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# Crustal structure of the southwestern margin of the Kuril arc sited in the eastern part of Hokkaido, Japan, inferred from seismic refraction/reflection experiments

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Hokkaido Island in northern Japan is located at the intersection of the Kuril and Northeast Japan arcs. Eastern Hokkaido represents the southwestern margin of the Kuril arc that evolved as an oceanic island arc, but its crustal structure has remained unclarified. In the summer of 2000, a highly dense onshore-offshore integrated seismic experiment was conducted in order to reveal the entire crustal section covering from the Kuril Trench to the back-arc basin of the Okhotsk Sea crossing eastern Hokkaido. High-quality refraction/wide-angle reflection data collected for the onshore survey line delineated a detailed crustal structure of easternmost part of Hokkaido Island. Our seismic velocity model shows a good correlation with the surface geology along the profile. The notable feature of the velocity model is the existence of a 6.0-km/s layer beneath the onshore survey line. A reflective middle to lower crust is also found beneath eastern Hokkaido. These results indicate that the southwestern margin of the Kuril arc is in a matured state of oceanic island-arc crust.

**Key words:** Crustal structure, southwestern margin of the Kuril arc, seismic refraction and reflection experiments, mature oceanic island-arc crust.

## 1. Introduction

Hokkaido Island in northern Japan is located at the intersection of two active island arc-trench systems: the Northeast Japan arc-Japan Trench and the Kuril arc-Kuril Trench systems (Fig. 1). Hokkaido is divided into three major provinces with regard to its pre-Neogene geology whose general trend is oblique to these arc-trench systems: Western, Central, and Eastern Hokkaido (e.g. Arita *et al.*, 1998). Western Hokkaido is the northern extension of the Northeast Japan arc, which consists of a late Jurassic accretionary complex with Cretaceous granitic intrusion (e.g. Kawamura *et al.*, 1986; Kimura, 1994). Central Hokkaido is composed of Cretaceous to early Paleogene accretionary complexes often covered by forearc basin sediments (e.g. Kimura, 1994). Eastern Hokkaido represents the southwestern margin of the Kuril arc, whose oldest rocks are the latest Cretaceous andesitic volcanic and sedimentary rocks, which probably form the ancient Kuril arc (Kimura and Tamaki, 1985; Kimura, 1994). The crustal evolution of Hokkaido Island has been dominated by a series of accretion and collision processes occurring from the late Jurassic to the present. Several tectonic models for Hokkaido have been proposed. Kimura (1994, 1996) suggested that the ancient Kuril arc situated along the southern margin of the Okhotsk Plate collided with the Asian continental margin in the late Eocene. After this collision, the northern part of the Okhotsk Sea was trapped, which caused the cessation of subduction along Hokkaido, Sakhalin, and the southern margin of Siberia, with only the Kuril trench surviving. Since the late Miocene, the Kuril forearc sliver has been migrating southwestward and colliding with the Northeast Japan arc (e.g. Kimura, 1986) due to the oblique subduction of the Pacific Plate along the Kuril trench. In order to investigate the structural variations and deformations associated with the accretion and collision tectonics in Hokkaido, several controlled-source seismic experiments including refraction and reflection profiling were conducted (e.g. Arita *et al.*, 1998; Iwasaki *et al.*, 1998; Tsumura *et al.*, 1999; Iwasaki *et al.*, 2004).

The easternmost part of Hokkaido is geologically considered to be the southern margin of the Kuril arc. Although the Japan National Oil Corporation (JNOC) conducted several seismic reflection experiments (JNOC, 1984, 1985, 1990, 2002) in this region, these studies shed light only upon a shallow crustal structure, and little is known about the entire crustal structure. In the summer of 2000, a highly dense onshore–offshore integrated seismic experiment was conducted in order to reveal the crustal section from the Kuril Trench to the back-arc basin of the Okhotsk Sea crossing eastern Hokkaido, which corresponds to the southwestern margin of the Kuril arc. Nakanishi *et al.* (2004) showed the detailed crustal and uppermost mantle structural model across the seismogenic zone of the southern Kuril Trench from off-shore data. In this paper we present the crustal structure beneath eastern Hokkaido obtained from on-shore data and discuss the maturity level of

