asian blocks (such as Junggar and Tarim Basins) due to the India-Asia collision (Huang & Zhao, 2022). The P-wave velocity model by Huang & Zhao (2022) showed low velocity down to 200 km depth beneath the Altay Mountains, suggesting that lithosphere-scale interactions may have caused it. Our S-wave model also shows the low-velocity anomalies along the northern Altay Mountain border, which can persist to about 100 km depth (Figures 5.1 (a, b, d, and f), 5.2 (a and c), and 5.3 (A)).

The Hentey Mountains show slower S wave speed anomaly at depths below around 120–150 km (Figures 5.3 (A, C, and F), C.7, and C.8). The notable low velocity found under the Hentey Mountains agrees well with a recent tomographic study by Chen et al. (2015a) who have reported the low-velocity anomaly may stretch from the uppermost mantle to 200 km depth at least. He et al. (2019) noted that the tectonic setting and

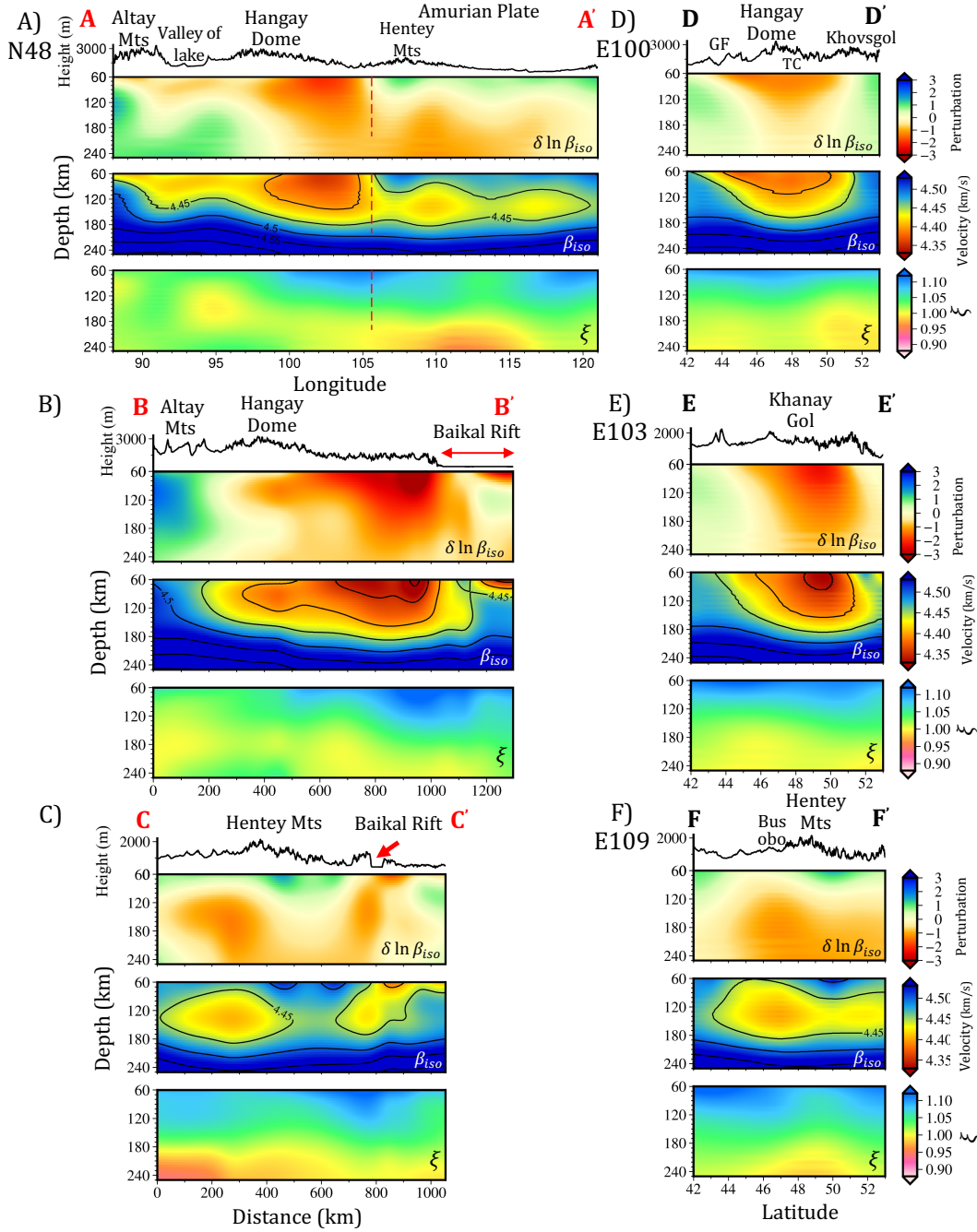


Figure 5.3: Cross-sections of β_{iso} perturbation (top), β_{iso} (middle), and ξ (bottom) along the profiles (A) A-A', (B) B-B', (C) C-C', (D) D-D', (E) E-E', and (F) F-F' shown in Figure 5.2 (g), with surface topography on top. The reference isotropic S wave speed profile (the average of entire Mongolia) used for the perturbation plot is shown as a solid green line in Figure 4.5(b). The surface topography on top is from ETOPO1 (Amante & Eakins, 2009). Major tectonic provinces are labeled on topography. The vertical red dashed lines show the vertical extension of the Amurian plate boundary by Bird (2003).

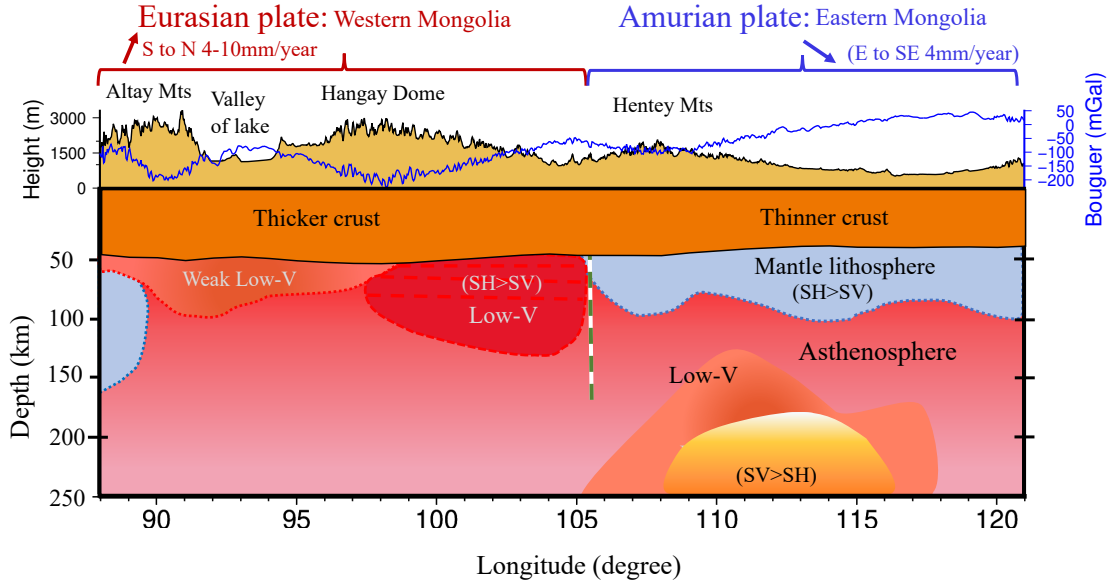


Figure 5.4: Schematic illustration of major structural features of our shear wave model across Mongolia (west-east cross section) at around 48°N. The Moho depth (dark orange part) is taken from the CRUST1.0 model (Laske et al., 2013). The surface topography on top is derived from ETOPO1 (Amante & Eakins, 2009), and the Bouguer gravity anomaly (blue curve on top) is from the WGM2012 model (Bonvalot et al., 2012). A vertical green dashed line indicates the vertical extension of the Eurasian-Amurian plate boundary from Bird (2003).

dynamic processes beneath the Hentey Mountains and Hangay Dome are distinctive. They also suggested that low-velocity anomalies beneath the Hentey Mountains might indicate partially molten or heated material in the upper mantle. Our radial anisotropy model suggests the low-velocity anomalies with radial anisotropy of $\beta_V > \beta_H$ beneath the Hentey Mountains below around 180 km might reflect mantle upwelling (Figures 5.2 (e–h) and 5.3 (A, C, and F)).

Figure 5.4 illustrates the schematic west-east cross section across Mongolia based on our tomography model with other geophysical information, indicating the major structural features and lateral variations in the upper mantle beneath Mongolia at around 48°N. The Moho depth (dark orange part) is taken from the CRUST1.0 model (Laske et al., 2013) and the Bouguer gravity anomaly (blue curve on top) is extracted from the WGM2012 model (Bonvalot et al., 2012). Western Mongolia is characterized by the high elevation, northward motion, thick crust, negative Bouguer anomaly, and the active seismic zone with low-velocity anomaly in the uppermost mantle. On the

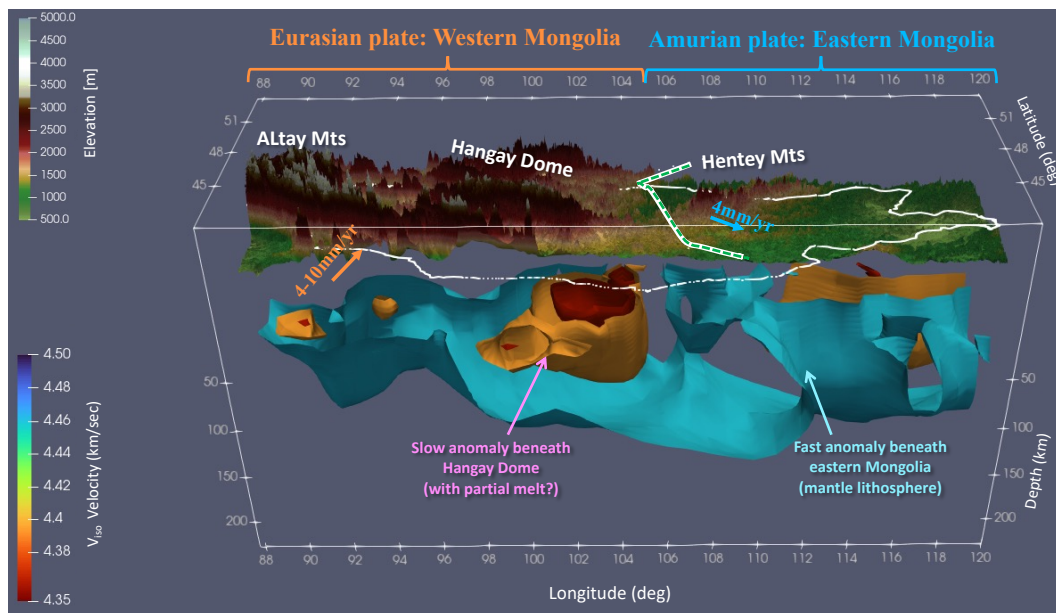


Figure 5.5: Three-dimensional visualization of major structural features of slow and fast shear velocity anomalies of the isotropic S-wave speed model in the Mongolian region.

contrary, eastern Mongolia is characterized by lower elevation, southeastward motion, thinner crust, positive (or weakly negative) Bouguer anomaly, and tectonically stable with high-velocity anomaly in the uppermost mantle.

A typical three-dimensional view of our isotropic S-wave model with isocontour plots is displayed in Figure 5.5, which corresponds to a bird-eye-view version of Figure 5.4. Major structural characteristics of the slow and fast shear velocity anomalies in the Mongolian region at around latitude 48°N is depicted in Figure 5.4; the mantle lithosphere beneath eastern Mongolia has a higher velocity (4.46 km/s contour); a weak low-V anomaly beneath the Altay Mountains (4.39 km/s contour); and a large-scale Low-V anomaly beneath the Hangay Dome (4.36 km/s contour). The distinctive structural variations between western and eastern Mongolia can be related to the differential motion between the Eurasian and Amurian plates. Rapid lateral variations of shear wave speed anomalies in the Mongolian region are reflected well as the major structural features displayed in Figure 5.5.

In Figure 5.6, other geophysical observations in the Mongolian region are displayed along with our 3-D S-wave model in the upper mantle. The crustal thickness taken from the CRUST1.0 model (Laske et al., 2013) is generally thicker in western Mongolia

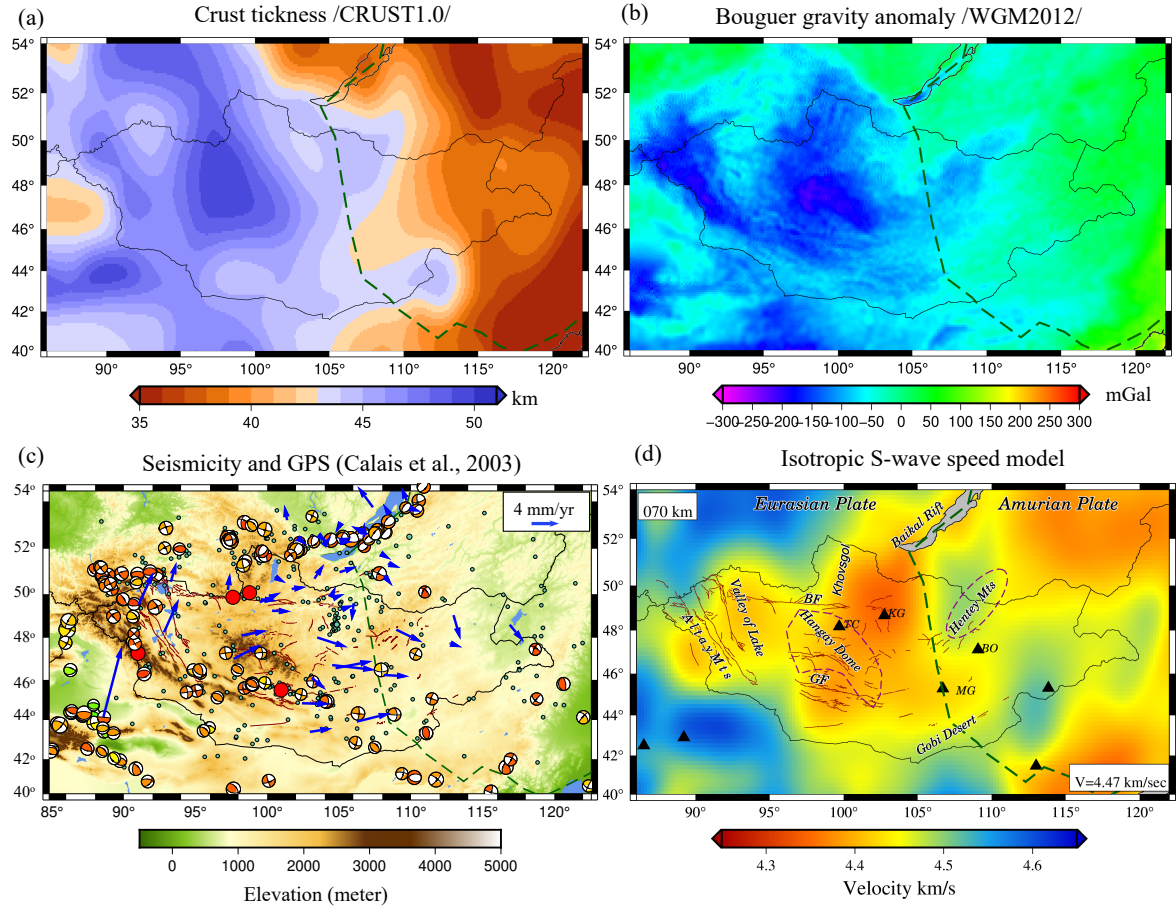


Figure 5.6: Comparison of various geophysical observations in the Mongolian region. (a) Crustal thickness of the Crust1.0 model (Laske et al., 2013), (b) The Bouguer gravity anomaly from the WGM2012 model (Bonvalot et al., 2012), (c) Plate motion from the GPS observation (Calais et al., 2003) and regional seismicity, and (d) isotropic S-wave speed map at the 70 km depth (this study).

(Figure 5.6 (a)), suggesting the large-scale variation of crustal structure between eastern and western Mongolia. The Bouguer gravity anomaly derived from the WGM2012 model (Bonvalot et al., 2012) (Figure 5.6 (b)) shows the significant negative anomalies in western Mongolia, while eastern Mongolia is characterized by weak or slightly positive anomalies, which may indicate the influence of thicker crust in western Mongolia. We can also see a good correlation between the negative gravity anomaly and a slow shear velocity anomaly in western Mongolia (Figure 5.6 (d)). GPS observation by Calais et al. (2003) (Figure 5.6 (c)) suggested that eastern and western Mongolia move in different ways. While central and eastern Mongolia moves about 4 mm/year eastward/southeastward, western Mongolia moves northward with varying speeds of

about 10 mm/year in the southern area and about 4 mm/year in the northern area.

Our S-wave speed model in the upper mantle (Figure 5.6 (d)) suggests fast anomalies in the stable eastern region and slower anomalies in the tectonically active west, which may be related to other geophysical information as above. Further investigations, including joint geophysical inversions of multiple data sets, may be worth doing to unravel the cause of seismic velocity anomalies found in our model.

In the subsequent sections, we will further discuss the major features shown in Figures 5.4 and 5.5, such as the low-velocity Hangay Dome region in western Mongolia (Eurasian plate) in Section 5.2, and relatively stable high-velocity region in eastern Mongolia (Amurian plate) in Section 5.3.

