Crust and uppermost mantle beneath the North China Craton, northeastern China, and the Sea of Japan from ambient noise tomography

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Abstract:

A 3-D shear velocity model of the crust and uppermost mantle to a depth of 100 km is presented beneath the North China Craton (NCC), northeastern China, the Korean Peninsula, and the Sea of Japan. Ambient noise Rayleigh wave tomography is applied to data from more than 300 broad-band seismic stations from Chinese provincial networks (CEArray), the Japanese F-Net, and the IRIS Global Seismic Network. Continuous data from 2007 to 2009 are used to produce group and phase velocity maps from 8 sec to 45 sec period. The model is motivated to constrain the distributed intra-plate volcanism, crustal extension, cratonic rejuvenation, and lithospheric thinning that are hypothesized for the study region. Numerous robust features are observed that

19   impose new constraints on the geometry of these processes, but discussion concentrates only on

20   four. (1) The North-South Gravity Lineament follows the ~40 km contour in crustal thickness,

21   and crustal thickness is anti-correlated with water depth beneath the Sea of Japan, consistent with

22   crustal isostasy for a crust with laterally variable composition. (2) The lithosphere is thin (~70

23   km) beneath the Songliao-Bohai Graben, but seismically fast. (3) Even thinner more attenuated

24   lithosphere bounds three sides of the eastern NCC (in a horseshoe shape), identifying a region of

25   particularly intense tectonothermal modification where lithospheric rejuvenation may have

26   reached nearly to the base of the crust. (4) Low velocity anomalies reach upward (in a Y-shape)

27   in the mantle beneath the eastern and western borders of the Sea of Japan, extending well into

28   continental East Asia in the west, and are separated by a ~60 km thick lithosphere beneath the

29   central Sea of Japan. This anomaly may reflect relatively shallow slab dehydration in the east

30   and in the west deeper dehydration and convective circulation in the mantle wedge overlying the

31   stagnant slab.

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1. Introduction

The goal of this study and companion papers by *Yang et al.* [*2010, 2011*], *Zheng et al.* [*2010b*]. and *Zhou et al.* [2011] is to advance toward an integrated, highly resolved shear velocity (Vs) model of the crust and uppermost mantle beneath China. *Zhou et al.* [2011] focus on Southeast China and *Yang et al* [*2010, 2011*] and *Zheng et al.* [*2010b*] on western China and Tibet. The complementary focus of the current paper is the Sino-Korean Craton, northeastern China, the Korean Peninsula, and the Sea of Japan. More than 320 seismic stations from Chinese provincial networks in northeastern China and surrounding areas, Japanese F-Net stations [*Okada et al.*, 2004], and IRIS GSN stations are the basis for this study (Fig. 1). The resulting station and path coverage that emerges is unprecedented in this region. In order to place strong constraints on crustal structure, ambient noise tomography is employed. As described below, ambient noise tomography has already been applied in other regions of China (e.g., Tibet), at larger scales (e.g., across all of China) at a lower resolution, or in a part of the study region (e.g., North China Platform, Korean peninsula), but a resulting integrated, high-resolution model of the crust and uppermost mantle of the entire region of study has never before been constructed. Such a model is desired to illuminate a set of interconnected tectonic problems that make the Sino-Korean Craton, northeastern China, and the Sea of Japan a particularly fertile area for seismic tomography.

Northeastern China is composed of a mosaic of tectonic blocks and lineated orogenic belts

(Fig. 2) that have been arranged and modified by a long, complex, and in some cases enigmatic history of subduction, accretion, and collision dating back to the Archean [e.g., *Sengor and Natal'in*, 1996; *Yin and Nie*, 1996; *Yin, 2010*]. In the south, the Sino-Korean Craton (SKC), delineated by solid red lines in Figure 2, is separated from the Yangtze Craton to the south by the Qinling-Dabie-Sulu orogenic belt [*Yin and Nie*, 1993]. The SKC itself consists of the Ordos Block, the North China Platform, Bohai Bay, the Korean Peninsula, as well as marginal and intruding mountain ranges. The Chinese part of the SKC is usually referred to as the North China Craton (NCC) and is divided here into western, central, and eastern parts delineated by dashed red lines. North of the SKC, the Xing'an-East Mongolia block is separated from the Songliao-Bohai graben by the North-South Gravity Lineament (NSGL). The NSGL extends southward into the SKC near the boundary between the central and eastern parts of the North China Craton. The Songliao-Bohai graben, stretching from the Songliao Basin to the Bohai Gulf, is flanked to the east by the Tancheng-Lujiang (Tanlu) Fault that extends into the SKC and forms the eastern boundary of the North China Craton. The Tanlu Fault also defines the western border of the Northeast Asia Foldbelt [e.g., *Sengor and Natal'in*, 1996; *Yin and Nie*, 1996].

Much of northeastern China has undergone extensive tectonism during the late Mesozoic and Cenozoic eras [e.g., *Yin*, 2010]. Northeastern China, bounded by the SKC to the south and the Sea of Japan back-arc basin to the east, is part of eastern China's Cenozoic volcanic zone [e.g., *Ren et al.*, 2002]. Episodic volcanism has been particularly prominent along three volcanic

76   mountain chains: Great Xing'an Range (GXAR), Lesser Xing'an Range (LXAR), Changbai

77   Mountains (CBM). Rifting and extension are believed to have begun in the late Mesozoic [*Tian*

78   *et al.*, 1992] and have led to the development of the Songliao Basin, although the Songliao

79   basement is traced back to the Archean [*Rogers and Santosh*, 2006]. It has also been

80   hypothesized that there is a physical linkage between the sequential openings of the Songliao

81   graben [e.g., *Liu et al.*, 2001] and the Sea of Japan [*Tatsumi et al.*, 1989; *Jolivet et al.*, 1994]. The

82   SKC formed largely in the Archean, but it is a paradigm of an Archean craton that has lost its

83   lithospheric keel. Petrological and geochemical evidence [e.g., *Menzies et al.*, 1993; *Griffin et al.*,

84   1998] suggests that typical cratonic lithosphere existed beneath the entire SKC until at least the

85   Ordovician, after which the SKC was reactivated and lithospheric thinning occurred at the

86   minimum beneath the eastern part of the NCC.

87   The physical mechanisms that have produced the highly distributed volcanism, crustal

88   extension, cratonic rejuvenation and lithospheric thinning that have occurred across parts of

89   northeastern China remain poorly understood [e.g., *Deng et al.*, 2007; *Chen*, 2010]. These issues

90   can be illuminated with seismic images of velocity heterogeneities, internal discontinuities, and

91   crustal and mantle anisotropy, yet the vast majority of the studies to date have been geochemical

92   or petrological in nature.   Increasingly, seismological studies are adding new information due in

93   part to the rapid expansion of seismic instrumentation in China over the past few years [e.g.,

94   *Zheng et al.*, 2010a]. These studies include body wave tomography of the mantle [e.g., *Lebedev*

95 *and Nolet*, 2003; *Huang and Zhao*, 2006; *Zhao*, 2009; *Tian et al*., 2009; *Xu and Zhao*, 2009; *Li*

96 *and van der Hilst*, 2010; *Santosh et al*., 2010] and of the crust [e.g., *Sun and Toksoz*, 2006]. Pn

97 tomography [*Li et al*., 2011], receiver functions analyses [*Zheng et al*., 2006; *Chen et al*., 2009],

98 and shear wave splitting studies have also been performed [*Zhao et al*., 2008; *Bai et al*., 2010; *Li*

99 *and Niu*, 2010]. Recent body wave results have focused mainly on the eastern part of the NCC,

100 including efforts to image its thin intact lithosphere as well as the potential remnants of the

101 delaminated lithosphere near 400 km depth [e.g., *Chen et al*., 2009; *Xu and Zhao*, 2009].

102   At least parts of northeast China have been imaged by larger scale teleseismic surface wave

103 dispersion studies [e.g., *Ritzwoller and Levshin*, 1998; *Ritzwoller et al*., 1998; *Villaseñor et al*.,

104 2001; *Yanovskaya and Kozhevnikov*, 2003; *Huang et al*., 2003, 2004; *Priestley et al*., 2006].

105 Recent regional scale teleseismic surface wave studies have also been conducted within North

106 China [e.g., *Tang and Chen*, 2008; *Huang et al*., 2009; *Zhou et al*., 2009; *He et al*., 2009; *Pan et*

107 *al*., 2011], in adjacent regions [e.g., *Yao et al*., 2006, 2008], and in the Sea of Japan [e.g.,

108 *Bourova et al*., 2010; *Yoshizawa et al*, 2010]. Some of these studies are discussed further in the

109 context of our 3D model in section 6.

110   Within the last few years a new method of surface wave tomography has emerged based on

111 using ambient seismic noise to extract surface wave empirical Green's functions (EGFs) and to

112 infer Rayleigh [e.g., *Sabra et al*., 2005; *Shapiro et al*., 2005] and Love wave [e.g., *Lin et al*.,

113 2008] group and phase speeds in continental areas. Compared with traditional earthquake

114 tomography methods, ambient noise tomography is relatively free of artifacts related to the

115 distribution of earthquakes as well as errors in earthquake locations and source mechanisms. The

116 dominant frequency band for ambient noise tomography lies between about 8 and 40 sec period.

117 Rayleigh waves in this band are sensitive to crustal and uppermost mantle structures. Ambient

118 noise tomography has produced phase and group velocity maps in various regions around the

119 world [e.g., *Moschetti et al.*, 2007; *Yang et al.*, 2007; *Villasenor et al.,* 2007; *Bensen et al.*, 2008]

120 and also is the basis for 3-D crustal and uppermost mantle models of isotropic shear velocity

121 structure [e.g., *Yang et al.*, 2008a, 2008b; *Bensen et al.*, 2009; *Moschetti et al.*, 2010b], radial

122 anisotropy [e.g., *Moschetti et al.*, 2010a], and azimuthal anisotropy [e.g., *Lin et al.*, 2011].

123 Ambient noise in east Asia has been shown to be sufficiently well distributed in azimuthal

124 content to be used for surface wave dispersion measurements [e.g., *Yang and Ritzwoller*, 2008]

125 and studies based on ambient noise have been conducted at large scales across all of China

126 [*Zheng et al.*, 2008; *Sun et al.*, 2010], in regions adjacent to northeastern China [e.g., *Yao et al.*,

127 2006, 2008; *Guo et al.*, 2009; *Li et al.*, 2009; *Huang et al.*, 2010; *Yang et al.*, 2010; *Zheng et al.*,

128 2010b; *Zhou et al.*, 2011], within the North China Craton [*Fang et al.*, 2010], on the Korean

129 Peninsula [*Kang and Shin*, 2006; *Cho et al.*, 2007], and in Japan [e.g., *Nishida et al.*, 2008].

130 Ambient noise tomography in northeastern China, the Korean Peninsula, the Sea of Japan,

131 however, is faced with an uncommon technical challenge – the existence of a persistent localized

132 microseismic source on Kyushu Island [*Zeng and Ni*, 2010] in the period band between about 8

133 and 14 sec, which has been explained to be caused by long period volcanic tremors beneath Aso

134 Volcano in the center of Kyushu [*Kawakatsu et al.*, 2011; *Zeng and Ni*, 2011]. This signal causes

135 a significant disturbance that is observable on cross-correlations of ambient noise, which, if left

136 uncorrected, would bias measurements of group and phase velocity in this period band of

137 considerable sensitivity to crustal structure. A principal focus of this paper, therefore, is to

138 identify this disturbance and minimize its effects on the estimated Rayleigh wave group and

139 phase velocity dispersion maps.

140 The present paper is based on Rayleigh wave group and phase velocity maps from 8 to 45 sec

141 period across the Sino-Korea Craton, northeastern China, Korea, and the Sea of Japan. Based on

142 these maps, a 3-D model of the crust and uppermost mantle is constructed by Monte-Carlo

143 inversion along with associated uncertainties. The resulting information complements existing

144 and emerging teleseismic body wave and surface studies by presenting new and much stronger

145 constraints on the structure of the crust and uppermost mantle. The data processing and quality

146 control procedures are described in section 2 and methods used to de-sensitize the data to

147 degradation caused by the persistent, localized Kyushu microseism are presented in section 3.

148 The Rayleigh wave group and phase velocity maps are described in section 4. Section 5 presents

149 a brief discussion of the Monte-Carlo inversion method. Section 6 describes the features of the

150 resulting model. In addition, there is a discussion of several key findings including the relation

151 between sea floor depth and crustal thickness beneath the Sea of Japan, the horseshoe-shape (in

152 map view) of the thinnest lithosphere that bounds the NCC, and the Y-shaped asthenosphere (on

153 vertical profiles) observed beneath the Sea of Japan and the Northeast Asian Foldbelt.

154 **2. Data Processing and Quality Control**

155 The data used in this study are continuous seismic waveforms recorded at broadband stations

156 that existed in and around northeast China and the Sea of Japan from August 2007 to July 2009.

157 Networks providing data include: (1) Chinese Provincial Networks in northeast China consisting

158 of 232 broadband seismic stations (referred to here as CEArray), (2) F-Net in Japan comprising

159 69 long period seismic stations, and (3) the IRIS GSN broadband network in northeast Asia

160 consisting of 22 stations. In total, two years of continuous waveform data have been acquired

161 that were recorded at the 323 stations denoted by solid triangles and squares in Figure 1. Only

162 vertical component data are processed, meaning only Rayleigh waves are studied.

163 The data processing procedures follow those of *Bensen et al*. [2007] and *Lin et al*. [2008].

164 After removing the instrument responses, all records are bandpass filtered between 5 and 150 sec

165 period. We apply both temporal normalization and spectral whitening. Temporal normalization is

166 applied in an 80 second moving window. Cross-correlations are performed daily between all

167 pairs of stations and then are stacked over the two-year time window. Figure 3 presents example

168 cross-correlation record sections among Chinese stations, among F-Net stations, and inter-station

169 pairs between F-Net stations and Chinese stations.

170   Data quality control is discussed here and in section 3 and consists of five principal steps,

171   denoted A-E. Most of these steps are based on procedures summarized by *Bensen et al.* [2007]

172   and *Lin et al.* [2008], but because of the mixture of instrument types used in this study and the

173   existence of the Kyushu microseism we add extra steps to ensure the reliability of the resulting

174   dispersion measurements. Step A. A dispersion measurement is retained for a cross-correlation at

175   a given period only if signal-to-noise ratio (SNR) > 15 at that period, where SNR is defined by

176   *Bensen et al.* [2007]. Step B. We remove the effects of the Kyushu microseism, which we discuss

177   further in section 3. Step C. We retain an observation at a given period only if both the group and

178   phase velocities are measured by the automated frequency-time analysis method [*Bensen et al.*,

179   2007]. Group and phase velocity are separate measurements and are not constrained to agree

180   even though they are related theoretically [e.g., *Levshin et al.*, 1999]. Step D. We identify and

181   discard stations with bad instrument responses. Step E. This step is broken into two parts. First,

182   we only accept dispersion measurements with path lengths $\geq$ 3 wavelengths. Second, a

183   measurement is retained only if the misfit determined from the final dispersion map is less than

184   12 sec for group travel time and less than 5 sec for phase travel time, which is somewhat more

185   than twice the standard deviation of the final misfit. Group and phase velocity dispersion

186   measurements of Rayleigh waves are obtained on the symmetric component of inter-station

187   cross-correlations except for paths identified as affected by the Kyushu microseism, as discussed

188   further in section 3.

189   Because the seismic instruments used in this study differ in origin between China, Japan, and

190   the US, and instruments can vary within CEArray between provinces in China, it is important to

191   identify errors and inconsistencies in response files. For the IRIS network data, F-Net data and

192   most of the Chinese stations, full response (RESP) files including both analog and digital filter

193   stages are available. For a small number of Chinese stations we possess only pole-zero response

194   files missing the digital filtering stages. We find that for the CEArray instruments, the analog

195   pole-zero responses sometimes differ from the full responses computed from the RESP files for

196   stations that have both types of response files. Therefore, we discard the Chinese stations for

197   which we have not been able to acquire RESP files. This affected 36 stations, none of which are

198   included in the 232 Chinese stations shown in Figure 1. Using these response files, all data are

199   converted to velocity prior to cross-correlation.

200   Two other procedures are applied to find other instrument response errors. First, we identify

201   polarity errors (π phase shift) that may represent a units error in the instrument by comparing

202   P-wave first motions observed across the array following deep, distant teleseisms. We also look

203   for half-period misfits based on the final tomography maps at each period. These procedures

204   identified seven stations with polarity errors that are discarded. Second, we compare the phase

205   and group times measured on the positive and negative lags of all cross-correlations, which

206   identifies timing errors as long as both lags have a high SNR [*Lin et al.*, 2007]. These procedures

207   identified and discarded three Chinese stations.

208   The 2π phase ambiguity inherent in phase velocity measurements is resolved iteratively, first

209   based on phase velocities predicted by the 3D model of *Shapiro et al*. [2002] and then later on

210   with increasingly refined phase velocity maps that are determined in this study.

211   With more than 300 stations, in principal about 50,000 inter-station cross-correlations could

212   be obtained. Quality control procedures reduce this number appreciably as Table 1 illustrates.

213   After applying the selection criteria, we obtain between about 10,000 to 30,000 group and phase

214   velocity measurements for tomography at periods ranging from 10 to 45 sec, although the

215   number of paths drops sharply below 10 sec period.

216   **3. The Effect of the Localized Persistent Kyushu Microseism**

217   In northeast China, a strong disturbance appears on the cross-correlation waveforms. The

218   record sections shown in Figures 3b and 3c show this disturbance, which appears as precursory

219   signals (identified by white dots) in addition to the expected surface wave part of the empirical

220   Green's function. Another example is presented in Figure 4a, where the disturbance appears at

221   positive correlation lag between 100 and 150 sec, whereas the desired Rayleigh wave signal

222   arrives much later and is seen clearly only on the negative correlation lag. The arrival of these

223   disturbances near to the Rayleigh wave wavepackets interferes with the ability to measure

224   Rayleigh wave speeds accurately. In fact, the effect tends to bias Rayleigh waves fast,

225   particularly for Rayleigh wave group velocities.