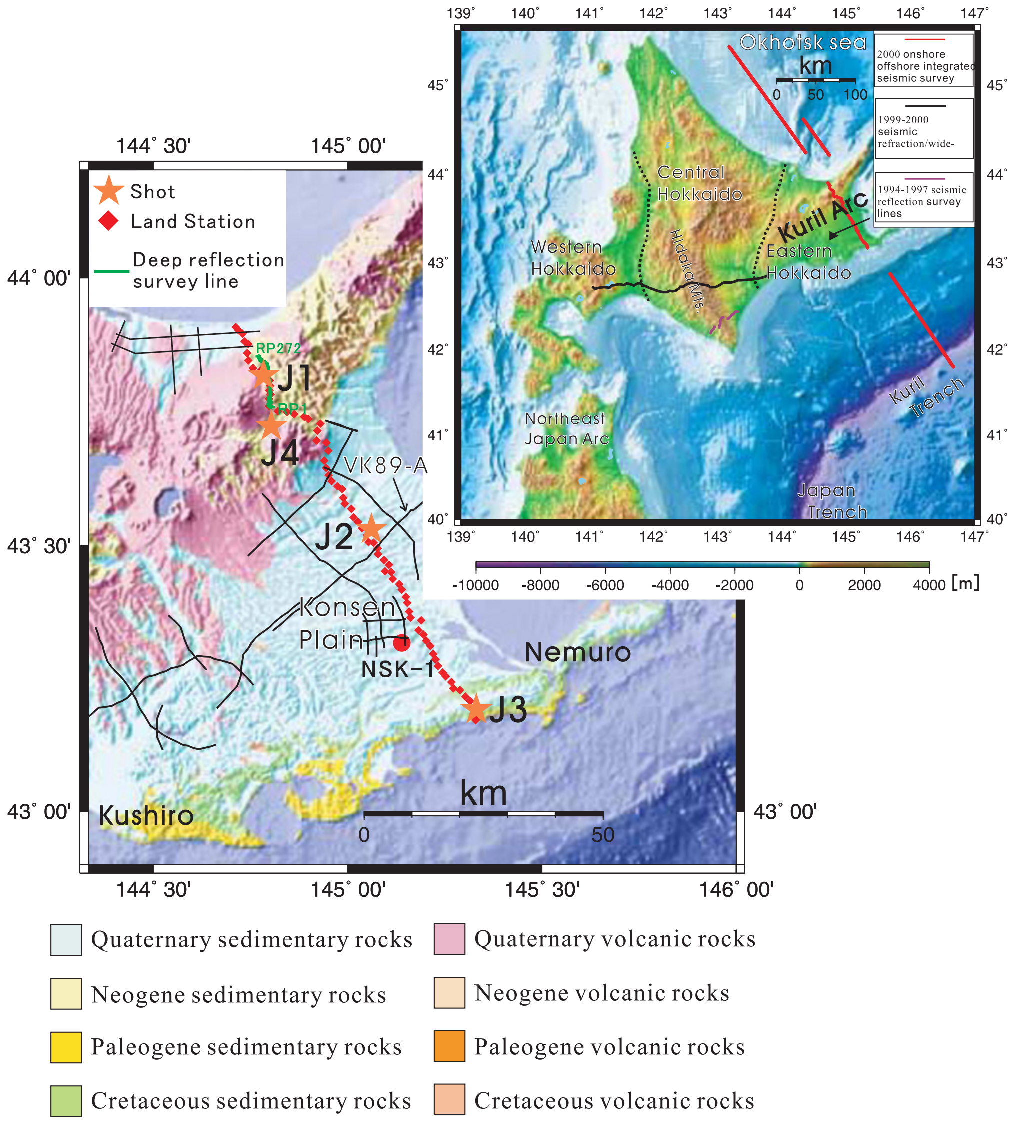


Fig. 1. Location map of the 2000 onshore–offshore integrated seismic experiment across eastern Hokkaido, Japan. The survey lines are shown in red. The reflection lines carried out from 1994 to 1997 (Tsumura *et al.*, 1999) and refraction/wide-angle reflection lines carried out in 1999–2000 (Iwasaki *et al.*, 2004) are also shown. The explosive shots (star symbols) and the land seismic stations (diamond symbols) are shown with a simplified geological map (Geological Survey of Japan, 1992). The green line denotes the deep seismic reflection survey line using a digital telemetry recording system. The red circle symbol labeled 'NSK-1' denotes the drilling site (JNOC, 2002). The black lines denote previous seismic reflection survey lines conducted by the Japan National Oil Corporation (JNOC, 1984, 1985, 1990, 2002). The seismic reflection profile along the VK89-A was reported by JNOC (1985).

the southwestern margin of the Kuril arc crust. According to Kimura (1996), the Kuril arc is an oceanic island arc. However, previous wide-angle seismic investigations in the eastern half of the Hokkaido Island provide no positive evidence of an oceanic island arc. The easternmost part of the Hokkaido has been less affected by the collision than previous survey areas, therefore, its crustal structure is expected to contain some information on the crustal evolution on the southwestern margin of the Kuril arc.

## 2. Experiments and Data

In the summer of 2000, a highly dense onshore–offshore integrated seismic experiment was conducted in order to reveal the crustal section from the Kuril Trench to the back-arc basin of the Okhotsk Sea crossing eastern Hokkaido (Fig. 1). Seventy-four portable seismographs were deployed onshore along a 95-km-long line in the north-northwest-south-southwest direction at intervals of 1–1.5 km. Each seismograph consisted of a 4.5-Hz three-component seismometer and a long-term low-power digital audio tape (DAT) recorder (Shinohara *et al.*, 1997). Waveforms were recorded at a sampling rate of 100 Hz. Each DAT recorder included a global positioning system (GPS) receiver to maintain the accuracy of its internal clock. Four explosives were fired as a controlled seismic source on the onshore survey line. The charge sizes were 300 kg at J1 and J3 and 100 kg at J2 and J4. Moreover, we conducted deep seismic reflection profiling in order to reveal the distribution of the deep reflections at the northern part of the onshore survey line (Fig. 1). The seismic signals from the explosive sources (J1-J4) were also recorded with a digital telemetry seismic recording system with 272 channels at a