5.2 Hangay Dome: Western Mongolia

The Hangay Dome in western Mongolia is considered one of the most significant tectonic units in this region. Significant earthquakes have not been recorded in the Hangay Dome itself, but some major earthquakes occurred around this area, e.g., the Mw 7.1 earthquake (1967) on the Mogod fault to the east of the Hangay Dome, Mw 8.3 earthquake (1905) on the Bolnay fault to the north, and Mw 8.1 Gobi-Altay earthquake (1957) on the Bogd faults to the south. As mentioned earlier, existing seismic tomography models showed significantly slow shear wave speeds in the upper mantle beneath the Hangay Dome and its surroundings (e.g., Tiberi et al., 2008; Chen et al., 2015a; Comeau et al., 2018; He et al., 2019; Wu et al., 2021; Feng, 2021; Wang et al., 2022; Huang & Zhao, 2022).

Chen et al. (2015a) found a large-scale slow anomaly in the upper mantle beneath the Hangay Dome, the Khovsgol rift system, parts of the Gobi-Altay Mountains, and the Sayan Mountains between 50 and 90 km depth, although the data set used in their tomography model lacks local seismic data in Mongolia. With low shear velocity (V_{SV}) of less than 4.2 km/s, they proposed that partial melting might spread over the area beneath the Hangay Dome and its surroundings. However, a recent shear velocity (V_{SV}) model by Feng (2021), who employed more extensive local data sets in this region, demonstrated that the partial melting zones are distributed not directly beneath the Hangay Dome. The partial melting and convective loss of the lithosphere

may result in the uplift of Hangay Dome and may cause such slow anomalies. Recent P-wave tomography by Huang & Zhao (2022) suggested a velocity anomaly around 250 km depth, indicating that the lithosphere beneath the Hangay Dome may have been delaminated (Zhao et al., 2021).

Numerical experiments by Elkins-Tanton (2005) on continental delamination suggested the necessity of ductile mid-lithosphere to allow delamination, implying that the delamination can occur in regions that have experienced high heat flow from volcanism or mantle upwelling. Based on thermo-mechanical numerical models incorporating geological, geochemical, and geophysical data, Comeau et al. (2021) have investigated the conditions for lithospheric removal by delaminations. They have found that central Mongolia satisfied the conditions for lithospheric delamination, supported by lines of evidence such as the high and dome-shaped surface topography, thin lithosphere, and elevated temperature at the crust-mantle interface. These findings suggest that mechanical delamination or thermal processes may be to blame for the lithosphere beneath the Hangay Dome being significantly weaker than the surrounding regions (Fullea et al., 2012).

Vertical cross-sections along profiles A-A', B-B', and D-D' in Figure 5.3 display the structure below the Hangay Dome. Anomalous low velocity beneath the Hangay Dome in our model is consistent with recent results by Feng (2021), showing low-velocity anomalies down to about 150 km depth, mainly beneath the northeast of the Hangay Dome, Khanuy Gol volcanic field. These low-velocity anomalies persist northeastward in the upper mantle beneath the Khovsgol and the south of Baikal rift regions, reaching a depth of about 200 km (Figure 5.3 (B)). The slow anomaly tends to be confined at shallower depths above 80 km right beneath the Baikal rift, though the lateral resolution of our model is limited in the north of Mongolia.

Beneath the Hangay Dome and on the northeastern side of the Hangay Dome, Khanuy Gol volcanic field, our 3-D model (Figures 5.1 and 5.2) shows slower SV wave speed anomalies with positive radial anisotropy ($\beta_H > \beta_V$), particularly at depths of about 70 – 150 km (Figure 5.3 (A), (B), and (D)). Such features may indicate the presence of layered or laminated heterogeneity, possibly including partially molten layers, which can generate significant positive radial anisotropy ($\beta_H > \beta_V$) accompanied

by slow shear velocity anomaly on average (e.g., Aki, 1968). Thus, our results suggest that the Hangay Dome region may be accompanied by partially molten layers in the top 100 km depth, where strong positive radial anisotropy can be seen in our model. The lithosphere here may be significantly weaker or may have been lost or delaminated, as indicated by Feng (2021), Zhao et al. (2021), Huang & Zhao (2022), and Wu & Bao (2023).

5.3 Amurian plate and its western boundary: Eastern Mongolia:

Zonenshain & Savostin (1981) proposed the "Amurian plate" covering eastern Mongolia, north China, and southeastern Russia in the extensional areas southeast of Lake Baikal. The Amurian plate and its geometrical boundary in East Asia have been studied extensively using space geodetic observations (e.g., Takahashi et al., 1999; Heki et al., 1999; Jin, 2013), a six-plate analysis of earthquake slip vectors and NUVEL-1 data (Wei & Seno, 1998), kinematics of the active faults (Sankov et al., 2022), and topography and seismicity (e.g., Bird, 2003). The southernmost edge of the Amurian plate is unclear compared to other areas. Heki et al. (1999) suggested that the south boundary of the Amurian Plate may be the Qinling Fault because it is consistent with the Euler vector and predicts left-lateral movement along its boundary with the South China block. However, Jin (2013) defined the tectonic boundary between the North China and Amurian plates as the Yin Shan–Yan Shan Mountain belts with about 2.4 mm/yr extension. Because of the sparse and limited data, these earlier investigations of tectonics in this region are inconclusive. Thus, the existence of the Amurian plate and the geometry of its boundary remains to be determined.

Figure 4.8 shows the isotropic S wave speed (β_{iso}) and radial anisotropy (ξ) from 80 to 200 km depths in the broader areas covering the whole Amurian plate. Corresponding cross-sections along the selected profiles in Figure 4.8 (g) are displayed in Figure 4.9, which also indicate the location of the Amurian plate boundary by Bird (2003).

The arrows in Figure 4.8 (a) show the relative plate motion; the Indian plate moves northward at a rate of about 44 mm/year, the Pacific, Okhotsk, and Philippine Sea plates move northeastward at about 93 mm/year, 84 mm/year and 67 mm/year, re-

spectively, and the Amurian plate moves southeastward at about 5 mm/year, for the fixed Eurasian plate based on the MORVEL plate model (DeMets et al., 2010). The GPS measurements in Mongolia by Calais et al. (2003) are displayed with black arrows in Figure 5.2 (c). In western Mongolia, we can see northward shortening in the Altay area, which varies from 10 mm/year in the south to 4 mm/year in the north. However, central and eastern Mongolia moves eastward or southeastward at a rate of about 4 mm/year.

Owing to a large number of local broadband seismic stations in Mongolia, our new 3-D model shows distinct lateral variations near the expected boundary between the Eurasian and Amurian plates (Figures 5.3 (A), 5.4, 5.5, 4.8 (a, c, e), and 4.9 (A, B, E, F)), especially near the northern and western edges of the Amurian plate. Still, the southern border of the Amurian plates in China is unclear in our model, where our station coverage is limited. We can observe a notable velocity contrast between the Eurasian and Amurian plates along the Stanovoy range in the depth slices from 80 to 150 km (Figure 4.8). The Baikal Rift is one of the surface evidence that divides the Amurian and Eurasian plates, where the Amurian plate moves southeastward relative to the Eurasian plate at a rate of 3.4–5 mm/yr (Calais et al., 2003). The Baikal Rift Zone is characterized by a low-velocity anomaly at shallow depths above 70–80 km as seen in the cross-sections of Figures 4.9 (A), 5.3 (B, C). The western area of the Amurian plate (in eastern Mongolia) is characterized by fast anomaly which is likely to represent a comparatively thin (around 70–100 km) but relatively stable lithosphere, as seen in the cross-sections (Figure 4.9 (A, B, E, F) and Figure 5.3).

Figure 4.9 (A, E) indicates that the Amurian lithosphere beneath eastern Mongolia is much thinner than the surrounding cratonic lithosphere, such as the North China craton (e.g., Xu et al., 2022) and Siberian craton (e.g., Lebedev et al., 2006), whose thickness reaches about 200 km. Based on seismological evidence of upper-mantle tomography and seismic anisotropy, Lebedev et al. (2006) argued that this asymmetry between the thick cratonic lithosphere in Siberia (~ 200 km) and the thinner lithosphere in eastern Mongolia (60–70 km) may affect the intra-continental volcanism in the Baikal Rift area, accompanied by subhorizontal flow in the shallower asthenosphere beneath eastern Mongolia. This hypothesis is consistent with our model, showing the $SH > SV$

radial anisotropy with a slow anomaly in the asthenosphere beneath the thinner eastern Mongolian lithosphere.

The lateral variations of the S wave model at 75 km depth (Figure 5.2 (a)) show slow velocities in seismically active regions in western Mongolia (1964 – 2023 $M_l \geq 3.5$) and fast velocities in relatively stable eastern Mongolia, which is likely to reflect the tectonic and structural differences between the two regions.

Note that, although our 3-D model well constrains the large-scale ($> 2^\circ$) upper mantle structure below 60 km depth, the shallower structure, including the crust, cannot be fully constrained due to the limited vertical resolution of the dispersion data used in this study. Also, the long-wavelength surface waves tend to blur the lithosphere images, limiting the lateral resolution of surface wave tomography. Additional seismic data, such as ambient noises and H/V ratios of Rayleigh waves, may help constrain the shallower structure. Moreover, body-wave travel times from local and teleseismic events, as well as receiver functions, will also help in higher-resolution imaging of mantle structure. Such further seismological constraints will enable us to clarify more detailed tectonic and dynamics processes in the enigmatic western border of the Amurian plate.

Chapter 6

Conclusions

6.1 Summary of this thesis

We constructed a new radially anisotropic 3-D S-wave model in and around the Mongolian region from multimode surface waves using the three-stage inversion technique of surface-wave tomography (Yoshizawa & Kennett, 2004). By combining broadband seismograms from the permanent and temporary seismic networks in Mongolia as well as neighboring GSN, PASSCAL, and NECESSArray stations, we could obtain the high-resolution 3-D S wave model under the whole areas of Mongolia, which provides new insight into large-scale tectonics of the Mongolian region, including the western border of the Amurian plate.

The data sets and multimode phase speed measurements – the initial step of the three-stage inversion technique – were presented in Chapter 2. We applied the method of nonlinear waveform fitting with the Neighborhood Algorithm. Through waveform fitting, phase speed dispersion curves for the fundamental and higher modes were extracted by matching synthetic and observed seismograms for each path. Utilizing many large earthquakes that occurred in the world, we analyzed seismic waveform data from many seismic stations distributed in and around Mongolia to measure the multimode phase speeds of Rayleigh and Love waves. We could collect many reliable path-specific phase speeds of approximately 70,000 Rayleigh wave paths, and 40,000 Love wave paths for the fundamental mode, most of which are in a reasonable range of standard deviations around the average dispersion curves.

In Chapter 3, we presented the second stage of the three-stage tomography, in which the collected dispersion curves were used to reconstruct multimode phase speed

maps as a function of frequency. We used the linearized inversions for surface-wave tomography taking account of the finite-frequency effects. We obtained phase speed maps from the fundamental to 4-th higher modes in the period range between 30 and 200 s. The general patterns of the fundamental-mode maps of Rayleigh and Love waves are consistent with the global model GDM52, but the lateral resolution of our model in the target region is significantly higher than the global model.

Chapter 4 described the construction of 3-D SH and SV wave structures using the multimode phase speed maps of Rayleigh and Love waves from the previous step. Following Yoshizawa (2014), we employed the method of generalized nonlinear inversions (Tarantola & Valette, 1982) for radially anisotropic S-wave speed models. The local dispersion curves were inverted for 1-D SH and SV wave speed models at each grid point of the region. The 3-D SH and SV wave speed models were constructed by assembling the local 1-D profiles. The multimode phase speed models over a broad period range enable us to construct the continental upper mantle structure of this region.

In Chapter 5, we discussed the relationship between the current tectonics and mantle dynamics under Mongolia, particularly focusing on the Amurian Plate and Hangay Dome based on our 3-D radially anisotropic shear wave model. The retrieved S-wave speeds in the upper mantle beneath Mongolia are generally slower in the tectonically active west and faster in the stable east above 100 km depth. The significant velocity contrast between western and eastern Mongolia corresponds well with existing geophysical studies in this region, which is likely to reflect the enigmatic plate boundary between the Eurasian and Amurian plates. In western Mongolia, two major low-velocity anomalies are found beneath the northern Altay Mountains and the Hangay Dome. The former persists to 100 km depth, which may relate to the northward movement of Asian blocks due to the India-Eurasia collision (Huang & Zhao, 2022). The latter continues down to about 150 km depth beneath the Hangay Dome region, particularly the Khanuy Gol volcanic field in the northeast of the Hangay area, accompanied by significant positive radial anisotropy with slower SV in the top 100 km which might indicate the presence of partially molten layers. In eastern Mongolia, located in the western Amurian plate, high velocities in the upper mantle suggest the lithospheric signature of the Amurian plate. The fast anomaly in eastern Mongolia indicates the

relatively stable lithosphere with a thickness of around 70–100 km, which is not seen in the west of the expected Amurian plate border. The new 3-D S-wave model exhibits significant velocity contrast between the Eurasian and Amurian plates and structural differences in the underlying upper mantle, implying the differential motions in eastern and western Mongolia can be related to such structural variations and local mantle dynamics.

6.2 Future directions

Seismic waves provide a variety of information on the Earth's internal structure with a better spatial resolution than other geophysical observations, such as gravity and geomagnetic data. Many kinds of seismological observations and analysis methods, such as surface wave dispersion, body wave travel times, ambient noises, receiver functions, and the ellipticity (or H/V ratio) of Rayleigh waves, have been employed to scrutinize the structure of the crust and upper mantle in continental areas. Yet, these seismological observables have different sensitivities and lateral and vertical resolutions. Each of these observed data and methods of analysis has some advantages and disadvantages, so the joint use of multiple seismological observations generally provides us with a stable estimation of the structure of the Earth. Here, we summarize some ideas for further studies to expand our knowledge of the seismic structure in the Mongolian region.

6.2.1 Improved constraints on the crustal structure

Using the multimode surface wave dispersion data in a period range from 30 – 200 s, the large-scale ($> 2^\circ$) upper mantle structure below 60 km depth has been well constrained by our 3-D model in this study. However, the vertical sensitivity of the dispersion data in this period range tends to be limited to the upper mantle (Figure A.1), which prevents constraining the shallower structure, including the crustal velocity and the Moho depth. Thus, additional seismological observables that may be more sensitive to the shallower structure, such as the short-period dispersion curves (e.g., down to about 10 s period or less) from ambient noises (e.g., Bao et al., 2015) and Rayleigh-wave ellipticity (H/V ratios) (e.g., Attanayake et al., 2017), might be useful in constraining the crustal structure. Such surface-wave observables can be jointly used with our

longer-period phase speed data sets, which will enable us to construct the crust and upper mantle models with improved vertical resolution.

6.2.2 Joint use of body and surface waves

Body-wave travel times from teleseismic and local events will enable us to retrieve higher-resolution mantle structure images since the wavelength of body waves is generally shorter than that of the long-period surface waves used in this study. Body wave data are also helpful in retrieving the deeper mantle structure at the mantle transition zone or below (e.g., Zhang et al., 2017). With the help of such additional seismological constraints, we will be able to better understand the tectonic and dynamic processes in this area.

6.2.3 Seismic discontinuities in the crust and upper mantle

Surface wave dispersion data used in this study are not very sensitive to discontinuities in the Earth. On the contrary, receiver functions that represent the converted body wave phases are more sensitive to the impedance contrast across the interface, which can constrain the depth of internal discontinuities and relative velocity differences (e.g., Taira & Yoshizawa, 2020). The joint inversions of multimode surface waves and receiver functions will enable us to retrieve the crust and upper mantle structure with internal discontinuities, such as the MLD (Mid-Lithosphere Discontinuity) and LAB (Lithosphere-Asthenosphere Boundary) beneath the continental lithosphere, which can be of great use for better constraining the 3-D anisotropic S-wave models and further detailed arguments on tectonic evolution in this region.

Appendix A

Methods of Three-stage Surface Wave Tomography

In this study, we have used the three-stage inversion method for multimode surface wave tomography developed by Yoshizawa & Kennett (2004). This appendix summarizes the essence of the practical methods in each stage.

A.1 1st-stage: Multimode phase speed measurements

In Chapter 2, we employed a fully nonlinear waveform fitting method by Yoshizawa & Ekström (2010), which is an extended version of the original technique developed by Yoshizawa & Kennett (2002b) for the multimode phase speed measurements. To match the observed synthetic seismograms, we utilize the Neighbourhood Algorithm (NA) by Sambridge (1999), which allows us to explore a model parameter space efficiently. Here we briefly explain the fundamental aspects of this method.