226   These precursory signals are due to the persistent, localized Kyushu microseism that has been

227   identified and located by *Zeng and Ni* [2010] on the island of Kyushu, Japan, within our study

228   region. *Kawakatsu et al.* [2011] explain the signal as having originated from long period volcanic

229   tremor beneath Aso Volcano on Kyuhsu Island. The period band of this microseism is

230   dominantly between 8 and 14 sec, and it is somewhat reminiscent of the longer period persistent

231   26 sec microseism located in the Gulf of Guinea [e.g., *Shapiro et al.*, 2006]. We are interested in

232   minimizing its interference with surface wave dispersion measurements across the study region.

233   To do so, we have re-located it, confirming the location of *Zeng and Ni* [2010], and have

234   developed a data processing procedure that allows reliable dispersion curves to be obtained

235   across the Sino-Korean Craton, northeastern China, the Korean peninsula, and the Sea of Japan.

236   The effect of the procedure, however, is to reduce the number of measurements in the region

237   broadly surrounding Kyushu at periods between 10 and 18 sec.

238   **3.1 Re-location of the Kyushu microseism**

239   We use the envelope functions between periods of 10 and 12 sec for the Kyushu signal

240   observed on inter-station cross-correlations to locate the Kyushu microseism. An example

241   envelope function is shown in the bottom panel of Figure 4a. Based on an initial observed group

242   velocity map at 11 sec period, we predict the theoretical arrival time of the Kyushu signal for

243   each inter-station pair for each hypothetical source location on a broad map of the region. We

244   then take the observed amplitude of the envelope function at the predicted time and plot it at the

245   hypothetical source location. An example is shown in Figure 4b, where the hyperbola identifies

246   the set of possible locations for the Kyushu microseism for a particular inter-station pair. We

247   refer to this figure as the migrated envelope function. Finally, we stack over all migrated

248   envelope functions for cross-correlations involving the GSN stations INCN (Inchon, South

249   Korea) and SSE (Shanghai, China), with the paths shown in Figure 4c. The resulting stack of

250   migrated envelope functions is shown in Figure 4d, demonstrating that the re-location of the

251   Kyushu microseism is close to the location from *Zeng and Ni* [2010] (Fig. 4d) on Kyushu Island.

252   **3.2 Eliminating the effect of the Kyushu microseism**

253   Assuming that the Kyushu microseism is a point source, for each inter-station

254   cross-correlation we calculate the expected arrival time of the Kyushu signal as well as the

255   theoretical arrival time of the Rayleigh wave between the two-stations. The relative arrival times

256   of the Kyushu signal (white dots in Fig. 3) and the inter-station Rayleigh waves are highly

257   variable but systematic. When the arrival times of the Kyushu signals are close to the desired

258   surface waves it becomes difficult to separate the signals between 8 and 14 sec period on the

259   symmetric component of the cross-correlations. However, the systematics of the relative arrival

260   times can be exploited to separate the two waves and obtain reliable Rayleigh wave group and

261   phase velocity measurements for most paths by focusing on a single correlation lag. Figure 5

262   illustrates this point by showing how the Kyushu signal biases group and phase velocity

263   measurements when the Rayleigh wave and the Kyushu signal arrive nearly simultaneously on

264   the negative lag component of the cross-correlation (Fig. 5a). Figure 5b shows the symmetric

265   component of the cross-correlation. Figure 5c is the positive component of the cross-correlation,

266   which is free from the Kyushu disturbance. On frequency-time diagrams [e.g., *Ritzwoller and*

267   *Levshin*, 1998], large differences between positive and negative correlation lag times are

268   observed at periods shorter than 18 sec, especially in the group velocity dispersion curves. This is

269   illustrated by Figure 5d,e, in which the Kyushu signal disturbs the dispersion curve in Figure 5d

270   but not Figure 5e. The Kyushu signal causes the measured group velocity to bias towards higher

271   velocities (Fig. 5f). At periods longer than 18 sec, however, differences are quite small.

272   Therefore, to separate the desired inter-station Rayleigh wave from the Kyushu signal we

273   must measure the Rayleigh wave dispersion on the correlation lag opposite from the arrival of

274   the Kyushu signal in the period band of disturbance. The symmetric component, the average of

275   the cross-correlation at positive and negative lag, cannot be used if the Kyushu signal arrives

276   near the inter-station Rayleigh wave at either positive or negative correlation lag time.

277   In practice, we take the following steps. At periods greater than 18 sec, we ignore the Kyushu

278   disturbance and use the symmetric component for dispersion measurement. At periods less than

279   or equal to 18 sec, the relative arrival time of the Kyushu signal and the surface wave signal at 12

280   sec period guides the measurement. If the expected apparent speed of the Kyushu signal is

281   greater than 4.5 km/s and the length of the inter-station path is longer than three wavelengths,

282   then the Kyushu signal is well enough separated from the surface wave that phase and group

283 velocities can be measured on the symmetric component. If the expected apparent speed of the

284 Kyushu signal is less than 4.5 km/s, then we make group and phase velocity measurements on

285 the correlation lag opposite from the Kyushu signal.

286 As Figure 5a exemplifies, the correlation lag on the opposite side of the Kyushu signal may

287 have a smaller amplitude than the lag containing the Kyushu signal and will have a significantly

288 lower SNR than the symmetric component. Thus, this data processing procedure tends to lower

289 the SNR for many inter-station cross-correlations between 8 and 16 sec period. Because

290 observations are discarded if SNR < 15, this reduces the data set appreciably, as Table 1 shows.

291 We are still left with sufficient high quality measurements to perform tomography across much

292 of the study region, however.

293 If left untreated, the Kyushu signal would have vitiated our maps predominantly in South

294 Korea, the southern Sea of Japan, and in the Kyushu region from 8 to 14 sec period. Artifacts

295 would have been much stronger on the group velocity than the phase velocity maps. Above 16

296 sec period, however, even if left untreated, the Kyushu signal would have had only a weak effect

297 on the estimated maps.

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4. Rayleigh Wave Dispersion Maps

4.1 Construction of the dispersion maps

Surface wave tomography is applied to the selected dispersion measurements to produce Rayleigh wave group and phase speed maps on a 0.5°×0.5° grid using the ray-theoretic method of *Barmin et al*. [2001]. Being mostly determined over regional (non-teleseismic) inter-station distances, the dispersion measurements observed here will not be affected strongly by off-great circle effects [e.g., *Lin et al*., 2009] except for relatively long paths undergoing a continent-ocean transition. This will primarily affect paths from Japanese F-Net to Chinese stations and the maps of the Sea of Japan may be degraded somewhat by this effect. Finite frequency effects will also be weak in the period band of study [*Lin et al*., 2011]. The tomographic method of *Barmin et al.* is based on minimizing a penalty functional composed of a linear combination of data misfit, model smoothness, and model amplitude. The choice of the damping parameters is based on the optimization of data misfit and the recovery of coherent model features. Due to a shortage of measurements, we are unable to produce reliable maps below 8 sec period and above about 45 sec period. The resulting Rayleigh wave dispersion maps, therefore, are constructed between 8 and 45 sec on a 2 sec period grid.

During tomography, resolution is also estimated via the method described by *Barmin et al*. [2001] with modifications presented by *Levshin et al*. [2005]. Resolution is defined as twice the

319 standard deviation of a 2-D Gaussian fit to the resolution surface at each geographic node.

320 Examples of resolution maps are plotted in Figure 6 for the 16 sec and 35 sec period

321 measurements. Resolution is estimated to be about 100 km across most of the region of study but

322 degrades to 200-400 km near the boundary of our studied area where station coverage is sparse.

323 Histograms of data misfit using the dispersion maps at periods of 14 sec, 20 sec, 30 sec, and

324 40 sec are plotted in Figure 7 for phase and group velocity. Group travel time misfits are

325 typically 3-4 times larger than phase travel time misfits because phase velocity measurements are

326 more accurate [*Bensen et al*. 2007; *Lin et al*., 2008] and group velocity sensitivity kernels have a

327 larger amplitude. The range of standard deviations of the group travel time misfits runs between

328 about 4 and 5 sec and that of phase travel time between about 1 and 1.5 sec. Phase travel time

329 misfits of ~1 sec between 14 and 30 sec period are indicative of the quality of the data set, being

330 similar to misfits that result from USArray data [e.g., *Lin et al*., 2008].

331 **4.2 Description of the dispersion maps**

332 Group and phase velocity maps at periods of 12, 20, 30, and 40 sec are plotted in Figures 8

333 and 9. Only those areas where the spatial resolution is better than 400 km are shown. The 100 km

334 spatial resolution contour is also plotted as a continuous white line. Because the 3-D model is

335 discussed in considerable detail in section 6, discussion of the dispersion maps here is brief but is

336 illuminated by the sensitivity kernels shown in Figure 10. Both phase (solid lines) and group

velocity (dashed lines) kernels are shown, computed for 10 and 40 sec period at two locations in the final model constructed here: (a) Sea of Japan and (b) North China Craton.

As Figure 10 illustrates, at each period the group velocity measurements are sensitive to shallower structures than the phase velocities. Phase velocity maps, therefore, generally should be compared with somewhat shorter period group velocity maps. Thus, comparison between Figures 8 and 9 reveals that the phase velocity map at 12 sec is similar to the group velocity map at 20 sec and the phase velocity map at 20 sec is quite like the group velocity pattern at 30 sec. Because of the lower uncertainty in the phase velocity measurements, however, the phase velocity maps generally are more accurate than the maps of group velocity.

The dispersion maps are sensitive to quite different structures between short and long periods and between continental and oceanic regions even at the same period as Figure 10 illustrates. The short period dispersion maps (12 sec group and phase velocity in Fig. 8 and 9) are primarily sensitive to upper crustal velocities in continental areas, which are dominated by the presence or lack of sediments. In regions with oceanic crust, sensitivity is predominantly to uppermost mantle structure. At intermediate periods (20 sec phase velocity, 20-40 sec group velocity), the maps are mostly sensitive to mid- to lower-crustal velocities beneath continents and continent-ocean crustal thickness variations. At long periods (30-40 sec phase velocity), the maps predominantly reflect crustal thickness variations on the continent and uppermost mantle conditions beneath oceans.

356 At short periods (12 sec), the dispersion maps in Figures 8-9 exhibit low velocity anomalies

357 where sediments are present [*Bassin et al.*, 2000]; e.g., the Songliao Basin (SLB), Huabei Basin

358 (HB), Bohai Bay (BH), North China Platform (NC), the Yellow Sea (ESYS), and the Tsushima

359 Basin (TB) in the southern Sea of Japan. The younger, Cenozoic age sediments of Bohai Bay and

360 the Huabei Basin of the North China Craton are much slower than the older, largely Mesozoic

361 age sediments of the Songliao Basin. Higher velocities are imaged in the mountains surrounding

362 the basins; e.g, the Great Xing'an Range (GXAR), Lesser Xing'an Range (LXAR), the Changbai

363 Mountains (CBM), the Yinshan Mountains (YSM), and the Taihang Mountains (THM), which is

364 consistent with the presence of crystalline rocks near the surface. Weak positive anomalies are

365 observed in the Korean Peninsula.

366 At intermediate periods (20 sec phase velocity, 20-30 sec group velocity), relatively low

367 wave speeds are still observed for the deeper basins: Songliao, Huabei, Bohai Bay, and Yellow

368 Sea. Group and phase velocity are very high in the Sea of Japan reflecting oceanic mantle

369 lithospheric shear wave speeds that clearly outline the continental boundary of the oceanic crust.

370 It should be noted that off-great-circle effects that may exist are not severe enough to distort the

371 borders of the oceanic crust badly. A significant low velocity anomaly appears in Xing'an-East

372 Mongolia. The anomaly does not develop along the Tancheng-Lujiang Fault, however, but rather

373 to the west of the Songliao Basin, encompassing the Great Xing'an Range and the Yinshan

374 Mountains north of the Ordos Block. To the east of the North-South Gravity Lineament the crust

375   is much thinner than to its west [e.g., *Xu*, 2007].

376   For group velocity at 40 sec and phase velocity at periods of 30-40 sec, four principal

377   observations are worth noting. First, the high wave speed anomalies of the Sea of Japan diminish

378   from 30 sec to 40 sec period on the phase velocity maps. This reflects relatively thin lithosphere

379   beneath the Sea of Japan and increased sensitivity to the underlying asthenosphere by the longer

380   period waves. Second, the Songliao, Huabei, and Bohai basins are situated within the

381   Songliao-Bohai Graben, which is believed to have formed by back-arc extension and potential

382   rifting [*Liu et al.*, 2001]. This graben, extending from the Songliao Basin into Bohai Bay, is

383   associated with a continuous high velocity anomaly, which is clearly observable on the 30 and 40

384   sec phase velocity maps. Third, on the longest period phase velocity maps, low wave speeds are

385   found along the entire length of the Northeast Asia Foldbelt, extending into the Korean Peninsula.

386   Such clear continuous low wave speeds are not apparent on the 40 sec group velocity map

387   because of the shallower sensitivity of group velocities than phase velocities. Apparent

388   discrepancies between phase and group velocities such as this one are reconciled by the 3-D Vsv

389   model, as discussed in section 5. Finally, a belt of low velocities is present on the longest period

390   phase velocity maps in the far eastern Sea of Japan bordering Hokkaido and northern and central

391   Honshu, extinguishing in southern Honshu, and then re-initiating along Kyushu and the Ryukyu

392   Islands.

393

5. Construction of the 3-D Model

This study is based exclusively on Rayleigh waves, which are primarily sensitive to vertically polarized shear wave speeds (Vsv) that may be slower than horizontally polarized wave speeds (Vsh) in regions where anisotropic minerals are preferentially aligned in the horizontal plane. The velocity difference between Vsv and Vsh is referred to as radial anisotropy. Thus, the model we produce is a Vsv model. In the presence of substantial radial anisotropy, Vsv can be several percent lower than the effective isotropic shear wave speed, Vs. Radial anisotropy is common in both the mantle [e.g., *Ekstrom and Dziewonski,* 1997; *Shapiro and Ritzwoller*, 2002] and the crust [e.g., *Shapiro et al*., 2004; *Moschetti et al*., 2010a,b], and is geographically variable. We will generally refer to our 3D model as being a Vsv model, but for simplicity will also refer to it as a shear wave speed or Vs model.

The 3D Vsv model is based on the Rayleigh wave phase and group speed maps from 8 to 45 sec period on a 0.5°×0.5° grid across the study region. Periods at which resolution is greater than 200 km are not included in the inversion. This determines the outline of the model and in oceanic regions local dispersion curves typically begin at 12 sec period. Local dispersion curves from the eight locations identified with red dots in Figure 2 are shown in Figure 11 with error bars. The group velocity and phase velocity curves shown in Figure 11 are mostly smoothly varying and are able to be fit by a vertically simple Vs model at each point within observational error. In particular, the group and phase velocity curves are reconcilable at each point. Local misfit is

413   presented for each point in Figure 11, where "RMS misfit" means the square root of the reduced

414   chi-square value:

$$\text{RMS misfit} = \left[\frac{1}{N} \sum_i \frac{d_i - p_i}{\sigma_i^2} \right]^{\frac{1}{2}} \tag{1}$$

416   where d_i is the observed group or phase velocity value, p_i is the value predicted from the model,

417   σ_i is the uncertainty for the observation, and N is the total number of phase plus group velocity

418   values along the dispersion curves.

419   Uncertainties in phase and group velocity curves are notoriously difficult to determine

420   reliably, as they are typically underestimated by most tomographic codes based on standard error

421   propagation [e.g., Barmin et al., 2001]. They can be estimated reliably by using the eikonal

422   tomography method [*Lin et al.,* 2009], but the method requires more uniform array spacing than

423   exists in the study region. Therefore, the uncertainties shown in Figure 11 are not from the data

424   set presented here, but, as discussed in section 2, misfit statistics in the present study are similar

425   to those that emerge in the western US based on the regularly spaced USArray Transportable

426   Array. For this reason, we take the average uncertainties estimated by eikonal tomography using

427   USArray in the western US, but to be conservative we double them [*Lin et al., 2009; Moschetti*

428   *et al.,* 2010a, 2010b].

429   The 3D model is constructed via a Monte-Carlo method that is similar to the methods of

430   *Shapiro and Ritzwoller* [2002] and *Yang et al.* [2008]. The starting model for the Monte-Carlo

431   search derives from the Vsv values of the global model of *Shapiro and Ritzwolle*r [2002] with

432   two principal modifications. First, we simplify the starting crustal model considerably: the two

433   sedimentary layers and three crystalline crustal layers of the model of *Shapiro and Ritzwoller* are

434   averaged separately to define only two constant velocity layers (not including a possible water

435   layer). Second, crustal thickness is not allowed to be less than 20 km in the Sea of Japan, as

436   discussed in section 6.1. Uniformly distributed perturbations in Vsv are considered using a single

437   sedimentary layer with variable thickness and shear velocity, four B-splines for Vsv in the

438   crystalline crust, and five B-splines in the mantle to a depth of 150 km, below which the model is

439   a constant velocity half-space. The resulting model is vertically smooth in both the crystalline

440   crust and mantle. Moho depth is allowed to vary in a uniform interval of ± 10 km relative to the

441   starting model. Vsv is constrained to increase monotonically in the crystalline crust and the depth

442   derivative of velocity directly below Moho is constrained to be positive (i.e., velocity increases

443   with depth right below Moho, but can decrease deeper into the mantle). Both constraints are

444   introduced to reduce the model space, in particular the magnitude of the trade-off between Moho

445   depth with structures at depths adjacent to Moho. The positivity constraint on the depth

446   derivative of velocity in the uppermost mantle typically introduces a thin low velocity "sill" right

447   below Moho, which is an artifact of the constraint.

