Anisotropic tomography of Hokkaido reveals delamination-induced flow above a subducting slab

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Abstract We present a new 3-D anisotropic seismic model for the crust and upper mantle beneath Hokkaido (Japan) based on tomographic inversion of P and S arrival times from a regional seismic database. The P model is parameterized with three parameters at each point that describe the azimuthal anisotropy; the S model is represented isotropically. The isotropic P and S velocity anomalies match nearly perfectly. In the crust, they show a prominent linear anomaly in central Hokkaido along the Kamuikotan and Hidaka Belts, which represents the area of eastward underthrusting of the Japan Block underneath the Kuril fore arc. We interpret the high-velocity anomaly beneath the Hidaka zone as being delaminated mafic crust and entrained mantle lithosphere, which developed due to crustal shortening in the collision zone. One of our vertical sections shows a very unusual configuration for a subduction zone: a low-velocity slab is overlain by a high-velocity body in the mantle wedge. We propose that the high-velocity delaminated material sinking along the slab surface prevents the escape of fluids and melts from the upper part of the slab, where they are generated due to phase transitions. As a result, a large portion of the fluids is entrained downward and lowers the seismic velocities in the slab. The azimuthal anisotropy in the crust clearly corresponds to the major tectonic units and delineates the major suture zones. In the mantle, the anisotropy has a fan-shaped configuration and most likely represents the deviating of flows starting in southern Hokkaido and splitting into three directions. The western and eastern flows proceed toward the two volcanic groups on Hokkaido, and they may carry with them additional material to trigger the characteristic caldera-forming eruptions in these groups.

1. Introduction

Hokkaido is the second largest of Japan’s islands and is located in northeastern Japan. This island experiences strong seismic activity: events of magnitude 7 or greater occur an average of every 40 years. The most recent strong earthquakes in Hokkaido occurred in 1970 (Mj = 6.7) and in 1982 (Mj = 7). Potentially strong eruptions posing serious hazards to the population and infrastructure of Hokkaido may occur in several active volcanic complexes. The presence of calderas with the diameter of 10–20 km indicates the large explosive potential of these volcanoes. All of these tectonic and volcanic activities are directly or indirectly connected with the ongoing subduction of the Pacific Plate. Understanding the links between the processes involving the subducting slab and surface tectonics is a challenging task that attracts the attention of many scientists from various domains of the geosciences. Advances in solving these problems are impossible without studying the deep structures, and seismic tomography is one of the most powerful instruments for this purpose.

Japan is densely covered by permanent and temporary seismic stations, which provide large amounts of data that can be used for tomographic studies. Although the average coverage in Hokkaido is less dense than in other parts of Japan, the seismic network in Hokkaido remains one of the best in the world, and the resulting data are well suited for seismic tomography, which explains the large number of seismic studies in Hokkaido. The crustal and mantle seismic structures beneath Hokkaido have been extensively studied for decades [Takanami, 1982; Miyamachi and Moriya, 1984; Miyamachi et al., 1994; Katsumata et al., 2006; Kita et al., 2006; Wang and Zhao, 2005]. Recent studies by Kita et al. [2010a, 2010b] benefited from the dense seismic networks in Hokkaido deployed during recent years, which recorded a large number of seismic events. Based on detailed analysis of the tomographic results in the area of the Hidaka Belt, Kita et al. [2012] identified an elongated high-velocity body below the suture zone, which was interpreted as a result of lithospheric thrusting due to collision between the Japan and Kuril fore-arc segments.
In addition to earthquake tomography, active source studies have also been performed on Hokkaido since the 1970s. Several seismic profiles were shot mostly in the mountain area of Hidaka, and the collected refraction and reflection data were used to construct detailed images of the upper crust that generally extended down to a depth of 20 km and in few cases extended down 40 km [e.g., Moriya et al., 1998; Iwasaki et al., 1998, 2004; Tsumura et al., 1999; Ito, 2000, and references herein]. In the same area, Arita et al. [1998] performed an analysis of vibroseismic reflection data together with gravity modeling.

Seismic anisotropy is an additional parameter that provides valuable information on dynamic processes, such as the distribution of deformations in the crust and material displacement in the mantle (e.g., see overview in Long [2013]). The crust and upper mantle in subduction zones appear to be strongly anisotropic. This conclusion has been demonstrated by numerous studies of shear wave splitting [e.g., Anglin and Fouch, 2005; Pozgay et al., 2008; Audoine et al., 2004; Abt and Fischer, 2008; Long and van der Hilst, 2005; Levin et al., 2004]. There are a number of studies based on shear wave splitting in the area of Japan that demonstrate the presence of regular features. For example, Long and van der Hilst [2005] presented shear splitting results for the entire area of Japan. In Ryukyu, they observed clear trench-parallel structures, whereas below large islands, they clearly distinguished the presence of both trench-parallel and trench-perpendicular anisotropy.

Using the much denser network of stations in the central part of Honshu, Nakajima and Hasegawa [2004] reported clear trench-parallel features based on data from most stations along the eastern coast and trench-perpendicular anisotropy below the central and western parts of the island. In another model involving northern Honshu and Hokkaido, Nakajima et al. [2006] reported scattered orientations of anisotropy, but after averaging and generalization, they reported a similar switch in anisotropy orientation from trench parallel to trench perpendicular. Note that shear wave splitting provides integral information regarding anisotropy along raypaths but cannot locate anisotropic bodies in space.

Anisotropic seismic tomography provides the possibility of studying the 3-D distribution of the anisotropy parameters. There are several examples of studies of subduction zones based on anisotropic tomography inversions [e.g., Eberhart-Phillips and Henderson, 2004; Koulakov et al., 2009a; Rabbel et al., 2011]. There are also several anisotropic tomographic models of various parts of Japan [e.g., Ishise and Oda, 2005; Huang et al., 2010; Wang and Zhao, 2008, 2010].

Wang and Zhao [2009] developed an anisotropic tomographic model of the area of Hokkaido, as in this study. In our opinion, the anisotropy orientations in the model by Wang and Zhao [2009] are too scattered: these orientations appear to vary greatly from one node to the next. As was found in several anisotropic tomography studies [e.g., Koulakov et al., 2009a; Rabbel et al., 2011], the inversion of anisotropy parameters is much less robust than that of isotropic inversion. When we observe so large variation in anisotropy directions, we suspect that they are merely the result of an unstable inversion solution. We believe that an additional study of this anisotropy using a different algorithm would be helpful in confirming or disconfirming the previous studies. Another problem with the model by Wang and Zhao [2009] is that the isotropic portion does not fit with the models of the same area developed by other authors [e.g., Murai et al., 2003; Kita et al., 2010b, 2012]. Because of these inconsistencies between the various tomographic models, many questions related to deep structures beneath Hokkaido remain. Therefore, in our opinion, despite the large number of existing models, additional tomography studies of this region using different algorithms are warranted. In this paper, we present a new anisotropic tomographic model of the crust and mantle wedge beneath Hokkaido and propose a new interpretation of the tectonic and volcanic processes in Hokkaido.