Based on the WKBJ approximation (e.g., Dahlen & Tromp, 1998), a synthetic seismogram in the frequency domain can be given as follows,

$$u(\Delta, \omega) = \sum_{j=0}^N R_j(\omega) \exp \{i(k_j^0(\omega) + \delta k_j(\omega))\Delta\} S_j(\omega), \quad (\text{A.1})$$

where Δ is the epicentral distance, ω is the angular frequency, R_j is the receiver term for the j -th mode, which includes attenuation and geometrical spreading terms, and S_j is the source term for the j -th mode derived from a given moment tensor. The exponential term represents the surface-wave propagation of the j th mode with a wavenumber k_j^0 for the reference model, and the wavenumber perturbation δk_j due

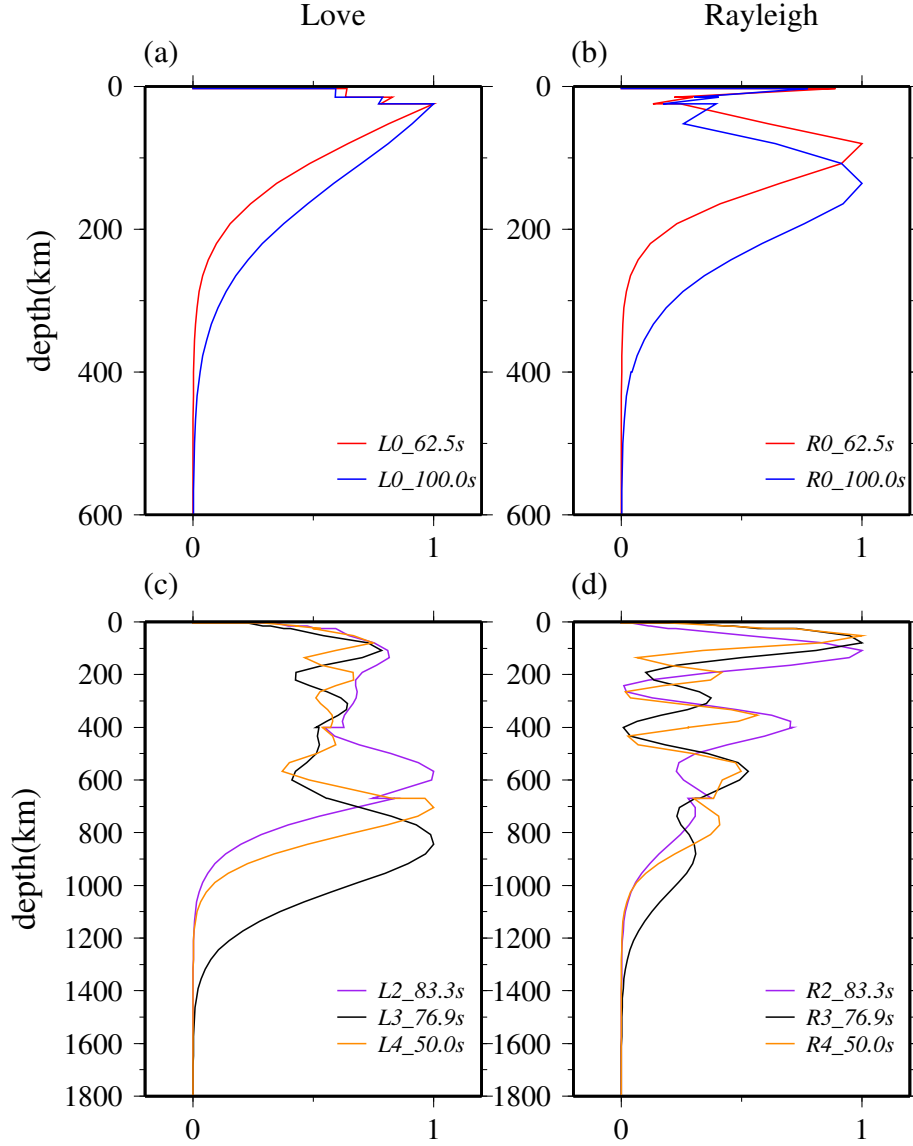


Figure A.1: Normalized sensitivity kernels of (a, c) Love waves and (b,d) Rayleigh waves for (a, b) fundamental mode and (c, d) higher modes calculated for PREM.

to a lateral heterogeneity. N is the number of modes used to compute full synthetic seismograms.

δk_j can be represented by the vertical integration of model parameters (perturbations of S-wave speed $\delta\beta$, P-wave speed $\delta\alpha$, and density $\delta\rho$), multiplied by the sensitivity kernels for each parameter, as follows,

$$\delta k_j(\omega) = \int_0^{R_E} \{K_\alpha^j(r, \omega)\delta\alpha(r) + K_\beta^j(r, \omega)\delta\beta(r) + K_\rho^j(r, \omega)\delta\rho(r)\}dr, \quad (\text{A.2})$$

where R_E is the Earth's radius, K_α^j , K_β^j , and K_ρ^j represent the sensitivity kernels of

Table A.1: Threshold values of the reliability parameters (for the fundamental and higher modes) and normalized radiation amplitude as a function of waveform misfit range, which are used to select reliable multimode phase speeds following Yoshizawa & Ekström (2010).

Waveform misfit	Minimum reliability		Minimum radiation
	Fundamental mode	Higher modes	
0.00 – 0.01	8.0	2.0	0.5
0.01 – 0.02	9.0	2.5	0.6
0.02 – 0.03	10.0	3.0	0.7

surface-wave phase speed of j -th mode for P-wave speed α , S-wave speed β , and density ρ , respectively. Figure A.1 displays examples of K_β for several modes. Surface-wave phase speeds are primarily sensitive to S-wave speed, with weak influence from P-wave speed and density. In general applications, the S-wave speed perturbation is employed as an independent model parameter, and P-wave speed and density perturbations are often disregarded or scaled to the S-wave speed perturbation. In this study, we considered the influence of P-wave speed and density through the conventional scaling relationship with the S-wave speed perturbations (e.g., Panning & Romanowicz, 2006).

We employed a set of B-spline functions, $B(r)$, as a function of radius r , to express the S-wave speed perturbation $\delta\beta(r) = \sum_{i=1}^M b_i B_i(r)$, where b_i are the model parameters to be searched by the NA that enables us to find models with minimum misfits (or global minima) in the model parameter space. The resultant best 1-D model, which provides the minimum misfit between the observed and synthetic seismograms, is used to compute the multimode phase speed curves as shown in Figures 2.3 and 2.4. The reliability parameters are computed by evaluating the waveform misfit and relative power (amplitude) of each mode with respect to the full synthetic seismograms, providing us with the reliability of measured phase speeds for each mode and frequency. To select the reliable phase speeds that will be used in the subsequent tomographic inversions, we have employed multiple quantitative information, including the waveform misfit, the reliability parameter, and the relative radiation amplitude of each mode from the seismic source. The threshold values for our data selection criteria are the same as the original method by Yoshizawa & Ekström (2010), summarized in Table A.1.

A.2 2nd-stage: Inversions for multimode phase speed maps

For the multimode phase speed mapping in Chapter 3, we employed the linearized tomographic inversions considering approximate influences of finite frequency (Yoshizawa & Kennett, 2002a, 2004). In this section, the methodology for the finite-frequency tomography is briefly summarized. We consider a ray-centered coordinate system (s, n) , where s is the distance along a path and n is a perpendicular axis to the path. The relationship between a phase perturbation of seismograms $\delta\psi$ and a two-dimensional distribution of phase speed perturbation $\delta c(s, n)$ may be expressed as a linearized form,

$$\delta\psi = \int_{path} ds \int_{width} dn K_{\psi}(s, n) \frac{\delta c(s, n)}{c_0}, \quad (\text{A.3})$$

where K_{ψ} is the two-dimensional sensitivity kernels for phase speed, and $\delta c/c_0$ is a fractional phase speed perturbation to be obtained by the inversion. The observed phase perturbation $\delta\psi^{obs}$ for each path can be related to the observed phase speed perturbation $\langle\delta c\rangle^{obs}$ as, $\delta\psi^{obs} = -k_0\Delta\langle\delta c\rangle^{obs}/c_0$, where k_0 and c_0 are respectively the reference wavenumber and phase speed, and Δ is the distance along a path.

The linearized equation (A.3) may be represented by a general vector form, $\mathbf{d} = \mathbf{G}\mathbf{m}$, where $\mathbf{d}$ is a data vector, $\mathbf{G}$ is the kernel matrix, and $\mathbf{m}$ is a model parameter vector. The components of $\mathbf{m}$ can be given as $m_j (j = 1, 2, \dots, N)$, where N is the number of model parameters. We use spherical B-spline functions $F_j(\theta, \phi)$ (Wang & Dahlen, 1995) at the rectangular geographical grids to express the phase speed perturbation on a sphere,

$$\frac{\delta c(\theta, \phi)}{c_0} = \sum_{j=1}^N m_j F_j(\theta, \phi). \quad (\text{A.4})$$

Thus, our model parameter m_j corresponds to the coefficient for the j -th B-spline function F_j . The grid (or knot) spacing of this study depends on the ray-path coverage for each mode and frequency; i.e., 2.0° for the fundamental mode with more than 30,000 paths, 3.0° for more than 15,000 paths, 4.0° for more than 8,000 paths, and 5.0° for all modes with more than 5000 paths. We also considered additional criteria for the minimum grid size for each mode and frequency so that the grid size must not be smaller than half of the wavelength.

We solved the linear equation iteratively, using a damped least-squares approach

based on the LSQR method by Paige & Saunders (1982). To apply weight on each phase speed datum, we used the reliability parameters for the individual measurements in the previous section. Damping parameters that control the trade-off between the model roughness and the data misfit are carefully checked for each phase speed model via visual checks of trade-off curves for each mode and frequency. Then, the best-compromised models are selected to be used in the final stage to construct the 3-D S-wave model.

A.3 3rd-stage: Inversions for local 1-D S wave models

We employed the generalized nonlinear least-squares inversion of Tarantola & Valette (1982) to construct the radially anisotropic 3-D S wave speed model in Chapter 4, following Yoshizawa (2014). To represent the radial anisotropy (or transverse isotropy with vertical symmetry axis), we consider six independent model parameters (ρ , α_H , α_V , β_H , β_V , η), where, ρ is density, α_H and α_V represent horizontally and vertically polarized P wave speeds, β_H and β_V represent horizontally and vertically polarized S wave speeds, η is an anisotropic parameter that controls the incident-angle dependence of seismic wave speeds.

The relationship between observed phase speed perturbation and model parameter perturbations can be represented in a linearized form as follows (e.g., Takeuchi & Saito, 1972; Dahlen & Tromp, 1998),

- For Rayleigh wave:

$$\frac{\delta c_R^{(n)}}{c_0} = \int_0^{R_E} \left[K_{\alpha_H}^{(n)} \frac{\delta \alpha_H}{\alpha_{H_0}} + K_{\alpha_V}^{(n)} \frac{\delta \alpha_V}{\alpha_{V_0}} + K_{\beta_H}^{(n)} \frac{\delta \beta_H}{\beta_{H_0}} + K_{\beta_V}^{(n)} \frac{\delta \beta_V}{\beta_{V_0}} + K_{\eta}^{(n)} \frac{\delta \eta}{\eta_0} + K_{\rho}^{(n)} \frac{\delta \rho}{\rho_0} \right] dr, \quad (\text{A.5})$$

- For Love wave:

$$\frac{\delta c_L^{(n)}}{c_0} = \int_0^{R_E} \left[K_{\beta_H}^{(n)} \frac{\delta \beta_H}{\beta_{H_0}} + K_{\beta_V}^{(n)} \frac{\delta \beta_V}{\beta_{V_0}} + K_{\rho}^{(n)} \frac{\delta \rho}{\rho_0} \right] dr, \quad (\text{A.6})$$

where R_E is the Earth radius, $\delta c_R^{(n)}$, $\delta c_L^{(n)}$ are phase speed perturbations of the n th-mode Rayleigh and Love waves. $K_{\rho}^{(n)}$, $K_{\alpha_V}^{(n)}$, $K_{\alpha_H}^{(n)}$, $K_{\beta_H}^{(n)}$, $K_{\beta_V}^{(n)}$ and $K_{\eta}^{(n)}$ are the vertical

sensitivity kernels for each model parameter (or partial derivatives of phase speed with respect to model parameters).

Rayleigh-wave phase speeds are the most sensitive to SV wave speeds (β_V), while the Love-wave phase speeds to SH wave speeds (β_H). Thus, we jointly invert the multimode phase speeds of both Rayleigh and Love waves for the SV and SH wave speeds simultaneously. Since the sensitivity to P-wave speeds and density is lower than that to S-wave speeds, we employ a conventional scaling relation for density and P-wave speeds in proportion to S-wave speeds, in the same way as Yoshizawa (2014). Therefore, we consider only the perturbations of SV and SH wave speed as model parameters for inversions.

Appendix B

Phase Speed Maps

In this appendix, we show many phase speed maps and the results of additional model recovery experiments to confirm the validity of our phase speed data sets and 2-D phase speed maps, which are the basis of the final 3-D S-wave model.

B.1 Phase speed maps of the whole region of inversion

In this section, we summarize ray-path coverage (Figure B.1) and phase speed maps (Figure B.2) for some selected modes and periods, which display the whole areas of inversions that encompass all the seismic events and stations used in our tomographic inversions.

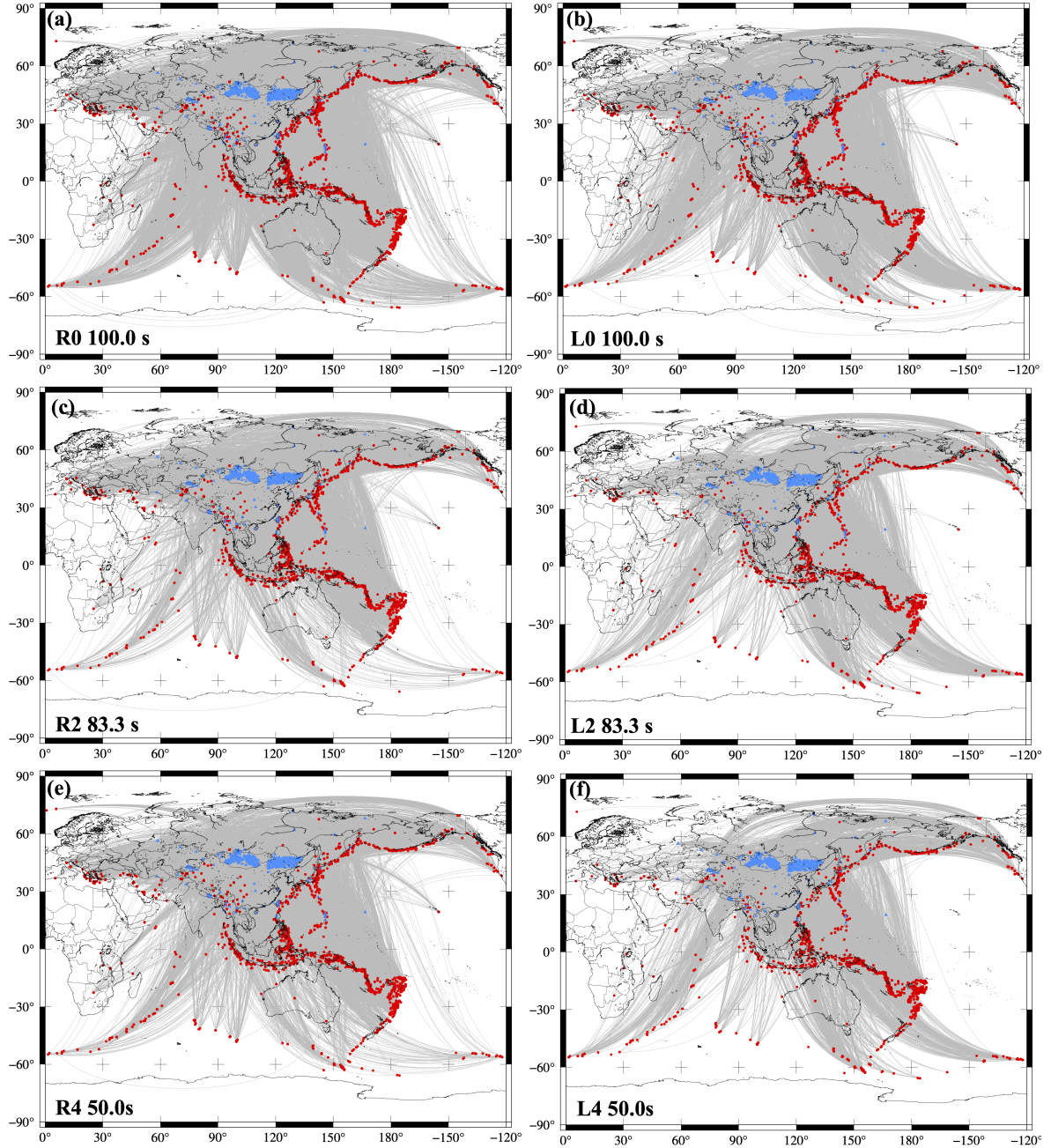


Figure B.1: The ray-path coverage of the selected periods of the fundamental and higher modes of Rayleigh and Love waves. Blue triangles represent all employed seismic stations, and red dots seismic events.

