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Crust and uppermost mantle structure beneath central Japan inferred from receiver function analysis

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We apply the receiver function method to estimate the structure of the crust and the uppermost mantle at an area that traverses central Japan including the Niigata-Kobe Tectonic Zone (NKTZ). The resultant receiver function images show clear seismic discontinuities, such as the subducting Philippine Sea plate, the Moho in the overriding plate, and other discontinuities inside the crust around the NKTZ. We also address station corrections for shallow structures using a synthetic receiver function. Crustal discontinuities seem to be complicated at the south side from the northern limit of the NKTZ. The dip of the discontinuities changes around the Atotsugawa active fault located in the NKTZ. The Moho discontinuity in the overriding plate is continuous and gradually dips to the south. The depths of the Moho discontinuity in the receiver function image exceed 40 km at the southern part of the profile line, and are 5–10 km deeper than that indicated by an explosion analysis of the same profile line. It seems that the differences between the estimated depths obtained by the two methods indicate complicated structures around the Moho discontinuity.

Key words: Crustal structure, the Niigata-Kobe Tectonic Zone, receiver function analysis.

1. Introduction

Inland earthquakes occurring at active faults near urban areas threaten our safety, even if their magnitudes are middle-sized. Predicting these earthquakes is important to mitigate future disasters. However, it is difficult to observe one cycle of active faults because their recurrence intervals are long, probably extending from several hundreds to several thousands of years. Thus, the mechanism is understood less than the mechanism of plate boundary earthquakes.

The Niigata-Kobe Tectonic Zone (NKTZ) is a high strain rate zone extending from Niigata to Kobe, which was recently identified from analyses using a spatially dense GPS network (e.g., Sagiya *et al.*, 2000) (Fig. 1). This suggests that the zone plays important roles in concentrating and accumulating strains in the land area of the Japanese Islands as well as other major tectonic lines. Many models have been proposed to explain the causes of the high strain rate at the NKTZ (e.g., Shimazaki and Zhao, 2000; Iio *et al.*, 2002; Hyodo and Hirahara, 2003; Yamasaki and Seno, 2005). This zone is interpreted to be the inland plate boundary of the Japanese Islands (e.g., Heki and Miyazaki, 2001; Miyazaki and Heki, 2001) or an internal deformation zone forming the Japanese Islands (Mazzotti *et al.*, 2000; Iio *et al.*, 2002; Yamasaki and Seno, 2005). In spite of such differences of interpretation, most researchers think viscosity heterogeneities located deeper than the seismogenic zone contribute to the high strain rates. Therefore, understanding the structure at the NKTZ is important to identify the mechanism of strain accumulation in the Japanese inland area.

The Atotsugawa fault is a right-lateral strike-slip fault located in the NKTZ. This fault is one of the most active in Japan. Seismic activities along this fault are not uniformly distributed along the fault (e.g., Ito and Wada, 2002). They are active at the eastern and western part of the fault. On the other hand, the shallow part of the central area of the fault has low seismic activity. A M 7 class earthquake also occurred along this fault in 1858. Furthermore, it is reported that creep-like movements of about 1.0 to 1.5 mm/yr were detected at the middle of the Atotsugawa fault (Geological Survey Institute, 1997). Kato *et al.* (2006, 2007) identified a large heterogeneous structure in the uppermost crust using the seismic tomography method and discussed the relationships between local crustal heterogeneity and seismic and geodetic activities. The heterogeneous structure and heterogeneous distribution of seismicity are the keys to understand the stress and strain concentration process. This area is also important for identifying the generation mechanisms of inland earthquakes.