2. Geologic Setting

The Japan arc system is composed of four main sections: the northeastern Honshu, southwestern Honshu, Ryukyu, and Kyushu arcs. The growth of this arc system along the continental margin of Asia has occurred primarily since the Permian [e.g., Taira, 2001]. Interactions among the five plates, namely, the Eurasian, Amur, Okhotsk, Pacific, and Philippine Sea Plates, control the current tectonics of the Japanese arc system [Taira, 2001; Isozaki, 1996]. The Pacific Plate, which is 120–150 Ma in age, approaches the Japan arc from the east-southeast at a rate of 8–10.5 cm/yr [Uto and Tatsumi, 1996; Isozaki, 1996] and subducts nearly perpendicularly under the Japan arc and obliquely under the Kuril and Izu-Bonin arcs. According to Isozaki [1996], Japan was detached from mainland Asia at approximately 20 Ma due to the back-arc opening of the Japan Sea.
Hokkaido is the northernmost large island of the Japan arc and is located on the southern end of the Kuril arc. The evolution of the Hokkaido Central Belt started in the Late Jurassic to the Early Cretaceous and is attributed to the collision between the Okhotsk and the Kamuikotan-Nukabira paleolands [Jolivet, 1986]. Hokkaido is located at a junction between the former and new subduction zones. The former subduction is presumed to be located along the Cretaceous-middle Eocene Susunai-Kamuikotan metamorphic belt passing through Hokkaido and Sakhalin [e.g., Kimura et al., 2014; Zharov, 2005]. Many structures in Hokkaido are prolonged to Sakhalin (Figure 1a). For example, similar low-pressure metamorphic rocks as in the Hidaka Belt are traced in the Aniva Belt in Sakhalin, whereas the Kamiukotan high-pressure belt with ophiolite fragments corresponds to Susunai Belt in Sakhalin [e.g., Zharov, 2005; Ueda and Miyashita, 2005]. The coexistence of two nearly perpendicular structures related to the former and present subduction zones possibly determines the rhombic shape of the Hokkaido Island.

The oblique component of subduction to the NE of Hokkaido causes southeastward displacement of the Kuril fore arc and leads to a collision with the Japan fore arc along the Hidaka Belt [e.g., Jolivet, 1986; Kimura, 1996]. Since the late Miocene, tectonic activity in the Hidaka Belt and adjacent areas is mostly controlled by this arc-arc collision. The strongest historic seismic activity in Hokkaido has been observed along this belt. The linear structure of the Hidaka Belt extends to the junction between the Kuril and Japan trenches (Figure 1b).

In Hokkaido, there are three clearly separated volcanic centers, namely, the western, central, and eastern volcanic centers (WVC, CVC, and EVC, respectively, in Figure 1b). In the WVC and EVC, several large calderas are present. For example, the 13 × 15 km Shikotsu caldera in the WVC was formed during one of Hokkaido’s largest Quaternary eruptions approximately 31,000–34,000 years ago [e.g., Yamagata, 1993]. The eastern volcanic group (EVC) is dominated by the Akan volcano, one of the most active in Hokkaido, which includes a large 24 × 13 km diameter caldera and a group of younger, partly Holocene, andesitic cones [Goto et al., 2000; Hasegawa et al., 2012]. Another large volcano in the EVC is Mashu, which exhibits a 7 km wide caldera that formed approximately 7000 years ago [Hasegawa et al., 2012]. In contrast, no caldera-forming eruptions were recorded in central Hokkaido. For example, the Tokachi zone in the central volcanic group (CVC) includes a group of andesitic stratovolcanoes and lava domes [e.g., Nishimura, 1995]. Frequent but mostly small to moderate phreatic eruptions occurred in this group during historic time. The explanation for the clustering of volcanoes in Hokkaido is still debated. It is unclear why the largest caldera-forming eruptions occur in eastern and western Hokkaido, whereas less volcanic activity occurs in central Hokkaido, which is crossed by the slab boundary. These and other questions related to the volcanic activity in Hokkaido are actively debated and have not yet been definitively answered. We expect that our new tomographic model will help to shed light to these and other questions related to the tectonics and volcanism of Hokkaido.
3. Data and Algorithms

In this study, we used open-access data provided by the Japan Meteorological Agency. We selected a circular area with a radius of 200 km centered at 43.5°N and 143°E, which includes most of Hokkaido (Figure 1c). It was shown in previous studies [e.g., Koulakov et al., 2009a] that the reconstructions of anisotropy parameters are much less stable than those of an isotropic model. Therefore, an anisotropic tomographic inversion is very sensitive to the amount and configuration of the data. To reduce the trade-off problems related to the uncertainties of source locations, it is preferable to use events with the largest possible number of picks per event. Due to the large amount of data in the Japan Meteorological Agency (JMA) catalog, we were able to select events with more than 40 recorded picks. We used the picks for the P and S data with residuals less than 1.5 s and 2 s after the initial locations in the starting 1-D velocity model, respectively.

Another criterion used in this study was that the event should be located a horizontal distance of no more than 100 km from the nearest station to avoid events located far outside the network, meaning that we used some events located outside the outer perimeter of the station network. When defining this criterion, we took into account results of modeling by Koulakov [2009a], who proved the importance of using out-of-network events for improving the results of tomography inversion. The existence of rays from such events increases the variability of the ray orientations in the study volume that is in favor to the resolution improvement. Although the location accuracy of such events is lower, the relative residuals, which are used for tomography, are not strongly affected by location errors. These relative residuals provide the information on seismic anomalies in the study area beneath seismic stations.

No other criteria (e.g., magnitude, gap, and amount of S picks) were used in this study. In total, we used the data from 113 stations, which are denoted by blue triangles in Figure 1c. The red dots in Figure 1c denote the locations of the 4561 events used in this study. The final data set used in the tomography contained 116,742 P and 99,556 S arrival times (average of ~47 picks per event).

We used the 1-D reference model developed in the previous tomographic study by Kita et al. [2012], except that instead of layers with constant velocities, we used a linear interpolation between velocities at different depths. In this case, we used the constant value of Vp/Vs ratio equal 1.75 and P velocities defined in several depth levels, as given in Table 1.

For the tomographic inversion, we used the ANITA algorithm, which is a modified version of the LOTOS code [Koulakov, 2009b]. Older versions of the ANITA code were used for anisotropic inversions in central Java [Koulakov et al., 2009a] and Central America [Rabbel et al., 2011]. The general workflow of the ANITA code includes the same steps of local earthquake tomography as those in the LOTOS code. The calculations start with the step of preliminary source location in the initial 1-D model. At this preliminary step of source location, we used approximate calculations of travel times based on previously computed tabulated values in the 1-D model. Using this approximation instead of performing ray tracing allowed for locating the events using the grid search method within a reasonable amount of calculation time.