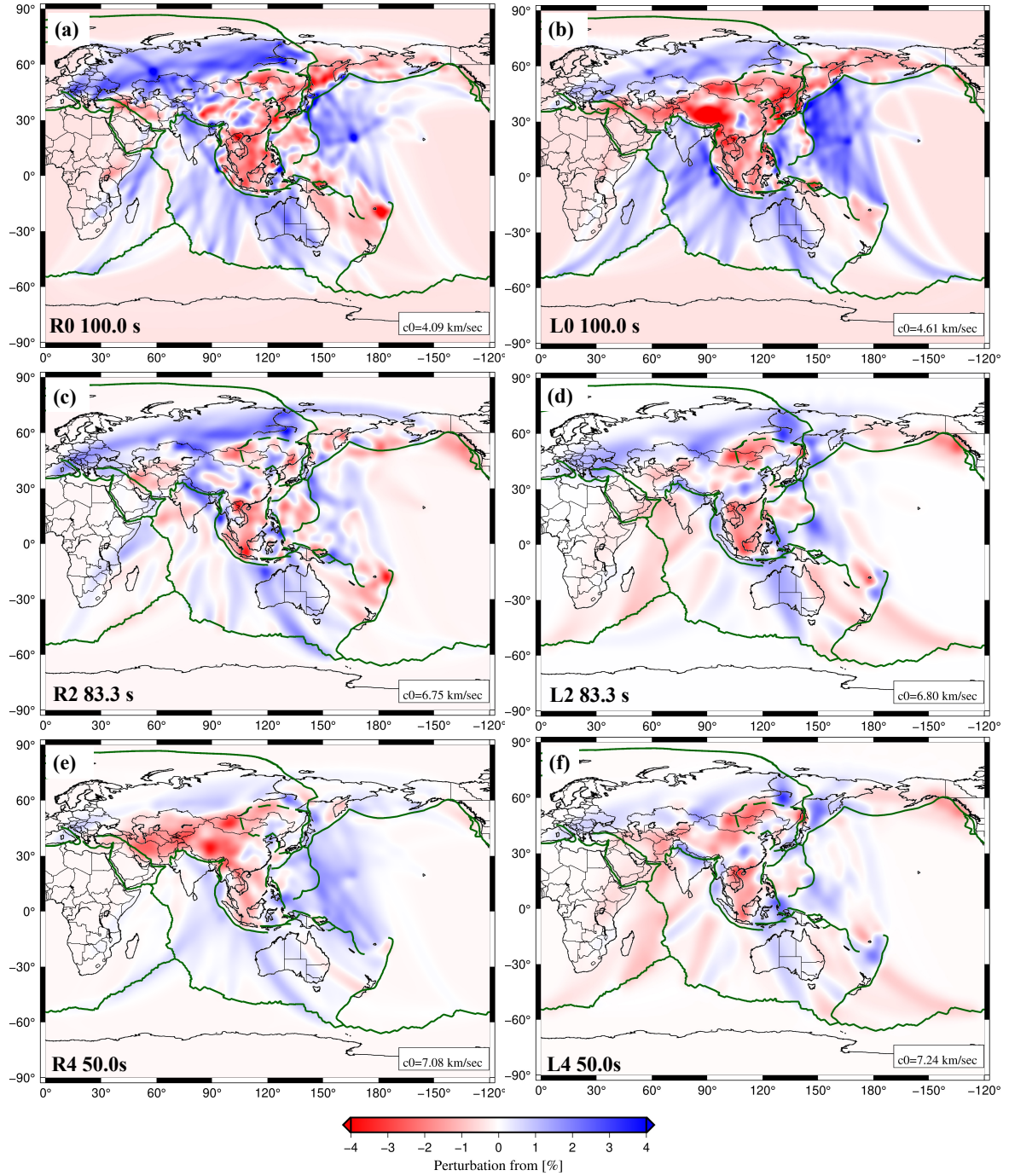


Figure B.2: Phase speed maps of the entire inversion area for the selected periods of the fundamental and higher modes of Rayleigh and Love waves. The corresponding ray paths are shown B.1.

B.2 Model recovery tests

To check the validity of the tomography models and the plausible artifact that may be caused by biased ray coverage, we performed model recovery tests, following similar procedures as we have done in the checkerboard resolution tests. Instead of using input checkerboard models, we have employed pre-computed phase speed maps as an input model. Synthetic phase speeds are computed along the great-circle paths using the input phase speed model. These synthetic phase speed data are treated as observed data, and tomographic inversions are performed to reconstruct the phase speed model by assuming the great-circle path propagation. The model parameterization and damping parameters are the same as those used in real data inversions.

Unlike the checkerboard tests, which consider a specific scale of regular-grid heterogeneity that includes only the structural information for a specific wavelength (or a wavenumber), this experiment, incorporating the realistic heterogeneity models, allows us to visually confirm the retrieval of the multi-scale heterogeneity over a wider wavenumber range.

The results of the model recovery tests are shown in Figures B.3 and B.4. In all cases, the large-scale model patterns and heterogeneity strength are fairly well recovered despite some minor differences in the smaller-scale heterogeneity.

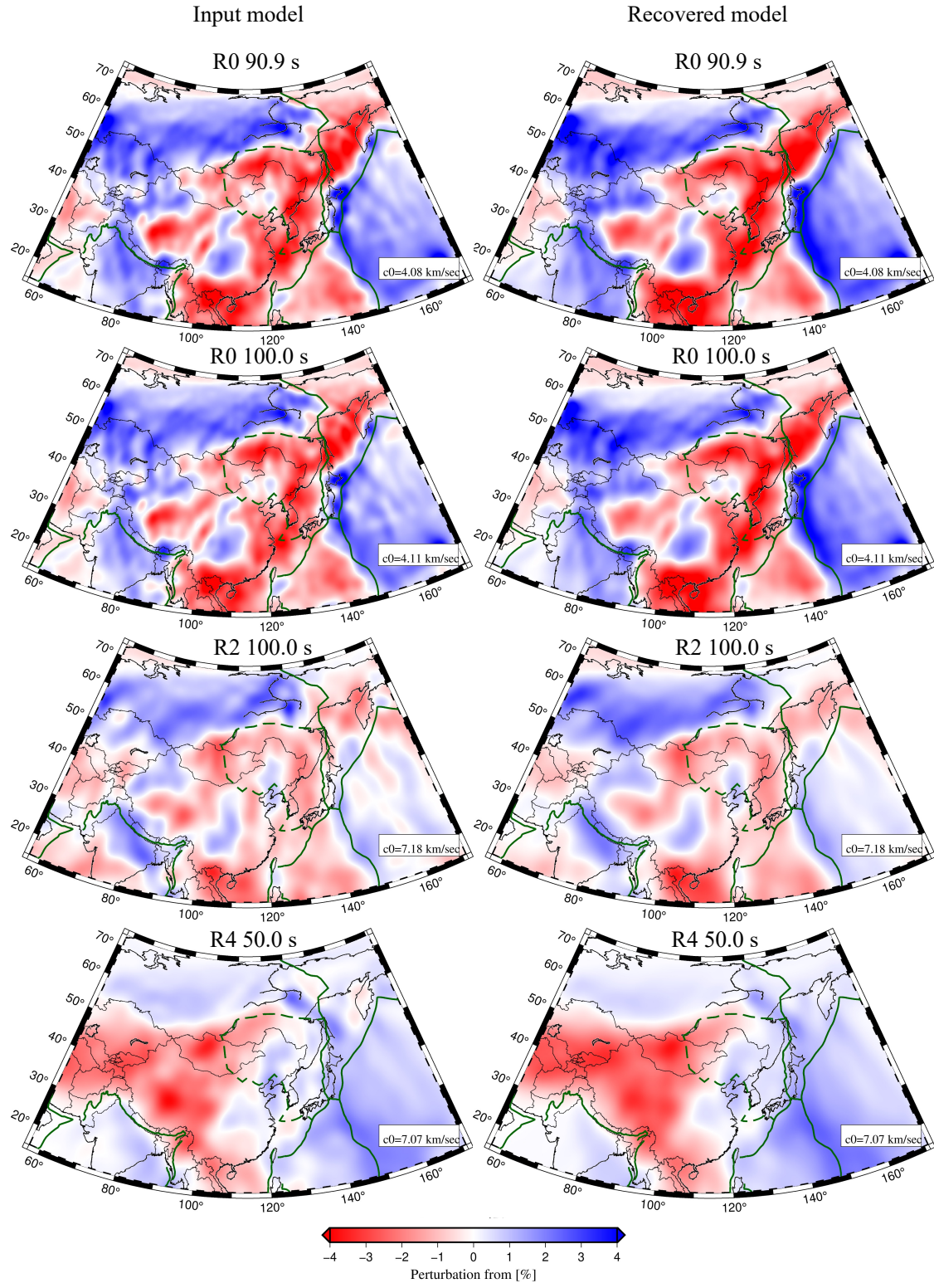


Figure B.3: Model recovery tests for multimode Rayleigh waves. (Left) Input models (the phase speed maps obtained in this study), and (right) recovered models.

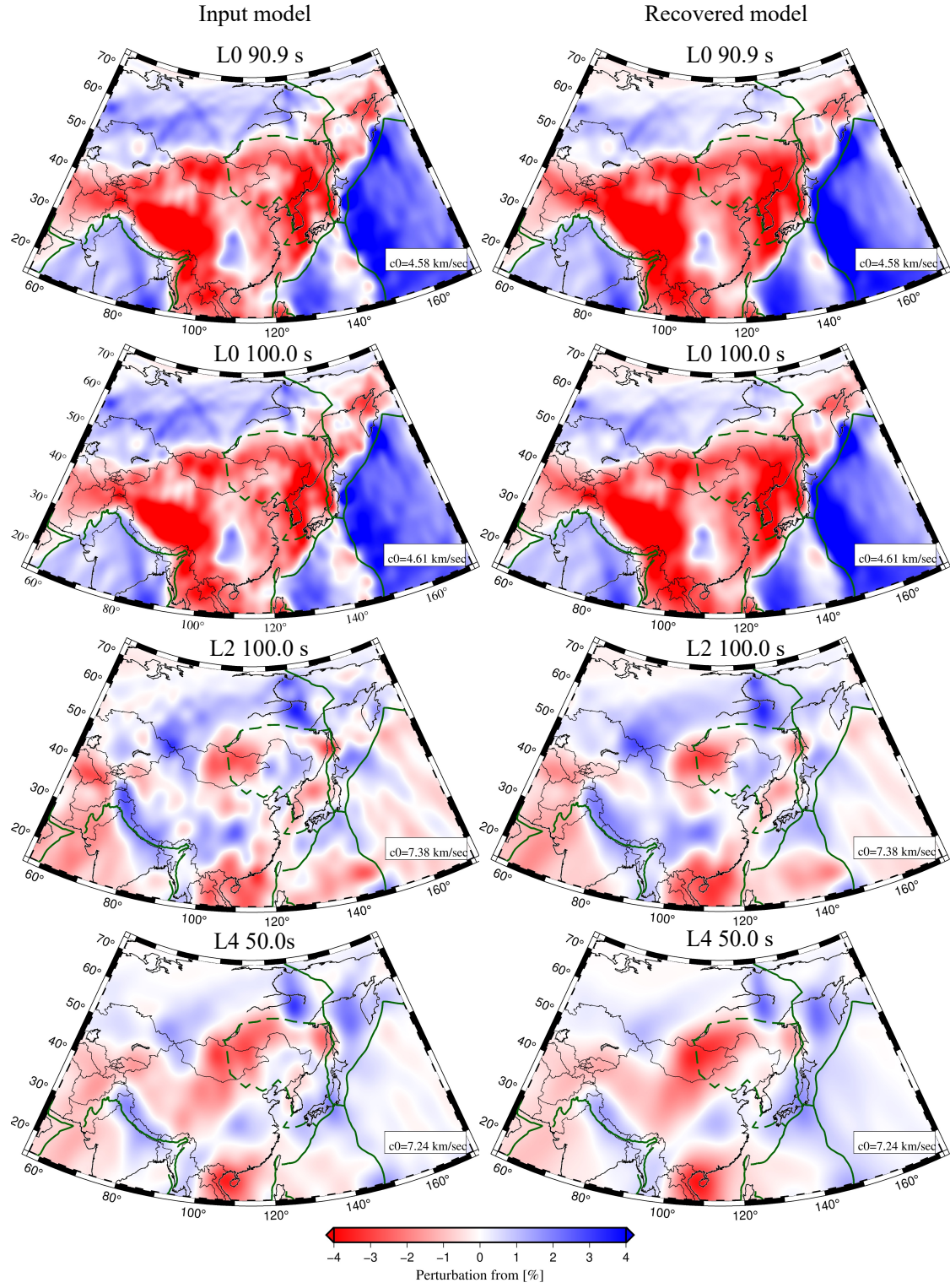


Figure B.4: Same as Figure B.3, but for multimode Love waves.

B.3 Phase speed maps gallery

In this section, we summarize selected multimode phase speed maps (Figures B.5–B.8), which have been used for the construction of the radially anisotropic 3-D S-wave model of this study.

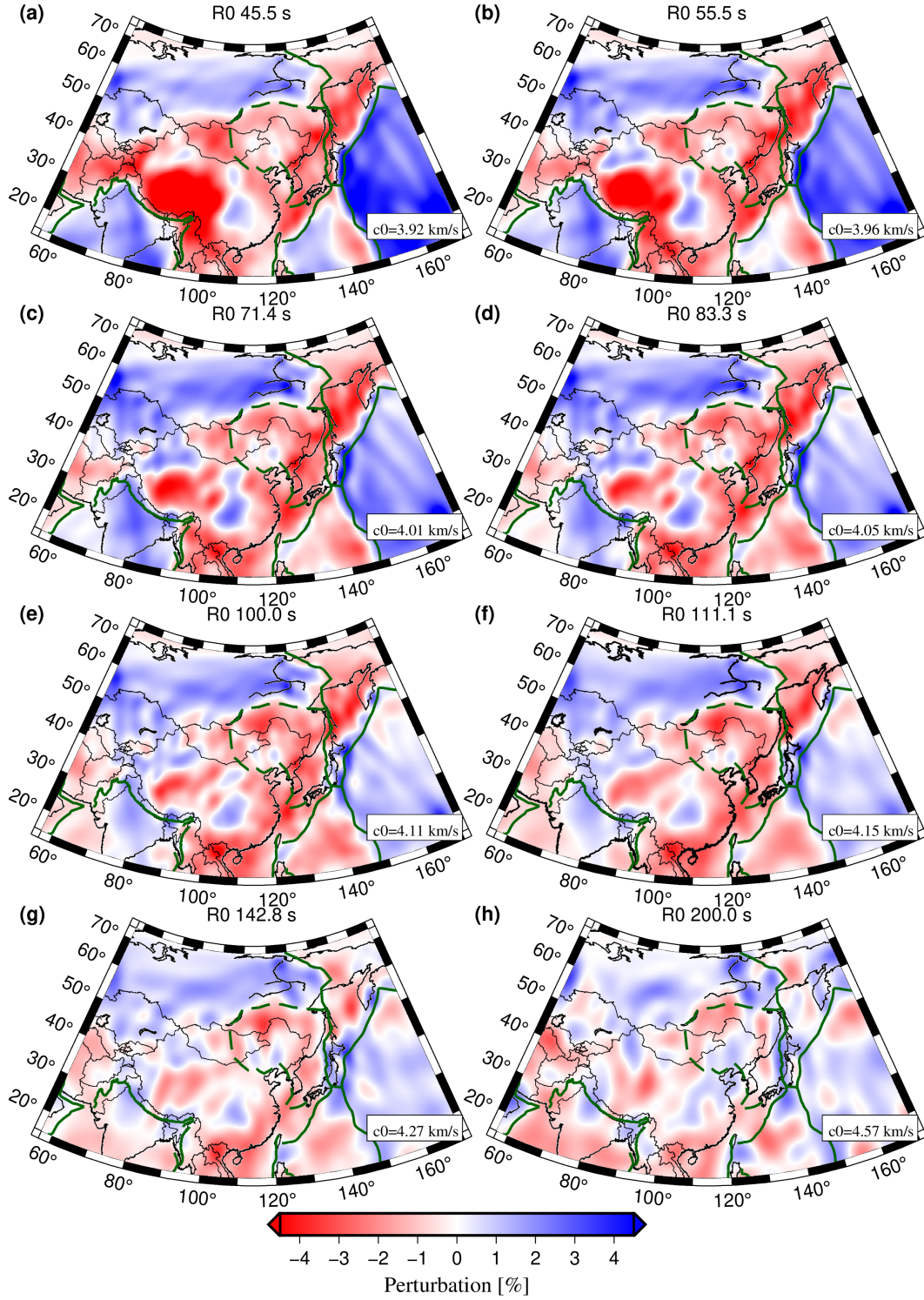


Figure B.5: Phase speed maps for the fundamental-mode Rayleigh waves.