Geophysical joint observations, which include seismic, geodetic, and electromagnetic observations, at the NKTZ including the Atotsugawa fault have been conducted by the Japanese University Groups since 2004 (The Japanese University Group of the Joint Seismic Observations at NKTZ, 2005). The purpose of seismic observations and research in this project is to understand seismicity and seismic structure. The groups installed 73 seismic stations—63 stations

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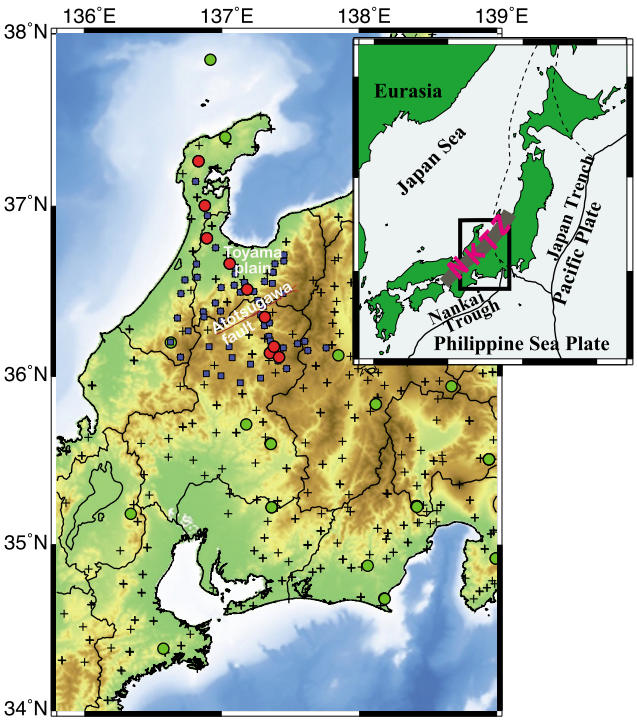

Fig. 1. Locations of seismic stations with a telemetry system in and around the Niigata-Kobe Tectonic Zone (NKTZ in the inset). The short-period and middle-period seismic stations deployed by the Japanese University Group of the Joint Seismic Observation are shown by blue squares and red circles, respectively. Telemetered short-period and broadband seismic stations, which were deployed by Earthquake Research Institute, University of Tokyo, National Research Institute for Earth Science and Disaster Prevention, and Japan Meteorological Agency, are shown by plus symbols and green circles, respectively. The Atotsugawa fault is indicated by a brown line. The research area is shown in the inset.

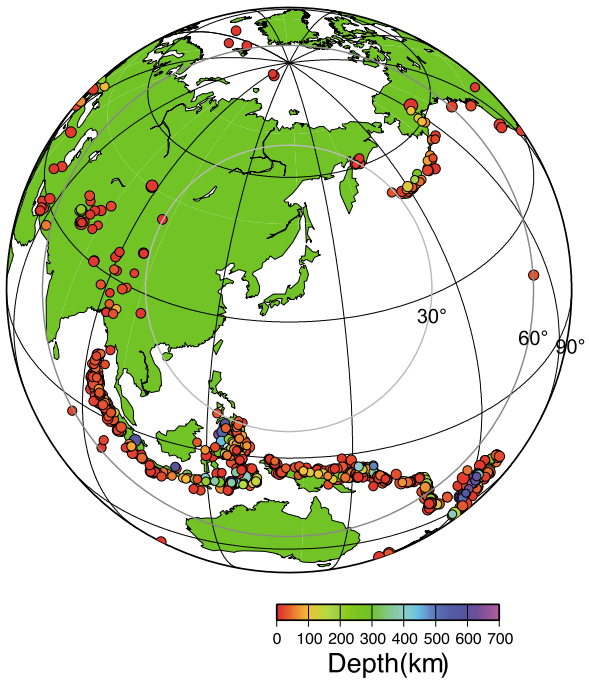

Fig. 2. Hypocenter distribution used in this study. Colors change by hypocenter depth.

using a telemetry system and 10 stations using a portable recorder. The seismic stations are distributed within a 100-km rectangle that includes the Atotsugawa fault (Fig. 1).

In this study, we use the dense seismic stations and apply the receiver function method using teleseismic waveform data to determine the structure of the crust and the uppermost mantle at the NKTZ. Receiver functions calculated by the multiple-taper correlation method can resolve up to several Hz (e.g., Park and Levin, 2000). The receiver functions with the high frequency band image fine seismic structures with short wavelengths of several kilometers. Therefore, we investigate not only geometries of the subducting Philippine Sea plate and the Moho in the overriding plate discussed in previous studies (Yoshimoto *et al.*, 2004; Shiomi *et al.*, 2008) but also seismic discontinuities in the crust.

