1 Layered anisotropy within the crust and lithospheric 2 mantle beneath the Sea of Japan

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10 Abstract

Continental rifting during the Oligocene to mid-Miocene caused the opening of the Sea of Japan and the separation between the Japanese Islands and the Eurasian Plate. The tectonic evolution in the Sea of Japan is important for understanding the evolution of back-arc regions in active convergent margins. Here, we use data from the seismic stations surrounding the Sea of Japan to map the Rayleigh-wave azimuthal anisotropy in the crust and lithospheric mantle beneath the Sea of Japan. We explore the variations of Rayleigh-wave phase-velocity beneath the Sea of Japan in a broad period range (30-80 s). Rayleigh-wave dispersion curves are measured by the two-station technique for a total of 231 interstation paths using vertical-component broad-band waveforms at 22 seismic stations around the Sea of Japan from 1,411 global earthquakes. The resulting maps of Rayleigh-wave phase velocity and azimuthal anisotropy allow the examination of azimuthal anisotropy at specific periods. They exhibit several regions with different isotropic and anisotropic patterns: the Japan Basin displays fast velocities at shorter periods (30 and

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Preprint submitted to *Journal of Asian Earth Sciences*

June 28, 2016

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40 s) with NNE-SSW anisotropy, whereas at 60 s and longer, the velocities become slow even if the anisotropy remains NE-SW; the East China Sea shows fast velocities at all periods (30 to 80 s) with constant **NW-SE anisotropy**. Trench-normal anisotropy beneath the Japanese Islands is found at short periods (30-40 s) and become trench-parallel at periods of 60 s and longer. Overall, our model resolves two layers of anisotropy, the shallowest and deepest layers being potentially related to frozen deformation due to recent geodynamic events, and asthenospheric flow, respectively.

11 *Key words:* Seismic tomography, Surface waves and free oscillations, Wave
12 propagation, Sea of Japan.

13 **1. Introduction**

14 The Sea of Japan is a marginal sea of the western Pacific Ocean. It
15 comprises three major basins (Figure 1): the Japan Basin in the north,
16 the Yamato Basin in the southeast, and the Tsushima Basin (later referred
17 as Ulleung Basin, Yoon et al. (2014)) in the southwest. The thickness of
18 the crust beneath the Sea of Japan changes drastically: the Moho depth
19 is about 22 km beneath the Yamato Rise (Kurashimo et al., 1996), 15 km
20 beneath the Ulleung Basin (Kim et al., 1994), and only 9 km beneath the
21 Japan Basin (Sato et al., 2004). There is a good correlation between the
22 Moho depth and the sea-floor topography, which is widely influenced by the
23 tectonic history of back-arc spreading that formed the Sea of Japan from 32
24 to 10 Ma (Tamaki, 1992; Pouclet et al., 1994; Choi et al., 2013; Zahirovic
25 et al., 2013, 2014; Yoon et al., 2014). Magnetic studies (Kobayashi and
26 Isezaki, 1976) and GPS measurements (Takahashi et al., 1999) in this region

27 constrained the regional flow, the location of the transform faulting as well
28 as the spreading centers. In addition, IODP experiments (Tamaki, 1992) and
29 geological investigations (Lallemant and Jolivet, 1986; Takahashi et al., 1999;
30 Yoon et al., 2014) helped constrain the ages of the structures, which were
31 later used to unravel the tectonic evolution of this region (Tapponnier and
32 Molnar, 1976; Fryer, 1996; Yin, 2010). The spreading and dynamic setting
33 of the marginal basins in the western Pacific (Sea of Japan, East China
34 Sea, South China Sea, Philippine Sea, as well as those in southern Pacific)
35 have also been discussed (Miyashiro, 1986; Ren and Li, 2000), as well as the
36 tectonic features (He and Tsukuda, 2003).

37 **The updated plate boundary model of Bird (2003) indicates that**
38 **the Sea of Japan and its surroundings are complex regions, com-**
39 **prising four small plates (Figure 1). The Okhotsk plate, associated**
40 **with the North America plate, includes the northern half of Hon-**
41 **shu, Hokkaido, and Shakhlin and is bounded to the east by the**
42 **Japan subduction zone. The Amur Plate covers most of the Japan**
43 **Basin and Korea, and extends northwards. South of Kyushu, the**
44 **Okinawa plate is bounded to the east by the Ryukyu Arc and cov-**
45 **ers the eastern part of the East China Sea. Finally, the Yangtze**
46 **plate extends west of the Okinawa plate and Taiwan. Both the**
47 **Okinawa and Yangtze plates are associated with the Eurasia plate.**

48 The 3D seismic structure of the Japanese Islands has been investigated
49 extensively. First results by Aki and Kaminuma (1963) and Aki (1968)
50 have been updated later by numerous studies (Hirahara and Mikumo, 1980;
51 Kurashimo et al., 1996; Honda and Nakanishi, 2003; Sato et al., 2004; Mat-

subara et al., 2005, 2008; Hirose et al., 2008; Bourova et al., 2010; Yoshizawa et al., 2010; Liu et al., 2013; Asamori and Zhao, 2015; Liu and Zhao, 2016). By comparison, the Sea of Japan is less well understood, despite several geological investigations (Lallemand and Jolivet, 1986; Takahashi et al., 1999; Yoon et al., 2014), IODP experiments (Tamaki, 1992; Tobin and Kinoshita, 2006), magnetotelluric studies (Kobayashi and Isezaki, 1976; Pouclet et al., 1994), seismic experiments (Ludwig et al., 1975; Hirata et al., 1989; Kim et al., 1994; Kurashimo et al., 1996; Choi et al., 2012, 2013), noise tomography (Spindel et al., 2003; Zheng et al., 2011), seismic attenuation tomography (Liu and Zhao, 2015), and surface-wave analyses (Abe and Kanamori, 1971; Evans et al., 1978; Sato et al., 2004; Zhao et al., 2011a).

