**ABSTRACT**

Seismic, geodetic, and tsunami data of the 2011 Tohoku-oki earthquake (offshore Japan; moment magnitude, \(M_w 9.0\)) have revealed that large coseismic slip reached the trench axis. Moreover, a clear, depth-dependent variation in the source location between high- and low-frequency seismic energy radiation was observed. However, depth-varying structural features in the rupture zone have not been well examined. We therefore processed seismic reflection data acquired along five profiles in the rupture zone and examined depth-varying structural characteristics. In the resultant seismic images were interpreted a low-velocity frontal prism, a reflective zone at the trenchward tip of the continental block, and subducted horst and graben structures. The frontal prism, which was imaged as a low-velocity (Vp 2.0–3.5 km/s) wedge-shaped unit with seafloor widths of 13.5–18 km north of 37.5°N, changed abruptly to an elongate sedimentary unit south of 37.5°N. Landward of the frontal prism, 30–80 km from the trench axis, a reflective zone was imaged above the subducted oceanic basement. Subducted horst and graben structures were clearly imaged beneath the frontal prism and the reflective zone, and they could be found to a depth of 25 km. The throws of the normal faults delineating the horst and graben structures become larger landward to as much as 2 km. Comparison of the seismic images, earthquake seismicity, and slip behaviors showed that slips of tsunami earthquakes occur along the plate interface where the frontal prism is well developed. Background seismicity along the plate interface may extend downward to the landward end of the frontal prism and it becomes active around 25 km depth extending down the subduction zone.

**INTRODUCTION**

The rupture process of the 2011 Tohoku-oki earthquake (offshore Japan) has been examined extensively using data from state of the art observational networks deployed both globally and locally. In the Japanese islands, these include dense seismic, geodetic, and tide-gauge station networks (e.g., Fujii et al., 2011; Ide et al., 2011; Lay et al., 2011; Sato et al., 2011; Iinuma et al., 2012). Although rupture models differ in detail among studies, a common feature of the slip distribution is that fault displacement of more than 50 m occurred in the shallowest part of the subduction interface beneath the middle slope of the Japan Trench. In addition to earthquake, tsunami, and geodetic studies, marine geological and geophysical studies, including time-lapse bathymetry and controlled source seismic studies, have presented clear evidence that the rupture along the plate boundary reached the seafloor at the trench axis (Fujiwara et al., 2011; Kodaira et al., 2012; Nakamura et al., 2013).

Moreover, a clear difference in source location between high- and low-frequency seismic energy radiation has been reported (e.g., Hara, 2011; Ishii, 2011; Koketsu et al., 2011; Koper et al., 2011; Simons et al., 2011). For example, inversion studies of tsunami waveforms, which reflect the source region of low-frequency radiation, show a concentration of large slip immediately landward of the trench axis (e.g., Fujii et al., 2011; Satake et al., 2013), whereas back-projection studies in which teleseismic data were used to map the source region of high-frequency radiation have shown that high-frequency energy primarily radiated from the deeper part of the rupture zone to the west of the hypocenter (e.g., Ishii 2011; Wang and Mori, 2011).

Similar spatial variation in seismic wave radiation sources has also been observed in the rupture zones of other large megathrust earthquakes, such as the 2004 Sumatra-Andaman (moment magnitude, \(M_w 9.2\)) and the 2010 Chile (\(M_w 8.8\)) earthquakes; that is, the source region of high-frequency radiation was distributed mostly in the deeper portion of the megathrust fault, whereas large slip occurred in the shallow portion where little high-frequency seismic energy was emitted (Lay et al., 2012). Along the Japan Trench, the 1896 Sanriku earthquake is another well-known example; during this earthquake, low levels of short-period seismic wave radiation emanated from the area close to the trench axis where the large fault slip occurred (e.g., Kanamori, 1972). Even though these depth-varying slip behaviors have been well documented, the structural features that control them have not been well examined. We therefore used newly processed prestack depth migrated images for deep seismic reflection data across the rupture zone of the 2011 Tohoku-oki earthquake and examined structural features along and around the plate boundary (Fig. 1).