Other groups in the Joint Seismic Observations at the NKTZ are studying seismicity, seismic tomography (Nakajima and Hasegawa, 2007), source mechanism, reflection and scattering studies, shear-wave splitting (Iidaka *et al.*, 2009), low-frequency earthquake, and attenuation structure using data obtained by this project. The results of these analyses will contribute greatly to reveal the mechanisms of inland earthquakes.

2. Data and Method

Seismic stations with telemetry systems operated by the Japanese University Groups of the Joint Seismic Observations, and Earthquake Research Institute, University of Tokyo (ERI), National Research Institute for Earth Science and Disaster Prevention (NIED), and Japan Meteorological Agency (JMA) are used. The time period of data at the networks of ERI, NIED, and JMA is from August 2002 to April 2007. Seismic data from the network operated by the Japanese University Groups of the Joint Seismic Observations are used for the period from January 2005 to April 2007. At most of the seismic stations, short-period seismographs (natural frequency is 1 Hz) are used. Middle-period seismometers (frequency band is 0.033 Hz to 50 Hz) and broadband seismometers (lower limit of frequency band is 0.01 Hz or less) are set up at several seismic stations, which are shown by red and green circles, respectively (Fig. 1). We used events with magnitudes larger or equal to 5.5 and epicentral distances between 30 and 90 degrees reported in the PDE or QED catalogue by the United States Geological Survey (USGS). Furthermore, we selected waveform records with a good signal-to-noise ratio around the direct P wave and its reverberation from events. The total number of events used in this study is 961 (Fig. 2).

The receiver function method is used to estimate the crustal structure beneath each station (Langston, 1979; Ammon, 1991). Receiver functions are time series deconvolved ground motions of horizontal (radial) component by vertical component. From this process, we can extract phases of converted Ps phases at any seismic discontinuity beneath each station. Here, we apply the receiver function analysis using the multiple-taper correlation method for the deconvolution process (Park and Levin, 2000). We fixed the time-bandwidth product and the number of the used Slepian

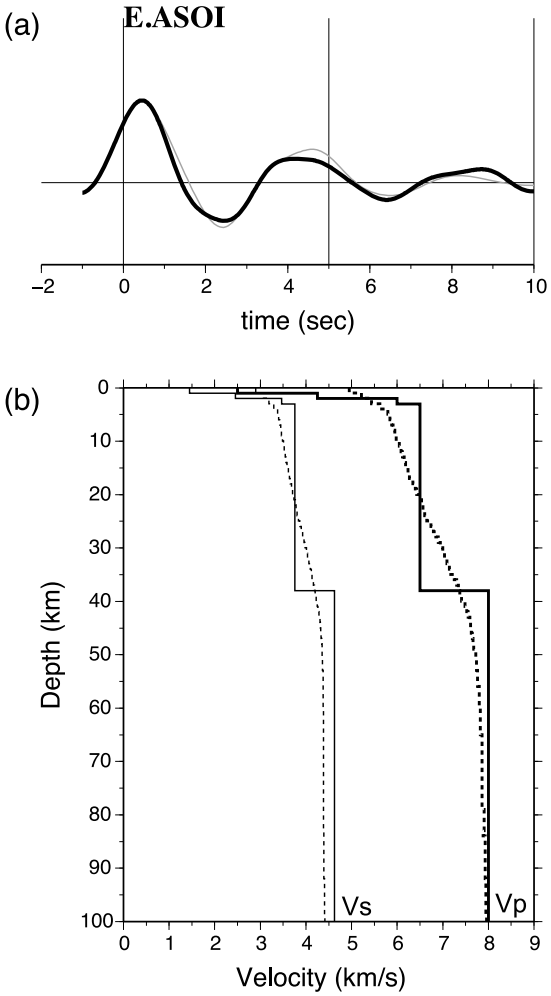

Fig. 3. (a) An example of the observed (thick black line) and synthetic receiver function (thin gray line) of E.ASOI station located near the Toyama plain. Both traces are normalized by peak amplitudes of direct P wave. (b) Structure models from this synthetic trace and JMA2001 (Ueno *et al.*, 2002) are also shown by solid lines and dashed lines, respectively. Thick and thin lines indicate P wave and S wave velocity, respectively.