448   The model has no radial anisotropy, thus Vs = Vsh = Vsv. Also, we assume that the

449   crystalline crust and mantle is a Poisson solid and set Vp = 1.73Vs, in the sediments we use Vp =

450   2.0Vs, and for density we use the scaling relation advocated by *Christensen and Mooney* [1995]:

451    $\rho = 0.541 + 0.3601$Vp, where ρ is in g/cm^3 and Vp is in km/s.   We apply a physical dispersion

452   correction [*Kanamori and Anderson, 1977*] using the Q model from PREM [*Dziewonski &*

453   *Anderson 1981*], and the resulting model is reduced to 1 sec period. Offshore, the water depth is

454   re-calculated based on a 0.2°×0.2° average of bathymetry at each point.

455   Models are chosen randomly guided by a Metropolis algorithm [e.g., *Mosegaard and*

456   *Tarantola*, 1995] and are accepted if the reduced χ^2 misfit to the dispersion curves is less than

457   twice the minimum misfit, χ^2_{min}, at each location. Reduced χ^2 misfit is the square of RMS misfit

458   defined by equation (1). RMS misfit averages less than 1.0 across the region of study. Much of

459   this misfit results from group velocities at short periods.

460   As examples, the procedure yields the eight ensembles of models presented in Figure 12

461   derived from the eight pairs of Rayleigh wave dispersion curves shown in Figure 11. The

462   ensemble is represented by the grey shaded region, which presents two standard deviations (2σ)

463   around the mean at each depth in each direction. The dispersion curves predicted by the

464   best-fitting model are shown in Figure 11 as the blue (group velocity) and red (phase velocity)

465   lines. At each depth, the Vsv model that we plot in subsequent figures and its uncertainty are

466   defined by the middle and one standard deviation of the ensemble, respectively. The

467   quarter-width of the ensemble is approximately one standard deviation (1σ). The uncertainties

468   in the model are largest where shear wave speeds trade-off effectively with boundary topography,

469   which occurs near free boundaries in the model: Moho and the base of the sedimentary layer.

470   Thus, uncertainties typically grow near the top of the crystalline crust and both above and below

471   Moho.

472   The eight examples of dispersion curves and the ensembles of models that fit them, presented

473   in Figures 11 and 12, demonstrate how vertically smooth models with two discontinuities can fit

474   the data well across the study region. A closer inspection of the model profiles, however, reveals

475   the limitations of inversion methods based exclusively on surface waves. In particular, the model

476   is affected by the starting model around which the Monte-Carlo method samples, particularly

477   sedimentary thickness and crustal thickness.

478   On the continent, input model-dependence is most important beneath sedimentary basins

479   where our simple parameterization of sedimentary velocities may not faithfully represent local

480   structure. A potential example of this is shown in Figure 12e, beneath the Songliao Basin. Model

481   velocities beneath the basin are very low (~3.2 km/s) in the crystalline upper crust, which causes

482   a large vertical crustal velocity gradient that may not be realistic. Thus, errors in the input model

483   and the simplicity of parameterization of sedimentary structure may bias Vs low in the upper

484   crystalline crust. For this reason, we do not interpret the resulting 3D model at depths above the

485   lower crust. Imposition of constraints from receiver functions would help to resolve this issue,

486   but is beyond the scope of this work.

487   In the ocean, in contrast, crustal thickness in the starting model is more poorly known than on

488   the continent. Surprisingly, however, the resulting model is fairly independent of the starting

489   model. This is illustrated by Figure 13, which shows the resulting ensemble of models for a point

490   in the Sea of Japan with two different starting models in which Moho is at 8 km (Fig. 13a) or 20

491   km (Fig. 13b), respectively. In both cases the surface wave data are fit admirably and nearly

492   identically (e.g., Fig. 12h). With different starting Moho depths of 8 km and 20 km, estimated

493   mantle velocities at 20 km depth are nearly identical in the two ensembles. Crustal velocities and

494   crustal thicknesses differ somewhat (average = 13 km and 15 km, respectively), but are within

495   estimated error. Although crustal thickness may vary from 10 km [*Sato et al.*, 2004] to more than

496   20 km [*Kurashimo et al.*, 1996] across the Sea of Japan, by setting the starting Moho depth to 20

497   km we are able to estimate mantle velocities reliably and recover crustal velocities and

498   thicknesses to within estimated errors. Further discussion of crustal thickness beneath the Sea of

499   Japan is presented in section 6.2, which describes the observed anti-correlation between crustal

500   thickness and sea floor depth and provides further support for the argument that crustal thickness

501   is relatively well determined beneath the area.

502   A terminological issue may require clarification. By "crustal thickness" we mean the

503   thickness of the crust including surface topography and a water layer if it exists. By "Moho

504   depth", we mean the depth to Moho below the geoid. By "solid crustal thickness" we refer to the

505   thickness of the crust not including a water layer.

506   Results from the individual vertical profiles are aggregated into the 3-D model: the middle of

507   the ensemble of accepted models defines the model profile at each point and the spread at each

508   depth yields the uncertainty. Figure 14 shows the Vsv model at four depths: (a) in the lower crust

509   at 80% of the depth to Moho, (b) 40 km, (c) 60 km, and (d) 80 km beneath the surface. Because

510   the dispersion curves extend only up to 45 sec period, the model is not reliable beyond a depth of

511   ~100 km. Estimated Moho depth and its uncertainty are shown in Figure 15. Figure 15 also

512   presents the estimated velocity jump across Moho and its uncertainty. Uncertainties in shear

513   wave speed at the four depths shown in Figure 14 are presented in Figure 16. Five vertical slices

514   of the model are presented in Figure 17 along the profiles identified in Figure 15a:: A-A'

515   extending northwest-southeast through the Songliao Basin and the Sea of Japan, B-B' running

516   northeast-southwest through the Songliao Basin and the eastern North China Craton, C-C'

517   extending roughly west-east from the Ordos Block through the North China Craton, the Bohai

518   Gulf, Korean Peninsula, and the southern Sea of Japan, D-D' running northwest-southeast from

519   Xing'an-East Mongolia through the Yellow Sea to the East China Sea, and E-E' extending

520   northwest-southeast from the Ordos Block, through the eastern NCC, to the East China Sea. The

521   model is discussed in detail in section 6.

522   Model uncertainties are presented in Figure 15b for Moho and in Figure 16 corresponding to

523   the horizontal depth slices in Figure 14. Model uncertainties are largest near Moho, i.e., in the

524   lower crust and uppermost mantle as Figure 16a,b reflect. At 40 km depth, large uncertainties

525 follow the North- South Gravity Lineament, which marks the point at which Moho depth

526 approximately crosses 40 km. Uncertainties at 40 km are smallest in the Sea of Japan because the

527 crust is thinnest and this depth is most distant from Moho. Uncertainties diminish as one moves

528 deeper beneath Moho, but grow again in the deeper parts of the model as data constraints weaken

529 (Fig. 16c,d). Uncertainties in crustal thickness are nearly homogeneous across the study region,

530 but diminish somewhat beneath the Sea of Japan. Conversely, uncertainties in the velocity jump

531 across Moho are largest beneath the Sea of Japan.

532 **6. Discussion**

533 **6.1 Structural features of the 3D model**

534   As discussed in section 5, the trade-off between sedimentary thickness and shear velocities in

535 the upper crust is strong beneath the Songliao and Huabei Basins; thus, we do not present or

536 discuss upper crustal velocities here. Lower crustal velocities are much more robust, however,

537 and they are shown in Figure 14a, which is a profile taken at 80% of the Moho depth. On the

538 continent, lower crustal velocities average about 3.75 km/s, but bifurcate. Velocities are higher,

539 averaging about 3.90 km/s, beneath the Korean peninsula, the Northeast Asian Fold Belt and the

540 Songliao Basin but are slower elsewhere, including the North China Craton, the Xing'an-East

541 Mongolia, and the Yellow Sea, averaging about 3.80 km/s. The high velocity lower crust beneath

542 far eastern Asia and the Songliao Basin may represent compositional differences between these

543   regions and the tectonically distinct North China Craton and Xing'an-East Mongolia. The

544   slowest lower crustal velocities are observed beneath the Sea of Japan, running between about

545   3.50 to 3.60 km/s. The largest uncertainties in lower crustal velocities are for the Sea of Japan,

546   however (Fig. 16a).

547   We use the term "mantle lithosphere" here qualitatively and relatively, and it may be

548   substituted with the term "seismic lid". We take it to mean the high velocity zone in the

549   uppermost mantle more or less directly below Moho. Thus, in our usage, the base of the

550   lithosphere is the depth at which velocities begin to decrease beneath a higher velocity overriding

551   mantle.   With this definition, on the vertical profiles in Figure 17 the base of the lithosphere

552   occurs at the depth at which blue colors give way to green colors (~4.5 km/s) if the highest

553   velocities are blues. If the highest velocities are greens (e.g., beneath the Sea of Japan), the base

554   of the lithosphere occurs where the greens give way to white (~4.4 km/s). The term

555   "asthenosphere" is defined more quantitatively, as the region of the mantle where $Vs < 4.25$ km/s,

556   or where red colors are found in Figure 17. With these definitions, for example, the lithosphere is

557   70-80 km thick beneath the Songliao Basin (Fig. 17, B-B') and is 50-60 km thick beneath the

558   Sea of Japan (Fig. 17, A-A'). The asthenosphere exists on the vertical profiles only beneath some

559   of the North China Craton (Fig. 17, B-B', C-C'), the regions bordering the Sea of Japan (Fig. 17,

560   A-A', C-C'), and the East China Sea (Fig. 17, D-D', E-E'). Thus, with these definitions, across

561   most of the region the asthenosphere is not reached by 100 km depth. Across most of the region

562 the lithosphere is underlain by a zone that is transitional between lithospheric and asthenospheric,

563 which appears mainly as green and yellow colors on the vertical profiles.

564 In terms of deeper structures, we describe four principal features: (1) crustal thickness

565 variations across the region, (2) the lithospheric structure of the Songliao – Bohai graben, (3) the

566 uppermost mantle beneath the Sea of Japan and the East China Sea, and (4) the uppermost

567 mantle beneath the North China Craton.

568 (1) Crustal thickness is presented in Figure 15a. Although internal boundaries such as Moho

569 are notoriously difficult to estimate reliably from surface waves, particularly beneath continents,

570 crustal thickness is about 40 km along the North-South Gravity Lineament (NSGL). The region

571 west of the North-South Gravity Lineament is in the crust at 40 km depth, whereas 40 km lies in

572 the mantle beneath the rest of the region. Thus, the lowest velocities near the western edge of the

573 map in Figure 14b simply reflect crustal wave speeds.

574 Beneath the Sea of Japan, as Figure 13 illustrates, the ability to estimate crustal thickness

575 improves relative to the continent. A blow up of the crustal thickness is presented in Figure 18a.

576 As expected, the crust is thinnest where the sea is deepest, principally in the Japan Basin, which

577 defines the northern and western parts of the Sea of Japan. The crust thickens, for example,

578 beneath the Yamato Rise, the topographic rise in the central Sea of Japan. Crustal thickness

579 beneath the Sea of Japan is discussed further in section 6.2, where the anti-correlation observed

580   between crustal thickness and water depth is used to support the reliability of the crustal

581   thickness estimates in oceanic regions.

582   (2) At 40 km depth east of the North-South Gravity Lineament relatively high wave speeds

583   (>4.4 km/s) lie beneath the Songliao Basin and run to the Bohai Sea, as Figure 14b shows. The

584   high velocities beneath the Songliao Basin form a lid that extends to 60-70 km but then

585   terminates, being underlain by more normal continental uppermost mantle wave speeds beneath

586   the Basin by 80 km depth. This can be seen clearly on vertical profiles A-A' and B-B' of Figure

587   17, in which the relatively thick fast lithosphere beneath the Songliao graben is contrasted with

588   the thinner slower lithosphere of the North China Craton and the Sea of Japan. The mantle

589   lithosphere, however, beneath the Songliao Basin is not cratonic in nature, contrasting with the

590   much thicker lithosphere beneath the Ordos Block (e.g., Fig. 14d). Although the Ordos Block is

591   only partially imaged in this study, high velocities are seen to underlie it in the uppermost mantle

592   extending at least to 100 km depth (and presumably deeper) as vertical profile C-C' in Figure 17

593   illustrates. Although the lithosphere underlying the Songliao Basin is not as thin as beneath the

594   North China Craton, it is thin for a region believed to have a basement of Archean age.

595   (3) At 40 km depth, relatively high wave speeds (>4.4 km/s) also underlie the Yellow and

596   Japan Seas. Although similar in the uppermost mantle, below 40 km the mantle structures

597   beneath these seas lose their similarity. The uppermost mantle beneath the Yellow Sea is

598   continental in character as profile D-D' in Figure 17 shows. Much lower wave speeds exist in the

599   back-arc spreading region beneath the Sea of Japan, as can been seen on Figure 14c,d. Starting at

600   about 60 km depth, low velocities bound the eastern margin of the Sea of Japan and also to a

601   somewhat lesser extent the western margin, extending onto the continent. Low velocities also

602   bound the eastern edge of the East China Sea, east and southeast of the Yellow Sea near the

603   Ryuku Islands. By 80 km depth, the bifurcated low velocity zone across the Sea of Japan has

604   largely merged into a more homogeneous low velocity asthenophere beneath the entire sea, and

605   the western edge of the anomaly has spread further inland to underlie the Northeast Asia Foldbelt,

606   including the volcanic region of the Changbai Mountains. This is presented most clearly in

607   Figure 19, which displays two low velocity contours at 80 km depth: $Vs < 4.25$ km/s and 4.25

608   km/s $\leq Vs < 4.35$ km/s. This also can be seen clearly in profile A-A' of Figure 17, which extends

609   across the center of the Sea of Japan, but is also visible in profile C-C' which run through the sea

610   farther south. We refer to the low velocities bounding the Sea of Japan as a "Y-shaped" anomaly

611   on vertical profiles, which is apparent on profile A-A'. A relatively high velocity 50-60 km thick

612   lithosphere is observed beneath the central Sea of Japan, as profiles A-A' and C-C' illustrate. This

613   relatively thin lithosphere should be contrasted with the thicker faster lithosphere beneath the

614   Songliao Basin, which is clearly seen on profile A-A'. The bordering Y-shaped asthenospheric

615   anomaly and the thin lithosphere underlying the Sea of Japan are discussed further in section 6.3.

616   The East China Sea lies near the southeastern corner of our study region, but like the Sea of

617   Japan its eastern margin near the Ryuku Islands is underlain by low velocities in the uppermost

mantle. This is apparent in Figure 19, Figure 14d and vertical profile D-D' in Figure 17. Unlike

the Sea of Japan, however, there is no "Y-shaped" anomaly beneath the East China Sea as the

western part of the sea merges with fast continental lithosphere beneath the Yellow Sea with no

low velocity feature in the uppermost mantle.

(4) The North China Craton deviates from most of the rest of continental northern and

northeastern China by possessing much slower wave speeds at depths greater than about 60 km.

This can be seen clearly on profile B-B' of Figure 17, which shows that the mantle lithosphere

beneath the North China craton is thinner and slower than beneath the Songliao Basin. Profile

C-C' illustrates, however, that the lithosphere extends deeper than beneath the Sea of Japan, with

underlying asthenosphere that is not as slow. However, asthenospheric wave speeds do underlie

the lithosphere beneath the North China Craton below about 80 km depth. These low wave

speeds in the uppermost mantle are horseshoe in shape on horizontal cross-sections across the

eastern North China Craton as seen in Figure 19, in which the low velocities follow the western

and eastern flanks of the eastern part of the North China Craton. The middle of the eastern part of

the North China Craton is not slow, however. The seismic structure of the North China Craton is

discussed further in section 6.4.

6.2 Seafloor depth and crustal thickness beneath the Sea of Japan

636   The crustal structure beneath the Sea of Japan is highly heterogeneous [e.g., *Taira,* 2001].

637   The Japan Basin is principally oceanic crust, while most of the Yamamoto and Tsushima basins

638   are extended continental crust, the Korean Plateau and Yamamoto Rise are composed mainly of

639   continental crust, and the eastern and southern boundaries of the Sea are rifted continental crust

640   [*Tamaki et al.,*1992]. Thus, crustal thickness varies appreciably across the basin in a way that we

641   expect to be related to seafloor depth, assuming isostatic compensation. Figure 18b presents

642   seafloor depth (bathymetry) across the Sea of Japan plotted versus solid crustal thickness from

643   our 3-D model. Seafloor depth and our estimate of crustal thickness (Fig. 18a) are correlated,

644   although there is substantial scatter. Assuming that water depth, H_W, is the dependent variable

645   and solid crustal thickness, H_C, is the independent variable, the least-squares linear fit, shown as

646   the solid red line in Figure 18b, is $H_W = -0.22\ H_C + 6.07$, where H_W, H_C, and the constant term

647   are in km. The linear fit provides a variance reduction of 57%.

648   The observed derivative $dH_W/dH_C = -0.22$ provides information about the nature of isostatic

649   compensation in the region. For crustal Airy isostacy, with mantle density equal to 3.32 g/cm^3

650   and a water density of 1.05 g/cm^3, dH_W/dH_C can be computed for different crustal densities. For

651   an average continental crustal density of $\rho_{cont} = 2.75$ g/cm^3, $dH_W/dH_C \approx -0.25$, and for an average

652   oceanic crustal density of $\rho_{oc} = 2.9$ g/cm^3, $dH_W/dH_C \approx -0.19$. Thus, the estimated value of

653    $dH_W/dH_C = -0.22$ is understood to be an average of the values expected if the Sea of Japan were

654   composed homogeneously of either oceanic or continental crust. In fact, much of the scatter

655 around the linear fit seen in Figure 18b is due to crustal heterogeneity across the sea. However,

656 local isostasy may not be dominated by the crystalline crust alone, but may also be affected

657 strongly by sediments and thermal buoyancy from the mantle, both of which have been ignored

658 in the analysis here.