In oceanic regions, where the distribution of seismic stations is uneven and sparse, surface waves play an important role to recover the seismic structure of the lithosphere. Surface-wave analyses can therefore be considered as the best option to resolve the seismic structures beneath the Sea of Japan. Evans et al. (1978) used group speeds of Love and Rayleigh waves to re-examine the original work of Abe and Kanamori (1970), and obtained an average 1D structure along paths north of the Sea of Japan. They showed a lithospheric thickness of about 40 km under the Sea of Japan, which is thinner than typical oceanic lithosphere, and suggested that a simple isotropic model is sufficient to explain the observed dispersion curves of both Love and Rayleigh waves. Aki and Kaminuma (1963) revealed the existence of radial anisotropy in the upper mantle beneath the Japanese Islands from the observation of Love and Rayleigh waves. Aki (1968) then proposed a laminated mantle model to explain the discrepancy between the observed dispersion curves of

77 the Love waves and those estimated from an isotropic shear-wave model.

78 In the past decade, studies based on acoustic-wave tomography for mon-
79 itoring the Sea of Japan (Spindel et al., 2003; Lee et al., 2015), body-wave
80 tomography (Hirahara and Mikumo, 1980; Hirata et al., 1989; Zhu et al.,
81 2002; Honda and Nakanishi, 2003; Matsubara et al., 2005, 2008; Hirose et al.,
82 2008; Zheng et al., 2011), surface wave tomography (Friederich, 2003; Huang
83 et al., 2003; Lebedev and Nolet, 2003; Huang and Zhao, 2006; Huang et al.,
84 2009; Legendre et al., 2015b), and other seismic experiments (Choi et al.,
85 2012) provided additional details on the complexity of the structure beneath
86 the Sea of Japan. Recent studies suggested that the formation of the Sea
87 of Japan is related to back-arc extension whereas its evolution is linked to
88 the deep subduction of the Pacific slab as well as the mantle flow induced by
89 the slab subduction (Tatsumi et al., 1990; Lei and Zhao, 2005; Zhao et al.,
90 2009, 2011b; Lei et al., 2011, 2013; Wu et al., 2016). Anisotropic analyses
91 (Montagner and Tanimoto, 1991; Trampert and Woodhouse, 2003) give fur-
92 ther constraints on the evolution of the lithosphere beneath the Sea of Japan.
93 The anisotropic pattern at long periods has also been compared with the
94 Absolute Plate Motion of the region (DeMets et al., 1990, 1994).

95 Azimuthal anisotropy can map the deformation at depth, and may be
96 related to past and present geodynamical events. Surface waves allow for re-
97 solving the azimuthal distribution of anisotropic structure (if any) at depth,
98 and may thus be used to reconstruct deformation history of the lithosphere in
99 the study region. Here, we build maps of isotropic and anisotropic anomalies
100 of the fundamental mode of Rayleigh-wave phase-velocity using teleseismic
101 data from seismic stations around the Sea of Japan. We observe two indepen-

102 dent anisotropic layers; the shallower one being associated with the frozen
103 sea-floor spreading anisotropy in the subducting Pacific slab, and the deeper
104 one with current asthenospheric flow.

105 **2. Seismic data and phase-velocity measurements**

106 To measure the Rayleigh-wave phase-velocities beneath the Sea of Japan,
107 we use the waveforms recorded at 22 broadband seismic stations from lo-
108 cal and international networks between January 2007 and December 2011.
109 These include 8 stations from the China National Seismic Network (CNSN)
110 (Zheng et al., 2010), one station from the Incorporated Research Institutions
111 for Seismology (IRIS), 10 stations from the National Research Institute for
112 Earth Science and Disaster Prevention in Japan (NIED), and 3 stations from
113 the Korea Meteorological Administration (KMA). Despite a lack of seismic
114 stations in the northern part of the Sea of Japan, the stations we used in
115 this study are almost evenly distributed around our study region (Figure 2).
116 However, the lack of seismic stations deployed in the Sakhalin Oblast and in
117 the easternmost part of Russia (Figure 1) limits the coverage in the north-
118 eastern part of our study region.

119 The two-station method, which we use here to measure dispersion curves
120 of Rayleigh-wave phase velocity, was first introduced by Sato (1955) and has
121 been used to measure dispersion curves of surface waves (Knopoff, 1972). In
122 this study, we adopt the implementation of the cross-correlation approach of
123 Meier et al. (2004) to the two-station method and the inversion scheme of
124 Lebedev et al. (2006) to build anisotropic maps.

125 The two-station method requires a small angle between the great circle

126 connecting the given two stations and the circle connecting this specific event
127 to the station pair. The station pair and the event are then considered aligned
128 along the same great circle. In this study, we set an upper limit of 10° for
129 this angle. Another condition is that the epicentral distance between the
130 earthquake and the station pair is greater than the distance between the two
131 stations (the epicentral distance must be at least three times larger than the
132 distance between the two stations). This ensures that the difference between
133 the recorded waveforms at the two stations mainly results from the structure
134 between the two stations. In our data set, epicentral distances are between
135 10° and 170° , and all interstation distances are in the range of 100-2000 km.

136 Following these criteria, we extracted 26,312 vertical-component sismo-
137 grams from 1,411 selected events (Figure 3). For each interstation pair, all
138 phase-velocity measurements obtained from earthquakes are averaged to de-
139 rive the dispersion curve. With 22 stations, the maximum number of station
140 pairs is 231. However, due to differences in the recording times of the se-
141 lected seismic stations, only 198 could be used. In addition, only high-quality
142 measurements were kept, further decreasing to 168 the total number of in-
143 terstation paths.

144 The phase-velocity measurement can be divided into two steps: in the
145 first step, we measure the phase-velocity using broadband regional waveform
146 records to determine the individual interstation dispersion curve for a specific
147 path connecting two stations and for a specific event. For each selected event,
148 the vertical-component displacements recorded at the two stations in each
149 pair are cross-correlated (Figure 4).