The following iterative procedure started with the steps of source locations in the 3-D anisotropic model (which is isotropic and 1-D in the first iteration). The travel times in this step were calculated using the bending algorithm in the search for raypaths that provide the minimum travel times. In the first iteration, we constructed the parameterization grids, which remained unchanged in the following iterations. We used the node parameterization (in contrast to the cell parameterization used in the previous versions of ANITA). The nodes were designated in areas with sufficient data coverage along vertical lines of regular spacing in map view (10 km in our case). In the vertical direction, the nodes were installed in areas with sufficient data

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>P Velocity (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>4.79</td>
</tr>
<tr>
<td>5</td>
<td>6.644</td>
</tr>
<tr>
<td>10</td>
<td>6.322</td>
</tr>
<tr>
<td>15</td>
<td>6.524</td>
</tr>
<tr>
<td>20</td>
<td>6.721</td>
</tr>
<tr>
<td>25</td>
<td>6.895</td>
</tr>
<tr>
<td>30</td>
<td>7.032</td>
</tr>
<tr>
<td>35</td>
<td>8.0</td>
</tr>
<tr>
<td>40</td>
<td>8.2</td>
</tr>
<tr>
<td>50</td>
<td>10.2</td>
</tr>
</tbody>
</table>

Table 1. One-Dimensional Model of P Velocity Used as a Starting Model for Computing the Tomography Results
coverage (more than 0.2 of the average ray density), and the spacing between nodes varied from 5 km to 20 km depending on the density of rays. To avoid any dependence on the grid orientation, we performed the inversion for four grids with different basic azimuths, i.e., 0°, 22°, 45°, and 67°, and then averaged the results.

The anisotropic inversion was performed only for the P velocity distribution. To assess the anisotropy of the S velocity, we would have needed more sophisticated information regarding the polarization of picked shear wave phases and their splitting, which was not provided by the standard JMA catalog. Therefore, the S velocity model was parameterized using a normal isotropic parameterization, as in the standard version of LOTOS.

In this study, for the P velocity model, we considered a simple approximation of azimuthal anisotropy, which is described by three parameters corresponding to the directions 0°, 60°, and 120°. The slowness along a ray, with the azimuth, α, and dip angle, β, (measured upward from the vertical axis) can be represented as follows:

\[
\sigma = \sigma_{\text{ref}} + (d\sigma_{\text{hor}} \sin \beta + d\sigma_{\text{ver}} \cos \beta) / (\sin \beta + \cos \beta)
\]

where

\[
d\sigma_{\text{hor}} = (d\sigma_0 + d\sigma_{60} + d\sigma_{120})/3
\]

\[
d\sigma_{\text{ver}} = d\sigma_1 + d\sigma_2 + d\sigma_3
\]

\[
d\sigma_1 = \frac{1}{3} \left( \cos(2\alpha) + 1 \right) d\sigma_0
\]

\[
d\sigma_2 = \frac{1}{3} \left( \cos \left(2\alpha - \frac{\pi}{3}\right) + 1 \right) d\sigma_{60}
\]

\[
d\sigma_3 = \frac{1}{3} \left( \cos \left(2\alpha + \frac{\pi}{3}\right) + 1 \right) d\sigma_{120}
\]

where \(d\sigma_0\), \(d\sigma_{60}\), and \(d\sigma_{120}\) are the variations of slowness along the corresponding azimuth with respect to the reference slowness value, \(\sigma_{\text{ref}}\). This parameterization represents a pseudoellipse with the orthogonally oriented maximum and minimum values of slowness (\(d\sigma_{\text{max}}\) and \(d\sigma_{\text{min}}\)) and azimuth of maximum slowness orientation (\(\psi\)). The parameters \(d\sigma_0\), \(d\sigma_{60}\), and \(d\sigma_{120}\) can be easily converted to \(d\sigma_{\text{max}}\), \(d\sigma_{\text{min}}\), and \(\psi\) and vice versa.

Note that when using the new version of the algorithm, we did not determine the vertical slowness variation as an independent parameter, as was done by Koulakov et al. [2009a] and Rabbel et al. [2011]. The vertical velocity anomaly appeared to be too sensitive to the a priori definitions of weighting and damping parameters, and it may take opposite values depending on the user settings; therefore, we removed it from the inversion.

The first derivative matrix, which represents the time variation along the \(i\)th ray due to the unit variation of slowness corresponding to \(j\)th parameter, was integrated along the \(j\)th raypath \(\gamma_j\) as follows:

\[
A_{ij} = \int_\gamma \Delta \sigma_j ds
\]

The value of \(\Delta \sigma_j\) at each point along the ray corresponded to the slowness variation due to the unit variation of \(j\)th parameter (node and one of the three directions). The slowness at the node was calculated based on the angles \(\alpha\) and \(\beta\) at the current ray point using formulas (1) and (2). The corresponding value of \(\Delta \sigma_j\) was computed numerically by bilinear interpolation, assuming that there was a unit nonzero value of anomaly at only the node corresponding to the \(i\)th parameter.

The amplitude and smoothing of the anisotropic model is controlled by additional matrix blocks. The amplitude damping assumed the addition of a diagonal matrix corresponding to a system of trivial equations:

\[
D^{\text{amp}} \Delta \sigma_j = 0
\]

where \(D^{\text{amp}}\) is the amplitude-damping parameter.

To smooth the distribution of anomalies in space, we used another matrix block, which can be represented by the equations

\[
D^{\text{sm}}(\Delta \sigma_k - \Delta \sigma_m) = 0
\]

where \(k\) and \(m\) represent the parameters in neighboring nodes corresponding to the same orientation; \(D^{\text{sm}}\) was the smoothing parameter. In this block, we listed all combinations of neighboring nodes in the grid for the three orientations of 0°, 60°, and 120°.
Finally, we controlled the anisotropy strength by damping the differences of parameter values corresponding to the same node and different directions as follows:

\[
\begin{align*}
D_{\text{anis}}(\Delta \sigma_0^m - \Delta \sigma_60^m) &= 0 \\
D_{\text{anis}}(\Delta \sigma_0^m - \Delta \sigma_{120}^m) &= 0
\end{align*}
\]

where \(D_{\text{anis}}\) is the parameter of anisotropy damping; \(\Delta \sigma_0^m, \Delta \sigma_60^m, \text{ and } \Delta \sigma_{120}^m\) are the slowness anomalies in different directions at the same node.