659 Many different lines could, in fact, fit the data in Figure 18b nearly as well as the linear fit

660 discussed above, which is presented with the solid red line. For example, if we assume that water

661 depth, H_W, is the independent variable and solid crustal thickness, H_C, is the dependent variable,

662 the least-squares linear fit, shown with the dashed red line in Figure 18b, would be $H_W = -0.41$

663 $H_C + 9.07$. Thus, a more careful analysis that considers lateral crustal heterogeneity, thermal

664 buoyancy from the mantle, the effects of sediments, and uncertainties in the crustal thickness

665 estimates is needed, but is beyond the scope of this paper. The results presented here demonstrate

666 the reliability of the crustal thickness estimates across the Sea of Japan and also provide evidence

667 that further analysis of the relation between seafloor depth and crustal thickness is warranted.

668 **6.3 Lithospheric and asthenospheric structure beneath the Sea of Japan**

669 Our 3-D model images what we refer to as a "Y-shaped" anomaly on vertical profiles with

670 low velocity arms reaching upward beneath the eastern and western borders of the Sea of Japan

671 extending well into continental East Asia on the west side (e.g., Fig. 17, A-A'), with an overlying

672 ~60 km thick lithosphere beneath the Sea of Japan. The location of the lowest wave speeds at 80

673 km depth can be seen in Figure 19. In recent years, several other tomographic studies have been

674 performed that are complementary to the ambient noise tomography presented here, which have

675 imaged the mantle beneath the Sea of Japan. These studies possess greater depth penetration

676 although lower resolution than our study and provide validation and assist with interpreting our

677 results.

678   Based on Rayleigh and Love wave phase velocities obtained from teleseismic earthquakes

679 using the two-station method, *Yoshizawa et al.* [2010] present a 3-D model of shear wave speeds

680 from 40 to 200 km depth. Although their resolution is best beneath Japan and fades by the central

681 Sea of Japan, their results in regions of good data coverage are similar to ours in several respects.

682 First, they report an average lithospheric thickness beneath the Yamato Rise and the Japan Basin

683 of 60 ±10 km, which is similar to what we observe (e.g., Fig. 17, A-A', C'C'). Second, they also

684 observe low velocity anomalies at 75 km depth bordering Honshu beneath the eastern Sea of

685 Japan and the East China Sea, west of the Ryuku Islands. Thus, they also image the eastern arm

686 of the Y-shaped anomaly that we report here and, like our model, observe it to migrate westward

687 with depth. Their data coverage does not extend into East Asia. However, the teleseismic body

688 wave study of *Zhao et al.* [2007] presented a large-scale image of the upper mantle beneath east

689 Asia, which shows the stagnation of the subducted Pacific Plate beneath the Sea of Japan and

690 clear low velocity anomalies in the back-arc region. This study concentrates interpretation on the

691 release of volatiles from the slab at several depths to produce the low velocity anomalies

692   observed in the back-arc region.

693   Thus, *Yoshizawa et al.* [2010] serves to validate our model and *Zhao et al.* [2007] provides a

694   means to interpret the Y-shaped anomaly based on what they call the "Big Mantle Wedge" model

695   of the Japanese back-arc and the devolatilization of the subducted slab. In their interpretation, the

696   western arm of the Y-shaped anomaly beneath the western Sea of Japan and the East Asian Fold

697   Belt would be caused by dehydration reactions of the stagnant slab in the mantle boundary layer

698   (410-660 km). These warm hydrated upwellings are the cause of volcanism in East Asia (e.g.,

699   Changbai volcano) and may have precipitated the crustal rifting that led to the opening of the Sea

700   of Japan. Presumably, the stagnant slab would generate hydrated plumes across the entire Sea of

701   Japan, but the stretched and thinned continental lithosphere remains intact beneath the central

702   Sea of Japan, as Figure 17 A-A', C-C' demonstrate, impeding upward migration of the hydrated

703   plumes. These plumes would then become part of a westward directed convecting system that

704   would amplify mantle hydration in the western Sea of Japan, near the rift margin. The eastern

705   arm of the Y-shaped anomaly would be caused by slab dehydration at shallower depths, possibly

706   caused by antigorite-related reactions. Thus, the Y-shaped anomaly reflects the details of shallow

707   and deep slab dehydration and convective circulation in the mantle wedge above a stagnant slab.

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709   **6.4 Lithospheric Rejuvenation of the North China Craton**

710   The Chinese part of the Sino-Korean Craton, called the North China Craton (NCC), is

711   composed of two Archean blocks, the eastern and the western NCC, which are separated by the

712   central NCC. The central NCC is a Paleoproterozoic orogen called the Trans-North China

713   Orogen. The three parts of the NCC are separated by dashed red lines in Figure 2. The NCC is

714   believed to have cratonized by about 1.85 Ga, but the eastern NCC has undergone significant

715   tectonothermal rejuvenation from the Ordovician to Cenozoic [*Menzies at al., 1993, 2007;*

716   *Griffin et al., 1998; Gao et al., 2002, 2004; Zhao et al., 2005, 2008*]. These conclusions have

717   been based on basalt-borne xenoliths and geophysical evidence for thin lithosphere (< 100 km)

718   with high heat flow in this region from the early Cenozoic to the present [e.g., *Menzies et al.,*

719   *1993; Griffin et al., 1998; Fan et al., 2000; Xu, 2001; Wu et al., 2005; Chen et al., 2006, 2008,*

720   *2009*]. Most studies agree that the central and western parts of the NCC have evolved differently

721   than the eastern NCC during Phanerozoic time [e.g., *Xu et al., 2004; Xu, 2007*], although some

722   recent seismic studies point to localized lithospheric thinning in the central and even western

723   NCC [e.g., *Huang et al., 2009; Chen et al., 2009*] and the generally high lateral heterogeneity of

724   the lithosphere underlying the NCC. The mechanism of Phanerozoic lithospheric rejuvenation

725   and its variability across the NCC remain intensely debated [e.g., *Chen, 2010* for a recent

726   summary]. Understanding of the Phanerozoic lithospheric tectonics would be improved with

727   clearer images of the current structure of the crust, mantle lithosphere, and asthenosphere across

728   the NCC.

729   To this debate our 3-D model presents new relatively high-resolution images of the crust and

730   uppermost mantle beneath most of the NCC. Crustal velocities beneath the eastern NCC are not

731   particularly distinct from surrounding regions (e.g., Fig. 14a), although the crust may be slightly

732   thinner, on average (Fig. 15a). More significantly, the eastern NCC hosts a "horseshoe" shaped

733   anomaly in map view in the uppermost mantle, which is identified at 80 km depth in Figure 14d

734   and, perhaps more clearly, by the blue contour in Figure 19. The velocities are lowest in the

735   northern part of the North China Platform near the intersection of the North-South Gravity

736   Lineament (NSGL) with the boundary between the central and eastern NCC. The low velocities

737   continue southward, flanking the eastern NCC along the NSGL in the west and the TLF in the

738   east. Low velocities also extend into small parts of the central NCC and the western NCC near

739   the Datong Volcanoes. The south central part of the NCC is not particularly slow in the

740   uppermost mantle as Figure 17 profile E-E' illustrates, which reveals the considerable variability

741   of lithospheric structure even within the eastern NCC.

742   Within the "horseshoe" shaped anomaly bounding three sides of the eastern NCC (Fig. 19),

743   the lithosphere can be seen to be about 60 km thick on vertical profiles B-B' and C-C' of Figure

744   17. The flanks of the eastern NCC are underlain by asthenospheric low velocity anomalies below

745   70 km depth as can be seen on profiles B-B', C-C', and E-E'. Profile B-B' illustrates that the

746   lithosphere is not simply thin; the velocities are lower than beneath the Songliao Basin to the

747   north and the Ordos Block to the west. Thus, we hypothesize that the low velocity anomalies that

748 bound three sides of the eastern NCC identify a region of more intense tectonothermal activity

749 than in south-central part of the eastern NCC. Indeed, lithospheric rejuvenation may have

750 reached nearly to the base of the crust in the most intensely altered zones.

751 **7. Conclusions**

752 This paper presents a high-resolution 3-D shear velocity model of the crust and uppermost

753 mantle with associated uncertainties across the North China Craton, northeastern China, the

754 Korean Peninsula, and the Sea of Japan constructed from the dispersion characteristics of

755 Rayleigh waves obtained from ambient noise cross-correlations. Broadband stations from

756 Chinese Provincial Networks, the Japanese F-Net, and the IRIS GSN have produced two-year

757 continuous time series from 2007 into 2009 that form the basis for these results. Phase and group

758 velocity maps from 8 sec to 45 sec period are first constructed, which constrain earth structures

759 in the crust and uppermost mantle to a depth of about 100 km. The dispersion maps are

760 geologically coherent, displaying the signatures of sedimentary basins, mountain roots,

761 ocean-continent variations, and anomalies due to rifting and volcanism. The maps at different

762 periods and between phase and group velocities are consistent with one another, and vertically

763 relatively smooth Vsv models fit the observations.

764 To generate the dispersion maps presented here, the effects of the persistent localized

765 Kyushsu microseism had to be eliminated in the dispersion measurements. We present a method

766 based on locating the microseism (thereby confirming the location of *Zeng and Ni* [2010]),

767 predicting the arrival of the disturbing signal on the ambient noise cross-correlations, and

768 obtaining the dispersion measurements by using the Green's functions at the opposite

769 cross-correlation lag times for the station pairs in which the apparent speed of the Kyushu signal

770 is less than 4.5km/s.

771 Ambient noise tomography places unique constraints on the structure of the crust and

772 uppermost mantle, providing information complementary to body wave travel times, receiver

773 functions, longer period teleseismic surface waves, gravity, and other types of data. The

774 motivation for the 3-D model is to produce new constraints on the volcanism, crustal extension,

775 cratonic rejuvenation, and lithospheric thinning that are hypothesized for the region of study. The

776 model presented here provides new information about these processes, particularly about the

777 geometry (the geographic distribution and depth extent) of the observed features in the top 100

778 km. A detailed interpretation of the model is inconsistent with the scope of this paper, but we

779 focus discussion on four principal model features: crustal thickness variations, particularly those

780 associated with the North-South Gravity Lineament (NSGL) and beneath the Sea of Japan, the

781 prominent but thin lithosphere underlying the Songliao-Bohai Graben, the "Y-shaped" mantle

782 low velocity anomaly (on vertical profiles) bordering the Sea of Japan, and a "horse-shoe"

783 shaped asthenospheric anomaly (in map view) bounding three sides of the eastern North China

784 Craton.

785 (1) The North-South Gravity Lineament is observed to follow approximately the 40 km

786   crustal thickness contour. Beneath the Sea of Japan, crustal thickness is anti-correlated with

787   water depth, and the relationship is consistent with expectations based on crustal isostasy for a

788   laterally compositionally variable crust. (2) The Songliao-Bohai Graben is underlain by a high

789   velocity lower crust and a fast mantle lithosphere that extends to about 70 km depth. The

790   lithosphere is not attenuated, that is seismic velocities are high, but it is thinner than expected for

791   a region in which parts are traceable back to the Archean. (3) Low velocity anomalies reach

792   upward beneath the eastern and western borders of the Sea of Japan, extending well into

793   continental East Asia on the west side. The arms of this Y-shaped anomaly are separated by the

794   intervening overlying lithosphere, which is ~60 km thick and may represent stretched continental

795   lithosphere. We argue that the Y-shaped structure in the east reflects relatively shallow slab

796   dehydration and in the west deeper dehydration and convective circulation in the mantle wedge

797   overlying the stagnant slab. (4) Very thin, attenuated lithosphere bounds three sides of the eastern

798   North China Craton (NCC), but somewhat thicker faster lithosphere underlies the central and

799   southern parts of the eastern NCC. These low velocity anomalies identify a region of intense

800   tectonothermal modification where lithospheric rejuvenation may have reached nearly to the

801   base of the crust.

802   In the future, assimilating similar results from a companion paper for southern China [*Zhou*

803   *et al*., 2011] will allow the 3-D model to extend across all of eastern China. The geometry of the

804   observing array makes the application of differential travel time methods for earthquakes

805 difficult, such as eikonal tomography [*Lin et al*., 2009], Helmholtz tomography [*Lin and*

806 *Ritzwoller*, 2011], or two-plane wave tomography [e.g., *Yang et al*., 2008b]. However, the

807 success of such methods would allow dispersion maps to be extended to longer periods and,

808 therefore, for the model to reach greater depths in the mantle. In addition to the production of a

809 3-D Vs model of the crust and uppermost mantle, other research envisioned includes the

810 estimation of azimuthal anisotropy, which would place complementary constraints on the

811 physical processes that motive this study. In addition, future understanding of the dispersion

812 characteristics of Love waves would allow estimation of radial anisotropy, which reflects

813 deformation undergone by the crust and uppermost mantle.

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Table 1: Number of measurements retained after each step in data quality control.

Period (s)	Step A	Step B	Step C	Step D	Step E (phase)	Step E(group)
10	20816	14563	14492	13567	11053	10180
14	31356	25770	25740	24082	22606	21593
20	33573	33573	33564	31555	29170	29515
25	28990	28990	28982	27398	25209	25161
30	25495	25495	25491	24210	22096	21623
35	21976	21976	21972	20985	18903	18247
40	18172	18172	18165	17488	15303	14688
45	14503	14503	14495	14031	11810	11352

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Figure Captions

1111 **Figure Captions**

1112 **Figure 1**. The broadband seismic stations used in this study. The blue triangles are the Chinese

1113   provincial broadband seismic stations (CEArray), the red triangles are the F-Net long

1114   period stations and the red squares are other broadband stations, mostly from the IRIS

1115   GSN. The black box defines the study region.

1116 **Figure 2.** The geological setting of this study. Solid red lines mark the boundaries of the

1117   Sino-Korean Craton (SKC), the solid white line identifies the North-South Gravity

1118   Lineament (NSGL), the dashed black line is the Tancheng-Lujiang fault (TLF), and

1119   black lines outline blocks and tectonic boundaries (Zhang, et al., 2003). The North

1120   China Craton is the part of the SKC in China, and is sub-divided by dashed red lines into

1121   western, central, and eastern parts. Feature names are abbreviated as follows: PS:

1122   Phillipine Sea Plate, OK: Okhotsk Plate, NCP: North China Platform, YB: Yangtze

1123   Block, SLB: Songliao Basin, JB: Japan Basin, TB: Yamato Basin, TB: Tsushima Basin,

1124   SB: Sichuan Basin, OR: Ordos Block, BH: Bohai Bay, GXAR: Great Xing'an Range,

1125   LXAR: Lesser Xing'an Range, CBM: Changbai Mountain Range, YSM: Yinshan

1126   Mountain Range, ESYS: East Shandong Yellow Sea Block, NEFB: Northeast Asian

1127   Fold Belt, NSGL: North-South Gravity Lineament, and HB: Huabei Basin. Red dots

1128   with letters mark locations of the dispersion curves and 1D shear velocity profiles shown

1129   in Figs. 11 and 12.

1130   **Figure 3.** Record sections of cross-correlations obtained from two years of waveform data (a)

1131   between Chinese provincial network stations, (b) between F-Net stations, and (c)

1132   between the Chinese stations and F-Net stations. The white dots identify the expected

1133   arrival times for the Kyushu persistent microseism.

1134   **Figure 4**. Process to locate the Kyushu microseism. (a) Top Panel: raw cross-correlation (CC)

1135   between station SSE and INCN (blue triangles in (b)). Bottom Panel: The envelope

1136   function of the CC filtered between 10 and 12 sec period. The Kyushu microseismic

1137   signal arrives at ~130 sec. (b) The migrated hyperbola from the envelope function in (a),

1138   normalized so its maximum value is 1. (c) Paths used to locate the Kyushu microseism

1139   are plotted with black lines. We use all paths associated with stations SSE and INCN

1140   (yellow triangles). (d) Stack of migrated hyperbolas to locate the Kyushu microseism.

1141   Blue circle is the location of *Zeng and Ni* [2010].

1142   **Figure 5**. Cross-correlations and corresponding frequency-time diagrams illustrating the effect of

1143   the persistent Kyushu microseism. (a) Cross-correlation between stations HXQ of

1144   CEArray and STM of F-Net, where the expected arrival time of the signal from the

1145   Kyushu microseism (marked "Kyushu") is close to the arrival time of the surface wave

1146   (marked "Surface") on the negative lag component. (b) Example waveform disturbed by

1147   the Kyushu microseism: symmetric component cross-correlation in which the positive

1148   and negative components in (a) are averaged. (c) Example waveform undisturbed by the

1149   Kyushu microseism: positive component of the cross-correlation in (a).   (d) & (e)

1150   Frequency-time (group velocity) diagrams from (b) and (c), respectively. (f) Comparison

1151   between the group velocity curves of cross-correlations shown in (b) and (c). Red

1152   triangles and stars are the phase and the group velocity curves from disturbed signal.

1153   Blue symbols show the group and phase velocity curves of the undisturbed signal.

1154   **Figure 6**. Resolution estimates at periods of 16 and 35 sec, presented in units of km and defined

1155   as twice the standard deviation of a 2-D Gaussian fit to the resolution surface at each

1156   geographic node [*Barmin et al., 2001*]. The black box shows the region of study.

1157   **Figure 7.** Misfit histograms at periods 14, 20, 30, and 40 sec, respectively. The top figures show

1158   the group velocity misfit and the lower figures are the phase velocity misfits. The

1159   standard deviations are shown on each panel.

1160   **Figure 8**. Group velocity maps at periods of 12, 20, 30, and 40 sec. Group velocities are plotted

1161   as perturbations relative to the average velocity (U_0) within the 400 km resolution

1162   contour and the thick white line is the 100 km resolution contour. Other lines show

1163   tectonic and geological boundaries identified in Fig. 2.

1164   **Figure 9**. Same as Fig. 8, but for phase velocity maps at periods of 12, 20, 30 and 40 sec.