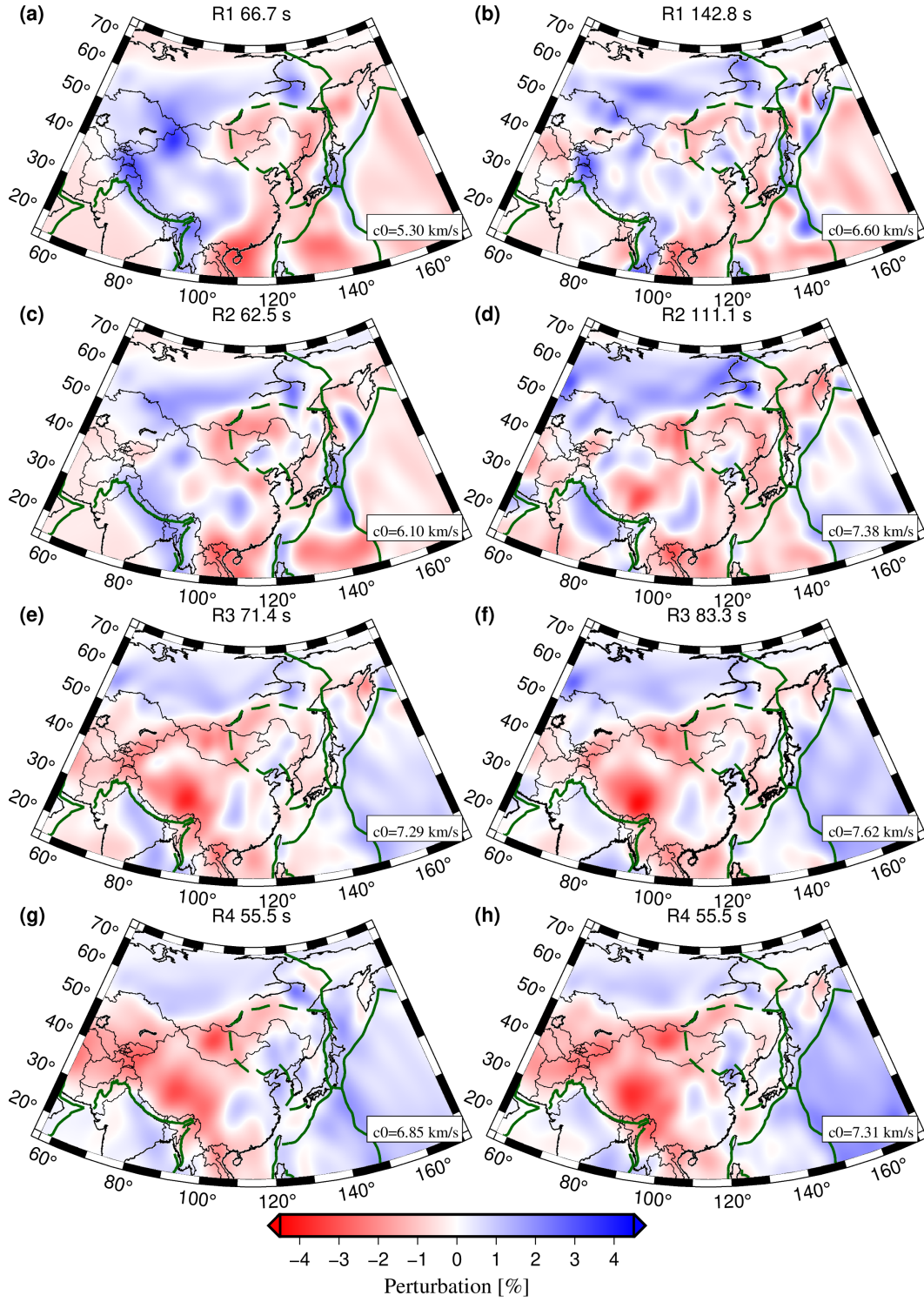


Figure B.6: Phase speed maps for the higher-mode Rayleigh waves (1st–4th higher modes).

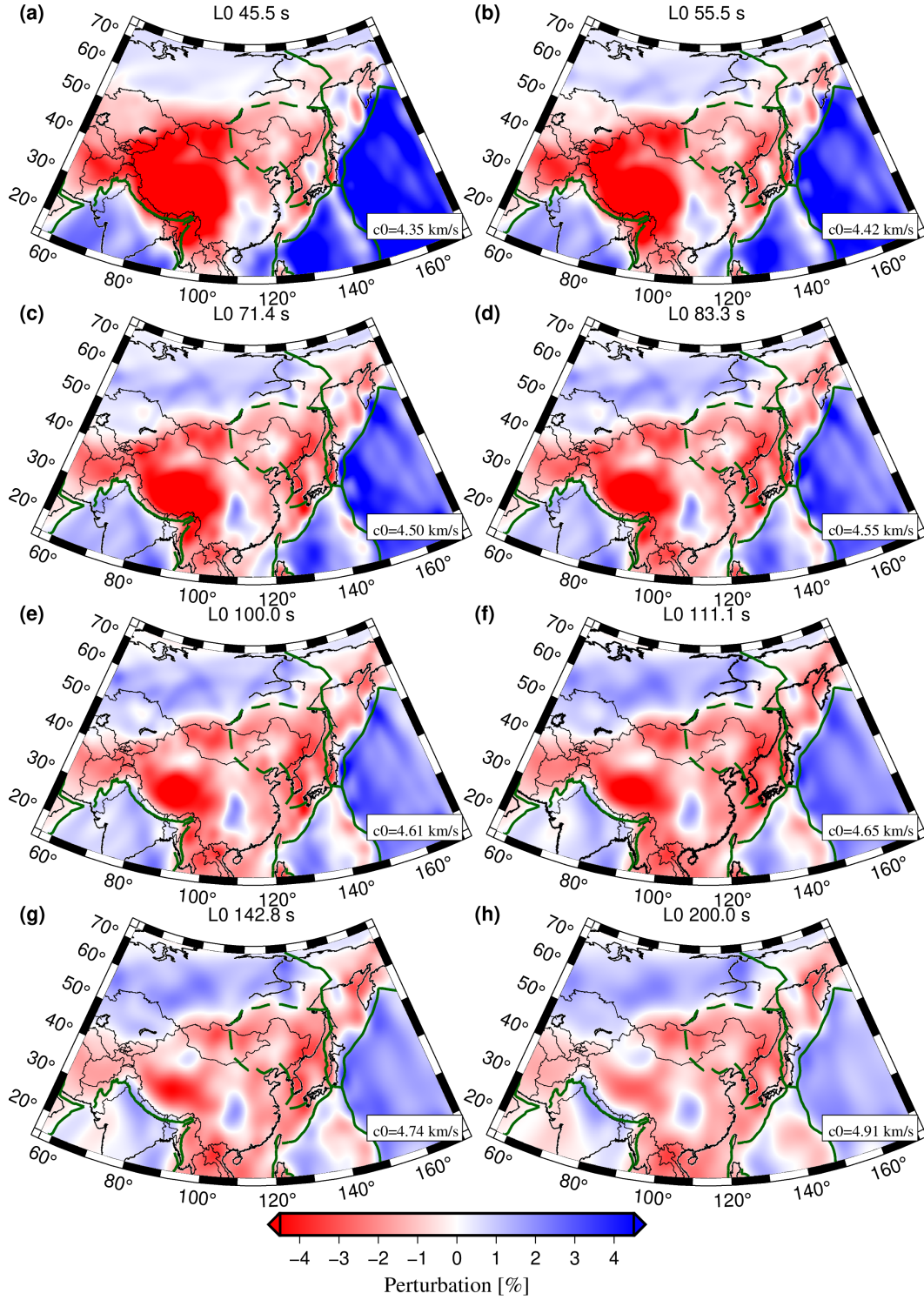


Figure B.7: Phase speed maps for the fundamental-mode Love waves.

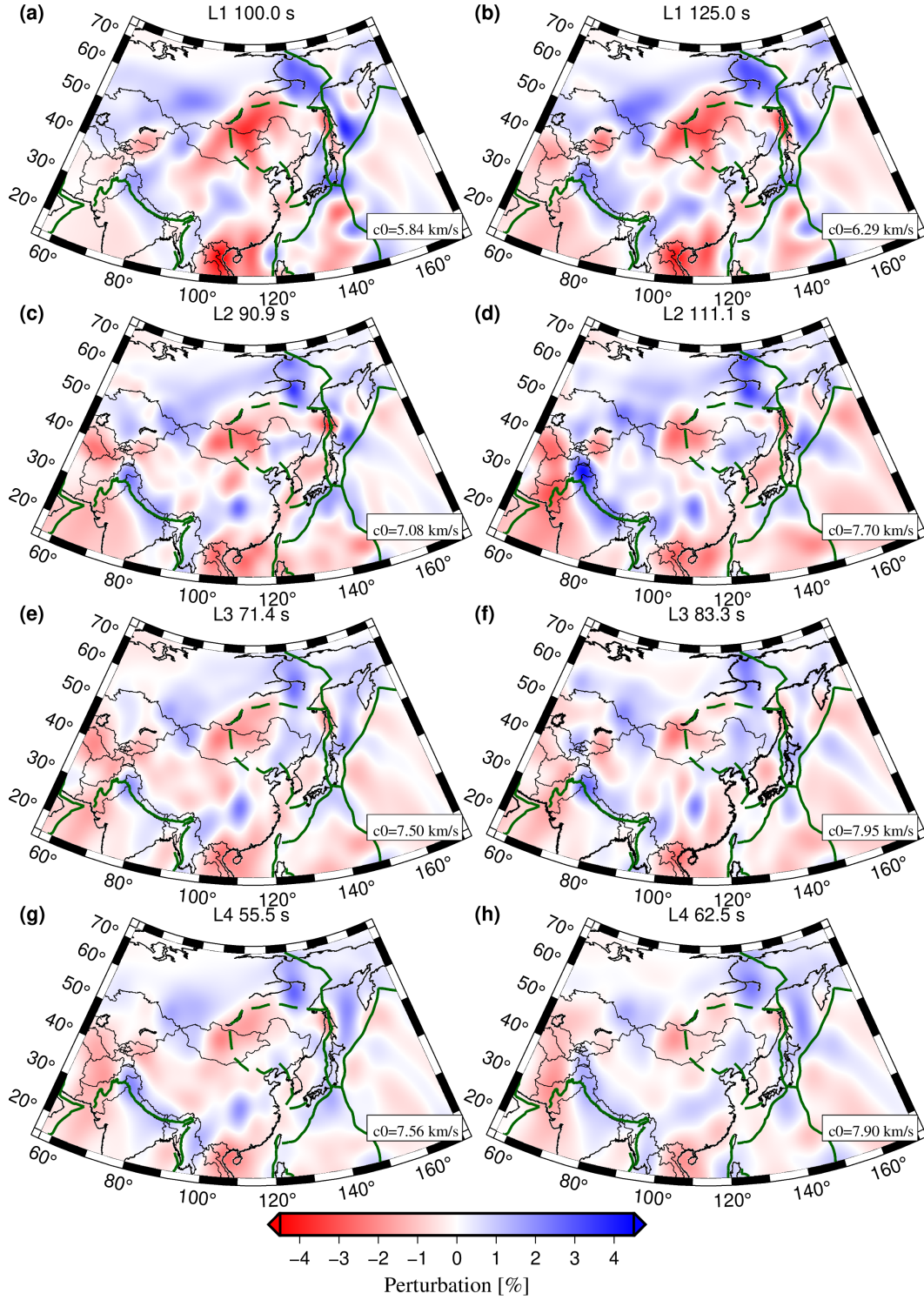


Figure B.8: Phase speed maps for the higher-mode Love waves (1st–4th higher modes).

Appendix C

3-D S-wave model gallery

In this appendix, we display a number of cross-sections of the radially anisotropic 3-D S-wave model developed in this study for broader areas in Eastern Eurasia (Appendix C.1) and the Mongolian region (Appendix C.2).

C.1 3-D S-wave model in a broader region in Eastern Eurasia

In this section, we show many cross-sections in the Eastern Eurasian region (Figures C.2–C.5) along the lines shown in Figure C.1.

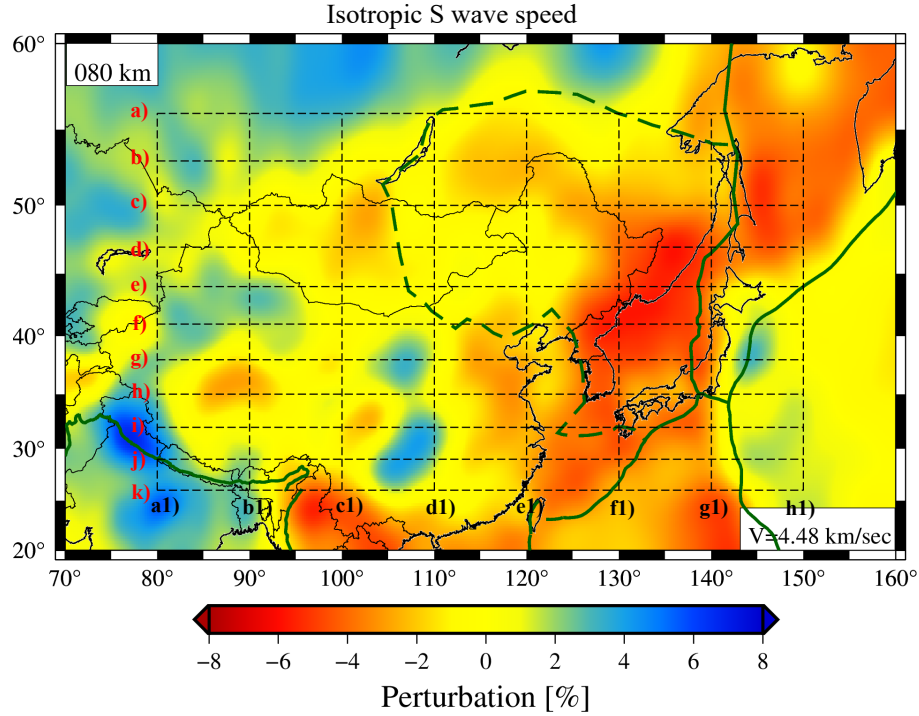


Figure C.1: Isotropic S wave speed model at 80 km depth, indicating the locations of cross sections in Figures C.2–C.5.

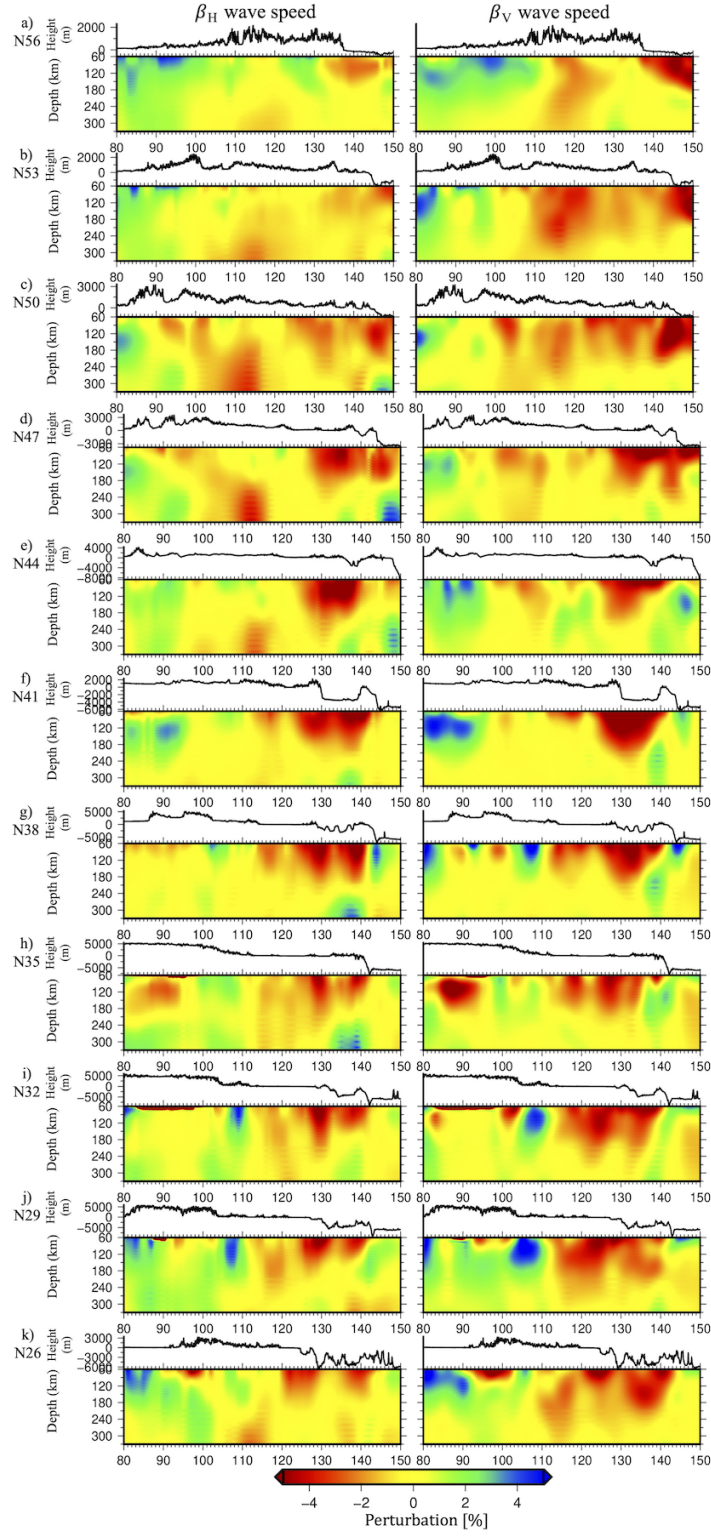


Figure C.2: Vertical W-E cross-sections of SH (β_H) and SV (β_V) wave speeds at latitudes from a) 56°N to k) 26°N. The reference SH and SV wave speed profiles (modified PREM) are shown in Figure 4.5 (a, dashed blue and red lines).