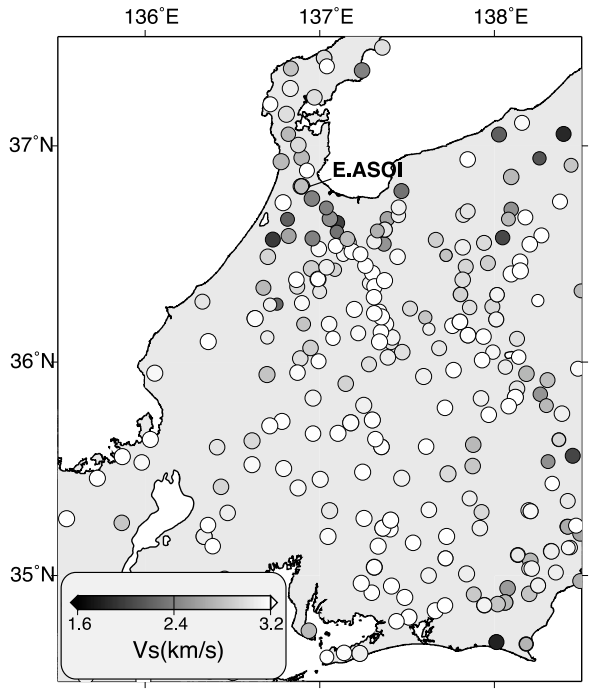

Fig. 4. Average S-wave velocities from ground surface to a depth of 5 km used for station correction of receiver functions are shown by a gray scale.

tapers for the multiple-taper to be 2.5 and 3, respectively.

Receiver function traces were converted from differential time to depth at each station after low-pass filtering with a squared cosine-taper with a cut-off frequency of 3 Hz. The observed dominant frequency of the traces is roughly 1–1.5 Hz. We checked the similarities among different types of sensor at this frequency, although broadband seismometers are useful for analyzing at frequencies lower than 1 Hz. The traces are stacked at 20-degree bin in back-azimuth with moving windows overlapping at 10-degree intervals. The amplitude data of stacked-traces are projected at the cross-section of the averaged incident angle and direction. In the time-depth conversion process, the one-dimensional velocity model JMA2001 (Ueno *et al.*, 2002; dashed line in Fig. 3(b)) was used.

3. Station Correction for Depth Conversion

We applied station correction to compensate for the altitude of seismic stations and the effects of the low-velocity sediment layer beneath each station.

To correct for the latter effect, we estimated simple velocity structures beneath each station presented by Igarashi (2009). First, we calculated many synthetic receiver function traces using the reflectivity algorithm (Levin and Park, 1997) in advance. Multiple reflected phases are considered for this calculation. The structural models consist of four layers (three seismic discontinuities), representing sediment layer, upper crust, lower crust, and homogeneous mantle. Seismic velocities of the sediment layer gradually increase from the ground surface to the top of the crustal layer. Velocities in crustal layers are constant. P wave velocities in the top of the sediment layer and two crustal layers were set to 1.0–6.0 km/s, 5.0–7.0 km/s, and 6.0–7.5 km/s at 0.5 km/s intervals, respectively. The P wave velocity was fixed in the mantle at 8.0 km/s. V_p/V_s ratio assumed was to be 1.73. Boundary layers were changed at 1-km intervals. We considered positive velocity step, and adapted structures with an average P wave velocity of 6.0–7.0 km/s from the ground surface to a depth of 60 km. The epicentral distance for the calculation was assumed at 45 degrees, which is the average distance of events used in this study.