We performed the simultaneous inversion for the anisotropic P model, isotropic S model, and source parameters (four parameters representing coordinates and origin time of each event). The inversion of the entire matrix was performed using the LSQR method [Paige and Saunders, 1982; Nolet, 1987].

A crucial problem is related to the selection of optimal weighting and damping parameters for the inversion. Normally, we would estimate the values of parameters based on the results of synthetic testing. In the present case, we performed calculations for dozens of parameter sets based on both real and synthetic data. Finally, we found that inversions with large variations in various parameters lead to very similar configurations in the resulting models. In particular, amplitude damping, \(D_{\text{amp}}\), provided an effect similar to that of smoothing; therefore, in the final calculations, we removed the amplitude damping (Table 2).

After performing the inversions corresponding to four parameterization grids with different orientations, we recalculated the 3-D models corresponding to a regular grid. These anisotropic P and isotropic S models were used in the next iteration for relocations of sources based on the 3-D anisotropic ray tracing. The total number of iterations was coupled with the amount of damping: strong damping required more iterations; weak damping caused the solution to become unstable with a small number of iterations. Usually, we performed five iterations as a compromise between the calculation time and the need to account for nonlinear effects. Then we tuned the damping parameters to achieve the optimal results mainly based on synthetic reconstructions with checkerboard and realistic patterns.

### 4. Inversion Results and Testing

The resulting anisotropic P velocity model is presented in Figure 2 in four horizontal sections at depths of 10, 30, 60, and 90 km. The isotropic velocity anomalies, which are shown in color, were calculated using the formula

\[
dv_{\text{iso}} = \frac{1}{\pi} \int_{-\pi/2}^{\pi/2} dv_{\text{hor}}(\alpha) \, d\alpha
\]

where \(dv_{\text{hor}}(\alpha)\) is calculated based on formulas (1) and (2), assuming a dip angle of \(\beta = \pi/2\).

The maximum and minimum velocities (\(dv_{\text{max}}\) and \(dv_{\text{min}}\)) and the azimuth of the maximum velocity orientation (\(\psi\)) were derived numerically from the computed slowness parameters \(d\sigma_0^m, d\sigma_{60}^m, \text{ and } d\sigma_{120}^m\).

The anisotropy was calculated using the formula

\[
A_{\text{anis}}(\%) = 100 \left( dv_{\text{max}} - dv_{\text{min}} \right) / V_{\text{ref}}
\]

where \(V_{\text{ref}}\) is the reference velocity.
Figure 2. The resulting anisotropic P velocity model in horizontal sections. The color background denotes the isotropic velocity anomalies. The bars denote the orientations of high velocities; the lengths of bars denote the amplitude of anisotropy. The black lines denote the major geologic structures as in Figure 1b. The triangles denote the recent volcanoes. The red arrows indicate the possible flow directions in the mantle wedge.
The average residuals and variance reductions during the five iterations of the $P$ and $S$ models are presented in Table 3. In addition, this table presents the information for the isotropic $P$ velocity model, which will be discussed later. Note that the variance reduction in the $P$ and $S$ models is approximately 40–47%. Such relatively high values of reduction can be explained by the high quality of the data and strong heterogeneities in the models with respect to the initial 1-D velocity model.

Table 3. Averages of Absolute Residuals and Variance Reduction (With Respect to the First Iteration of the Model) of Anisotropic and Isotropic $P$ Models and Isotropic $S$ Model After Five Iterations

<table>
<thead>
<tr>
<th>Iteration</th>
<th>$P$ Model, Anisotropic</th>
<th>$P$ Model, Isotropic</th>
<th>$S$ model, Isotropic</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Average Residual (s)</td>
<td>Variance Reduction (%)</td>
<td>Average Residual (s)</td>
</tr>
<tr>
<td>1</td>
<td>0.311</td>
<td>0</td>
<td>0.311</td>
</tr>
<tr>
<td>2</td>
<td>0.242</td>
<td>22.0</td>
<td>0.254</td>
</tr>
<tr>
<td>3</td>
<td>0.202</td>
<td>35.0</td>
<td>0.221</td>
</tr>
<tr>
<td>4</td>
<td>0.178</td>
<td>42.8</td>
<td>0.200</td>
</tr>
<tr>
<td>5</td>
<td>0.162</td>
<td>47.7</td>
<td>0.186</td>
</tr>
</tbody>
</table>

Figure 3. $S$ velocity anomalies in horizontal sections. The black lines denote the major geologic structures as in Figure 1b. The triangles denote the recent volcanoes.
The resulting anisotropy distribution is denoted by bars in Figure 2, which are based on the derived values of $A_{\text{anis}}$ and $\psi$. Note that at all four depths, the anisotropy exhibits regular patterns that appear to be controlled by the major geologic structures. The maximum anisotropy is 15%; however, one should be careful when evaluating these values because they depend directly on the anisotropy damping. We substituted several values for this parameter and observed general increases and decreases in the anisotropy, but the anisotropy orientations remained nearly unchanged. Thus, we trust most of the anisotropy orientations but not their scalar magnitudes.

Figure 4. Five vertical sections showing the $P$ velocity anomalies. The black dots denote the events from the JMA catalog located less than 30 km away projected horizontally onto the sections. The triangles denote the projected locations of volcanoes. The dashed lines with numbers denote the intersections with other sections. Exaggerated topography is shown above each section. The map shows locations of the profiles and events from the JMA catalogue classified by color according to depth. The red dotted line denotes the curved alignment of the cluster of events in the depth interval of 50–80 km.
Horizontal sections of the $S$ velocity isotropic model are shown in Figure 3. The most striking feature is a nearly perfect correlation with the isotropy in the $P$ model. The same semblance can be observed in the vertical sections of the $P$ and $S$ velocity models shown in Figures 4 and 5. Note that for the $P$ velocity model, we do not show the anisotropy in the vertical sections because in our model, we considered only the azimuthal anisotropy; in the vertical sections, the model was assumed to be isotropic. In Figures 4 and 5, we present the seismicity distribution in horizontal and vertical sections through Hokkaido based on the entire data set from the JMA catalog to facilitate the following discussion.

We explored the effect of anisotropy on the resulting anomalies and data fit. Assigning a very large value to $D_{\text{anis}}$ in formula (6) in the $P$ model (for example, 1000 instead of 1.5) results in very small anisotropy values that approximate the isotropic solution for the $P$ model. If the other parameters remain unchanged, we can compare the results of the isotropic and anisotropic inversions and assess the effects of anisotropy on
improving the data fit. The results of the inversion with the excessively damped anisotropy are shown in Figure 6. The results of this inversion are very similar to that of the anisotropic inversion shown in Figure 2. There are minor differences, which are mostly related to the amplitudes of the positive anomalies. An analysis of the a posteriori residuals presented in Table 3 indicates a considerably larger reduction in the average absolute residuals in the case of the anisotropic inversion. The final value in the anisotropic case was 0.162 s, whereas that in the isotropic $P$ model was 0.186 s. Thus, there was a 13% reduction due to the anisotropic effect. In other regions such as central Java and Central America, which were studied using a similar algorithm, the anisotropic effect was much smaller. The reduction observed in our study is a clear indication of the strong anisotropy present in the crust and mantle wedge beneath Hokkaido.