150 The cross-correlation function is first filtered with a frequency-dependent

151 Gaussian band-pass filter (Meier et al., 2004) to minimize the effects of noise
 152 and interferences. Side lobes caused by interaction of the fundamental mode
 153 with scattered waves and higher modes are down-weighted by the applica-
 154 tion in the time domain of a frequency-dependent Gaussian window to the
 155 filtered cross-correlation function. A special effort was made on the quality of
 156 the measurements to avoid scattered waves, multi-pathing or off-great-circle
 157 propagation which bias our measurements (Legendre et al., 2015c). This
 158 approach to filtering and windowing is effective as long as the fundamental
 159 mode has the largest amplitude in the seismogram and therefore dominates
 160 the contribution to the cross-correlation function (Lebedev et al., 2006; En-
 161 drun et al., 2011; Legendre et al., 2012), which is the case of our dataset.
 162 The cross-correlation is then transferred to the frequency domain (Figure 5),
 163 and its complex phase is used to calculate the Rayleigh-wave phase-velocity
 164 $C(\omega)$, as detailed in previous studies (Meier et al., 2004; Lebedev et al., 2006;
 165 Deschamps et al., 2008; Endrun et al., 2011; Legendre et al., 2014a,b, 2015a).

166 In the second step, for each selected station pair, all measurements from
 167 available earthquakes are averaged, resulting in a single path-specific disper-
 168 sion curve (Figure 6). Averaged dispersion curves that are constrained by
 169 at least 5 single event-specific dispersion curves are kept for further analy-
 170 ses, and the rest are discarded. We repeated this procedure for each station
 171 pair, and obtained a collection of 168 phase velocity dispersion curves for the
 172 fundamental mode Rayleigh wave (Figure 7).

173 As the interstation distance is relatively large (on average, around 1,500 km),
 174 measurements at short periods (30 s and shorter) are less robust. Because
 175 here we focus on the structure of the upper lithospheric mantle beneath the

176 Sea of Japan, which is sampled by periods in the range 20-100 s, this limi-
 177 tation alters only very slightly our results. Our data set allowed measuring
 178 dispersion curves between 5 and 300 s, which we then invert for maps of
 179 Raleigh-wave velocity anomalies at selected periods between 30 and 80 s, for
 180 which we have a good regional and azimuthal coverage (Figure 8). Accord-
 181 ing to depth-sensitivity kernels (Figure 9), this range of period samples the
 182 whole crust and the lithospheric mantle.

183 3. Inversion for phase-velocity maps

184 To recover maps of Rayleigh-wave velocity anomalies, we then invert our
 185 collection of dispersion curves for both isotropic and azimuthally anisotropic
 186 (2ψ and 4ψ) Rayleigh-wave phase-velocity maps at selected periods. At
 187 each point of the model, the total velocity anomaly can be parameterized
 188 with 5 coefficients: one for the isotropic phase-velocity variation, δC_{iso} , 2 for
 189 the 2ψ -variation, $A_{2\psi}$, and $B_{2\psi}$, and 2 for the 4ψ -variation, $A_{4\psi}$, and $B_{4\psi}$
 190 (Deschamps et al., 2008; Endrun et al., 2011):

$$\delta C = \delta C_{iso} + A_{2\psi} \cos(2\psi) + B_{2\psi} \sin(2\psi) + A_{4\psi} \cos(4\psi) + B_{4\psi} \sin(4\psi), \quad (1)$$

191 where ψ is the local azimuth of the ray. The amplitudes of azimuthal velocity
 192 variation (Λ) and the directions of fast propagation (θ) of the 2ψ - and 4ψ -
 193 anisotropy are then given by:

$$\begin{cases} \Lambda_{2\psi} = \sqrt{A_{2\psi}^2 + B_{2\psi}^2} \\ \theta_{2\psi} = \frac{1}{2} \arctan(\frac{B_{2\psi}}{A_{2\psi}}) \end{cases} \quad and \quad \begin{cases} \Lambda_{4\psi} = \sqrt{A_{4\psi}^2 + B_{4\psi}^2} \\ \theta_{4\psi} = \frac{1}{4} \arctan(\frac{B_{4\psi}}{A_{4\psi}}) \end{cases} . \quad (2)$$

194 Note that in Eq. 2, the calculation of the fast directions for 2ψ and 4ψ
 195 anisotropy ($\theta_{2\psi}$ and $\theta_{4\psi}$, respectively), takes into account the π -periodicity
 196 of the arctangent function.

197 The model is parameterized on a triangular grid of knots (Wang and
 198 Dahlen, 1995) with a grid spacing of 50 km. Each dispersion curve yields the
 199 average phase-velocity along the path linking the two stations as a function
 200 of period. The total average phase-velocity anomaly along the i^{th} path may
 201 be written as the integral of local anomalies at each grid knot sampled by
 202 the given path,

$$\overline{\delta C_i} = \int_{\varphi} \int_{\theta} K_i(\varphi, \theta) \delta C(\varphi, \theta) d\theta d\varphi, \quad (3)$$

203 where the local anomalies $\delta C(\varphi, \theta)$ are weighted by the sensitivity kernels
 204 $K_i(\varphi, \theta)$. The kernels provide the contribution at each knot on a specific
 205 path to the total phase-velocity anomaly (Lebedev and Van Der Hilst, 2008).
 206 The phase-velocities obtained from the dispersion curves are averaged for
 207 each period, and the inversion is performed to obtain the phase-velocity
 208 perturbations relative to the regional average model.

209 Previous studies (Montagner and Tanimoto, 1990; Trampert and Wood-
 210 house, 2003) pointed out that the amplitude of 4ψ anisotropy is comparable
 211 to that of the 2ψ anisotropy, and should thus be taken into account in the
 212 inversion (Eq. 3).