On the other hand, the observed receiver function traces with a low-pass filter at 1 Hz were stacked without considering back-azimuth or epicentral distance. Finally, we searched for the best-correlated velocity structure model between observed trace and synthetic traces. We set the length of the time window at 10 seconds from direct P arrival. Observed receiver function, best-correlated synthetic function traces, and structure model at E.ASOI station are shown in Fig. 3. The velocity structures from ground surface to a depth of 5 km were replaced if the averaged S wave velocity in these depth ranges was below 3.1 km/s, which is slower than that of the JMA2001 structure model.

In this study area, stations around the Toyama plain and some localized areas showed extremely low velocities of 1.7–2.5 km/s at *S* wave (Fig. 4). If the average *S* wave velocity is 1.7 km/s, the estimated depths of discontinuities become about 6 km shallower than that of the JMA2001 structure model in depths of several tens of kilometers. Although this correction may be too simple to fit a receiver function, the results are consistent with the location of the low *P* wave velocity layer near the ground surface estimated from a seismic experiment with controlled sources (Iidaka *et al.*, 2003), *S* wave velocity structure estimated from ambient seismic noise (Nishida *et al.*, 2008), and the structure models to basement boundaries around the Toyama plain (e.g., Sutou *et al.*, 2004; Senna *et al.*, 2005).

4. Results

We show the cross-sections of receiver function traces along the profile lines in Fig. 5. Three profile lines in the northwest-southeast direction traverse the island arc of Japan from the south coast to the north coast (Fig. 5(a–c)). They also traverse the NKTZ. Another four southwest-northeast profile lines, which are perpendicular to the