To assess the resolution of the models, we performed a checkerboard synthetic test which is shown in Figure 7. Previous tomography results by Kita et al. [2012] demonstrated very good vertical and horizontal resolutions when resolving velocity anomalies, which can be explained by the large amount of data available in Hokkaido. In this paper, we mainly focus on the anisotropy. It is known from a number of studies [e.g., Koulakov et al., 2009a] that the anisotropic parameters in tomographic inversion are much less stable than the isotropic parameters. Therefore, to check the resolution of the anisotropic parameters, we

Figure 6. $P$ velocity anomalies resulting from isotropic inversion in horizontal sections. The black lines denote the major geological structures as in Figure 1b. The triangles denote the recent volcanoes.
developed a checkerboard model with a large size of patterns (75 km). It is also important that for a smaller spacing, the checkerboard itself behaves as an anisotropic medium, and such a configuration makes it difficult to discern anisotropy orientations inside the blocks. In every point of the synthetic model, we defined the azimuth of the main anisotropy axis and two velocity values oriented along orthogonal azimuths corresponding to fastest and slowest directions. In the “blue” cells, we defined the maximum and minimum velocities of 4% and 12%, respectively, and the fast velocity was oriented longitudinally. For the “red” cells, the anomalies were −4% and −12%, and the orientation was latitudinal. The change of the pattern signs occurred at a depth of 30 km. The synthetic data were computed for the same data as were used for calculation of the main model. To the synthetic data, we added random noise with the similar magnitude of 0.1 s for the $P$ and $S$ models. After calculating the synthetic data, we “forgot” the information regarding sources and started the full processing including the step of initial source locations in the 1-D model. All the inversion parameters were identical to those used for computing the main model. The restored anisotropic $P$ and isotropic $S$ models are shown in Figure 8. Below most of Hokkaido, the isotropic anomalies are well resolved. Sharp boundaries and angles between cells are evident, which means that sufficient resolution would be achieved using patterns with a smaller spacing, as reported in the previous tomography studies of

Figure 7. Checkerboard test for the anisotropic $P$ and isotropic $S$ velocity models. The configurations of the synthetic patterns is denoted by dotted lines. The synthetic anisotropy is oriented longitudinally in the red patterns and latitudinally in the blue ones. In the vertical direction, a change in sign occurs at a depth of 30 km.
Hokkaido. The reconstruction of the anisotropic patterns appears to be much less stable. High-quality reconstruction of the anisotropy orientations was achieved in only the central blocks. In the periphery, although the anisotropy orientation generally displays correct trends, it appears to be strongly smeared. Nevertheless, we can conclude from this test that the large anisotropy patterns with smooth variations in orientations identified in the primary result shown in Figures 2 are of the same size as the resolved anomalies in the checkerboard test, and thus, these findings appear to be reliable.

To assess the stability of the main structures in vertical sections, we have performed a series of tests with realistic configurations of synthetic patterns. The purpose of this test was to produce a model, which enables similar reconstructed anomalies as in the case of inversion of observed data. An important problem of tomography inversion is that because of smearing and damping, the amplitudes of anomalies in the resulting model are often strongly biased in respect to the true values. The synthetic modeling with realistic configurations of anomalies gives the possibility to take into account this effect and provide realistic values of anomalies [e.g., Koulakov et al., 2009b]. In Figure 8, we present two examples of such modeling for section A2-B2. In this case, the anomalies are defined as prisms with fixed shapes in the direction across the section and a predefined thickness (50 km in our case). We have tried many different amplitudes and shapes of anomalies to achieve the maximal semblance between the results of synthetic and observed data inversions for this section. In the first model (Figure 8a), there is a high-velocity slab with the amplitude of 13% (which does not affect the result), a low-velocity transition zone roughly corresponding the double seismic zone (−13%), and a high-velocity zone in the mantle wedge overlaying the slab (13%). In the mantle wedge, we define two anomalies representing the feeding paths of the volcanoes with the magnitudes of −11% and −6%. Second model (Figure 8d) is similar, except for the low-velocity layer above the slab which is twice thinner and more intensive compared to the first case. The results of the reconstruction for the P and S models for both models are shown in Figures 8b, 8c, 8e, and 8f. It can be seen that the restored models for the both models are generally similar to the results of observed data inversion for the same section shown in Figures 4 and 5. In particular, for the upper part of the model, none of these models provides the information on the high-velocity slab. At the same time, there is some difference in the lower partition of the area: the model with thinner low-velocity layer provides larger positive anomalies. From the better semblance of the first case with the real data inversion, we can conclude that the model with thicker low-velocity layer is more realistic for this profile. Based on

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**Figure 8.** Two synthetic tests with realistic distribution of anomalies in section A2-B2. (a and d) Configurations of the synthetic models. The numbers indicate the anomaly values. (b, c, e, and f) Reconstruction results. The black contours delineate the shapes of synthetic patterns. The dots depict the seismicity, same as in Figures 4 and 5.
this test, we can conclude that the amplitudes of anomalies in the slab and in the mantle wedge appear to be very large, more than 10%.

5. Discussion

5.1. Comparison With Previous Seismic Models

Although, the isotropic velocity anomalies were not the main focus of this study, it is interesting to compare our results with previous seismic models of Hokkaido constructed recently by other authors. Note that there is a serious misfit between some of these models and it is important to know which of these are more consistent with our models. We note a moderately good correlation between our results and the $P$ and $S$ velocity models of the same area by Katsumata et al. [2006], particularly with regard to the shallow structures down to a depth of 30 km. At greater depths, our models differ along the southern coast of Hokkaido. More differences are observed when comparing our models with the isotropic anomalies and anisotropic patterns in the model by Wang and Zhao [2009]. This difference can be partly explained by those authors’ predefined high-velocity slab, which strongly affected the final results. In addition, in our opinion, they used too small damping for the anisotropic parameters. In their model, the anisotropy orientations may be perpendicular in neighboring nodes, especially at shallow depths. However, it is clear that due to the trade-off between isotropic and anisotropic parameters, as well as due to source location uncertainty, the available data cannot enable so high resolution, making possible restoring sharp variations of anisotropy from node to node.