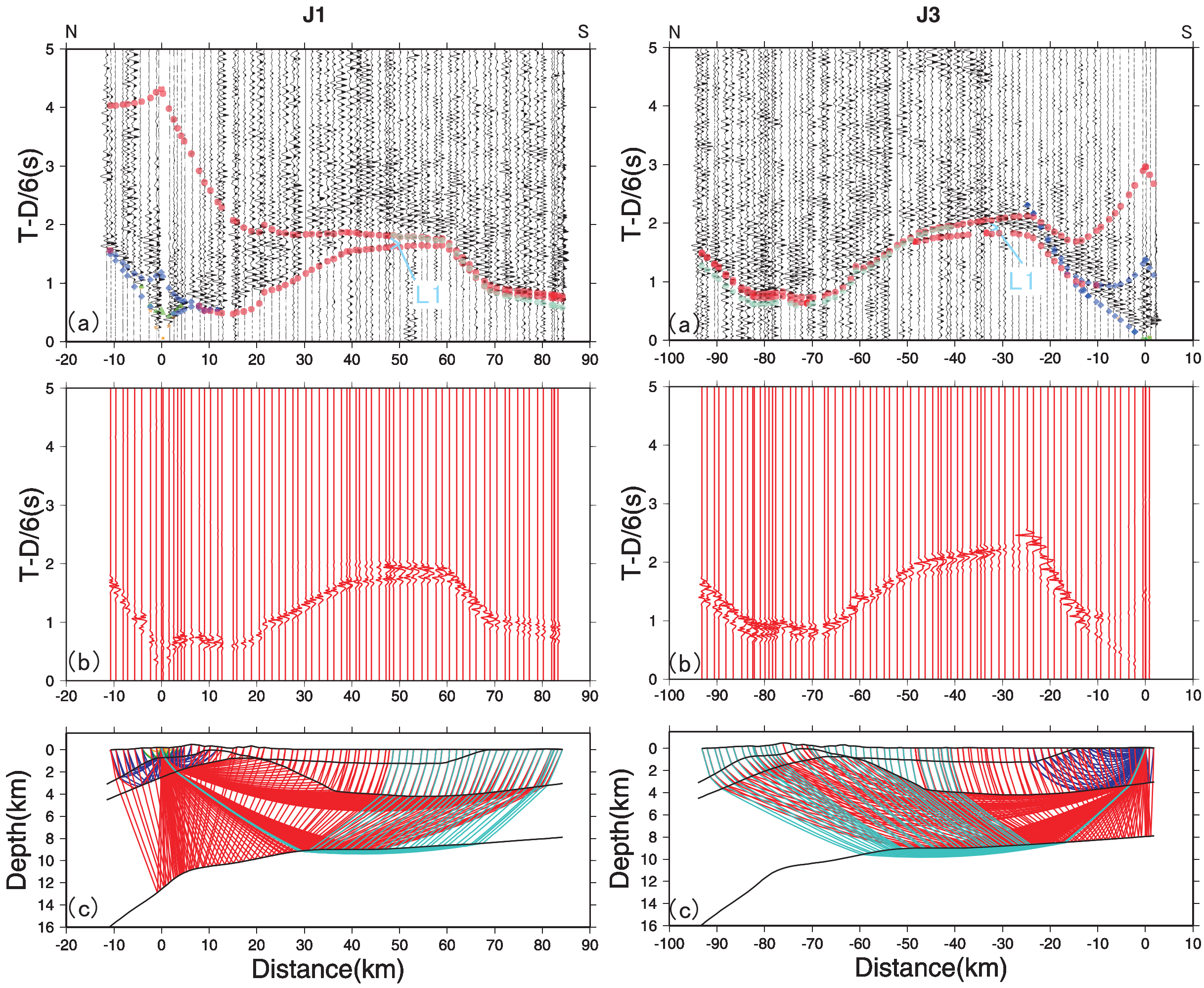


Fig. 2. Observed seismograms, calculated travel times, synthetic seismograms, and ray diagrams for shots J1 (left) and J3 (right). The horizontal axis represents the source-receiver offset. (a) Observed seismograms. Band-pass filter (1–25 Hz) is applied. Each trace is normalized by its maximum amplitude. The reduction velocity is 6 km/s. We obtained high signal-to-noise ratio data along the entire length of the onshore survey line. The calculated travel times for the final crustal model (Fig. 3) are superimposed on the observed seismograms. (b) Ray theoretical synthetic seismograms. (c) Ray diagram. The ray paths and calculated travel times correspond in color.

sampling rate of 250 Hz. A 50-m group interval of 10-Hz geophones was used. We obtained high signal-to-noise ratio data along the entire length of the onshore survey line (Fig. 2).

## 3. Analysis and Results

### 3.1 Two-dimensional velocity structure

The two-dimensional (2-D) velocity structure beneath eastern Hokkaido was derived by a 2-D ray-tracing method using the first arrivals and later phases observed on the survey line so as to determine the velocities and the geometry of the velocity interface. In order to construct an initial model of the ray-tracing analysis, the first arrivals from J1-J4 were inverted by using an extended time-term method (Iwasaki, 2002) so as to obtain the rough geometry of the surface layer and the basement velocity. Subsequently, the ray-tracing method was employed to calculate the wave field in the 2-D varying velocity model (Červený and Pšenčík, 1983). We obtained a model that could provide a reasonable explanation for both of the observed travel times within a 0.1-s error and the gross features of the amplitudes (Figs. 2 and 3). Our model indicates a lateral variation of the surface layer. Although its velocities range from 1.7 to 4.3 km/s, significant velocity changes are found at distances of 30 and 70–80 km (Fig. 3). The thickness of the surface layer is 0.2–3 km in the northern part of the profile, while it is nearly 4–5 km in the southern part. The velocity of the top of the uppermost crystalline crust is 6.0 km/s, and this 6.0-km/s layer exists beneath the entire onshore survey line. In Fig. 2, the later phase, L1, is a reflector from an interface located at a depth of 9 km beneath the central part of survey line, whose strong amplitudes required a velocity jump of 0.3 km/s at the corresponding interface (Fig. 3). The resultant velocity below the interface is 6.4 km/s. As described in Section 3.2, the data from the digital telemetry seismic recording system strongly indicate mid-crustal reflectors below a depth of 10–12 km in the northern part of the profile.