213 In our models, we also found that the amplitude of the 2ψ anisotropy (de-
 214 fined in Eq. 2, displayed in Figure 10) is larger than that of the 4ψ term by a
 215 factor of 2, consistent with previous studies using the same techniques (De-
 216 schamps et al., 2008; Legendre et al., 2014a,b, 2015a). In our inversions, we

217 consider the phase velocity variation to be in the form of Eq. 1, i.e. isotropic
 218 part, 2ψ part and 4ψ part. The inversion results show that these three parts
 219 have different amplitudes, with the isotropic part being dominant compared
 220 to the anisotropic contributions, and the 2ψ anisotropy being slightly stronger
 221 than the 4ψ anisotropy (Figure 10). We consider all three parts in the in-
 222 version because 2ψ and 4ψ parts can reduce the residuals. It is important to
 223 note that, including 4ψ anisotropy or not doesn't change the isotropic and
 224 2ψ anisotropy models. However, unlike 2ψ anisotropy, which can be related
 225 to Lattice Preferred Orientation (LPO) and/or Shape Preferred Orientation
 226 (SPO) and to regional deformation, there is no clear mechanism to which 4ψ
 227 anisotropy can be linked. Therefore, we do not interpret 4ψ result because it
 228 has smaller amplitude than 2ψ part (see discussion on the reduced χ^2 below).

229 To control the quality of our inversions, we performed a series of tests and
 230 corrections. First, we detect and reject the dispersion curves that showed ex-
 231 cessive deviations from the average (Legendre et al., 2014a). For this purpose,
 232 we calculated the average phase velocity at each period using all the disper-
 233 sion curves in Figure 7 and its first and second derivatives. Then, using
 234 a simple least-squares misfit function, we rejected a certain amount of our
 235 dataset to remove possible errors and inconsistent data, which might affect
 236 our final model. Note that we applied this procedure independently for each
 237 period. The results of these tests are displayed in Figure 11 and show that
 238 the dataset is very consistent. With only 40% of the dataset (Figure 11-c),
 239 the general pattern of the inversions does not change much compared to other
 240 models with less data rejection (Figure 11-a-b). For our preferred model, we
 241 selected a rejection threshold of 30%. Removing more than 30% of the less

242 consistent dispersion curves did not affect much the resulting model (except
243 the anisotropic pattern). Furthermore, similar structures to those obtained
244 with higher rejection rates may be obtained by increasing the smoothing of
245 the isotropic term.

246 We then investigated the effect of regularization parameters. Here, we
247 discuss the influence of norm damping and smoothing on the properties of
248 the resulting model. Resolution is mainly controlled by the cumulative sen-
249 sitivity of the available data to isotropic and anisotropic perturbations of
250 the medium, but also by additional *a priori* constraints such as norm damp-
251 ing and smoothing. The norm damping is used to constrain the amplitudes
252 of phase-velocity perturbations during the inversion. Therefore, investigat-
253 ing the influence of this parameter allows us to calibrate our inversion. As
254 shown in Figure 11(e-h), the isotropic damping does not affect the lateral dis-
255 tribution of the anomalies or the anisotropic part. Increasing norm damping
256 decreases the amplitude of the velocity anomaly. Therefore the parameter
257 has been set to ensure a good agreement between the resulting model and the
258 initial dispersion curves, from which the average velocity at selected periods
259 is known and the amplitudes of velocity perturbations can be estimated. Sim-
260 ilar tests have been performed to estimate the amount of damping necessary
261 for the anisotropic part of the model (Figure 11(i-l)).

262 Lateral smoothing penalizes the difference between the anomaly at a
263 model knot and the average of the anomalies at its neighboring knots by
264 minimizing the first and second spatial derivatives of velocity anomalies. The
265 lateral smoothing weight is set up independently for each period. We apply
266 a stronger smoothing with decreasing path coverage. Figure 11(m-p) shows

267 the effect of smoothing on the resulting velocity model at a specific period
 268 of 40 s. If the smoothing weight is too low (Figure 11-m), small-scale arti-
 269 facts may appear in the regions with poor data coverage (i.e. at the borders
 270 of the model). If the smoothing weight is instead too high, information on
 271 small-scale features is lost (Figure 11-p). Therefore, we imposed a smoothing
 272 weight large enough in order to avoid artifacts yet still allowing the resolu-
 273 tion of regional and local anomalies. Figure 11(m-p) further shows that the
 274 anisotropic pattern remains mostly unchanged, indicating the robustness of
 275 the inversion.

276 Similar tests have been conducted for the smoothing of the anisotropic pa-
 277 rameters (Figure 11(q-t)). During all those tests on isotropic and anisotropic
 278 smoothing and damping, we selected all parameters independently. The re-
 279 sulting smoothness of isotropic, 2ψ and 4ψ anisotropies are checked carefully,
 280 to ensure there is limited leaking from anisotropy to isotropic. The resulting
 281 L-curves are displayed Figure 12. This figure highlight the fact that there is
 282 a low trade-off between the isotropic and anisotropic terms during the inver-
 283 sion process, which is already suggested by the resolution tests on smoothing
 284 and damping parameters.

285 When the anisotropic smoothing factor increases, information on the local
 286 changes of fast direction of the anisotropy vanishes, and only the average
 287 direction of the anisotropy in the region (for a selected period) is visible.
 288 We then test the robustness of the structures imaged by our inversion, by
 289 performing a series of resolution tests (Figure 13). For each test, an input
 290 test model is first created with representative patterns for both isotropic
 291 and anisotropic anomalies (Figure 13, (a-c)). We then generate synthetic

292 data from these input models, and invert these synthetic data under the
 293 same conditions as the real data (Figure 13, (d-f)). In particular, the same
 294 amount of smoothing and damping, and the same path coverage are used for
 295 a given period. This allows testing for the reconstruction of both isotropic
 296 anomalies and anisotropic patterns. The retrieval of the isotropic anomalies
 297 should display correct amplitudes and locations. The anisotropic features
 298 should be retrieved with consistent directions and amplitudes. Note that
 299 the resolution tests presented here addressed the coverage by the data only
 300 and no additional noise has been added. The main sources of errors in the
 301 actual inversions are those in the dispersion curves and diffraction effects,
 302 which are not accounted for by the approximate sensitivity kernels $K_i(\varphi, \theta)$
 303 in Eq. 3 that we used. Given the very substantial redundancy of our coverage
 304 provided by the hundreds of crossing paths, random errors are unlikely to
 305 have any systematic impact on the results (Legendre et al., 2012). Inversion
 306 result shows only small deviation between input and output models both
 307 in the azimuth of fast direction and in amplitude. In addition, the output
 308 model did not show substantial isotropic perturbations.