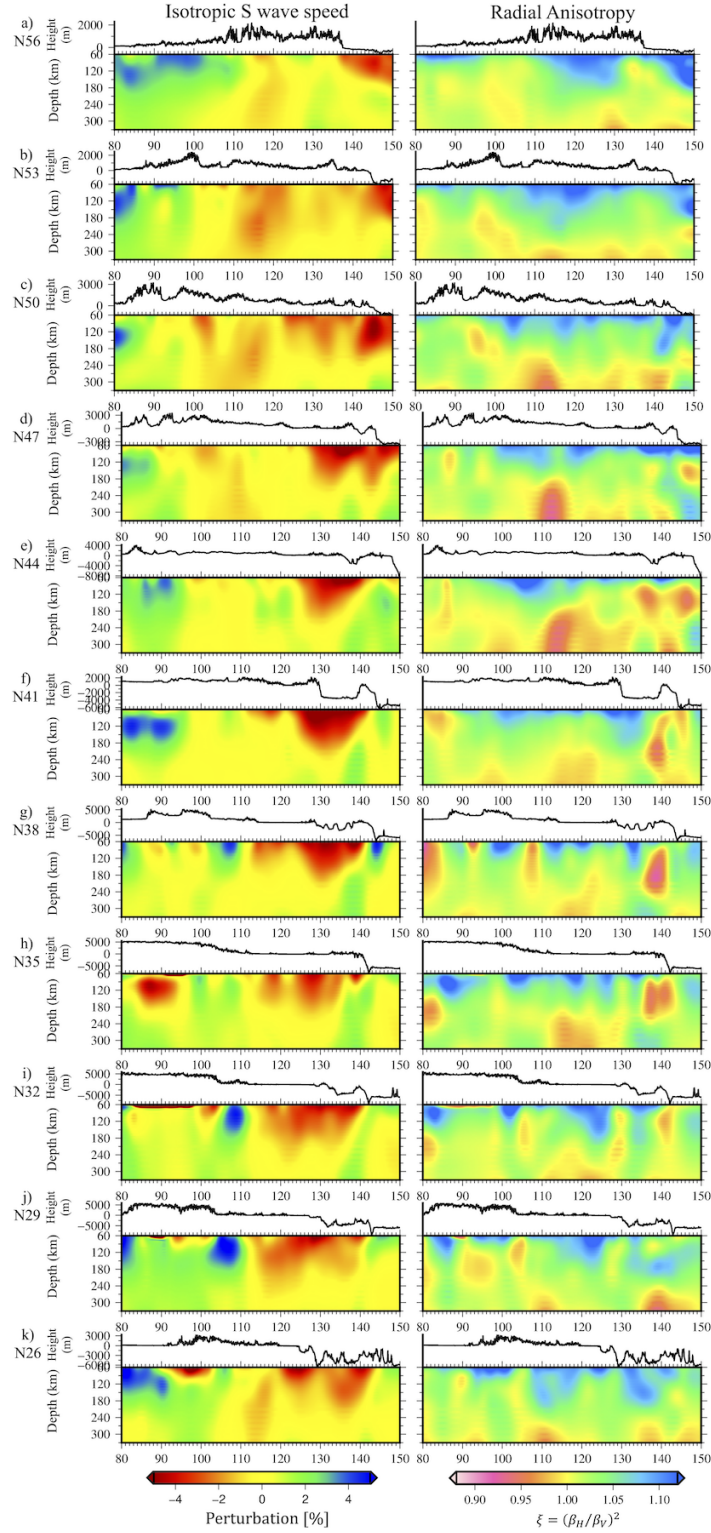


Figure C.3: Same as Figure C.2, but for isotropic S wave speed (β_{iso}) and radial anisotropy (ξ). The reference isotropic S wave speed profile (modified PREM) is shown in Figure 4.5 (a, dashed green line).

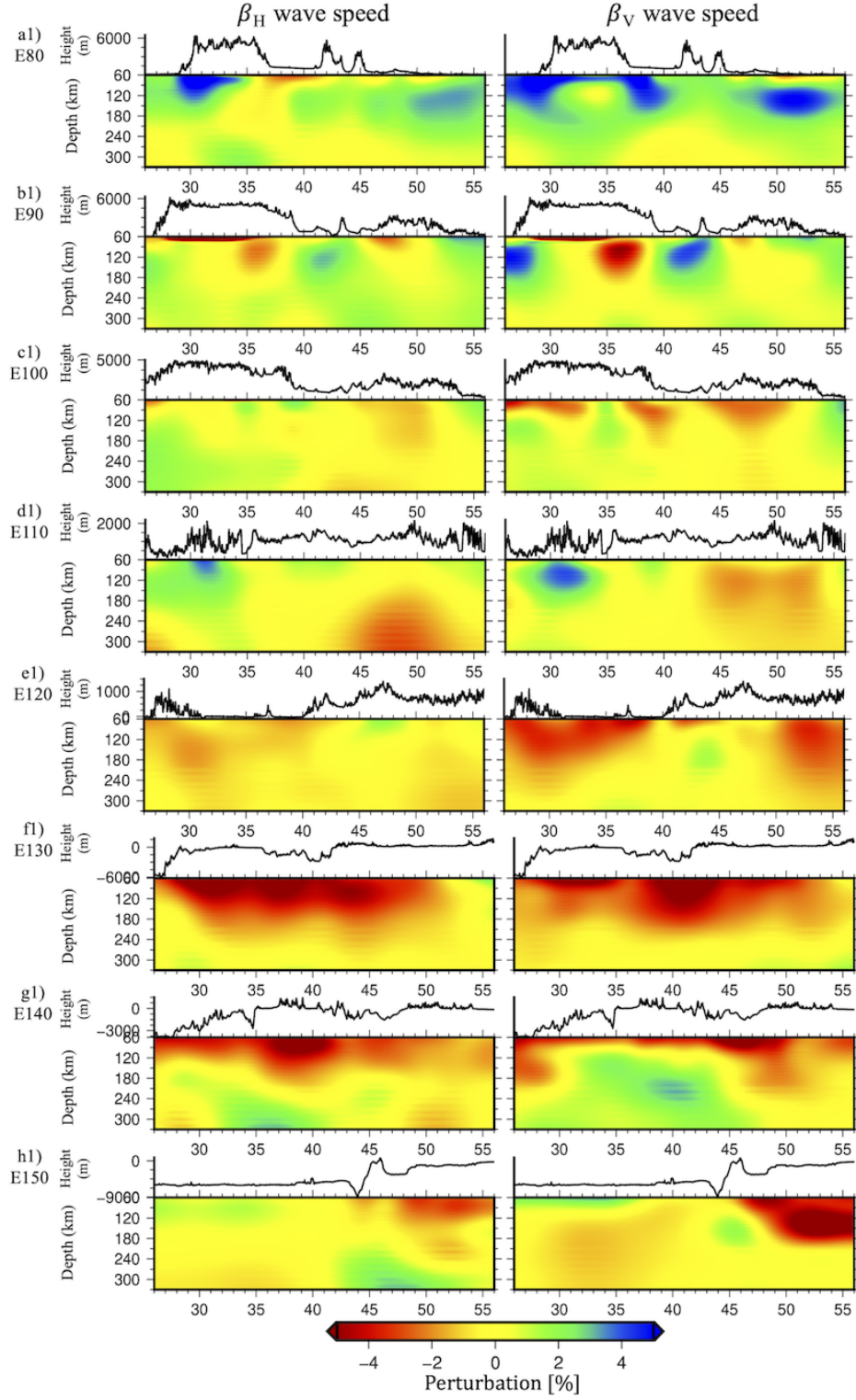


Figure C.4: Same as Figure C.2, but for vertical S-N cross-section at longitudes from a1) 80°E to h1) 150°E.

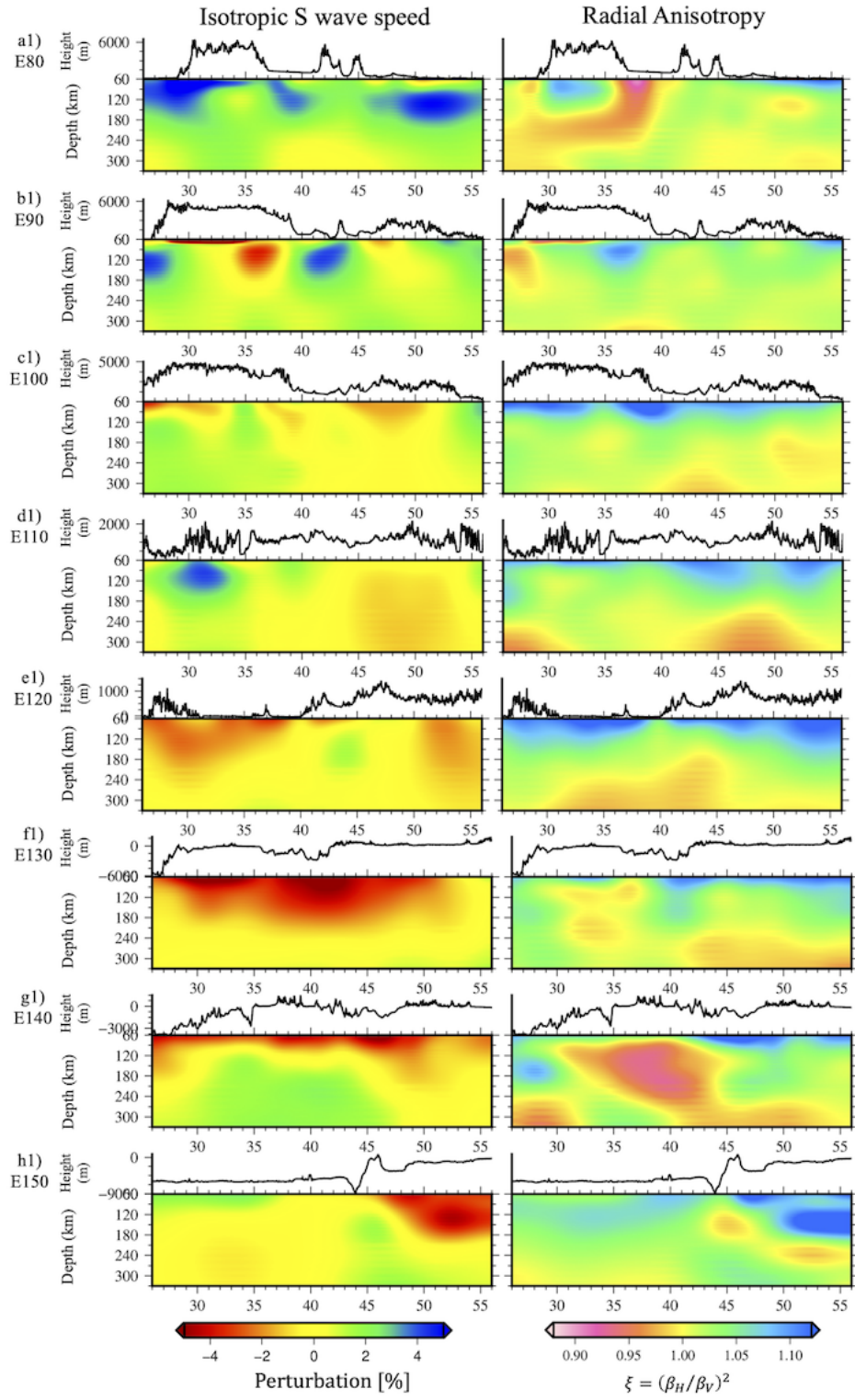


Figure C.5: Same as Figure C.3, but for vertical S-N cross-section at longitudes from a1) 80°E to h1) 150°E.

C.2 3-D S-wave model in the Mongolian region

In this section, we show many cross-sections in the Mongolian region (Figures C.7–C.8) along the lines shown in Figure C.6.

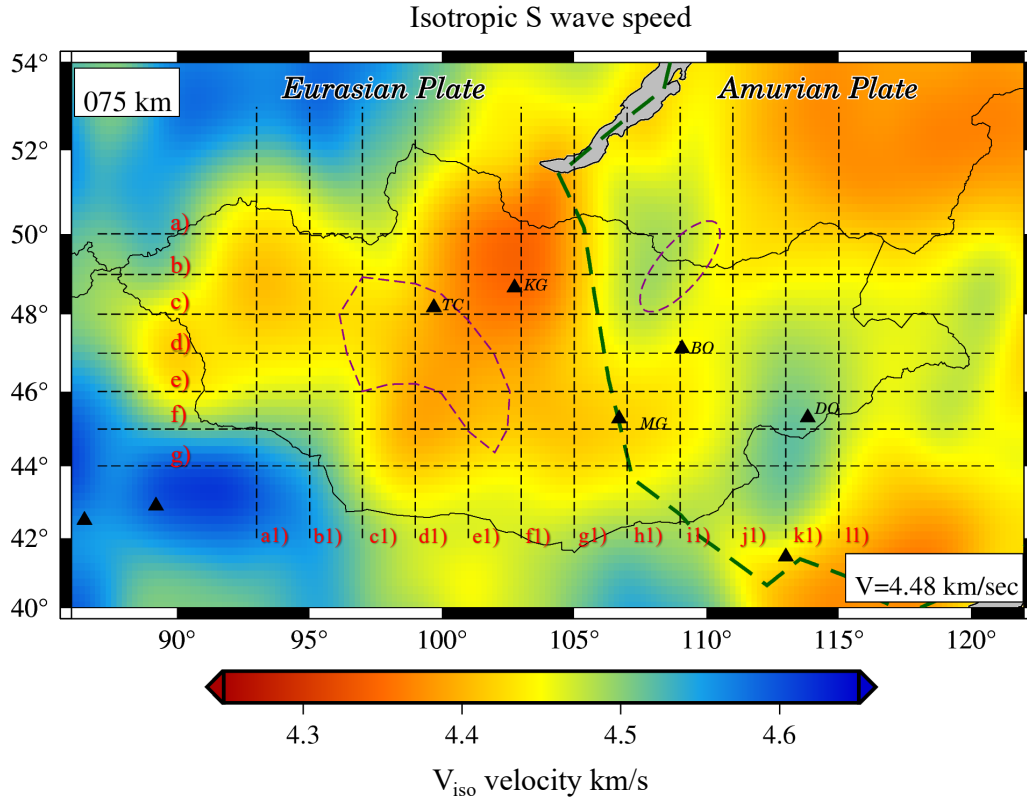


Figure C.6: Isotropic S wave speed model at 75 km depth, indicating the locations of cross sections in Figures C.7–C.10.

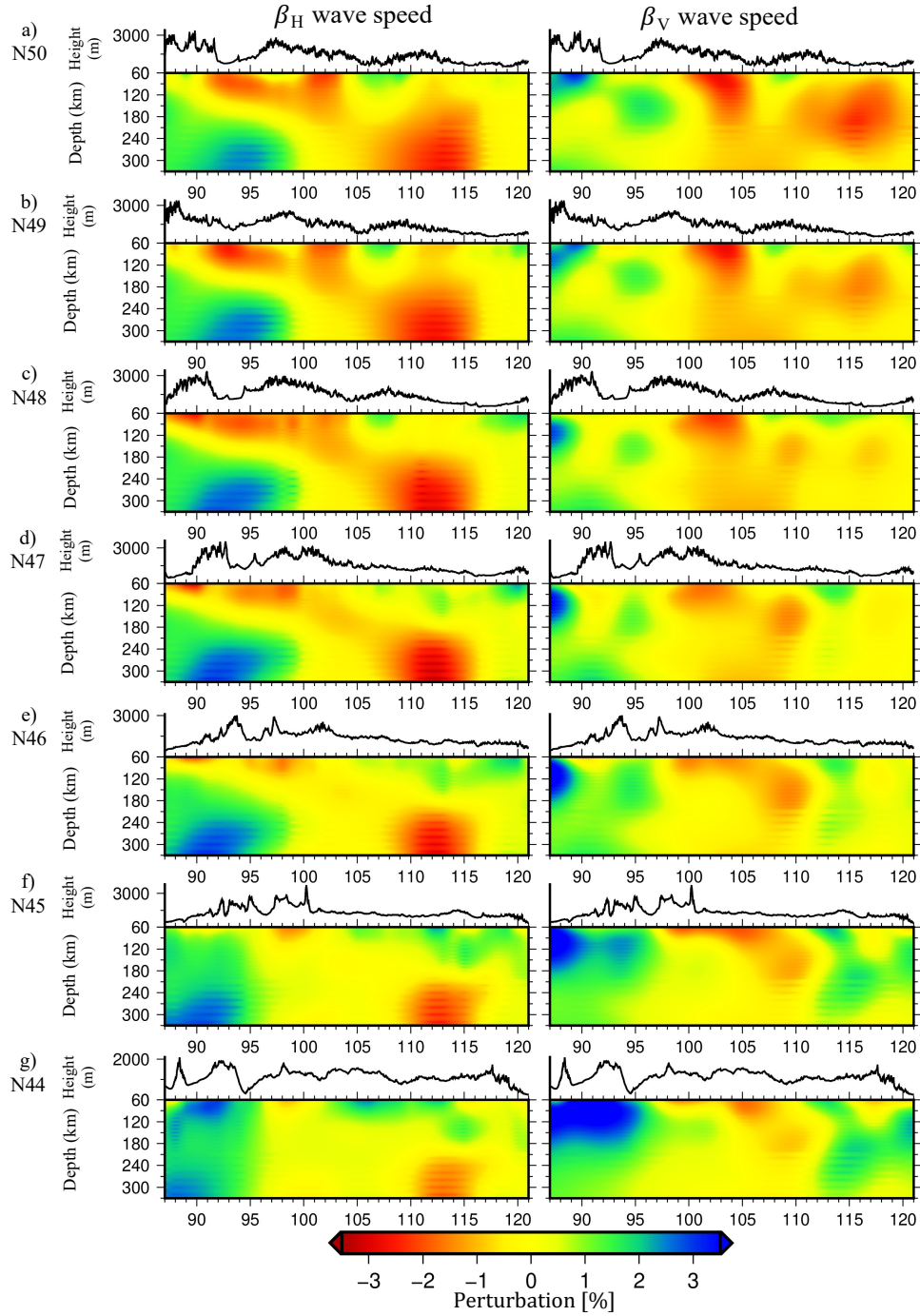


Figure C.7: Vertical W-E cross-sections of SH (β_H) and SV (β_V) wave speeds at latitudes from a) 50° to g) 44°. The reference SH and SV wave speed profiles (the average 1-D model in Mongolia) are shown in Figure 4.5 (b, solid blue and red lines).