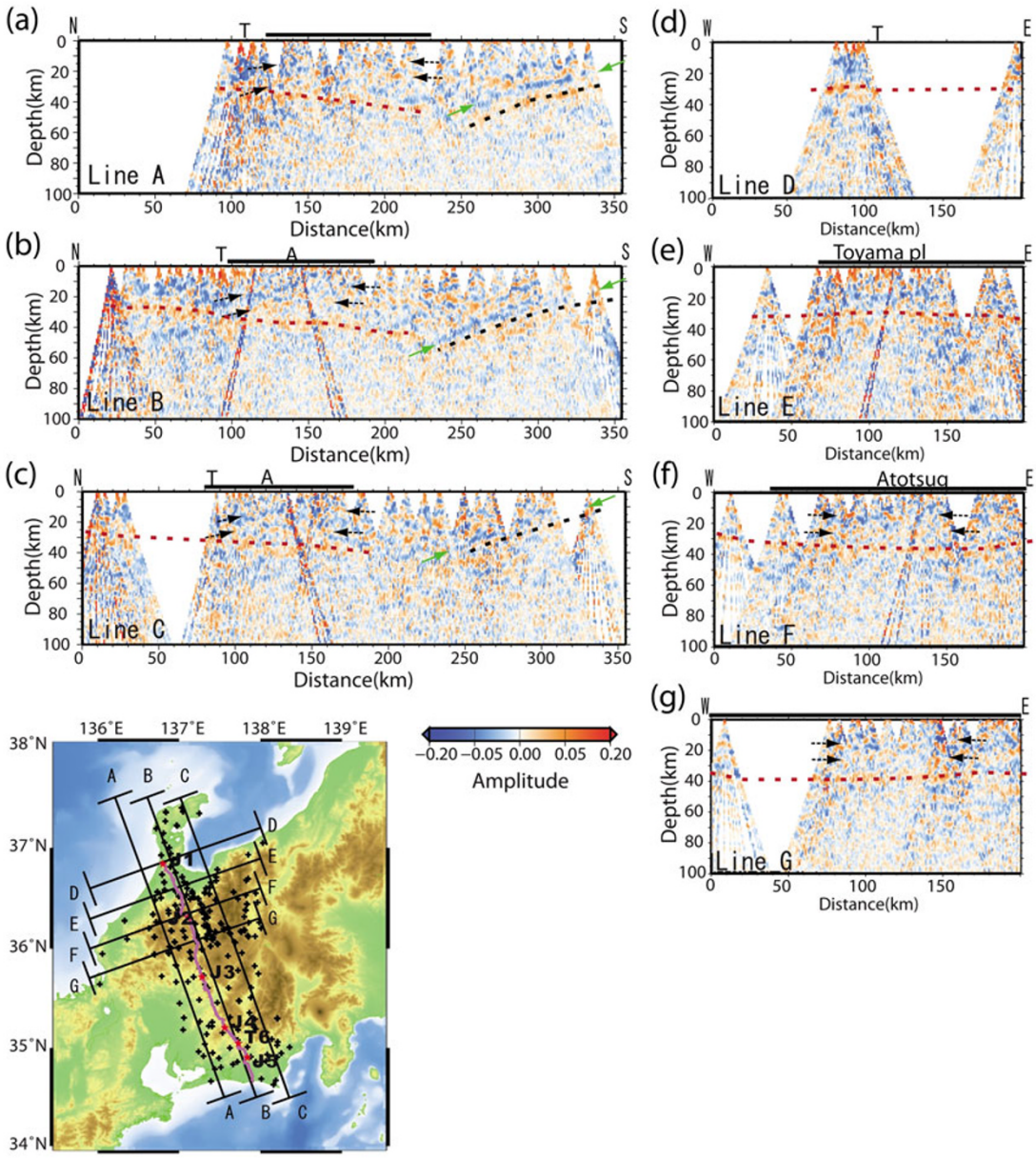

Fig. 5. Cross-sections of depth-converted receiver functions (dominant frequency is about 1 Hz) after correcting the time difference by altitude and low-velocity sediment-layer in shallow parts. (a–c) Profile lines in the northwest-southeast direction traversed the island arc of Japan from the south coast to the north coast. (d–f) Southwest-northeast profile lines, which are perpendicular to the northwest-southeast lines. The solid bar at the top represents the location of the NKTZ. Red and black dashed lines indicate the Moho discontinuity in the overriding plate and the subducting Philippine Sea plate, respectively. Green and black arrows indicate the upper boundary of the subducting Philippine Sea plate and discontinuities in crust of the NKTZ, respectively. Labels 'T' and 'Toyama pl' indicate the Toyama plain, and labels 'A' and 'Atotsug' indicate the Atotsugawa fault. Lines of cross-sections are shown in the inset. The profile line in the 2001 seismic experiment with six explosive sources (Iidaka *et al.*, 2003, 2004) is shown by a purple line. The shot points are shown by red stars. We stacked receiver function traces with a width of 40 km for each line. Stations used for cross-sections are shown by plus symbols.

northwest-southeast lines, are shown (Fig. 5(d–f)). Positive and negative amplitudes of receiver functions indicate velocity increase and decrease downward, respectively. They are shown by red and blue colors in this study. We can detect many discontinuities from these cross-sections.

First, distinct overlying blue and underlying red lines that dip northwest are shown in the distance range of 230–340 km and depths of 10–50 km (green arrows and black dashed lines in Fig. 5(a–c)). We interpret these to be the upper boundary of the subducting Philippine Sea plate and the oceanic Moho of the plate upon comparing them to the results of previous studies (e.g., Iidaka *et al.*, 2003; Kodaira *et al.*, 2004). The layers are characterized by overlying low-velocity oceanic crust and high-velocity mantle.

Next, we estimated the Moho discontinuity in the overriding plate. It is indicated from positive amplitude phases on stacked receiver function traces (red dashed lines in Fig. 5). The discontinuity is consistent with information of the crust-mantle boundary estimated from four-layered structure in Section 3 and the results of Zhao *et al.* (1992). The discontinuity is continuous and undulating. The depth is estimated to be about 25 km at the northern part and about 43 km at the southern part of the profile lines.