309 Finally, to test the ability of the anisotropy to explain the raw data,
 310 we performed a set of inversions in which anisotropic terms are neglected.
 311 Damping and smoothing are similar to those of the preferred model. Because
 312 the number of parameters is not the same among the inversions, comparing
 313 the variance reductions obtained for each inversion is meaningless. Therefore,
 314 we computed the reduced χ^2 , as defined by Trampert and Woodhouse (2003):

$$\chi^2 = \frac{1}{N_P - T_r} (d - Gm)^T C_d^{-1} (d - Gm) \quad (4)$$

315 where N_P is the number of data, T_r the trace of the resolution matrix, d
 316 the data, G the kernel matrix, m the model vector, and C_d the covariance
 317 matrix. Models with smaller reduced χ^2 explain the data better. To decide
 318 whether differences in reduced χ^2 are significant, we performed F-tests (Bev-
 319 ington and Robinson, 2003). For all periods, the reduced χ^2 is smaller when
 320 anisotropy is accounted for (Figure 14). However, the decrease in χ^2 is mod-
 321 est in the period ranges 30-70 s, and F-tests show that these differences are
 322 not significant. At periods of 80 s, the decrease in χ^2 is more relevant, and
 323 F-tests indicate that it is slightly more significant. The reduced χ^2 decreases
 324 further when 4ψ terms are accounted for, but compared to the model with
 325 isotropic and 2ψ terms only, differences are very small and F-tests show that
 326 this χ^2 reduction is not significant. Overall, these tests suggest that the con-
 327 tribution of the 2ψ anisotropic term and the 4ψ anisotropic term **may not**
 328 **be needed to explain** our data in all our period ranges, which is consistent
 329 with previous study by Abe and Kanamori (1970). However, because the
 330 amplitudes of the 2ψ and 4ψ anisotropy are not negligible compared to the
 331 isotropic anomalies, we accounted for these terms in the inversion. Further-
 332 more, we propose a possible interpretation for the 2ψ anisotropy (sections
 333 Results and Discussion) that is consistent with the identified tectonic and
 334 geodynamics events, thus giving support to the presence of anisotropy.

335 4. Results

336 Figure 15 plots our preferred model for selected periods in the range 30-
 337 80 s. The background color map represents the isotropic anomalies, and the
 338 direction and size of the bars indicate the direction of fast propagation and

339 amplitude of 2ψ anisotropy.

340 At periods of 30 s and 40 s (Figure 15 a and b), the pattern of isotropic
341 anomalies is dominated by two main regions. The central part of the Sea
342 of Japan mostly displays fast velocities (up to +5% velocity perturbations),
343 whereas the surrounding regions show slow velocities (up to -4% perturba-
344 tions in the east). **Interestingly, at 40 and 50 s, the separation be-**
345 **tween slow and fast velocities is well correlated with the boundary**
346 **between the Okhotsk and Amur plates (Bird, 2003).** Around 60 s
347 period, there is a change in the pattern. Fast velocities are found mostly to
348 the northern and eastern part of our model, whereas the central part of the
349 Sea of Japan displays slow velocities (up to -5%). Finally, at 70 and 80 s
350 period, the dominant feature is a dichotomy between the southwestern part
351 of the Sea of Japan (up to -4% perturbations) and its northeastern part (up
352 to +5% velocity perturbations) becomes the dominant feature.

353 The anisotropic pattern changes with period, but is regionally more co-
354 herent than the isotropic pattern. **At periods of 40 and 50 s, sampling**
355 **the lower crust and upper lithospheric mantle, the anisotropy in**
356 **the Sea of Japan is about 1% in amplitude, with a mostly NE-**
357 **SW direction of fast propagation. Note that this direction slightly**
358 **varies around the central part. Furthermore, it turns to a NW-SE**
359 **direction beneath the Honshu and Hokkaido Islands.**

360 At periods of 60 s and longer, anisotropy is different from that
361 at shorter period. **Beneath the Sea of Japan and the Japanese**
362 **Islands, the 2ψ anisotropy reaches 0.5% to 1% in amplitude, with**
363 **a NE-SW direction of fast propagation, i.e. roughly perpendicular**

364 to that at 30-40 s. Beneath south of Korea and east of Kyushu, by
365 contrast, the direction of fast propagation is NW-SE.

366 At periods of 50-70 s, the isotropic and anisotropic patterns
367 change slightly beneath the Sea of Japan (isotropic) and the Japanese
368 Islands (anisotropic). This suggests a transition zone between the
369 anisotropy found at shorter periods (around 30 s) and those found
370 at longer periods (70 s and longer).

371 5. Discussion

372 In this section, we discuss our results (Figure 15) and compare them
373 with previous studies. We also provide some tectonic and geodynamical
374 interpretations for the evolution of the lithosphere beneath the Sea of Japan.

375 5.1. Korean Peninsula

376 Beneath the Korean Peninsula, our model predicts positive isotropic ve-
377 locity anomalies in the period range 30-40s, but negative anomalies at longer
378 periods (50-80 s).