We note nearly perfect correlation between our model and the $P$ and $S$ velocity models of shallow layers below southern Hokkaido developed by Kita et al. [2012], which were based on much denser data coverage than was our model. As in our model, Kita et al. [2012] identified an elongated high-velocity anomaly at shallow depths along the Hidaka Belt. Starting at a depth of $\sim$30 km, this anomaly grades downward into a large isometric anomaly that extends eastward, as in our results. Kita et al. [2006, 2010] reported similar features but with poorer resolution.

Reports of a similar structural configuration were the result of numerous seismic reflection studies that were mentioned in the introduction and described in the compilation by Ito [2000]. These studies revealed a clear eastward dip of the western (Japan) block underneath the eastern (Kuril) block, which correlates with a system of thrust faults in the Hidaka Belt (Figure 9). Ito [2000] also reported strongly divergent west dipping structures of the lower crust, which he referred to as “shark’s jaws” and interpreted as evidence of delamination beneath the Kuril fore arc.

There are other nonseismic observations revealing the same contrasting feature beneath the Hidaka Belt. For example, a strong positive Bouguer gravity anomaly aligned approximately N-S corresponds very well to the location of the high-velocity anomaly revealed in this work and in previous studies [Arita et al., 1998]. This high gravity indicates that this pattern is due to the presence of anomalously high-density rocks, which is an important basis for the delamination concept discussed later in this paper.

5.2. A Low-Velocity Slab?

One of the striking aspects of our findings is that the slab shown in section A2-B2 is evident as a low-velocity anomaly overlain by a high-velocity anomaly in the mantle wedge. When studying subduction zones, we are accustomed to working with the opposite configuration: a high-velocity slab and a low-velocity mantle wedge. This preconceived notion is so prevalent that a few authors (including the first author of the present study, in Koulaev et al. [2006]) included a predefined high-velocity slab in their models to enhance the resulting images. In particular, Wang and Zhao [2009], in their tomographic study of...
Hokkaido, a priori included a high-velocity slab, which preserved the positive value of the anomaly in the final solution. Based on our findings, we conclude that at least in the Hokkaido region, such a strategy is not valid. We were strongly intrigued by this conclusion, and we therefore assigned dozens of sets of values to the inversion parameters (damping and source weighting). In all cases, the slab in this section consisted of a low-velocity anomaly; consequently, this structural interpretation appears to be robust.

The stability of the low velocities within the slab was also tested using synthetic modeling with realistic distributions of seismic patterns (Figure 8). It was shown that this structure might be caused by the existence of a low-velocity layer lying over the high-velocity slab. Unfortunately, the available data configuration cannot provide any information on the main high-velocity slab, which should exist there as follow from global and regional tomography models [e.g., Bijwaard et al., 1998; Koulakov et al., 2011]. The locations of events along the upper surface of the slab allow neither revealing the high-velocity slab nor estimating the thickness of the low-velocity layer. This statement is demonstrated by reconstructions of two synthetic models in Figure 8 with considerably different thickness of the low-velocity layer that give similar reconstructions.

We present our interpretation of this unusual configuration in Figure 10. We propose that the main explanation for the low-velocity anomaly in the slab is a high amount of fluids generated by phase transitions of hydrous mineral phases in the descending slab [e.g., Poli and Schmidt, 1995; Chepurov et al., 2012]. Water is accumulated in the oceanic lithosphere due to various processes taking place during its evolution from spreading to subduction. Approximately 90–95% of water that is released from the slab during the subduction is contained in hydrous minerals, such as serpentine (13 wt % H$_2$O), lawsonite (11 wt % H$_2$O), phlogopite (2 wt % H$_2$O), amphibole (2 wt % H$_2$O), chlorite, talc, and zoisite [Peacock, 2003]. Each of these minerals is metamorphosed under different P-T conditions that lead to continuous dehydration of rocks in the crustal and mantle parts of the slab in a wide range of depths starting from 10 km to ~300 km [Peacock, 2003]. Typical reactions in the subducting oceanic crust correspond to the consequent facies, such as the zeolite and prehnite-pumpellyte (10–30 km), blueschist (30–70 km), and eclogite (80–120 km) [e.g., Winter, 2001]. In the subducting slab, amphibole releases water at approximately 75 km depth. Phlogopite releases water at approximately 200 km depth. Transformation of lawsonite occurs at approximately 300 km depth, and it is the deepest dehydration reaction occurring in the slab. Dehydration of serpentine, which is the main source of water, occurs along the upper limit of the mantle part of the slab and is probably the main reason of the origin of double seismic zone in the subducting slab in a depth range from ~50 to 150 km [e.g., Peacock, 2001]. The details of the water-related processes during subduction are summarized in numerical simulations by Arcay et al. [2005].

Normally, in most subduction zones, most of the fluids escape from the slab due to dehydration at depths of 80–120 km. These fluids ascend through the mantle wedge and lower the melting temperature that leads to the origin of arc volcanoes. Typically, we observe arc volcanoes above the seismicity clusters in the same depth intervals marking the maximum water release [e.g., Poli and Schmidt, 1995].

We assume that beneath Hokkaido in section A2-B2, the ascent of fluids from the slab is blocked by a high-velocity body, which most likely consists of rocks that are more rigid than the mantle material in the mantle wedge. We will speculate about the origin of this high-velocity body in the following paragraphs. We propose that water is conserved in the slab and entrained downward to the lower edge of the high-velocity body, as shown in Figure 10. Then, fluids start to ascend from a depth of approximately 150 km, which is deeper than in the case of other volcanic centers on Hokkaido. We can trace the path of the ascending fluids beneath the CVG along the prominent low-velocity anomaly, which is clearly observed in both the P and S velocity models.

Note that the structures in sections 1 and 3 differ significantly from those in section 2. In section 3, the red slab and blue mantle wedge are evident, but the corresponding anomalies are much weaker and less prominent than those in section 2. In section 1, no prominent structures located within the double seismic zone are evident. Below 100 km depth, we observe alternation of positive and negative anomalies. For the shallow part of the slab, we observe a low-velocity pattern, which, however, is likely due to the downward smearing of slow anomalies in the mantle wedge. Based on these differences between neighboring sections, we conclude that the high-velocity body in the mantle wedge in section 2 is a local thin structure that is not evident in the other sections.
5.3. High-Velocity Anomaly Beneath the Hidaka Belt: Delamination or Seamount?

Beneath the Hidaka Belt in both the $P$ and $S$ velocity models, we observe a prominent high-velocity anomaly that has an elongated shape at shallower levels and becomes isometric and larger at deeper levels. The same anomaly was identified in several passive and active seismic studies discussed above. In Figure 9, we present the distribution of the main linear reflectors identified by compilation of active source seismic data by Ito [2000], imaged over $P$ velocity anomalies in section A5-B5. It can be seen that most reflectors are located to the left, to the right, and above the prominent positive anomaly. Inside this anomaly, the reflectivity does not have any predominant orientations. Note that the reflectors to the right and to the left sides of the high-velocity anomaly tend to deepen toward the location of this anomaly. In this section, we also plot the interpretation provided by Ito [2000], showing the shape of the delamination wedge and dipping the lower portion of the high-velocity body.