1165   **Figure 10**. Rayleigh wave phase (solid lines) and group (dashed lines) velocity sensitivity

1166   kernels for 10 sec (red lines) and 40 sec (black lines) period. (a) Kernels are based on

1167   our 3D model in the Sea of Japan (point H, Fig. 2). (b) Kernels are based on our 3D

1168   model in the North China Craton (point B, Fig. 2). The horizontal black lines denote

1169   local Moho depth at each point.

1170 **Figure 11.** Phase and group velocity observations at the eight locations identified with letters in

1171   Fig. 2, presented with error bars intended to be 1σ. The dispersion curves are predicted

1172   by the center of the ensemble of models shown in Fig. 12 for each location, with red

1173   lined being phase velocity and blue lines group velocity. The RMS misfits are calculated

1174   by equation (1).

1175 **Figure 12**. Shear wave (V_{sv}) velocity profiles at the eight selected locations identified with

1176   letters in Fig. 2 constructed to fit the corresponding dispersion curves shown in Fig. 11.

1177   The gray corridor is the 2σ (standard deviation) uncertainty and the dashed lines indicate

1178   Moho depth.

1179 **Figure 13**. Comparison between inversions performed in the Sea of Japan (point H, Fig. 2) using

1180   data from Fig. 11h with two starting models identified with blue lines: (a) Moho at 8 km

1181   and (b) Moho at 20 km. Horizontal lines mark the median Moho depths. The resulting

1182   ensembles of models are quite similar, with differences within model uncertainties.

1183 **Figure 14**. Four horizontal depth slices through the 3-D Vsv model: (a) in the lower crust at a

1184   depth of 80% of Moho depth, (b) 40 km, (c) 60 km, and (d) 80 km depth. Color bars

1185   beneath each panel give the range of shear wave speeds in km/s.

1186   **Figure 15**. (a) Estimated crustal thickness and (b) 1σ uncertainty in crustal thickness, both in km.

1187   The lines A-A', B-B', etc. in (a) are the location of the vertical profiles shown in Fig. 17.

1188   (c) Shear velocity jump across Moho (km/s) and (d) 1σ uncertainty in the shear velocity

1189   jump across Moho, also in km/s.

1190   **Figure 16**. Uncertainty (1σ) in Vsv at the four depths shown in Fig. 14. Uncertainties are in

1191   km/s.

1192   **Figure 17**. Four vertical profiles through the 3-D model along the lines shown in Fig. 15a.

1193   Crustal velocities are presented using an absolute scale and mantle velocities are defined

1194   relative to 4.35 km/s.

1195   **Figure 18**. (a) Blow-up of part of Fig. 15a, showing crustal thickness across the Sea of Japan (in

1196   km). Contours of bathymetry are over-plotted in increments of 1 km to illustrate the

1197   anti-correlation between crustal thickness and seafloor depth. (b) Plot of seafloor depth

1198   versus solid crustal thickness (crustal thickness minus ocean depth) with values on a

1199   0.5°×0.5° grid across the Sea of Japan shown as blue dots. The solid red line is the

1200   least-squares fit line in which seafloor depth is the dependent variable and the dashed

1201   red line is the fit in which crustal thickness is the dependent variable.

1202   **Figure 19.** Simplified representation of the 3D model in the upper mantle, showing the

1203   distribution of low velocities at 80 km depth with velocity intervals: red: $V_{sv} \leq 4.25$

1204   km/sec, blue: $4.25 < V_{sv} \leq 4.35$ km/s. The red contour identifies the "Y-shaped"

1205   anomaly that underlies the Sea of Japan and parts of northeastern Asia believed to arise

1206   from slab dehydration at several depths overlying the stagnant slab. The blue region

1207   surrounding the red contour is the large area of the uppermost mantle affected by

1208   back-arc spreading. The blue contour within the North China Craton identifies the

1209   "horseshoe" shaped anomaly that flanks the eastern North China Craton, believed to

1210   result from intense tectonothermal modification that may have reached as high as the