379 Shorter periods (30 s, sampling the deep crust) display fast velocities be-
380 neath Korea, consistent with previous surface-wave tomographic imaging by
381 Cho et al. (2007) and ambient noise cross-correlation study by Kang and Shin
382 (2006). At 30-40 s, the fast velocities are also in good agreement with tomo-
383 graphic studies at larger scale (Friederich, 2003; Lebedev and Nolet, 2003;
384 Gao et al., 2011; Legendre et al., 2015b), and are consistent with previous
385 observations and crustal models of mainland China (Zheng et al., 2008; Sun
386 et al., 2010; Zhou et al., 2012), receiver function analyses (Yoo et al., 2007;

387 Chen et al., 2010), gravity measurements (Xu, 2007), and surface-wave anal-
 388 ysis of the East China Sea and Korean Peninsula (Legendre et al., 2014b).
 389 Our phase-velocity maps beneath the Korean Peninsula are furthermore in
 390 agreement with previous studies and with characteristic features of the ge-
 391 ological map of the region, displaying high velocities in the regions where
 392 granitoid basement rock outcrops are located (Pouclet et al., 1994; Kang and
 393 Shin, 2006; Cho et al., 2007; Yoo et al., 2007; Choi et al., 2013; Lee et al.,
 394 2015).

395 The anisotropic pattern is fairly stable in the lithosphere beneath the
 396 Korean Peninsula (sampled by periods of 50 s and larger), displaying very
 397 small amplitudes. This coincides well with the timing of the volcanic activity
 398 during the Mesozoic and Cenozoic (Kim et al., 2014). The lack of recent
 399 volcanic activity, combined with the fast velocities found at periods of 30 and
 400 40 s, suggests the presence of a possible deep magmatic reservoir, inferred
 401 by slow velocities found at 50 s and longer, as already observed in Legendre
 402 et al. (2014b).

403 5.2. *Japan Basin*

404 In the northern part of our model, covering the Japan Basin (Figure 1),
 405 the path coverage is quite sparse (Figure 8) due to the lack of seismic sta-
 406 tions deployed in the north (Figure 2). In this region, our models display
 407 slow velocities beneath the Japan Basin in the period range 30-50s, and fast
 408 velocities for periods of 70 s and longer. This correlates well with the varia-
 409 tions of thickness of the crust, i.e. thicker in the south and thinner beneath
 410 the Japan Basin (Kim et al., 1994; Kurashimo et al., 1996; Sato et al., 2004):
 411 thick crust will display lower velocities in the shallow lithospheric depth be-

412 cause surface waves sensitive to that depth wave will sample more crustal
413 than lithospheric mantle due to the thick crust. The decrease in crustal
414 thickness towards the north was widely influenced by the tectonic history
415 of the back-arc spreading that formed the Sea of Japan from 32 to 10 Ma
416 (Tamaki, 1992; Pouclet et al., 1994; Choi et al., 2013; Zahirovic et al., 2013,
417 2014; Yoon et al., 2014).

418 At periods of 30-40 s, the direction of fast propagation in the Japan
419 Basin is mostly NE-SW, parallel to the direction of the opening of the Sea
420 of Japan. This area belongs to the North American Plate or Okhotsk Plate,
421 which displays slow velocities beneath the Japanese Islands and beneath the
422 Japan Basin at periods of 30-60s, and fast velocities at periods of 70 s and
423 longer. For periods longer than 60 s, sampling the top of the as-
424 thenosphere,, the anisotropy beneath the central part of the Sea
425 of Japan is parallel to the direction of the Absolute Plate Mo-
426 tion (DeMets et al., 1990, 1994; Bird, 2003) of the Amur Plate
427 (roughly NNE/SSW), as shown in Figure 1, suggesting that it re-
428 sults from current deformation related to asthenospheric flow. At
429 shorter periods (40-50 s), sampling the lower crust, the direction
430 of fast propagation is again NE-SW, but the possible origin of the
431 anisotropy is less clear. The fast direction does not fit the direction
432 of the opening of the Sea of Japan. Because it matches the APM
433 of the Amur Plate (Bird, 2003), it could be related to an earlier
434 motion of the Amur Plate, as the lithosphere was thinner.

435 5.3. Subduction zones

436 Beneath the Japanese Islands, slow velocities are found at periods in the
437 range 30-60 s, whereas fast velocities are seen at longer periods (70 – 80 s).
438 Beneath the Japanese Islands, the slabs are dipping **at an angle of about**
439 **30°** (Hasegawa et al., 2013; Liu et al., 2013), **and the slow velocity ob-**
440 **served at 30-60 s in this region may be related to crustal thickening**
441 **above the slab (Figure 16). Rayleigh waves at these periods would**
442 **then sample the crust beneath the Japanese Islands, and the litho-**
443 **spheric mantle west of these islands. It is also interesting to note**
444 **that the transition between the fast and slow velocity is well cor-**
445 **related with the boundary between the Okhotsk and Amur plates**
446 **(Bird, 2003).** The orientation of the fast axis of anisotropy is also very
447 different from the surrounding areas, indicating a different mechanism of de-
448 formation. Beneath the subduction zone, trench-normal anisotropy can be
449 seen at periods of 30-40 s, whereas a trench-parallel anisotropy is found at
450 longer periods (70 – 80 s).

451 **The trench perpendicular anisotropy at periods of 30-60 s could**
452 **be related to crustal slicing triggered by the descent of the slab.**
453 **Such a mechanism was suggested to explain seismic anisotropy ob-**
454 **served in the central Alps (Fry, 2010).** Trench-parallel anisotropy, as
455 we observe here in the range 70-80 s, has not been observed in all sub-
456 duction system, but Long and van der Hilst (2006) concluded that for the
457 Ryukyu subduction system the most likely explanation for the observations
458 of trench parallel anisotropy is the corner flow in the mantle wedge com-
459 bined with the B-type olivine fabric (Kao et al., 1998; Kneller et al., 2005).

460 Anisotropy beneath the Japan subduction zone has been investigated using
 461 SKS splitting measurements (Nakajima and Hasegawa, 2004; Tono et al.,
 462 2009; Huang et al., 2011), and P-wave azimuthal anisotropy (Ishise and Oda,
 463 2005; Wei et al.; Zhao et al., 2015). Most of these studies found a consis-
 464 tent anisotropic pattern beneath the Sea of Japan, with fast direction of the
 465 azimuthal anisotropy parallel to the trench in the lithosphere, and trench-
 466 normal in the asthenosphere. Behn et al. (2007) considered several models
 467 to explain the trench-parallel anisotropy under the volcanic front, including
 468 slab rollback, oblique subduction, and deformation of water-rich olivine, to
 469 explain trench-parallel anisotropy, and concluded that none of these mech-
 470 anisms are consistent with all observations. Instead, small-scale convection
 471 driven by the foundering of dense arc lower crust may explain for the trench-
 472 parallel anisotropy, even in settings with orthogonal convergence and no slab
 473 rollback.