### 3.2 Reflection image

To image the mid and lower crustal structures in the northern part of the profile, the data sets (J1 and J4) recoded by the digital telemetry seismic recording system were processed using the seismic reflection technique. The imaging was performed using conventional common midpoint

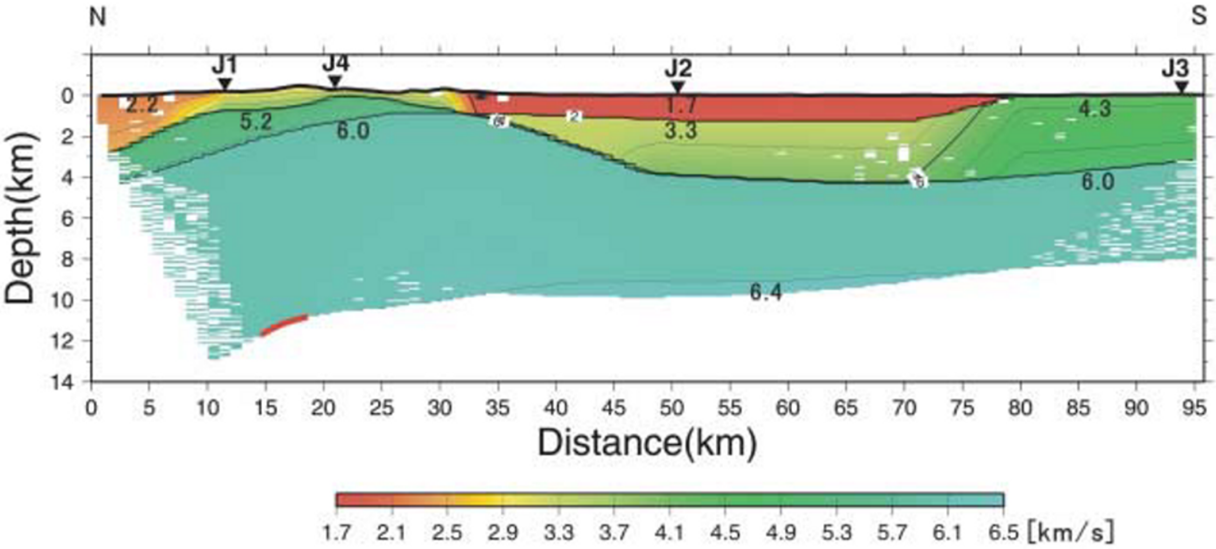


Fig. 3. *P*-wave velocity structure derived by a 2-D ray-tracing method. The horizontal axis represents the distance from the north end of the onshore survey line. The shot positions are denoted by solid triangles at the top of figure. The *P*-wave velocities are shown using a color scale. The numerals indicate the P-wave velocities in km/s. The solid counter lines denote the *P*-wave velocity in km/s. The counter interval is 0.4 km/s. The areas with no ray coverage are represented as blanks. The red line denotes the mid-crustal reflector (R1) obtained from the seismic reflection image (Fig. 5).