1211   base of the crust.
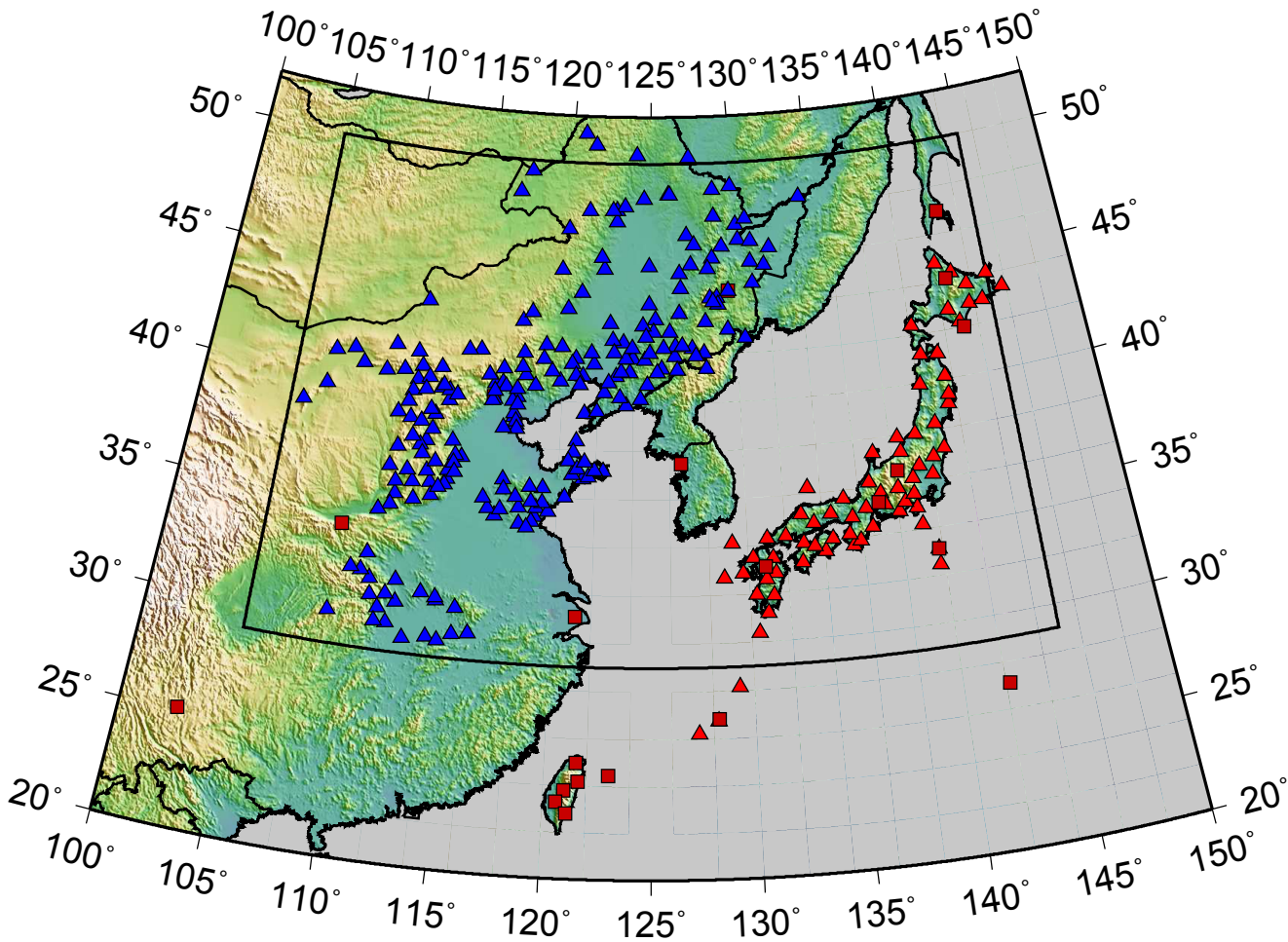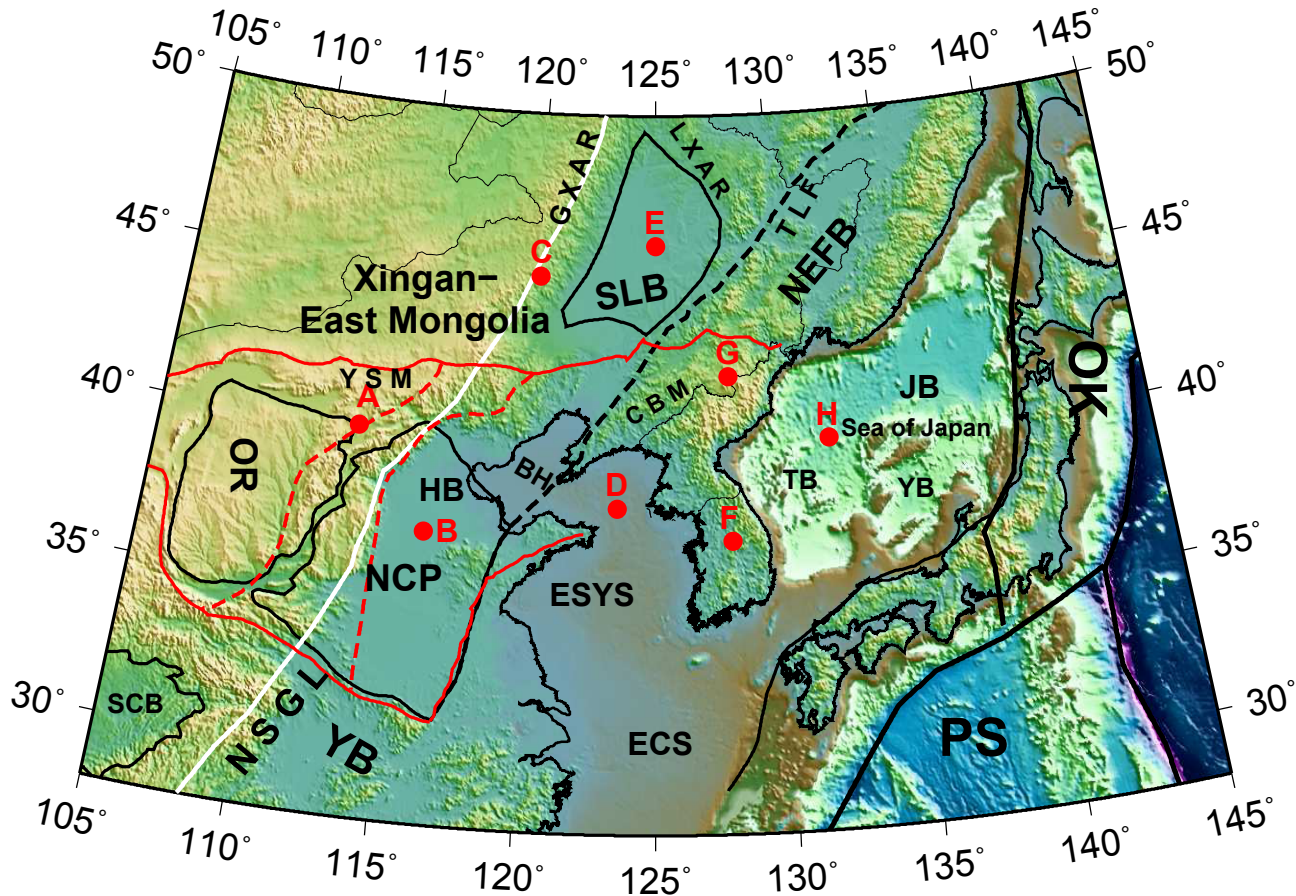
(a) NEH-CH   (b) WJM-FN   (c) WJM-CH
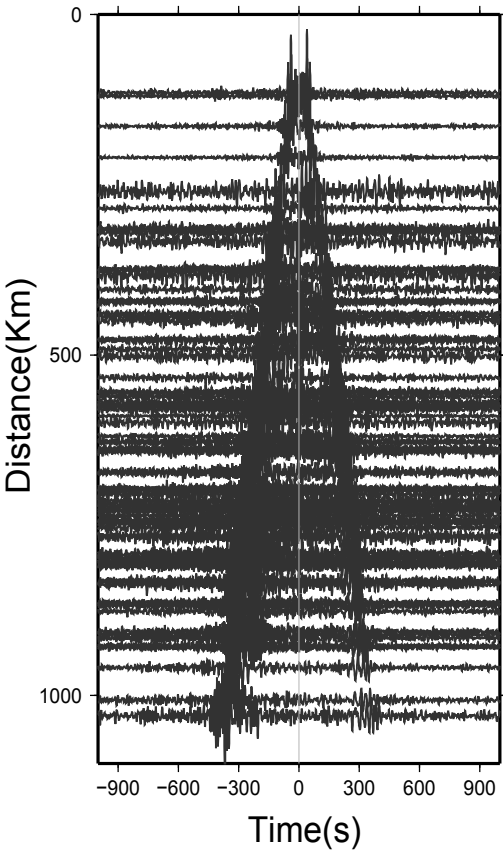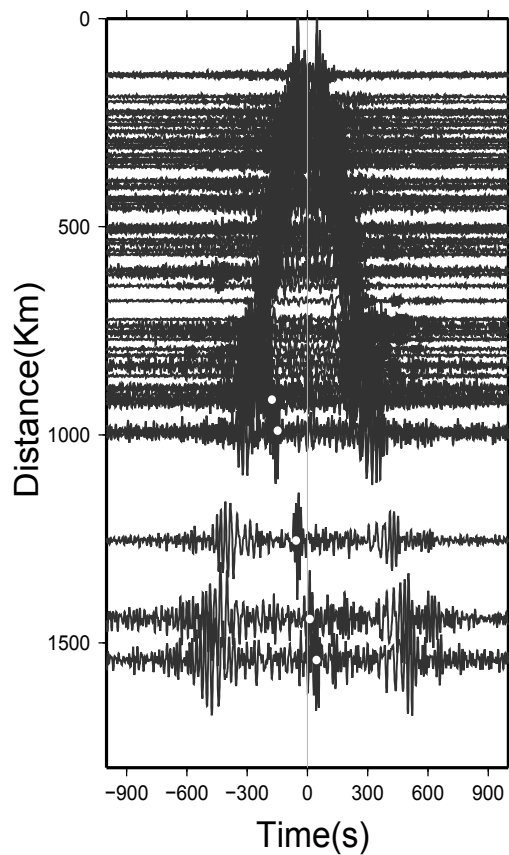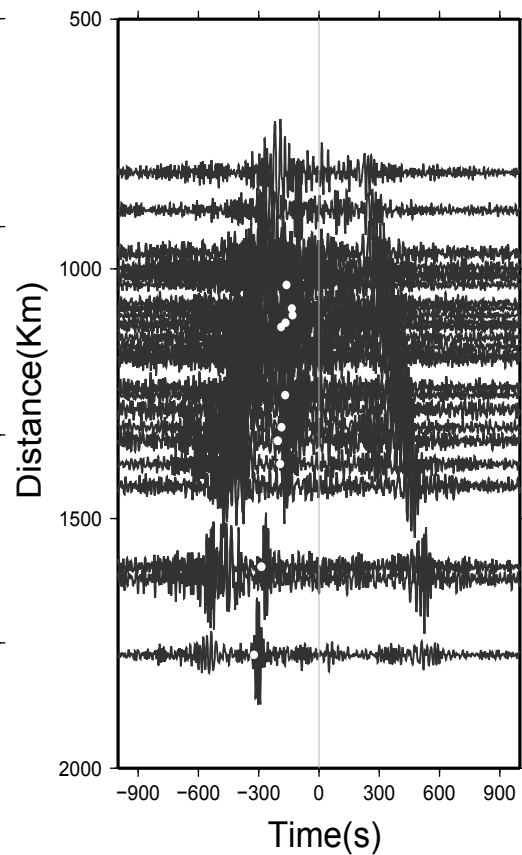
(a) (b) INCN−SSE
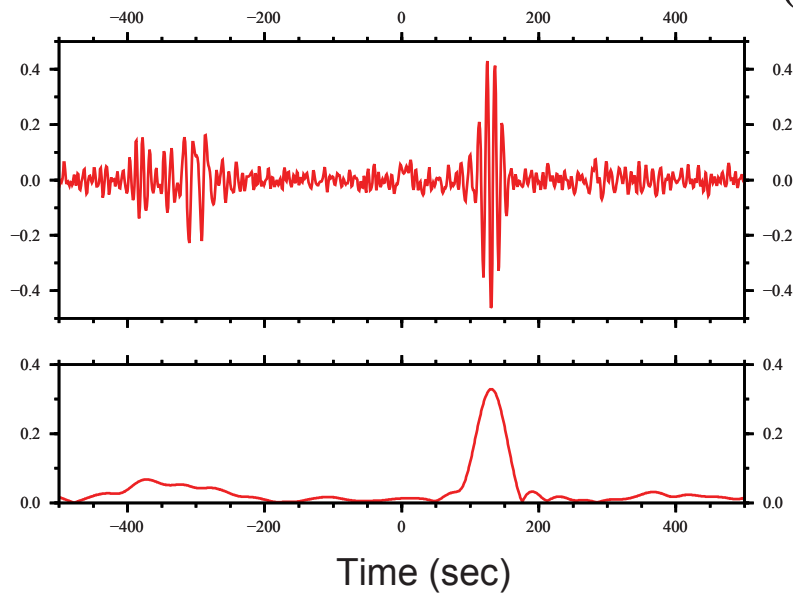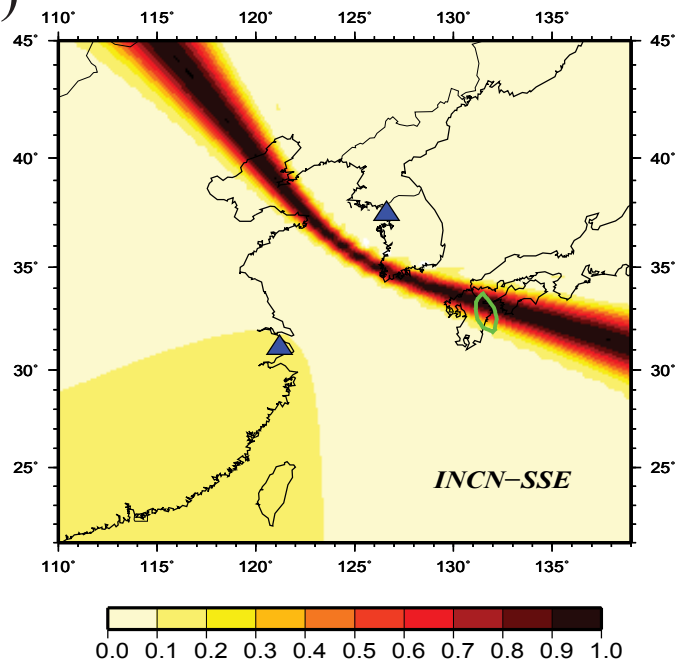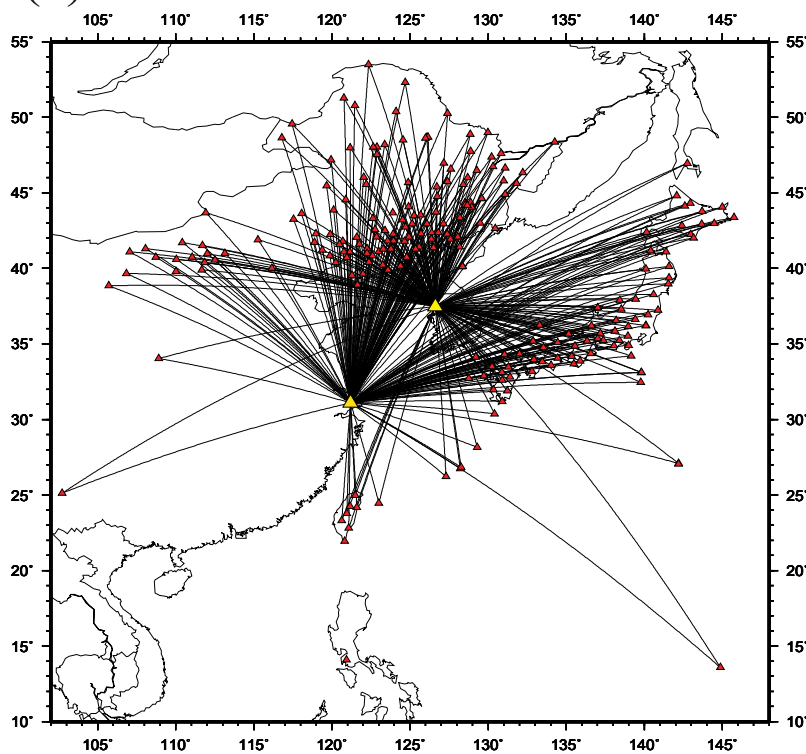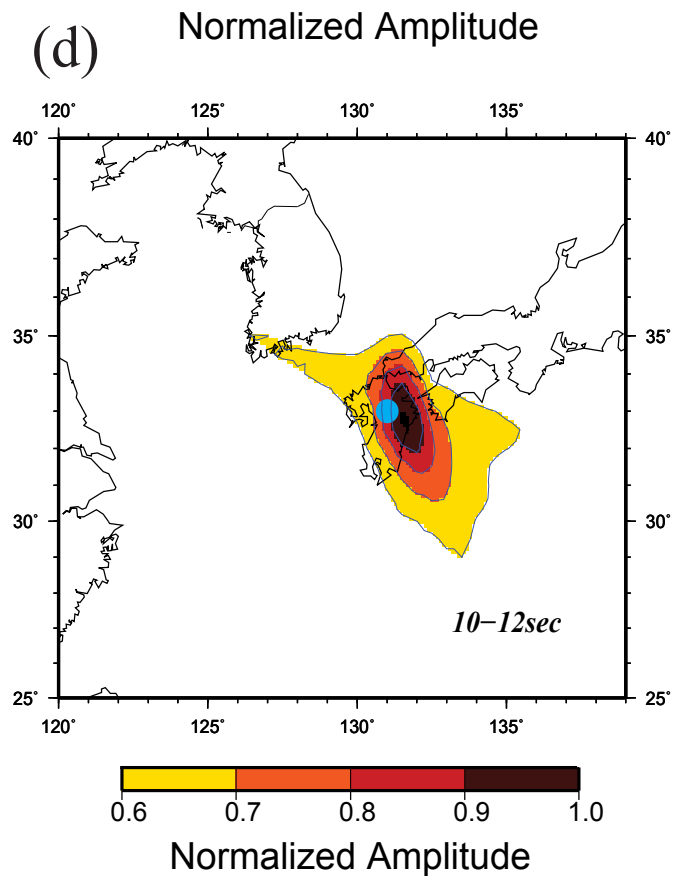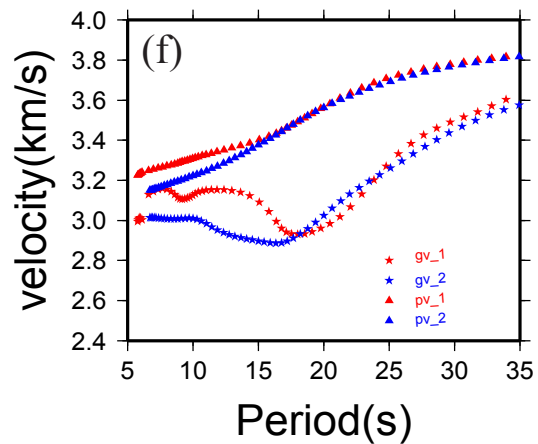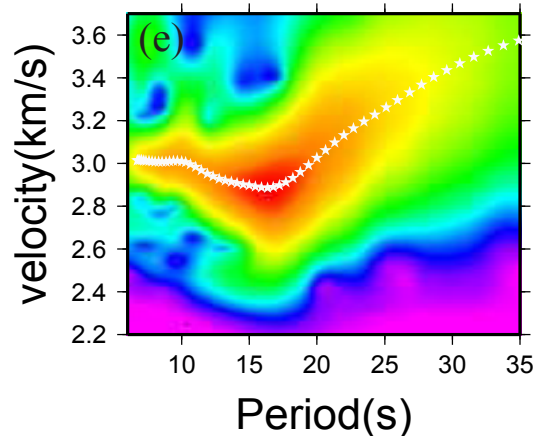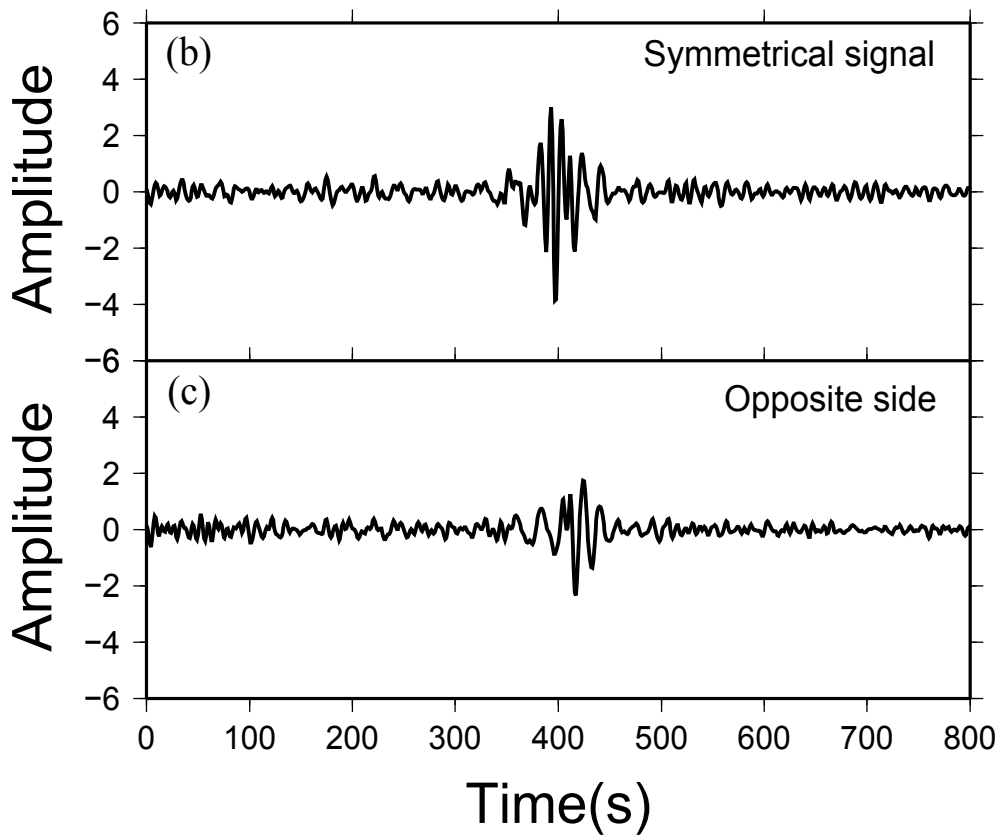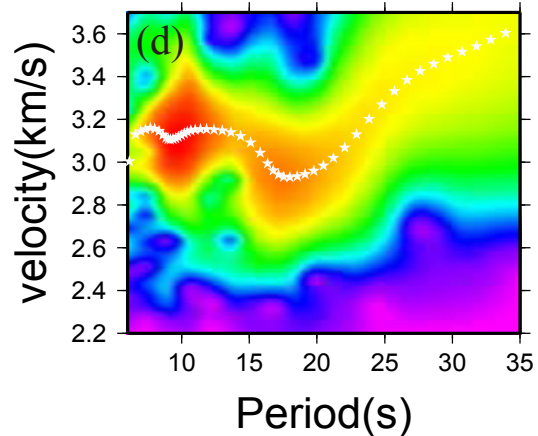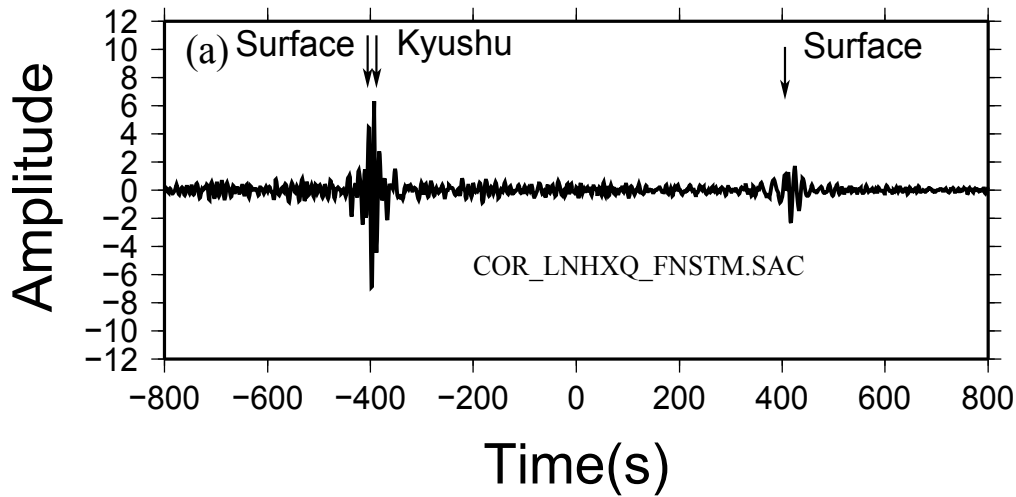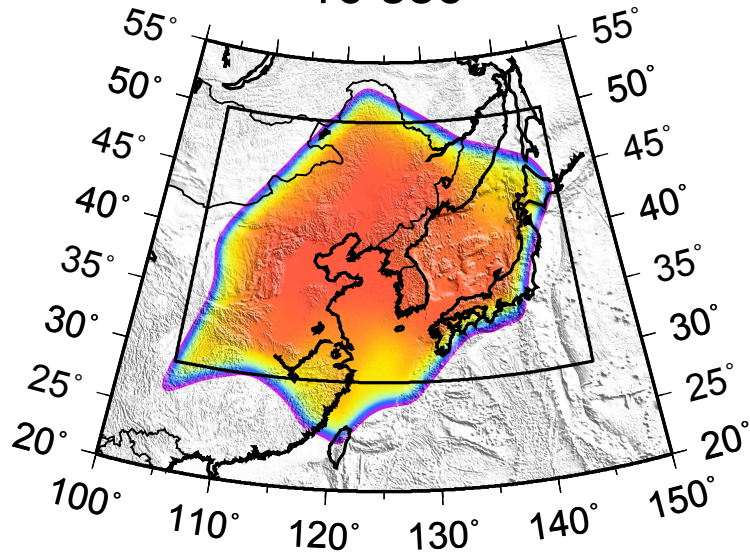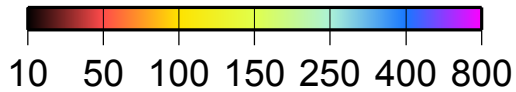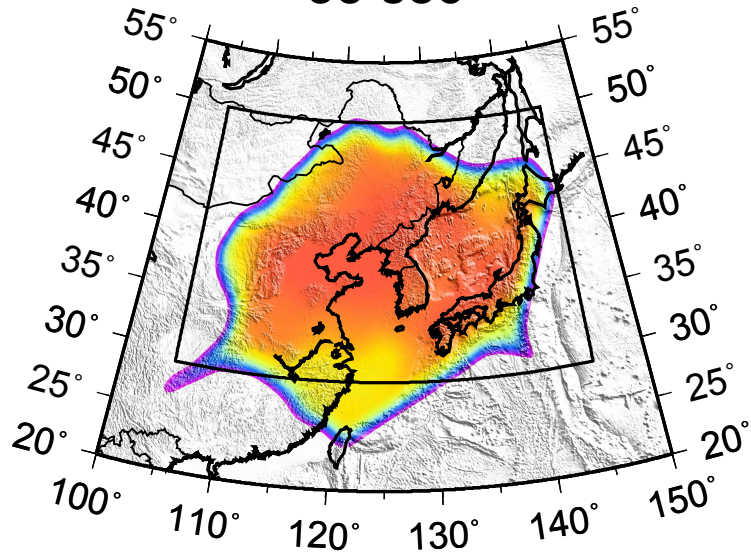
Normalized Amplitude
0.0 0.1 0.2 0.3 0.4 0.5 0.6 0.7 0.8 0.9 1.0

(c) (d) 10−12sec

Normalized Amplitude
0.6 0.7 0.8 0.9 1.0
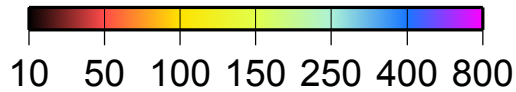
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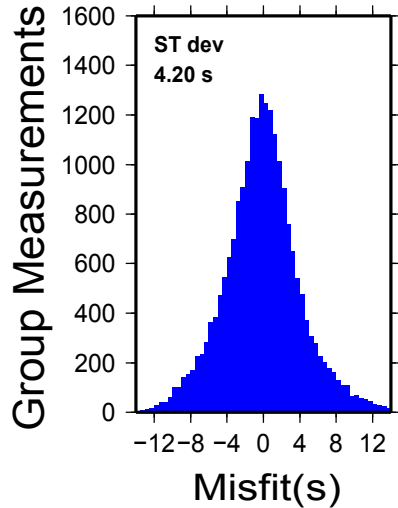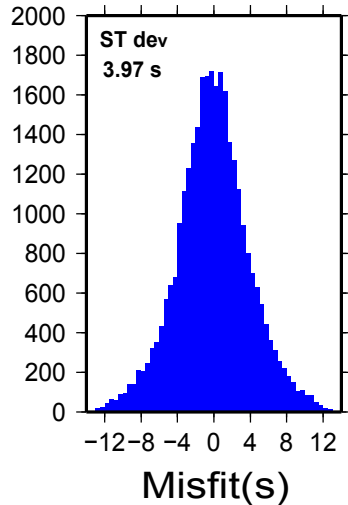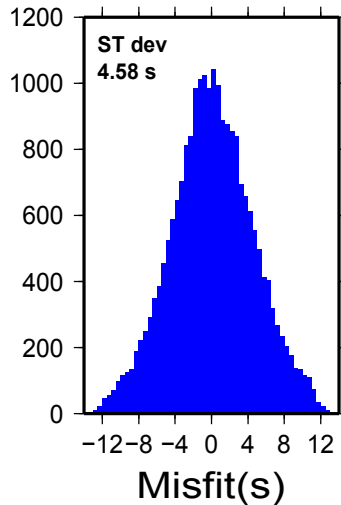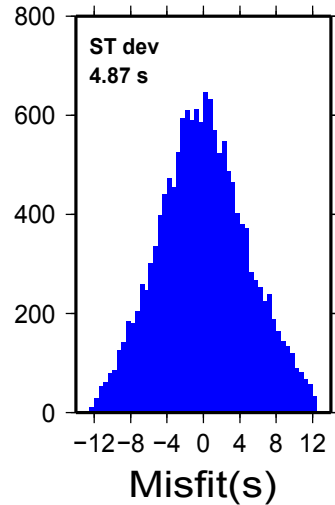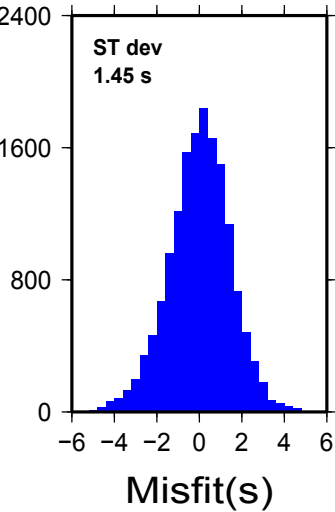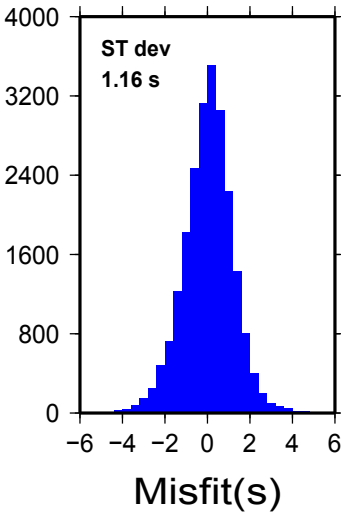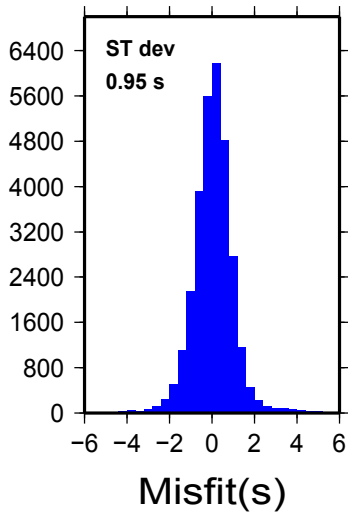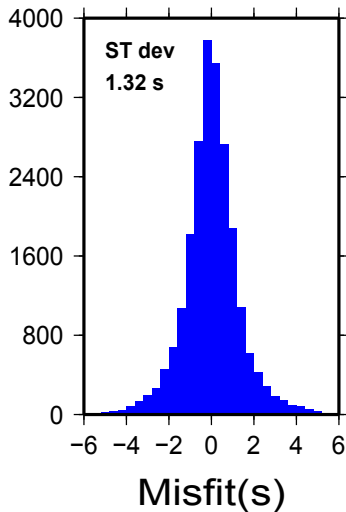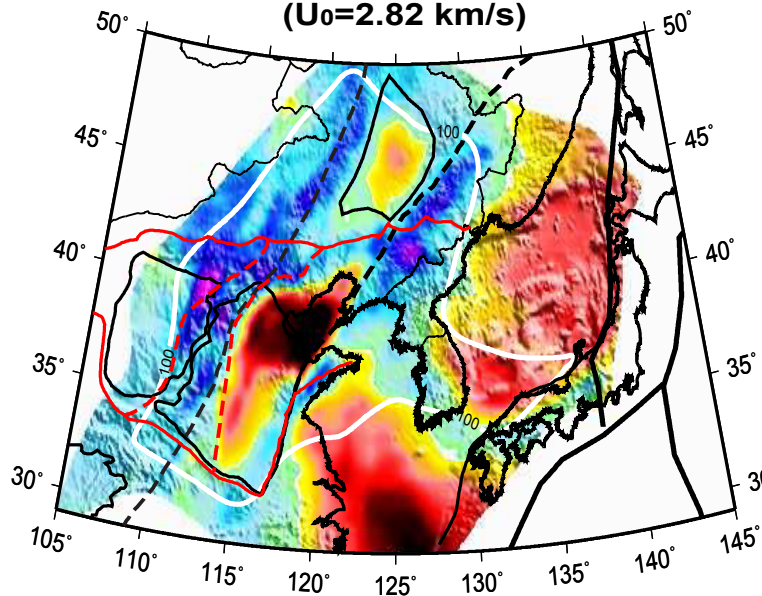
12 Sec
(U₀=2.82 km/s)