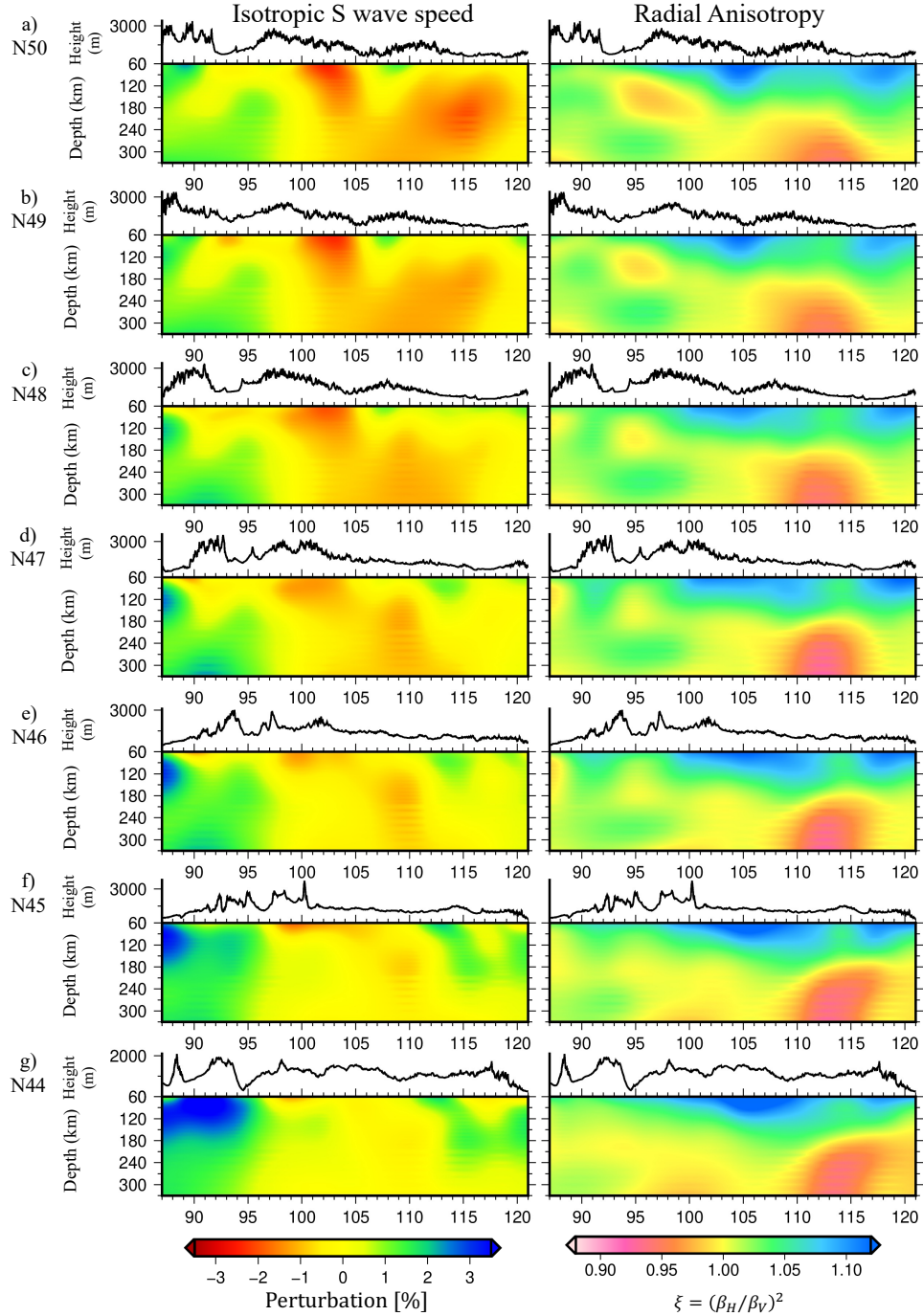


Figure C.8: Same as Figure C.7, but for isotropic S wave speed (β_{iso}) and radial anisotropy (ξ). The reference isotropic S wave speed profile (average 1-D model in Mongolia) is shown in Figure 4.5 (b, solid green line).

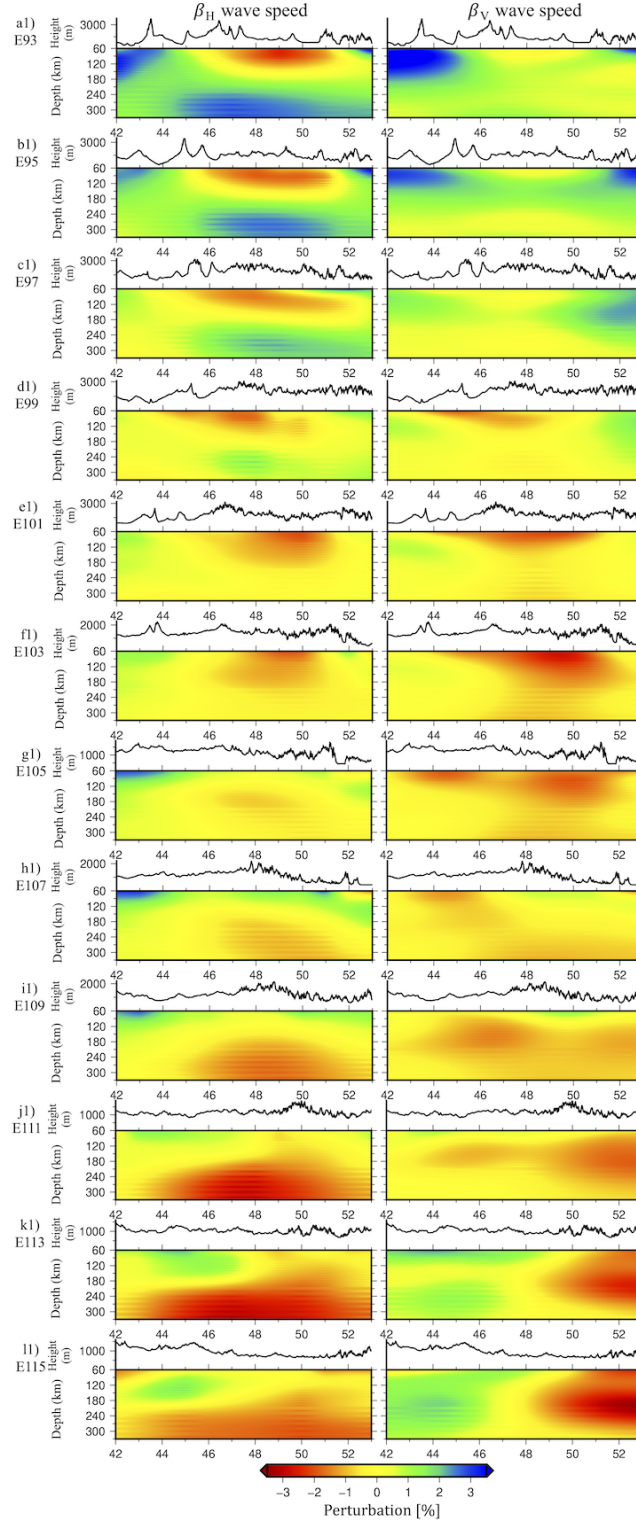


Figure C.9: Same as Figure C.7, but for vertical S-N cross-sections at longitudes from a1) 93° to l1) 115°.

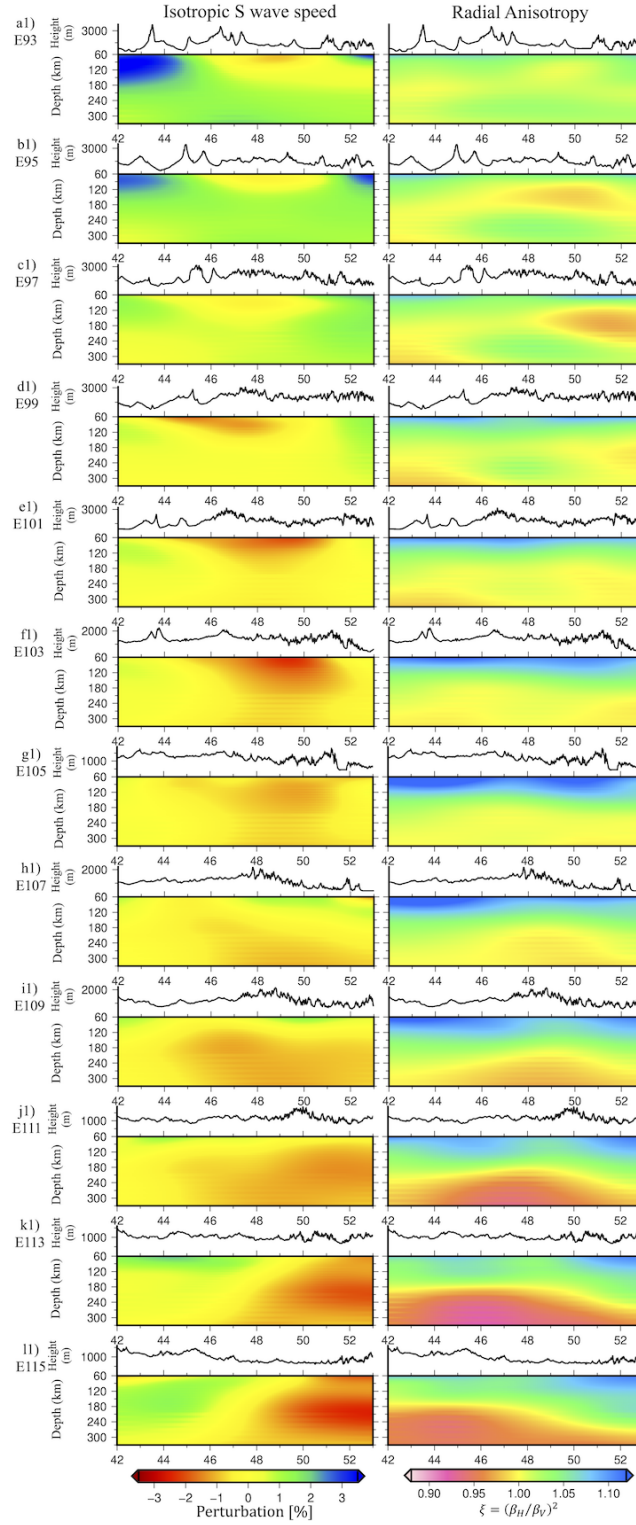


Figure C.10: Same as Figure C.8, but for vertical S-N cross-sections at longitudes from a1) 93° to l1) 115°.

Bibliography

- Aki, K., 1968. Seismological evidences for the existence of soft thin layers in the upper mantle under Japan, *Journal of Geophysical Research*, **73**(2), 585–594.
- Amante, C. & Eakins, B., 2009. ETOPO1 1 Arc-Minute Global Relief Model: Procedures, Data Sources and Analysis. NOAA Technical Memorandum NESDIS NGDC-24., *National Geophysical Data Center, NOAA*, Accessed: April 2024.
- Ashurkov, S., San'kov, V., Miroshnichenko, A., Lukhnev, A., Sorokin, A., Serov, M., & Byzov, L., 2011. Gps geodetic constraints on the kinematics of the amurian plate, *Russian Geology and Geophysics*, **52**, 239–249.
- Attanayake, J., Ferreira, A., Berbellini, A., & Morelli, A., 2017. Crustal structure beneath portugal from teleseismic rayleigh wave ellipticity, *Tectonophysics*, **712**, 344–361.
- Badarch, G., Cunningham, W., & F.Windley, B., 2002. A new terrane subdivision for mongolia: implications for the phanerozoic crustal growth of central asia, *Journal of Asian Earth Sciences*., **21**, 87–110.
- Bao, X., Song, X., Xu, M., Wang, L., Sun, X., Mi, N., Yu, D., & H., L., 2013. Crust and upper mantle structure of the North China Craton and the NE Tibetan Plateau and its tectonic implications, *Earth and Planetary Science Letters*, **369-370**, 129–137.
- Bao, X., Song, X., & J., L., 2015. High-resolution lithospheric structure beneath Mainland China from ambient noise and earthquake surface-wave tomography, *Earth and Planetary Science Letters*, **417**, 132–141.
- Barry, T. L. & Kent, R. W., 1998. Cenozoic magmatism in Mongolia and the origin of central and east Asian basalts, in Mantle Dynamics and Plate Interactions in

- East Asia, pp. 347–364, eds Martin F.J. Folower, Sun-Lin Chung, Ching-Hua Lo, & Tung-Yi Lee, American Geophysical Union, Washington D.C.
- Barry, T. L., Saunders, A. D., Kempton, P. D., Windley, B. F., Pringle, M. S., Dorjnamjaa, D., & Saandar, S., 2003. Petrogenesis of Cenozoic basalts from Mongolia: Evidence for the role of asthenospheric versus metasomatized lithospheric mantle sources, *Journal of Petrology*, **44**(1), 55–91.
- Bayasgalan, A., Jackson, J., & McKenzie, D., 2005. Lithosphere rheology and active tectonics in Mongolia: Relations between earthquake source parameters, gravity and GPS measurements, *Geophysical Journal International*, **163**(3), 1151–1179.
- Bird, P., 2003. An updated digital model of plate boundaries, *Geochemistry, Geophysics, Geosystems*, **4**(3), 1027, doi:10.1029/2001GC000252.
- Bonvalot, S., Balmino, G., Briais, A., M.Kuhn, Peyrefitte, A., Vales, N., Biancale, R., Gabalda, G., Reinquin, F., & Sarrailh, M., 2012. World gravity map. commission for the geological map of the world., *Eds. BGI-CGMW-CNES-IRD*.
- Brown, M. G., Lin, G., Matsuzawa, H., & Yoshizawa, K., 2022. Recovery of Love wave overtone waveforms and dispersion curves from single-station seismograms using time-warping, *Geophysical Journal International*, **233**(1), 70–83.
- Calais, E., Vergnolle, M., San'kov, V., Lukhnev, A., Miroshnitchenko, A., Amarjargal, S., & Déverchère, J., 2003. GPS measurements of crustal deformation in the Baikal-Mongolia area (1994-2002): Implications for current kinematics of Asia, *Journal of Geophysical Research: Solid Earth*, **108**(B10), 2501, doi:10.1029/2002jb002373.
- Chen, M., Niu, F., Liu, Q., & Tromp, J., 2015a. Mantle-driven uplift of Hangaï Dome: New seismic constraints from adjoint tomography, *Geophysical Research Letters*, **42**(17), 6967–6974.
- Chen, M., Niu, F., Liu, Q., Tromp, J., & Zheng, X., 2015b. Multiparameter adjoint tomography of the crust and upper mantle beneath East Asia: 1. Model construction and comparisons, *Journal of Geophysical Research: Solid Earth*, **120**(3), 1762–1786.

- Comeau, M. J., Käüfl, J. S., Becken, M., Kuvshinov, A., Grayver, A. V., Kamm, J., Demberel, S., Sukhbaatar, U., & Batmagnai, E., 2018. Evidence for fluid and melt generation in response to an asthenospheric upwelling beneath the Hangai Dome, Mongolia, *Earth and Planetary Science Letters*, **487**, 201–209.
- Comeau, M. J., Stein, C., Becken, M., & Hansen, U., 2021. Geodynamic Modeling of Lithospheric Removal and Surface Deformation: Application to Intraplate Uplift in Central Mongolia, *Journal of Geophysical Research: Solid Earth*, **126**(5), e2020JB021304, <https://doi.org/10.1029/2020JB021304>.
- Cunningham, W. D., 2001. Cenozoic normal faulting and regional doming in the southern Hangay region, Central Mongolia: Implications for the origin of the Baikal rift province, *Tectonophysics*, **331**(4), 389–411.
- Dahlen, F. A. & Tromp, J., 1998. *Theoretical global seismology*, Princeton University Press, Princeton, NJ.
- DeMets, C., Gordon, R. G., Argus, D. F., & Stein, S., 1990. Current plate motions, *Geophysical Journal International*, **101**(2), 425 – 478.
- DeMets, C., Gordon, R. G., & Argus, D. F., 2010. Geologically current plate motions, *Geophysical Journal International*, **181**(1), 1–80.
- Dziewonski, A. M. & Anderson, D. L., 1981. Preliminary reference Earth model, *Physics of the Earth and Planetary Interiors*, **25**(4), 297–356.
- Ekström, G., 2011. A global model of Love and Rayleigh surface wave dispersion and anisotropy, 25-250s, *Geophysical Journal International*, **187**(3), 1668–1686.
- Ekström, G., Nettles, M., & Dziewoński, A. M., 2012. The global CMT project 2004-2010: Centroid-moment tensors for 13,017 earthquakes, *Physics of the Earth and Planetary Interiors*, **200-201**, 1–9.
- Elkins-Tanton, L. T., 2005. Continental magmatism caused by lithospheric delamination, in *Plates, plumes and paradigms*, vol. 388, pp. 449–461, eds Foulger, G. R., Natland, J. H., Presnall, D. C., & Anderson, D. L., Geological Society of America.