474 Although seismic anisotropy in the upper mantle is generally attributed
 475 to the crystallographic-preferred orientation (CPO) of olivine crystals, the
 476 strong trench-parallel anisotropy observed in several subduction systems (as
 477 the Ryukyu Trench) is difficult to explain in terms of olivine anisotropy
 478 only. Katayama et al. (2009) showed that the lattice-preferred orientations
 479 (LPO) of serpentine, the main hydrous mineral in the upper mantle, can
 480 produce the strong trench-parallel seismic anisotropy observed in subduction
 481 systems because the anisotropic amplitude due to serpentine is much larger
 482 than that due to olivine. Recent shear-wave splitting measurements display
 483 constant NW fast direction in a broad region below the mantle wedge of
 484 the Japan subduction zone (Tono et al., 2009). The constant NNW fast

485 direction of the anisotropy would be consistent with the anisotropic pattern
 486 we observe in this region at periods of 30-40 s (sampling the mantle wedge).
 487 These measurements have been attributed to the fossil sea-floor spreading
 488 anisotropy in the subducting Pacific slab, due to the fact that such anisotropy
 489 may have been preserved during its subduction down to the transition zone
 490 (Tono et al., 2009).

491 *5.4. Southern part of the Sea of Japan and East China Sea*

492 **In the southern part of our model, south of the Kyushu Is-**
 493 **land and the Korean Peninsula (Figure 1), we see fast velocities**
 494 **at shorter periods (30 s) and slow velocities at longer periods (40-**
 495 **80 s). The anisotropy is relatively uniform in this area, with a NW-**
 496 **SE direction at all periods.** The low anisotropy at 30 s and the constant
 497 direction of the anisotropy at periods of 40 s and longer could indicate that
 498 this region has been less affected by surface deformation, with less rifting and
 499 no oceanization, leading to a more uniform pattern of anisotropy. The region
 500 beneath the Kyushu Island extending towards the Korean Peninsula might
 501 be the southern extent of the fast anomaly beneath the continental crust of
 502 the Sea of Japan (Ulleung and Yamato Basins). **In the east, slow veloci-**
 503 **ties are seen at periods of 30-60 s, probably due to the oceanization**
 504 **of the Japan Basin, whereas in the west, fast velocities are found**
 505 **at similar periods.**

506 *5.5. Comparison with previous studies*

507 Our model is generally in good agreement with tomographic studies at
 508 larger scale (Friederich, 2003; Lebedev and Nolet, 2003; Gao et al., 2011;

509 Legendre et al., 2015b). However the lack of resolution of those models does
 510 not allow for local interpretations. Recent noise tomography for monitoring
 511 the Sea of Japan (Yoshizawa et al., 2010; Lee et al., 2015) presents high-
 512 resolution tomographic images of the region for shorter periods (up to 30-
 513 40 s). These studies focus on the strait between Korea and Japan, and
 514 interestingly, we observe similar pattern to those in the isotropic part of our
 515 model in this region. Body-wave tomography (Hirahara and Mikumo, 1980;
 516 Hirata et al., 1989; Zhu et al., 2002; Honda and Nakanishi, 2003; Matsubara
 517 et al., 2005, 2008; Hirose et al., 2008; Zheng et al., 2011) and surface wave
 518 tomography (Friederich, 2003; Huang et al., 2003; Lebedev and Nolet, 2003;
 519 Legendre et al., 2015b) at a local scale also display similar features to our
 520 model, including the dichotomy between the continental and oceanic crust
 521 beneath the Sea of Japan.

522 The main contribution of this work is to present isotropic and azimuthally
 523 anisotropic Rayleigh-wave phase-velocity maps beneath the Sea of Japan.
 524 Those maps allow the examination of azimuthal anisotropy at different peri-
 525 ods (depths), which can be related to the evolution of the lithosphere in this
 526 region.

527 **6. Conclusions**

528 We have constructed isotropic and azimuthally anisotropic Rayleigh-wave
 529 phase-velocity maps for the Sea of Japan in the period range 30-80 s, and us-
 530 ing regional broadband records from multiple networks including the China
 531 National Seismic Network, the Korea Meteorological Administration, the Na-
 532 tional Research Institute for Earth Science and Disaster Prevention in Japan,

533 and the Incorporated Research Institutions for Seismology. This study dis-
 534 plays for the first time the examination of azimuthal anisotropy with depth
 535 beneath the Sea of Japan, which is summarized in Figure 16. The models
 536 we obtain reveal features changing both laterally and with period (i.e. with
 537 depth). **Anisotropy is not needed to explain the raw data. How-**
 538 **ever, if it is present, our models resolve two distinct anisotropic**
 539 **layers, which may have different origins. In the period range 30-**
 540 **50 s, anisotropy beneath the Japanese Islands may be related to**
 541 **crustal slicing. The origin of the anisotropy we observe at these**
 542 **periods in the Sea of Japan is more uncertain. At periods of 60 s**
 543 **and longer, sampling the top of the asthenosphere, anisotropy may**
 544 **be related to asthenospheric flow.** The anisotropic pattern coupled with
 545 the isotropic velocity perturbations can be used to sketch the tectonic units
 546 in the region: the boundary between continental and oceanic lithospheres,
 547 and the boundary between North American, Eurasian and Pacific plates.