20 Sec
(U₀=3.00 km/s)
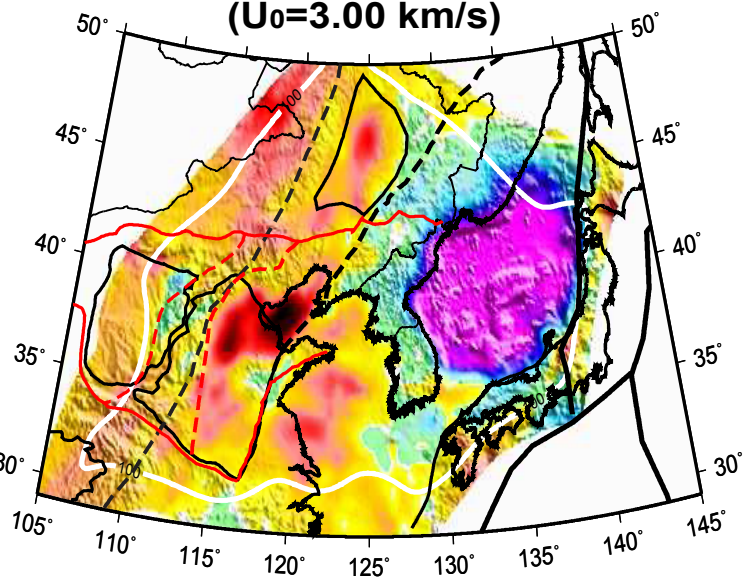
30 Sec
(U₀=3.44 km/s)

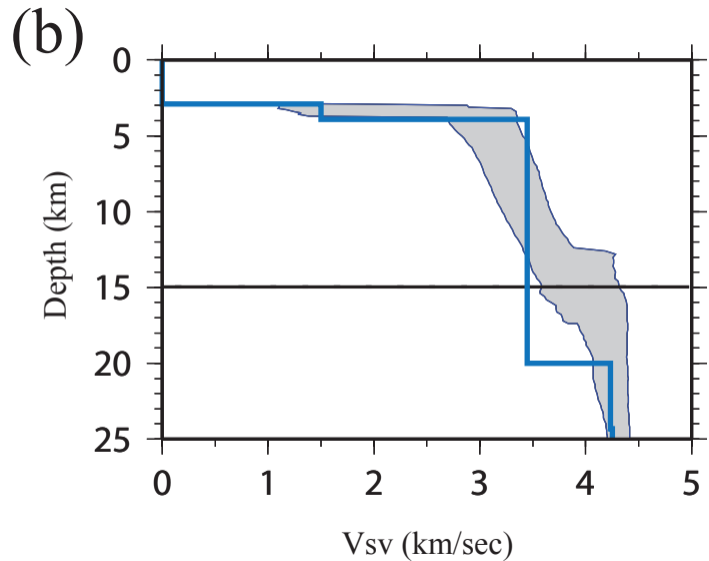**40 Sec**
(U₀=3.65 km/s)

−13.0 −10.0 −8.0 −5.0 −3.0 −1.0 −0.5 −0.2 0.2 0.5 1.0 3.0 5.0 8.0 10.0 13.0
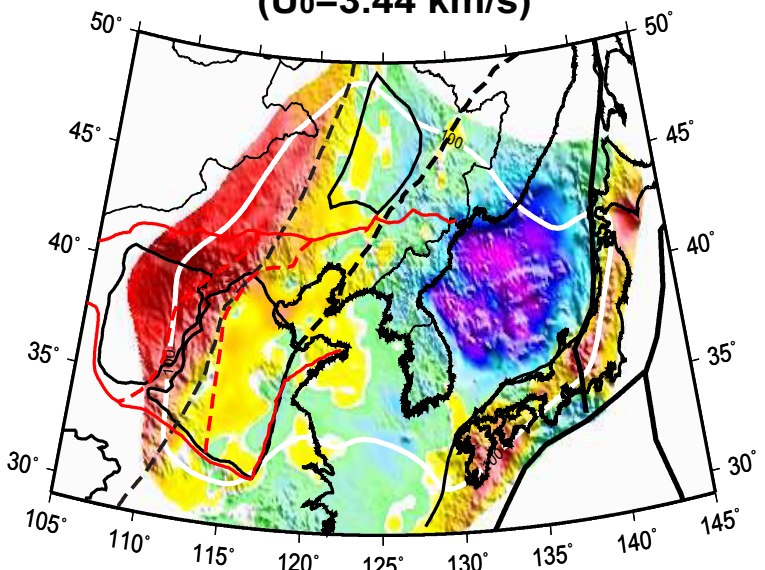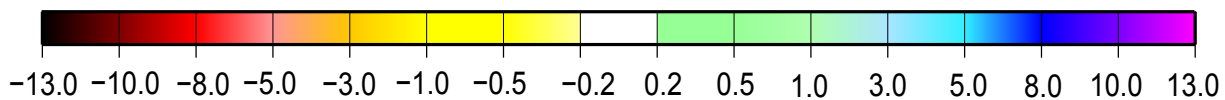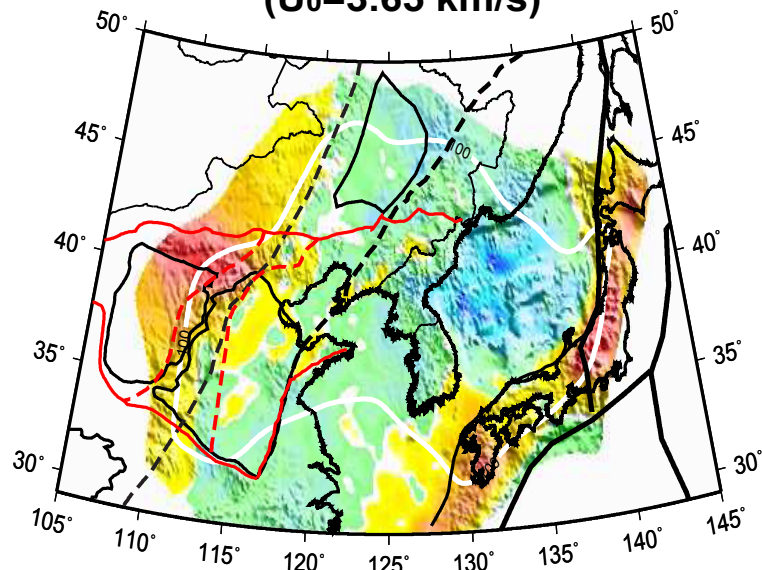
Group Velocity Perturbation (%)

12 Sec
(C₀=3.24 km/s)
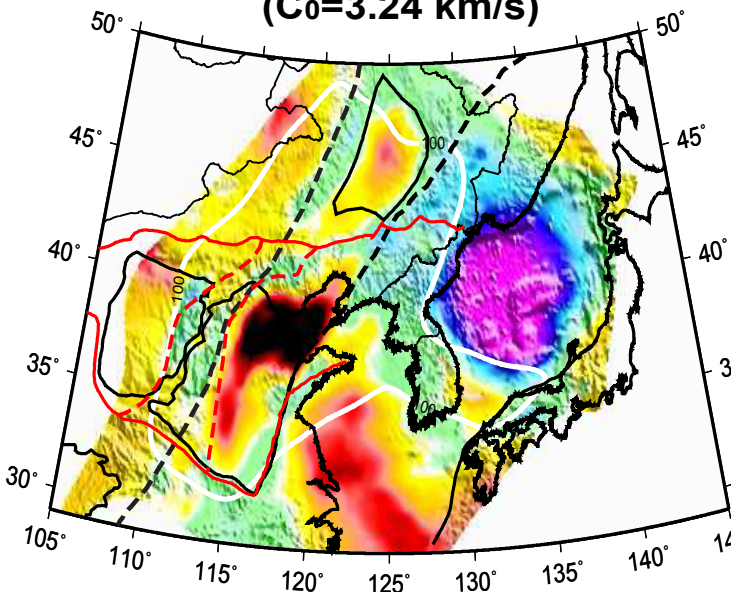
20 Sec
(C₀=3.54 km/s)
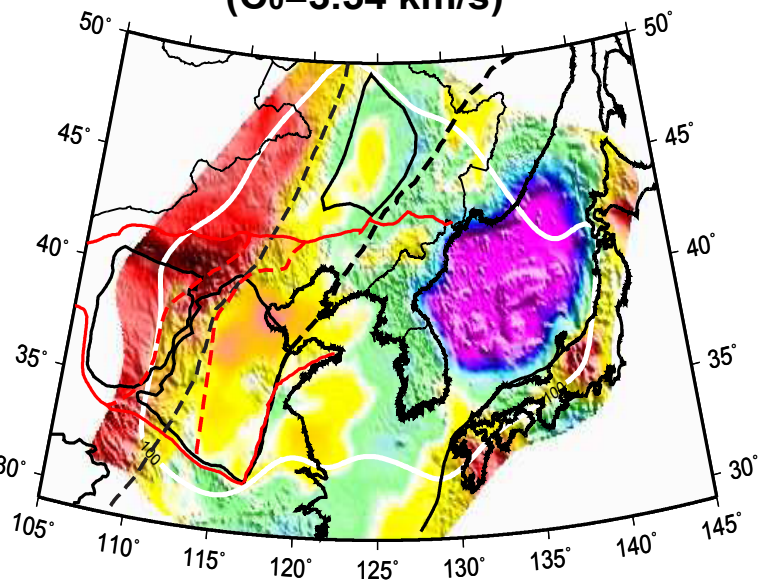
30 Sec
(C₀=3.76 km/s)
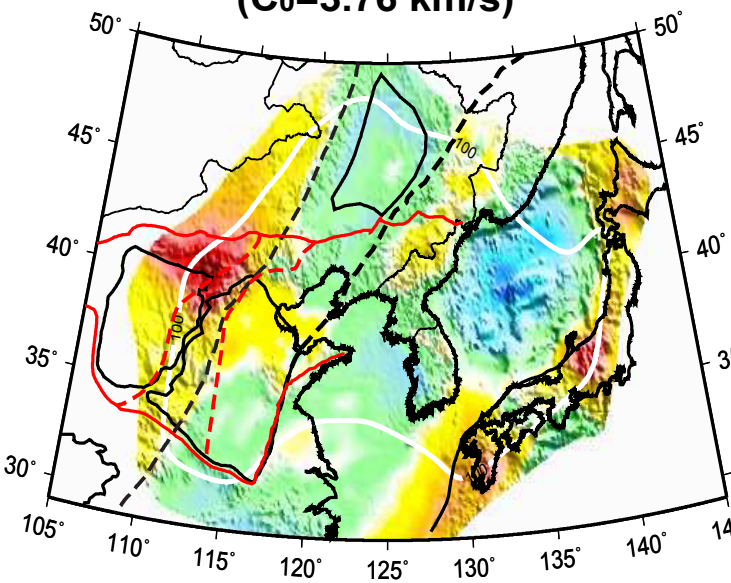
40 Sec
(C₀=3.84 km/s)
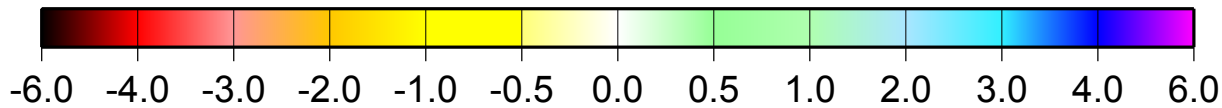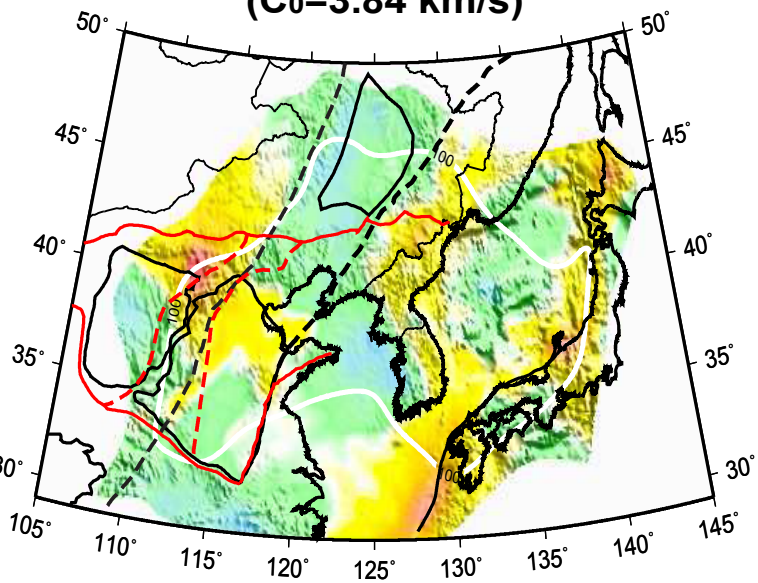
Phase Velocity Perturbation (%)