- Erdenesaihan, G., Ishiwatari, A., Orolmaa, D., Arai, S., & Tamura, A., 2013. Middle paleozoic greenstones of the Hangay region, central Mongolia: Remnants of an accreted oceanic plateau and forearc magmatism, *Journal of Mineralogical and Petrological Sciences*, **108**(6), 303–325.
- Feng, L., 2021. High-Resolution Crustal and Uppermost Mantle Structure Beneath Central Mongolia From Rayleigh Waves and Receiver Functions, *Journal of Geophysical Research: Solid Earth*, **126**(4), e2020JB021161, <https://doi.org/10.1029/2020JB021161>.
- Fullea, J., Lebedev, S., Agius, M. R., Jones, A. G., & Afonso, J. C., 2012. Lithospheric structure in the Baikal-central Mongolia region from integrated geophysical-petrological inversion of surface-wave data and topographic elevation, *Geochemistry, Geophysics, Geosystems*, **13**(1), Q0AK09, doi:10.1029/2012GC004138.
- Global Volcanism Program, 2023. [Database] Volcanoes of the World (v. 5.1.0; 9 Jun 2023). Distributed by Smithsonian Institution, compiled by Venzke, E.
- Gung, Y., Panning, M., & Romanowicz, B., 2003. Global anisotropy and the thickness of continents, *Nature*, **422**(6933), 707–711.
- Hall, R., 2002. Cenozoic geological and plate tectonic evolution of SE Asia and the SW Pacific: Computer-based reconstructions, model and animations, *Journal of Asian Earth Sciences*, **20**(4), 353–431.
- Hao, M., Li, Y., & Zhuang, W., 2019. Crustal movement and strain distribution in East Asia revealed by GPS observations, *Scientific Reports*, **9**(1), 16797, <https://doi.org/10.1038/s41598-019-53306-y>.
- He, J., Wu, Q., Sandvol, E., Ni, J., Gallegos, A., Gao, M., Ulziibat, M., & Demberel, S., 2016. The crustal structure of south central Mongolia using receiver functions, *Tectonics*, **35**(6), 1392–1403.
- He, J., Li, Y., Sandvol, E., Wu, Q., Du, G., Zhang, R., Yu, D., Liu, H., Lei, J., & Huang, J., 2019. Tomographic Pn Velocity and Anisotropy Structure in Mongolia

- and the Adjacent Regions, *Journal of Geophysical Research: Solid Earth*, **124**(4), 3662–3679.
- Heki, K., Miyazaki, S. N., Takahashi, H., Kasahara, M., Kimata, F., Miura, S., Vasilenko, N. F., Ivashchenko, A., & An, K. D., 1999. The amurian plate motion and current plate kinematics in eastern Asia, *Journal of Geophysical Research: Solid Earth*, **104**(B12), 29147–29155.
- Huang, J. & Zhao, D., 2006. High-resolution mantle tomography of China and surrounding regions, *Journal of Geophysical Research: Solid Earth*, **111**(B9), B09305, doi:10.1029/2005JB004066.
- Huang, Z. & Zhao, D., 2022. Seismotectonics of Mongolia and Baikal Rift Zone Controlled by Lithospheric Structures, *Geophysical Research Letters*, **49**(15), e2022GL099525, <https://doi.org/10.1029/2022GL099525>.
- Hunt, A. C., Parkinson, I. J., Harris, N. B., Barry, T. L., Rogers, N. W., & Yondon, M., 2012. Cenozoic volcanism on the Hangai Dome, Central Mongolia: Geochemical evidence for changing melt sources and implications for mechanisms of melting, *Journal of Petrology*, **53**(9), 1913–1942.
- Isse, T., Kawakatsu, H., Yoshizawa, K., Takeo, A., Shiobara, H., Sugioka, H., Ito, A., Suetsugu, D., & Reymond, D., 2019. Surface wave tomography for the Pacific Ocean incorporating seafloor seismic observations and plate thermal evolution, *Earth and Planetary Science Letters*, **510**, 116–130.
- Jin, S., 2013. GNSS Observations of Crustal Deformation: A Case Study in East Asia, in *Geodetic Sciences - Observations, Modeling and Applications*, pp. 267–284, IntechOpen.
- Kennett, B., Engdahl, E., & Buland, R., 1995. Constraints on seismic velocities in the Earth from traveltimes, *Geophysical Journal International*, **122**, 108–124.
- Kennett, B. L. N. & Yoshizawa, K., 2002. A reappraisal of regional surface wave tomography, *Geophysical Journal International*, **150**(1), 37–44.

- Khutorskoy, M. & Yarmoluk, V., 1989. Heat flow, structure and evolution of the lithosphere of Mongolia, *Tectonophysics*, **164**(2-4), 315–322.
- Kozakov, I. K., Kovach, V. P., Bibikova, E. V., Kirnozova, T. I., Zagornaya, N. Y., Plotkina, Y. V., & Podkovyrov, V. N., 2007. Age and sources of granitoids in the junction zone of the caledonides and hercynides in southwestern mongolia: Geodynamic implications, *Petrology*, **15**, 126–150.
- Lamb, M. A. & Badarch, G., 2001. Paleozoic sedimentary basins and volcanic arc systems of southern mongolia: New geochemical and petrographic constraints, *Memoris-Geol. Soc. Am*, **21**, 117–150.
- Laske, G., Masters, G., Ma, Z., & Pasyanos, M., 2013. Update on CRUST1.0—A 1-degree global model of Earth’s crust, *EGU General Assembly 2013*, **15**, 2658.
- Lebedev, S., Meier, T., & van der Hilst, R. D., 2006. Asthenospheric flow and origin of volcanism in the Baikal Rift area, *Earth and Planetary Science Letters*, **249**(3-4), 415–424.
- Li, M., Song, X., Li, J., & Bao, X., 2022. Crust and upper mantle structure of East Asia from ambient noise and earthquake surface wave tomography, *Earthquake Science*, **35**(2), 71–92.
- Matsuzawa, H. & Yoshizawa, K., 2019. Array-based analysis of multimode surface waves: application to phase speed measurements and modal waveform decomposition, *Geophysical Journal International*, **218**(1), 295–312.
- McDannell, K. T., Zeitler, P. K., & Idleman, B. D., 2018. Relict Topography Within the Hangay Mountains in Central Mongolia: Quantifying Long-Term Exhumation and Relief Change in an Old Landscape, *Tectonics*, **37**(8), 2531–2558.
- Meltzer, A., Stachnik, J. C., Sodnomsambu, D., Munkhuu, U., Tsagaan, B., Dashdondog, M., & Russo, R., 2019. The central Mongolia seismic experiment: Multiple applications of temporary broadband seismic arrays, *Seismological Research Letters*, **90**(3), 1364–1376.

- Meng, Q. R., Hu, J. M., Jin, J. Q., Zhang, Y., & Xu, D. F., 2003. Tectonics of the late Mesozoic wide extensional basin system in the China-Mongolia border region, *Basin Research*, **15**(3), 397–415.
- Molnar, P. & Tapponnier, P., 1975. Cenozoic tectonics of Asia: Effects of a continental collision, *Science*, **189**(4201), 419–426.
- Montagner, J. P. & Anderson, D. L., 1989. Petrological constraints on seismic anisotropy, *Physics of the Earth and Planetary Interiors*, **54**(1-2), 82–105.
- Nataf, H. C. & Ricard, Y., 1996. 3SMAC: An a priori tomographic model of the upper mantle based on geophysical modeling, *Physics of the Earth and Planetary Interiors*, **95**(1-2), 101–122.
- Nishimura, C. E. & Forsyth, D. W., 1989. The anisotropic structure of the upper mantle in the Pacific, *Geophysical Journal*, **96**(2), 203 – 229.
- Paige, C. & Saunders, M., 1982. Lsq: an algorithm for sparse linear equations and sparse least squares, *ACM Trans. Math. Software*, **8**, 43–71.
- Pan, J.-T., Wu, Q., Li, Y., Yu, D.-X., Gao, M., Ulziibat, M., & Demberel, S., 2015. Ambient Noise Tomography in Central-South Mongolia, *Chinese Journal of Geophysics*, **58**(8), 3009–3022.
- Panning, M. & Romanowicz, B., 2006. A three-dimensional radially anisotropic model of shear velocity in the whole mantle, *Geophysical Journal International*, **167**(1), 361–379.
- Petit, C. & Fournier, M., 2005. Present-day velocity and stress fields of the amurian plate from thin-shell finite-element modelling, *Geophysical Journal International*, **160**, 357–369.
- Petit, C., Deverchere, J., Hiudry, F., Sankov, V., & Melnikova, V., 1996. Present-day stress field changes along the baikal rift and tectonic implications, *Tectonics*, **15**, 1171–1191.

- Petit, C., Déverchère, J., Calais, E., San'kov, V., & Fairhead, D., 2002. Deep structure and mechanical behavior of the lithosphere in the Hangai-Hövsgöl region, Mongolia: New constraints from gravity modeling, *Earth and Planetary Science Letters*, **197**(3-4), 133–149.
- Petit, C., Tiberi, C., Deschamps, A., & Déverchère, J., 2008. Teleseismic traveltimes, topography and the lithospheric structure across central Mongolia, *Geophysical Research Letters*, **35**(11), L11301, doi:10.1029/2008GL033993.
- Ritzwoller, M. H. & Levshin, A. L., 1998. Eurasian surface wave tomography: Group velocities, *Journal of Geophysical Research: Solid Earth*, **103**(B3), 4839–4878.
- Sambridge, M., 1999. Geophysical inversion with a neighbourhood algorithm - I. Searching a parameter space, *Geophysical Journal International*, **138**(2), 479–494.
- Sankov, V. A., Parfeevets, A. V., Miroshnichenko, A. I., Sankov, A. V., Bayasgalan, A., & Demberel, S., 2022. Active faulting along the western boundary of the Amur plate (territory of Mongolia), *Earth Science Frontiers*, **29**(1), 245–265.
- Sengör, A. M. C., Natal'in, B. A., & Burtman, V. S., 1993. Evolution of the altaid tectonic collage and palaeozoic crustal growth in eurasia, *Nature*, **364**, 299–307.
- Taira, T. & Yoshizawa, K., 2020. Upper mantle discontinuities beneath Australia from transdimensional Bayesian inversions using multi-mode surface waves and receiver functions, *Geophysical Journal International*, **223**, 2085–2100.
- Takahashi, H., Kasahara, M., Kimata, F., Miura, S., Heki, K., Seno, T., Kato, T., Vasilenko, N., Ivashchenko, A., Bahtiarov, V., Levin, V., Gordeev, E., Korchagin, F., & Gerasimenko, M., 1999. Velocity field of around the Sea of Okhotsk and Sea of Japan regions determined from a new continuous GPS network data, *Geophysical Research Letters*, **26**(16), 2533–2536.
- Takeuchi, H. & Saito, M., 1972. *Seismic Surface Waves*, vol. 11, pp. 217–295, Academic Press, New York.

- Tang, Q., Sun, W., Yoshizawa, K., & Fu, L. Y., 2022. Anomalous Radial Anisotropy and Its Implications for Upper Mantle Dynamics Beneath South China From Multimode Surface Wave Tomography, *Journal of Geophysical Research: Solid Earth*, **127**(8), e2021JB023485, <https://doi.org/10.1029/2021JB023485>.
- Tao, K., Grand, S. P., & Niu, F., 2018. Seismic Structure of the Upper Mantle Beneath Eastern Asia From Full Waveform Seismic Tomography, *Geochemistry, Geophysics, Geosystems*, **19**(8), 2732–2763.
- Tarantola, A. & Valette, B., 1982. Generalized nonlinear inverse problems solved using the least squares criterion, *Reviews of Geophysics*, **20**(2), 219–232.
- Tiberi, C., Deschamps, A., Déverchère, J., Petit, C., Perrot, J., Appriou, D., Mordvinova, V., Dugaarma, T., Ulzibaat, M., & Artemiev, A. A., 2008. Asthenospheric imprints on the lithosphere in Central Mongolia and Southern Siberia from a joint inversion of gravity and seismology (MOBAL experiment), *Geophysical Journal International*, **175**(3), 1283–1297.
- van Heijst, H. J. & Woodhouse, J. H., 1997. Measuring surface-wave overtone phase velocities using a mode-branch stripping technique, *Geophysical Journal International*, **131**(2), 209–230.
- Walker, R. T., Nissen, E., Molor, E., & Bayasgalan, A., 2007. Reinterpretation of the active faulting in central Mongolia, *Geology*, **35**(8), 759–762.
- Wang, X., Wu, H., Wang, H., Wu, B., & Huang, Z., 2022. Rayleigh wave tomography of central and southern Mongolia, *Tectonophysics*, **836**, 229426, <https://doi.org/10.1016/j.tecto.2022.229426>.
- Wang, Z. & Dahlen, F., 1995. Spherical-spline parameterization of three-dimensional earth models, *Geophysical Research Letters*, **22**, 3099–3102.
- Watson, M., Hayward, A., Parkinson, D., & Z.M, Z., 1987. Plate tectonic history, basin development and petroleum source rock deposition onshore china, *Mar. Petrol Geology*, **4**, 205–225.

- Wei, D. & Seno, T., 1998. *Determination of the Amurian plate motion*, vol. 27, pp. 337–346, AGU, Washington.
- Wei, W., Xu, J., Zhao, D., & Shi, Y., 2012. East Asia mantle tomography: New insight into plate subduction and intraplate volcanism, *Journal of Asian Earth Sciences*, **60**, 88–103.
- Wilhem, C., F.Windley, B., & M.Stampfli, G., 2012. The altaids of central asia: A tectonic and evolutionary innovative review, *Earth-Sciences Reviews*, **113**, 303–341.
- Windley, B. F. & Allen, M. B., 1993. Mongolian Plateau: evidence for a late Cenozoic mantle plume under central Asia, *Geology*, **21**(4), 295–298.
- Windley, B. F., Alexeiev, D., Xiao, W., Kröner, A., & Badarch, G., 2007. Tectonic models for accretion of the Central Asian Orogenic Belt, *Journal of the Geological Society*, **164**(1), 31–47.
- Witek, M., Chang, S. J., Lim, D. Y., Ning, S., & Ning, J., 2021. Radial Anisotropy in East Asia From Multimode Surface Wave Tomography, *Journal of Geophysical Research: Solid Earth*, **126**(7), e2020JB021201, <https://doi.org/10.1029/2020JB021201>.
- Wu, H., Huang, Z., & Zhao, D., 2021. Deep structure beneath the southwestern flank of the Baikal rift zone and adjacent areas, *Physics of the Earth and Planetary Interiors*, **310**, 106616, <https://doi.org/10.1016/j.pepi.2020.106616>.
- Wu, Y. & Bao, X., 2023. Cenozoic Uplift and Volcanism of Hangai Dome, Central Mongolia Triggered by Lower Mantle Upwellings, *Geophysical Research Letters*, **50**(8), e2023GL102838, <https://doi.org/10.1029/2023GL102838>.
- Xu, Y., Zhang, Y., Yang, B., & Bao, X., 2022. Phanerozoic evolution of lithospheric structures of the north china craton, *Geophysical Research Letters*, **49**, e2022GL098341, <https://doi.org/10.1029/2022GL098341>.
- Yin, A., 2010. Cenozoic tectonic evolution of Asia: A preliminary synthesis, *Tectonophysics*, **488**(1-4), 293–325.

- Yoshizawa, K., 2014. Radially anisotropic 3-D shear wave structure of the Australian lithosphere and asthenosphere from multi-mode surface waves, *Physics of the Earth and Planetary Interiors*, **235**, 33–48.
- Yoshizawa, K. & Ekström, G., 2010. Automated multimode phase speed measurements for high-resolution regional-scale tomography: Application to North America, *Geophysical Journal International*, **183**(3), 1538–1558.
- Yoshizawa, K. & Kennett, B. L. N., 2002a. Determination of the influence zone for surface wave paths, *Geophysical Journal International*, **149**(2), 440–453.
- Yoshizawa, K. & Kennett, B. L. N., 2002b. Non-linear waveform inversion for surface waves with a neighbourhood algorithm-application to multimode dispersion measurements, *Geophysical Journal International*, **149**(1), 118–133.
- Yoshizawa, K. & Kennett, B. L. N., 2004. Multimode surface wave tomography for the Australian region using a three-stage approach incorporating finite frequency effects, *Journal of Geophysical Research: Solid Earth*, **109**(B2), B02310, doi:10.1029/2002jb002254.
- Yu, D. X., Wu, Q. J., Li, Y. H., Pan, J. T., Zhang, F. X., He, J., Gao, M. T., Ulziibat, M., & Demberel, S., 2015. Rayleigh wave tomography of the phase velocity in the central and southern Mongolia, *Acta Geophysica Sinica*, **58**(1), 134–142.
- Zhang, F., Wu, Q., Grand, S. P., Li, Y., Gao, M., Demberel, S., Ulziibat, M., & Sukhbaatar, U., 2017. Seismic velocity variations beneath central Mongolia: Evidence for upper mantle plumes?, *Earth and Planetary Science Letters*, **459**, 406–416.
- Zhao, H., Wang, P., & Huang, Z., 2021. Lithospheric structures beneath the western Mongolian Plateau: Insight from S wave receiver function, *Journal of Asian Earth Sciences*, **212**, 104733, <https://doi.org/10.1016/j.jseas.2021.104733>.
- Zheng, Y., Davis, G., Cong, W., B.J., D., & C., Z., 2000. Major mesozoic tectonic events in the yanshan belt and the plate tectonic setting, *Acta. Geology. Sinica*, **74**, 289–302.

- Zonenshain, L. P. & Savostin, L. A., 1981. Geodynamics of the Baikal rift zone and plate tectonics of Asia, *Tectonophysics*, **76**(1-2), 1–45.
- Zorin, Y. A., Novoselova, M. R., Turutanov, E. K., & Kozhevnikov, V. M., 1990. Structure of the lithosphere of the Mongolian-Siberian mountainous province, *Journal of Geodynamics*, **11**(4), 327–342.
- Zorin, Y. A., Turutanov, E. K., Kozhevnikov, V. M., Rasskazov, S. V., & Ivanov, A. V., 2006. Cenozoic upper mantle plumes in East Siberia and Central Mongolia and subduction of the Pacific Plate, *Doklady Earth Sciences*, **409**(1), 723–726.