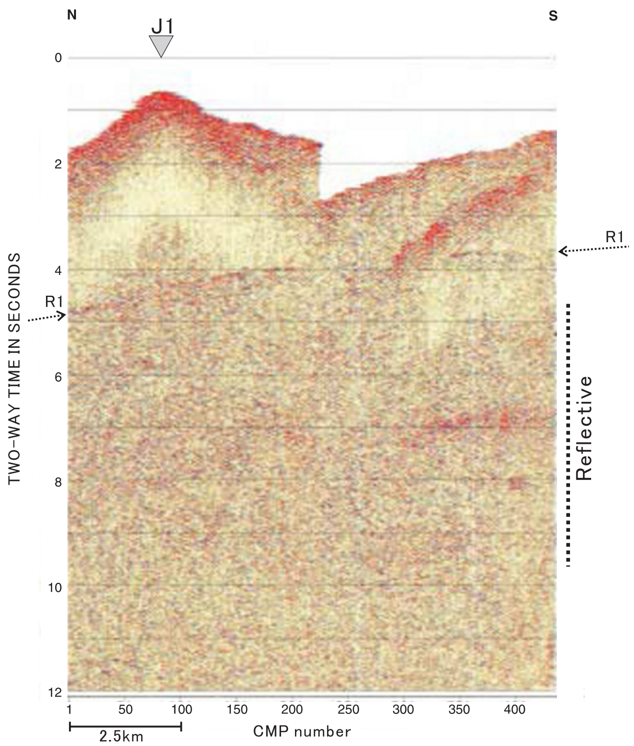


Fig. 4. Unmigrated stacked time section of the deep reflection survey using a digital telemetry recording system. See text for detailed explanation.

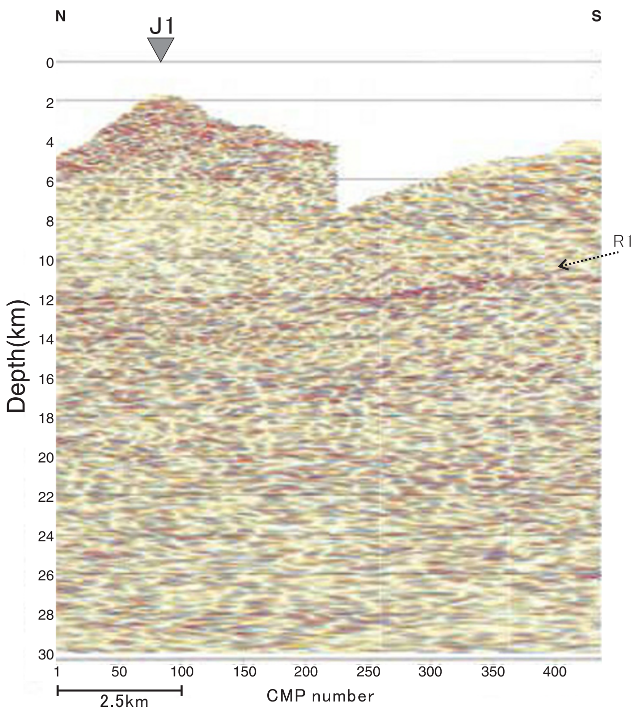


Fig. 5. Post-stack migrated depth section of the deep reflection survey using a digital telemetry recording system. See text for detailed explanation.

(CMP) processing steps, including post-stack migration and depth conversion. A static correction for the shallow low-velocity layer was applied based on the refraction analysis by using the extended time-term method (Iwasaki, 2002). The length of the automatic gain control operator is 4 s. The normal moveout and depth conversion velocities were based on the 2-D velocity structure (Fig. 3). In Figs. 4 and 5, the CMP stacked images show several features of the deeper part of the crust. In the unmigrated time section (Fig. 4), a clear reflector (R1) is found at a 4–4.4 s two-way travel time (TWT), below which a number of reflectors are distributed down to a 9–10 s TWT. Figure 5 shows the migrated depth section in which the northward dipping reflector (R1) was mapped at a depth of 10–12 km.

## 4. Discussion and Conclusions

The $P$-wave velocity structure beneath eastern Hokkaido (Fig. 3) correlates well with the surface geology (GSJ, 1992, see Fig. 1). Namely, a surface layer with a velocity of 2.9–4.0 km/s in the northern part of our model represents Miocene to Quaternary volcanic rocks. The Nemuro group, which is mainly composed of upper Cretaceous to Paleocene sediments in the southern part of our profile, has a velocity of 4.3 km/s. The 1.7-km/s layer beneath the Konsen Plain corresponds to a Neogene and Quaternary sedimentary layer. The analysis of drill cores at Nishibetsu (NSK1 in Fig. 1) indicates that the boundary between the upper Eocene and Cretaceous units is located at a depth of 900 m (JNOC, 2002), which is almost consistent with our results. A density model constructed from the Bouguer anomaly data beneath eastern Hokkaido (Morijiri *et al.*, 2000) indicates that the maximum thickness of the Cenozoic deposits is approximately 1.7 km in the Konsen Plain, showing good correspondence to our $P$-wave velocity model. In the Konsen Plain, several seismic reflection experiments were conducted (JNOC, 1985, 1990, 2002). The VK89-A line crosses our survey line (Fig. 1). On this profile, clear reflectors are recognized at a depth of 4–5 km beneath the point of intersection with our survey line. The top of the 6.0-km/s layer in our model is almost consistent with that of the reflector. The lower side of the reflector that corresponds to the 6-km/s layer is interpreted as the Cretaceous Nemuro group (JNOC, 1990).