We also detected several discontinuities in the crust. The boundaries appear at the south side from the northern limit of the NKTZ (black arrows in Fig. 5). There are continuous phases with depths of 13–20 km and 25–30 km. A shallower discontinuity is just beneath the seismogenic zone of crustal earthquakes. On the other hand, a deeper discontinuity, which seems to be located in the distance range from 100 km to 200 km with depths of 25–30 km at the cross-sections of Lines A and B (Figs. 5(a) and 5(b)), is clear. The boundary is shown by an orange color. The location is consistent with the area from the north end of the NKTZ to the south end of the NKTZ. The boundary is not flat; rather, it is bent. Depth decreases from the north end of the NKTZ to the central part of the NKTZ where the Atotsugawa fault is located. On the other hand, the depth is uniform from the Atotsugawa fault to the south end of the NKTZ. It seems to connect to the southern discontinuity above the Philippine Sea plate (Fig. 5(a)). Furthermore, we can see red and blue stripes dipping northwestward above the Philippine Sea plate (Fig. 5(a–c)).

The cross-section along the Atotsugawa fault is shown in Fig. 5(f). Many horizontal boundaries are found along the Atotsugawa fault. The boundaries are continuous within short distance ranges. The seismic structure in this area obtained by seismic tomography studies is heterogeneous (Kato *et al.*, 2006, 2007). The area seems to have a heterogeneous structure. We will reveal a fine structure from a comparison of seismic tomography and hypocenter distribution in further analyses.

5. Discussion

Several studies propose the possibility that the NKTZ is a plate boundary (e.g., Heki and Miyazaki, 2001). However, we do not interpret the NKTZ to be the plate boundary. It seems to be an internal deformation zone because the continuous Moho and heterogeneous crust are found over the whole area. The high strain rate at the NKTZ might cause the concentration and accumulation of strain in these crustal structures—east-west compression.

Iidaka *et al.* (2003, 2004) surveyed the crustal structure in the Tokai-Chubu region, central Japan, from an explosion analysis using controlled sources. Figure 5(b) is located along the survey line. We compared their seismic structures to the cross-section of receiver function traces (Fig. 6). The boundaries estimated from this study are substantially con-

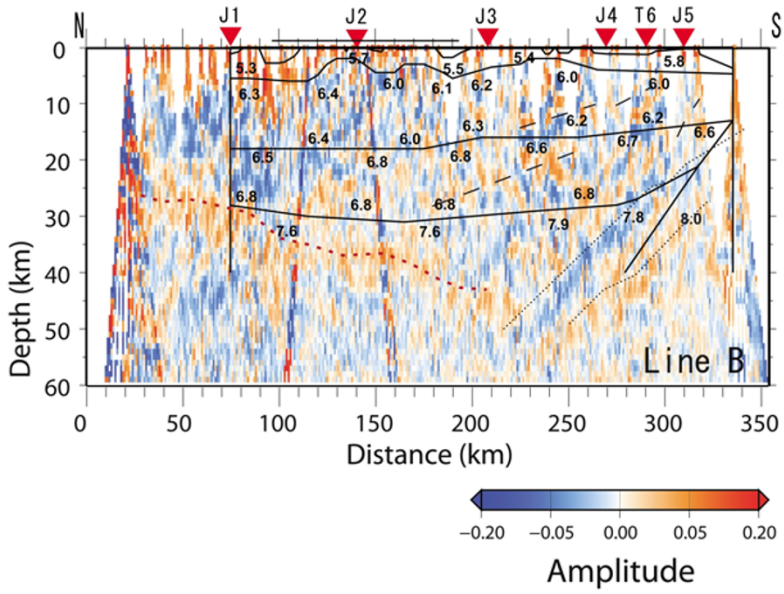

Fig. 6. Comparison of results for the same cross-section between P wave velocity structure model from the explosion analysis (Iidaka *et al.*, 2003, 2004) and depth-converted receiver functions from Fig. 5(b). We changed the horizontal to vertical ratio to 1:3. The red dashed line indicates the Moho discontinuity in the overriding plate estimated in this study. Red triangles, black solid lines, broken lines, and numbers indicate explosive source points, major seismic discontinuities, reflectors in the crust, and P-wave seismic velocities, respectively, by Iidaka *et al.* (2003, 2004). Dotted lines represent the top of the subducting Philippine Sea plate and oceanic Moho by Kodaira *et al.* (2004).