548 Acknowledgments

549 Three anonymous reviewers are acknowledged for their constructive com-
 550 ments that greatly improved this manuscript. This work was funded by
 551 Ministry of Science and Technology (MOST) of Taiwan under grants 103-
 552 2116-M-001-026-MY3, 101-2116-M-001-035, 104-2116-M-001-012, and 104-
 553 2811-M-002-063, and an Academia Sinica Career Development Award (AS-
 554 098-CDA-M02. Q.-F. Chen was funded by the National Science Founda-
 555 tion of China under grant 41130316. Waveform data for this study are
 556 provided by the Data Management Center of China National Seismic Net-

557 work at Institute of Geophysics, China Earthquake Administration (SEIS-
558 DMC, doi:10.7914/SN/CB), Korea Meteorological Administration, the Na-
559 tional Research Institute for Earth Science and Disaster Prevention in Japan
560 and the Incorporated Research Institutions for Seismology. Figures were
561 generated with the Generic Mapping Tools (Wessel and Smith, 1995).

Figure 1: Tectonic setting of the Sea of Japan. The thick pink lines represent the plate boundaries (Bird, 2003). Blue dots are active volcanoes. Tectonic features include SCC - South China Craton; EP - Eurasian Plate; Jb - Japan Basin; Jt - Japan Trench; Kp - Korean Peninsula; PP - Pacific Plate; PSP - Philippine Sea Plate; Sa - Sakhalin Oblast (Russia); SKp - South Korean Plateau; Ub - Ulleung Basin; Yb - Yamato Basin; Yr - Yamato Ridge. The Japan Peninsula is formed by four main islands: Ho - Hokkaido, Hn - Honshu; Sh - Shikoku; Ky - Kyushu. The yellow boxes indicate the presence of continental crust, whereas the green boxes indicate the presence of oceanic crust. The Absolute Plate Motion (Bird, 2003) is indicated with the brown arrow.

Figure 2: Stations (triangles) and interstation paths (black lines). The colors of the triangles indicate the network: red (CNSN), blue (KMA), green (FNet) and pink (IRIS). The red path indicates a specific station pair used as an example in Figure 4 to explain the method.

Figure 3: Seismic events used in this study. They occurred between January 2007 and December 2011. Circles represent the locations of the epicentres, with the colors indicating the depths of the events (red: 0-100 km; green: 100-300 km; blue: 300 + km). The sizes of the circles are proportional to the event magnitudes.

Figure 4: Vertical components of the displacement recorded at the pair of stations (top and center) and the cross-correlation function (bottom). The color plots show the time-frequency diagram of the energy, where warm colors in the images indicate higher energy, and the maximal energy is plotted with a thin white line.

Figure 5: Inversion of the cross-correlation function for a phase-velocity dispersion curve. Possible solutions are represented by an array of curves (blue lines). The model-predicted phase-velocity for the Rayleigh wave in Earth Model AK135 is displayed in dashed line. The red part of the curve is the dispersion range selected for this specific event.

Figure 6: Individual phase-velocity curves measured for different earthquakes (top) and the average (bottom) with standard deviation (blue lines).

Figure 7: Manually measured phase-velocity curves (black) for all station pairs. The orange line is the predicted phase-velocity calculated from a 1D reference Earth Model AK135 (Kennett et al., 1995), and the red line is the average curve for the study region.

Figure 8: Inter-station paths that provide phase-velocity measurements at specific periods: (a) 30s, (b) 40s, (c) 50s, (d) 60s, (e) 70s and (f) 80 s. **The red triangles represent locations of the seismic stations.**

Figure 9: Sensitivities of the Rayleigh-wave phase velocities to shear-wave speed as functions of depth for selected periods.

Figure 10: RMS of the amplitudes calculated for all periods for the isotropic terms (orange), the 2Ψ anisotropy (green) and the 4Ψ term (red).

Figure 11: Inversion tests. The models have been calculated for a specific period of 40 s with our preferred parameters (framed in dark blue in panels b, f, j, n and r), with decreasing percentage in the rejection of data, from (a) 0% to (d) 90% with an interval of 30% between each image, increasing damping values from (e) to (h) for isotropic components, increasing damping values from (i) to (l) for anisotropic components, increasing smoothing values from (m) to (p) for isotropic components, and increasing smoothing values from (q) to (t) for anisotropic components. The preferred parameter chosen for the final inversion is framed in dark blue. The number in green in the bottom right indicates the chosen value for the selected parameter.

Figure 12: Smoothness of the models calculated with respect to the selected parameters for the smoothing on the isotropic component (a), the damping on the isotropic component (b), the smoothing on the anisotropic component (c), and the damping on the anisotropic component (d). The blue curves are the model smoothness for the isotropic part of the model, the red curves are the model smoothness for the 2ψ anisotropy part of the model, and the green curves are the model smoothness for the 4ψ anisotropy part of the model. The black stars indicate the preferred chosen values for each parameter.

Figure 13: Resolution tests. (a-c) Input test models at periods of 40, 60 and 80 seconds; and (d-f) reconstructed models for respective periods. The reconstructions are performed to test the retrieval of both isotropic and anisotropic structures.

Figure 14: Reduced χ^2 as a function of period. Inversions are performed for isotropic anomalies only (orange stars), isotropic and 2ψ -anisotropy anomaly (green diamonds), and isotropic and full anisotropy (2ψ and 4ψ) (red circles). The other parameters (grid spacing, rejection, damping and smoothing factors) are the same as those of the preferred values.

Figure 15: Isotropic (background color) and azimuthally anisotropic (yellow lines) phase-velocity maps for specific periods: (a) 30 s, (b) 40 s, (c) 50 s, (d) 60 s, (e) 70 s and (f) 80 s.

Figure 16: Sketch of the possible deformation pattern in the region based on our anisotropic model. The average anisotropic patterns at different periods (orange: 30 s, sampling the upper lithosphere; blue: 60 s, sampling the lower lithospheric and asthenospheric mantle). The origin of the anisotropy is indicated by the green arrows (opening of the Japan Basin) and the brown arrow (slab interaction). The Absolute Plate Motion (Bird, 2003) is indicated with the red arrow.

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