The deeper structure beneath our profile is characterized by a 6.4 km/s layer in its southern part and mid-crustal reflective zone in its northern part. From our seismic data, however, it is not clarified that the top of the 6.4 km/s layer in the southern part of the profile corresponds to the reflector R1 in the northern part of the profile. It also remains unclear that the 6.4 km/s layer contains a reflective zone as found in the northern part of the profile.

A magneto-telluric (MT) survey conducted across eastern Hokkaido revealed a highly resistive zone at a depth of 10–20 km below the forearc side (Satoh *et al.*, 2001). The depth of the top of the highly resistive zone is almost consistent with that of the mid-crustal interface beneath the central part of our survey line. It is also noted that this highly resistive zone does not extend to the northern part of the line. A highly positive gravity anomaly is observed in and around the Nemuro Peninsula (Morijiri *et al.*, 2000). These data suggest the existence of structural differences between the northern and southern parts of easternmost Hokkaido. Granodiorite formed by Cretaceous–earliest Paleogene island–arc magmatism (54–79 Ma) was found in the southern part of eastern Hokkaido in the Kushiro-Nemuro area (Ogasawara *et al.*, 1997). Based on studies of K-Ar age data, volcanic stratigraphy, and physical volcanology, Hirose and Nakagawa (1999) concluded that subduction-related island-arc volcanism occurred at least up to 9 Ma in the northern part of eastern Hokkaido. These results suggest that the materials forming the northern part of the southwestern Kuril arc are younger than those of the southern part.

Finally, we compare our results with those from other island arcs. As stated above, the southwestern margin of the Kuril arc has a 6.0-km/s layer. This characteristic is similar to the case of the Izu-Bonin arc, which also evolved as an oceanic island arc. According to a wide-angle reflection survey (Suyehiro *et al.*, 1996; Tahahashi *et al.*, 1998), the Izu-Bonin arc has a mid-crustal layer with a velocity of 6.0–6.3 km/s. Two major volcanic episodes occurred in this arc before and after the back-arc spreading (Karig and Moore, 1975). Takahashi*et al.* (1998) indicate that the mid-crustal layer of the Izu-Bonin arc is probably a granitic layer formed in more than two volcanic episodes. On the other hand, the central Aleutian arc, which is an oceanic island arc built on old oceanic crust, has no mid-crustal layer with a velocity of 6.0 km/s (Holbrook *et al.*, 1999). According to Scholl *et al.* (1987), the arc magmatism in the Aleutians probably started in the early to middle Eocene (55–50 Ma). The Aleutian Islands are volcanically active, just like the Kuril Islands, and the Tertiary-Quaternary volcanic rocks are exposed (e.g. Beikman, 1980). However, a back-arc spreading zone has never developed in the Aleutians, which distinguishes it from the Izu-Bonin and southwestern Kuril arcs. In the central Aleutian, the reflective lower crust, which is often seen in continental crusts, is not recognized (Holbrook *et al.*, 1999). From these observations, Holbrook *et al.* (1999) proposed that the mid-crustal layer with a velocity of 6.0 km/s and seismic reflectivity are not native to oceanic island arcs and therefore additional processes are required. Namely, intra-crustal melting for forming a mid-crustal layer with a velocity of 6.0 km/s and delamination of a mafic lower crust must occur to become "continental" in character. These comparisons indicate that the southern margin of the Kuril arc is in a matured state of the oceanic island arc.

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