sistent with those of Iidaka *et al.* (2003, 2004), which are the boundaries between upper crust and lower crust, and reflectors in the crust. However, we estimate the depths of the Moho discontinuity is to be 5–10 km deeper than their results, although we can identify weak positive phases that correspond to the Moho discontinuity obtained by Iidaka *et al.* (2003, 2004). The location of the upper boundary of the subducting Philippine Sea plate obtained here does not fit. To explain the discrepancy, we considered the following effects—dipping plate boundary, differences in velocity structure, and differences among analyzed seismic waves.

First, we consider the effects of the dipping plate boundary. In this region, the Philippine Sea plate subducts northwestward, dipping with a dip angle of about 26 degrees from Iidaka *et al.* (2003, 2004) or about 17 degrees from Kodaira *et al.* (2004). The effects of depth errors of the receiver function on the dipping boundary were evaluated by Shiomi *et al.* (2004). The conversion depths were about 3 to 7 km shallower than that of the flat structure model at a depth range of 30 to 50 km in the case of these dip angles. However, the dipping angle, which we estimated from Fig. 5(b), is about 18 degrees, and our image of the plate boundary corresponds to that of Kodaira *et al.* (2004). The difference in the depth of the subducting Philippine Sea plate between our results and Kodaira *et al.* (2004) is about 2 km at maximum. It seems that differences in the relative locations and the dip angle of the boundary plane are small because the conversion points move southeastward at the same time. The conversion points of the Moho discontinuity in the overriding plate might be changed to the similar extent for this effect. However, the depth error is not so large.

Next, we calculated receiver functions from the one-dimensional velocity structure based on the structures of Iidaka *et al.* (2003, 2004) and converted the receiver function traces to the depth section using the JMA2001 one-dimensional structure model. As a result, the maximum difference between the two projections was about 2 km at the Moho discontinuity. The obtained Moho is deeper than that of the explosion analysis model. The depth error caused by the assumed velocity model is not sufficient to explain the observed discrepancy.

It seems that the differences between the depths estimated by the two methods indicate complicated structures around the Moho discontinuity. Our receiver function analysis basically estimated S wave structures using Ps converted waves with a dominant frequency of about 1 Hz, while the explosion analysis uses reflected and/or refracted P waves from controlled sources at the near ground surface with a dominant frequency of about 10 Hz. The used wavelength and its wave type differ from those of the receiver function analysis. The explosion analysis may be sensitive to the top portion of the Moho discontinuity. The disagreement regarding Moho depth is also explained by their sharpness. The existence of a weak positive phase corresponding to the Moho discontinuity from the explosion analysis is mainly obtained from the increase of the P wave velocity. It seems that the velocity discontinuities of the P wave are not in accord with those of the S wave. Further studies of the velocity transitions near the crust-mantle boundary will provide the physical and chemical properties in the formation process of the Island lithosphere.

6. Conclusions

We estimated the seismic discontinuities of the crust and the uppermost mantle in central Japan using the receiver function method with altitude and low-velocity sediment layer correction. The subducting Philippine Sea plate, the Moho discontinuity in the overriding plate, and some discontinuities in the crust were found in the receiver function images. Crustal discontinuities at the NKTZ seem to be complicated. In particular, the dips of discontinuities changed around the Atotsugawa fault.

The Moho discontinuity is found continuously at every cross-section. It dips gradually and is located at a depth of about 25 km at the northern part and about 43 km at the southern part of the profile lines. The depth is 5–10 km greater than that obtained by the explosion analysis. The results of this study are important to understand the structure from the ground surface to the uppermost mantle beneath central Japan. The heterogeneous structures in the NKTZ will give us important constraints to identify the mechanisms of strain accumulation in the Japanese island area and the mechanisms of inland earthquakes.

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