Section A2-B2 shows that this high-velocity anomaly deepens along the upper border of the slab, just above the upper seismicity belt. This anomaly appears in nearly identical fashion in the $P$ and $S$ models, which may indicate the reliability of this finding. We have stated that this configuration is highly atypical of subduction zones, and we will now propose two possible explanations for the presence of this high-velocity anomaly.

Based on the regional structural framework, we note that the Hidaka Belt corresponds to the junction on Hokkaido between the Kuril and Japan arcs and extends southeast (dotted line in Figure 1a) to the point where the corresponding trench abruptly changes its orientation. The correspondence between this location and the narrow high-velocity anomaly above the slab supports the hypothesis that this body might originate in the Pacific Plate and subduct underneath Hokkaido. It is known that large heterogeneities on the incoming plate, such as seamounts or remnant arcs, might strongly change the shape of the trench. The presence of a seamount or a remnant arc beneath Hokkaido is not as common as in southern Japan [e.g., Zhao et al., 1997] but is not excluded by certain geologists. For example, Ueda [2005] concluded that "the Lower Cretaceous Iwashimizu complex in the southern Kamuikotan zone, (Hokkaido) Japan, was formed by accretion of seamounts, which subducted to the lawsonite-albite to blueschist facies." The possible role of seamounts in affecting the present subduction configuration was also discussed by Zhao et al. [1997]. Evidence for possible accretion of a subducted remnant arc and its role in the evolution of the subduction beneath NE Japan is discussed by Ueda and Miyashita [2005]. However, it is still unclear whether the subducting seamount material behaves as a high-velocity anomaly. Furthermore, the expected total amount of material that might be brought to the mantle wedge is not expected to be as large as that observed in our tomographic images. Thus, after considering these pros and cons, we conclude that the seamount hypothesis is not highly likely.

Another scenario that may explain the particular structures beneath the central part of Hokkaido is based on the idea of delamination, and this idea is more widely accepted by the scientific community. As was stated earlier, the concept of delamination of the lower crust below Hokkaido was proposed in a number of papers based on the results of reflection seismic profiling (see compilation in Ito [2000]). These authors propose that westward displacement of the Kuril fore arc caused the Japan fore arc to underthrust eastward at dip angles of 40°–50°. A particular behavior of reflectors in the lower crust may indicate the delamination of the lower mantle material and its possible descent to the mantle wedge. In particular, Ito [2000]...
[2000] stated that “the upper half of the lithosphere (upper crust + upper portion of the lower crust) thrusts westward on the northeast Japan arc, whereas the lower half (lower portion of the lower crust + upper mantle) descends down.” Similar conclusions follow from another reflection seismic study by Tsumura et al. [1999], who proposed that “the Hidaka Collision Zone represents an active model for continental growth by arc accretion with delamination of lower arc crust.” This concept is also supported by geological observations indicating the peak metamorphic grades exposed in the Hidaka Belt [Osanai et al., 1986].

The delamination mechanism is one of the most often-cited explanations for lithosphere recycling during collision. In the case of collision of large continental plates, this mechanism might be responsible for the sinking of large volumes of mantle lithosphere, as was proposed for the regions of Vrancea [Koulakov et al., 2010], Caucasus [Koulakov et al., 2012], Pamir [Koulakov, 2011], and Tien Shan [Zabelina et al., 2013]. In subduction zones, collisional processes may cause the delamination at much smaller scales. For example, in the central Andes, there is a significant shortening caused by the displacement of the Brazilian shield toward the subduction complex. The possibility of delamination in this case was first proposed by Kay and Kay [1993] and was then confirmed using numerical modeling by Babeyko and Sobolev [2005] and Sobolev et al. [2006]. Here we adopt the same explanation for the geodynamic setting of Hokkaido.

The delamination scenario in the Hokkaido area is schematically presented in Figure 11. In the E-W oriented section A5-B5, the Hidaka Belt is located between areas of different types of crust. It is transitional between oceanic and continental crusts, with a predominantly mafic part to the west and thicker felsic crust to the east. Because of the convergent displacements, there is considerable shortening leading to thickening of the crust associated with mountain building. As was demonstrated by Sobolev et al. [2006], plunging of the lower mafic crust material may lead to a phase transformation to eclogite, which is depicted in Figure 11 in yellow. The accumulation of a heavy eclogitic drops below the shortening zone continues until it reaches a critical mass. Then, it descends and entrains part of the mantle lithosphere into the mantle wedge. This stage appears to be identical to the case in the central Andes described by Sobolev et al. [2006], who numerically simulated the detachment of the delaminated material and its further sinking along the upper surface of the slab. This scenario may explain most of the observations in Hokkaido. Underthrusting revealed by the reflection seismic surveys is a clear signature of crustal shortening and thickening. Positive velocity anomalies in our and previous tomographic studies indicate the areas of accumulation of mafic crust and their possible transformation to eclogite. Linear positive gravity anomalies [Arita et al., 1998] indicate high-density patterns that may be associated with heavy eclogite. Finally, the narrow positive anomaly above the subducting slab might indicate the trace of the descent of the delaminated material. As described by Sobolev et al. [2006], the eclogitic drops may trigger the delamination, but their total volume might be minor compared to the amount of the entrained mantle lithosphere material. The total volume of delaminated material above the slab, as estimated from the size of the high-velocity anomaly in our tomographic model, is compatible with that simulated by Sobolev et al. [2006] in their numerical modeling of the central Andes. After evaluating all of these arguments, we conclude that delamination is the most plausible explanation for the tectonic regime of Hokkaido.
5.4. Anisotropic Structures Reveal Flows in the Mantle Wedge?

Anisotropy is an important parameter providing information on the dynamic processes in the crust and in the mantle. In the crust, the anisotropy is mostly controlled by the orientations of regional tectonic structures [e.g., Bokelmann, 1995]. In particular, fault systems usually have prominent anisotropic properties, specifically high velocities oriented along faults [e.g., Zhang and Schwartz, 1994]. Anisotropy in the mantle is due to convection flows and is mostly explained by the orientations of olivine crystals [e.g., Karato and Wu, 1993; Fouch and Rondenay, 2006]. However, the relationship between the anisotropy and mantle flows is not always unambiguous. For example, in certain cases of high amounts of fluids and melts, the anisotropy may be orthogonal to mantle flows. The petrologic basis of such behavior was studied in laboratory experiments by Jung and Karato [2001].