(a) Yinshan Mtns
RMS misfit = 0.363

(b) N. China Platform
RMS misfit = 0.481

(c) Great Xing'an Range
RMS misfit = 0.363

(d) Yellow Sea
RMS misfit = 0.316

(e) Songliao Basin
RMS misfit = 0.524

(f) Korean Peninsula
RMS misfit = 0.469

(g) Changbai Mtns
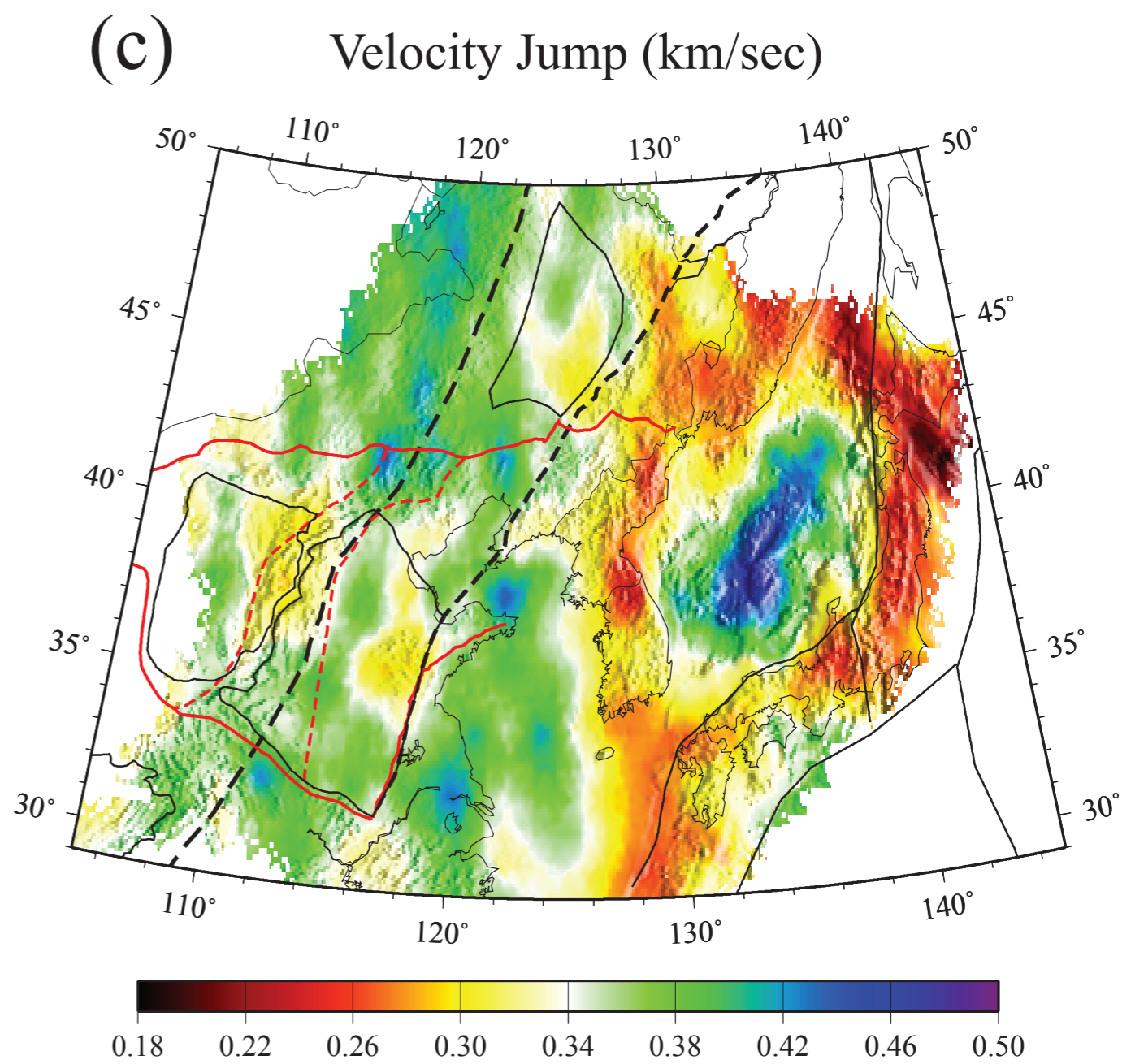RMS misfit = 0.613

(h) Sea of Japan
RMS misfit = 0.472

(a) Lower Crust

(b) Depth = 40 km
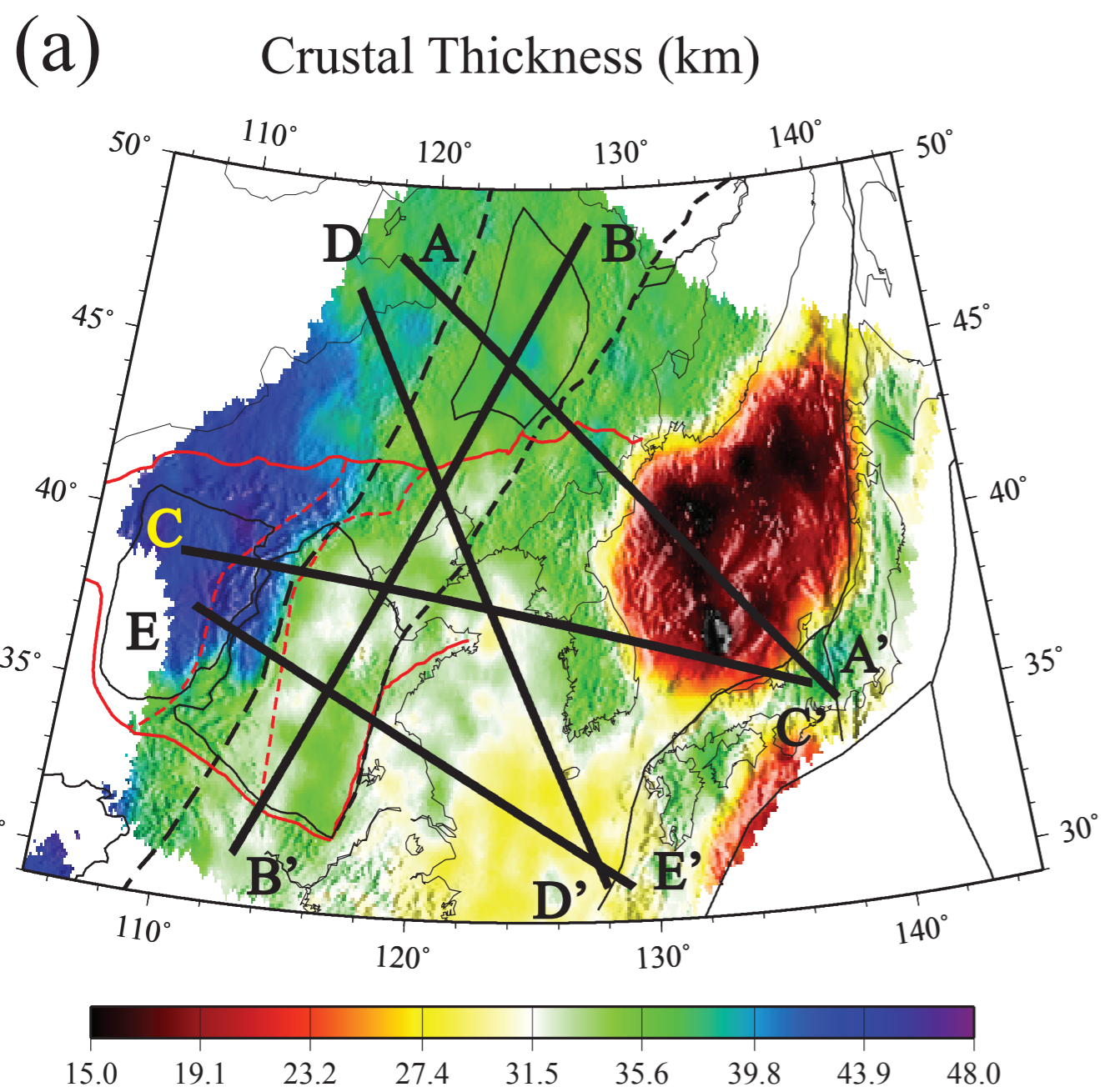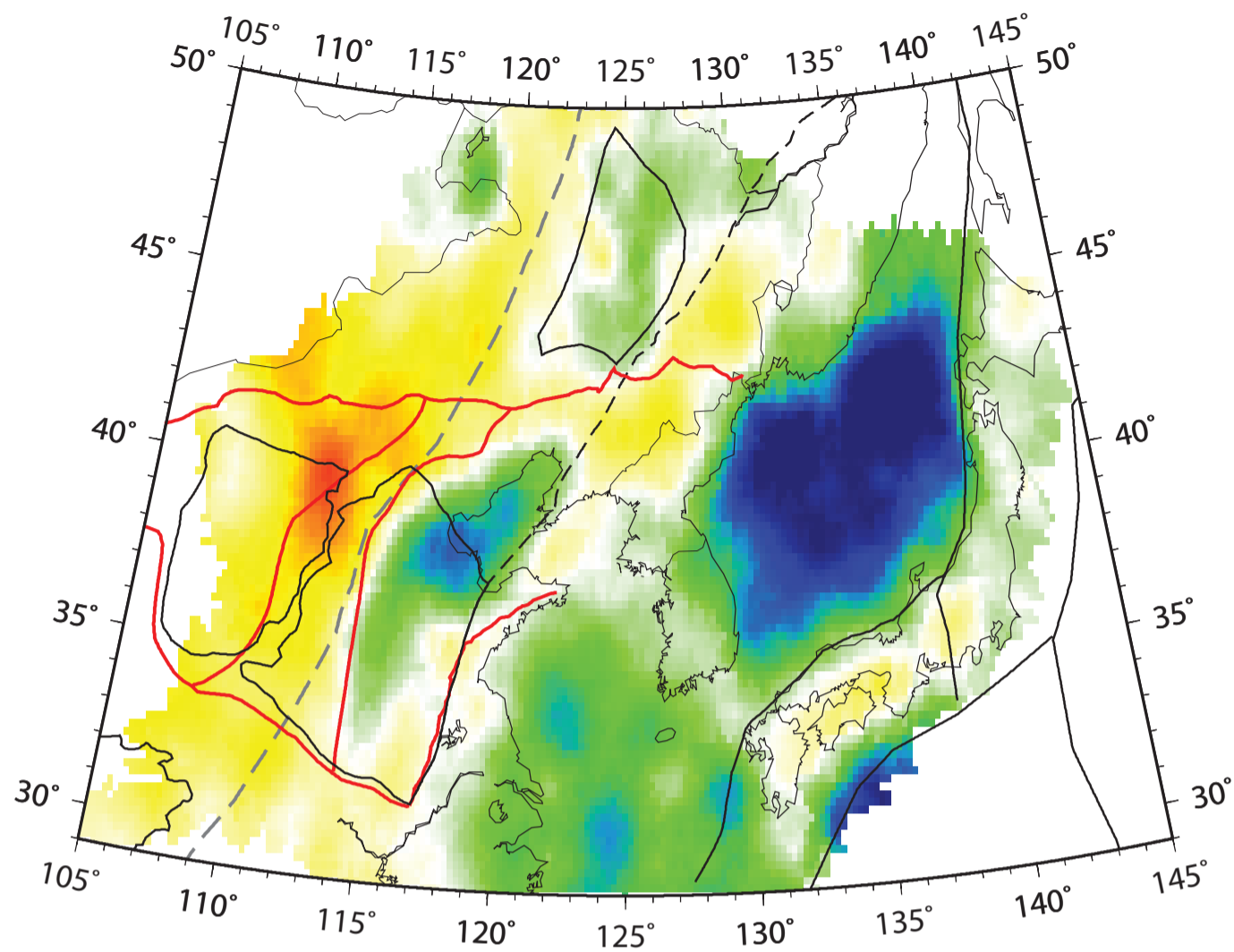
(c) Depth = 60 km

(d) Depth = 80 km

(a) Crustal Thickness (km)

15.0 19.1 23.2 27.4 31.5 35.6 39.8 43.9 48.0

(b) Uncertainties (km)

1.50 1.94 2.38 2.81 3.25 3.69 4.12 4.56 5.00

(c) Velocity Jump (km/sec)

0.18 0.22 0.26 0.30 0.34 0.38 0.42 0.46 0.50

(d) Uncertainties (km/sec)

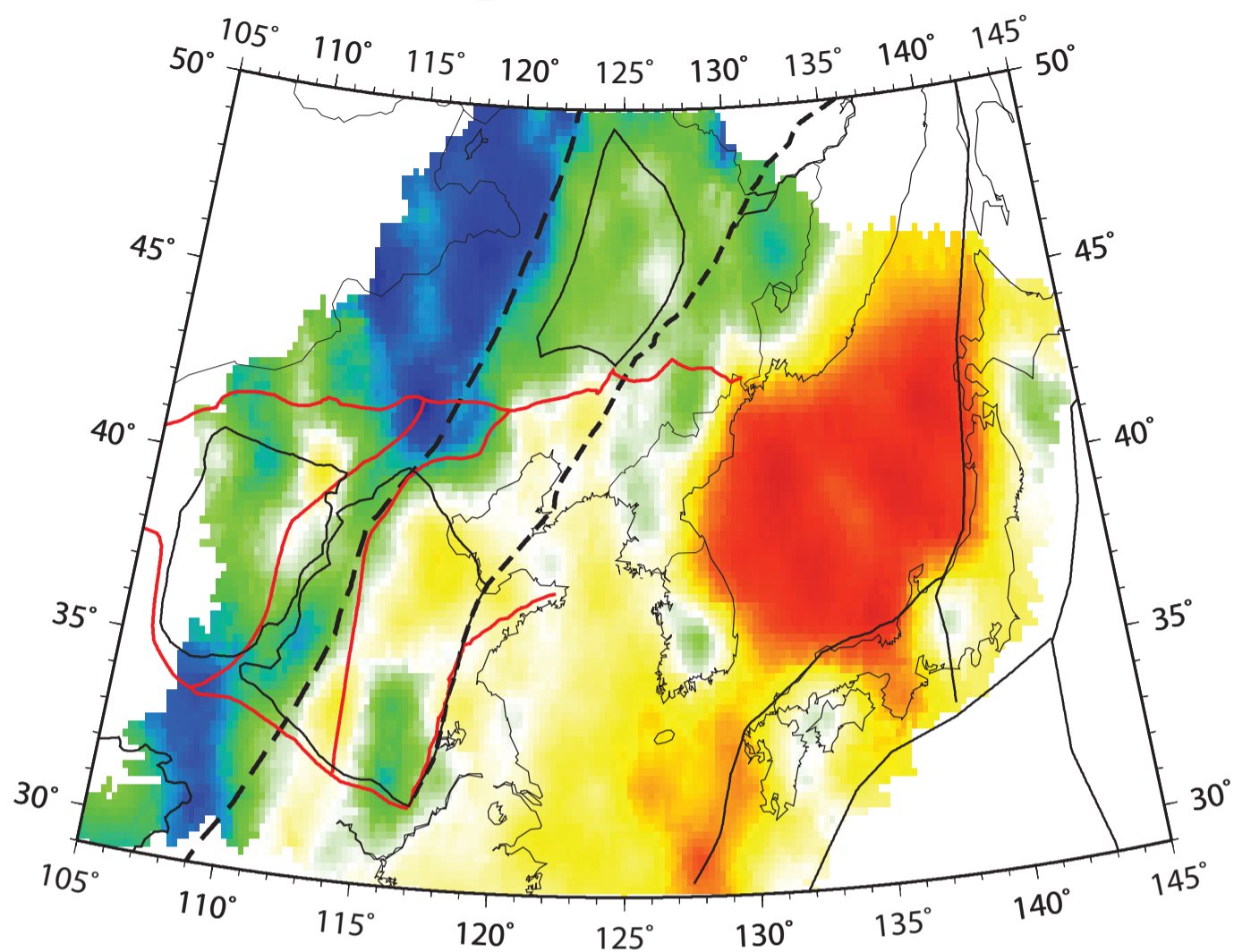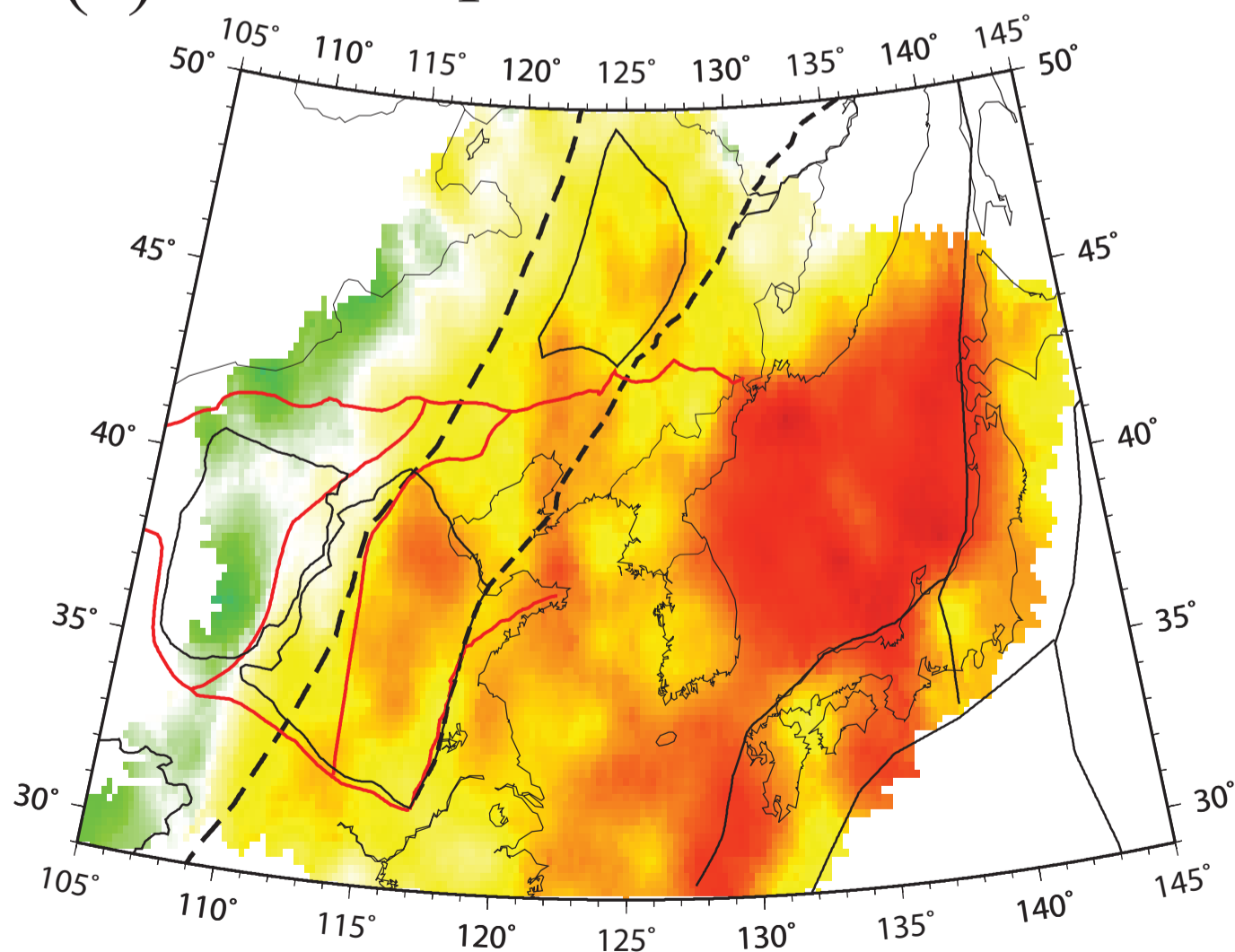0.050 0.075 0.100 0.125 0.150 0.175 0.200 0.225 0.250

(a) Lower Crust

(b) Depth = 40km

(c) Depth = 60km

(d) Depth = 80km

0.000 0.025 0.050 0.075 0.100 0.125 0.150 0.175 0.200

(a)

(b)

Bathymetry (km)

Solid Crustal Thickness (km)