Anisotropic features have been studied in many subduction zones. It is interesting that in most cases, the geometry is controlled by the orientation of trenches, but there are many cases of both trench-parallel and trench-perpendicular anisotropy. If the flow is trench perpendicular, such anisotropy is explained by A-type olivine in the return flow in the mantle wedge generated by the descending slab [e.g., Audoine et al., 2004]. In cases of trench-parallel anisotropy, the explanation is more complex but generally reduces to two hypotheses. The anisotropy can be related to trench-parallel flows caused by heterogeneities in the slab shape [e.g., Russo and Silver, 1994; Anderson et al., 2004]. Alternatively, trench-parallel anisotropy may be explained by B-type olivine fabrics [e.g., Jung and Karato, 2001; Kneller and van Keken, 2007].

Estimates of anisotropic structures have been performed in Japan and particularly in Hokkaido. Based on shear wave splitting analysis, Nakajima et al. [2006] estimated the anisotropy at many stations in northern Honshu and Hokkaido. They found a clear trend of mostly trench-parallel orientations of anisotropy in a narrow zone of fore arc and trench-perpendicular anisotropy in the remaining parts of islands. In general, the anisotropy observations by Nakajima et al. [2006] in Hokkaido appear to be consistent with our result in Figure 2. However, one problem with the shear wave splitting method is that it does not provide depth information regarding the anisotropic bodies.

An anisotropic tomography inversion for Hokkaido was previously performed by Wang and Zhao [2009]. Unfortunately, we do not note any correlation between our model and their results. The orientations of anisotropy in their model appear to vary from point to point and appear to be chaotic, and we cannot identify any regular feature amenable to reliable interpretation.

In our anisotropy results shown in Figure 2, we note several features that can be associated with tectonic structures of the region. At shallower depths (10 and 30 km), we note certain clear, prominent crustal structures. In central-southern Hokkaido, we observe clear N-S oriented anisotropy that corresponds to elongated tectonic structures of the Hidaka Belt. In the Kuril fore arc in the northeastern part of the island, we observe E-W oriented anisotropy in the crust. This finding is consistent with the displacement in this area caused by the oblique subduction. This southwestward displacement causes strong shear stresses and strike-slip faulting, which may control the orientation of the anisotropy. Similar orientations of anisotropy are evident in western Hokkaido; however, based on the results of the checkerboard test, the anisotropy reconstruction in this area is not highly robust. In northern Hokkaido, we observe a clear transition from the N-S oriented anisotropy, which is most likely associated with northward continuation of the latitudinally oriented structures in the Hidaka Belt, to E-W anisotropy in northwestern Hokkaido. In summary, the E-W anisotropy in the crust is observed in both the Kuril and NE Japan parts of Hokkaido beyond the latitudinally oriented collision zone, where the orientation of anisotropy is N-S.

In the mantle at a depth of 60 km, the anisotropy orientations appear to be similar to those in the crust, and this similarity might be due to the limited vertical anisotropy. However, based on the checkerboard test, the opposite directions of anisotropy in the crust and mantle can be clearly distinguished based on our algorithm and available data. The anisotropy distribution in the mantle appears to be fan shaped and tends to split into three directions: westward in the western part, northward in the center, and eastward in the eastern part. If we accept that the anisotropy is oriented along the flow direction, this configuration might be caused by complex 3-D orientations of flows in the mantle wedge. We propose that there are two return flows indicated with red arrows in 60 km section in Figure 2. Such deviation of flows from the trench-perpendicular configuration (which could be expected in the 2-D case of normal subduction) might be caused by a strong obstacle in the mantle wedge. We propose that in our case, the flows surround the
column of the delaminated materials below the Hidaka Belt, as shown in Figure 9. The eastward and westward flows proceed toward the eastern and western volcanic groups of Hokkaido. This action may bring additional elements (e.g., more water) to the volcanic areas, which would explain their strongly explosive behavior and the resulting development of large calderas. The central volcanic zone appears to be “protected” by a vertical column of delaminating material; thus, the behavior of these volcanoes is fully controlled by fluids released from the slab at great depths. This may explain the character of volcanism in the central volcanic group compared to that of the eastern and western groups.

It should also be mentioned that the central volcanic group is associated with a large low-velocity anomaly in the mantle and thus indicating the deep source of the magmatic activity. The eastern group is also associated with the negative anomaly in the mantle, but this anomaly appears to be much weaker. In addition, at a depth of 30 km, a prominent linear low-velocity anomaly evidently connects the southern edge of the island with this group. This may support our hypothesis that there is a lateral flow, possibly below the Moho interface, that brings the material to create the magmatic activity. The western group does not display such a pattern, most likely because it is located outside our area.

6. Conclusions

We have presented new anisotropic P and isotropic S models of the crust and upper mantle beneath Hokkaido based on tomographic inversion of arrival time data from the JMA catalog. The isotropic models generally exhibit features similar to those observed in previous studies of the region. The novel feature of this model is the anisotropic imaging, which exhibits regularity and depicts well the major tectonic units of Hokkaido. It is important that the inclusion of the anisotropy has significantly improved the data fit compared to the regular isotropic inversion. Based on the analysis of the isotropic and anisotropic patterns in our model, we propose several geodynamic scenarios to explain several tectonic phenomena in Hokkaido:

1. Delamination: Beneath the Hidaka Belt, we observe a high-velocity anomaly that starts as a linear anomaly in the crust and then follows above the upper limit of the slab down to ~130 km. We propose that this anomaly represents the delamination of the mafic crust and mantle lithosphere beneath the Hidaka area, where strong crustal shortening and thickening occur. Heavy delaminating material, most likely containing dense eclogite, sinks rapidly along the subducting slab.

2. Low-velocity slab: One of the sections shows an unusual configuration for a subduction zone: the low-velocity slab is overlain by a high-velocity body in the mantle wedge. We propose that reduced velocities in the slab might be related to high amounts of fluids and melts released from the oceanic plate due to phase transitions. However, the overlying delaminated material, which is imaged as a high-velocity anomaly, prevents the upward escape of fluids from the slab. This creates much more prominent low-velocity anomalies in the slab than is typically observed.

3. Mantle flows imaged using anisotropy: The anisotropy in the mantle is evident as fan-shaped structures that start in southernmost Hokkaido and split into three different directions. We propose that these represent lateral flows in the mantle wedge that are strongly affected by a column of delaminated material beneath central Hokkaido. The westward and eastward flows proceed toward the two volcanic groups in western and eastern Hokkaido. It is possible that these flows bring additional material (for example, water), which is responsible for the caldera-forming eruptions characterizing both of these volcanic groups.

Acknowledgments

Travel time data for this study were obtained from the Japan Meteorological Agency. I.K. and E.K. are supported by the Russian Scientific Foundation (grant 14-17-00430). The authors extend their appreciation to the Deanship of Scientific Research at King Saud University for funding the work through the research group project RG-1435-027.

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