**TECTONIC SETTING**

The structure and lithology of the Japan Trench subduction zone have been intensively investigated with seismic surveys and ocean drilling for the past three decades. In studies of seafloor topography and seismicity, the Japan Trench forearc region was divided into four areas: a deep-sea terrace, a steep
upper slope, a relatively flat middle slope, and a generally steep and rugged lower slope (von Huene and Cullotta, 1989; von Huene and Lallemand, 1990; von Huene et al., 1994) (Fig. 2). Seismic surveys and ocean drilling have shown that the shallowest part of the deep-sea terrace and the upper slope consist of Pliocene–Pleistocene and Miocene sediments that unconformably overlie the Cretaceous continental framework (von Huene et al., 1982, 1994; Tamaki et al., 1990; Jolivet and Tamaki, 1992; Tsuru et al., 2002) (Fig. 2). The unconformity between the Cretaceous prism and Miocene slope deposit is imaged as a strong reflector that is continuous from the deep-sea terrace to the middle slope. The seismic reflection data also show that the shallow Pliocene–Pleistocene strata are deformed by a normal fault that extends locally to the seafloor (von Huene et al., 1994; Tsuru et al., 2000, 2002). This normal fault structure in the forearc basin is interpreted to have formed as a result of continuous subsidence in the forearc caused by tectonic (or basal) erosion at the base of the overriding plate (von Huene and Lallemand, 1990; von Huene et al., 1994).

A prism-shaped low-velocity wedge (herein, the frontal prism) is imaged beneath the middle to lower slope, and a clear landward-dipping reflector, the backstop interface (Fig. 2), is the boundary between the frontal prism and the Cretaceous continental framework. Tsuru et al. (2002), who compiled size and distribution data of the frontal prism along seismic reflection profiles.

**Figure 1.** (A) Map showing the seismic profiles processed in this study (white lines). The dotted white line indicates a profile processed by Nakamura et al. (2014). The coseismic slip distribution obtained by tsunami waveform inversion (color scale and contours) and the epicenters of the 2011 Tohoku-oki earthquake (orange star), aftershocks in the initial 24 h (yellow circles), other aftershocks with magnitude greater than 7.0 (orange circles), and a foreshock (white circle) (Koketsu et al., 2011) are also shown. (B) Map showing common depth point (CDP) values along each profile. The yellow star shows the 2011 Tohoku-oki earthquake epicenter.
Figure 2 (on this and following page). Prestack depth migrated sections. In each panel, the vertical dotted line indicates the location of the trench axis. CMP—common midpoint. See the text for the data acquisition parameters. (A) D19. (B) D17. (C) D11. (D) TH03. (E) D02.

- Deep sea terrace
- Upper slope
- Middle slope
- Lower slope
- Trench axis
- Pliocene-Pleistocene sediments
- Cretaceous unconformity
- Cretaceous continental framework
- Top of oceanic crust
- Frontal prism
- Backstop interface
- Oceanic Moho
- Miocene sediments
- CMP number (6.25 m / CMP)

A: CMP numbers 27703 to 503
B: CMP numbers 1601 to 26951
C: CMP numbers 1601 to 25951

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perpendicular to the trench from 36°N to 41°N, reported that the frontal prism is developed primarily along the northern Japan Trench, whereas along the southern Japan Trench, an elongate sedimentary unit rather than wedge-shaped body characterized by low seismic velocities (3.4–4.0 km/s) extends in the downdip direction along the subduction interface. A core sample collected at Integrated Ocean Drilling Program Japan Trench Fast Drilling Project (IODP JFAST) Site C0019 (Fig. 1), which penetrated the trenchward tip of the frontal prism, shows that the frontal prism there consists of steeply dipping pelagic and/or hemipelagic mudstone, which has accreted above a thin (~5 m thick), highly fractured plate boundary fault zone composed of pelagic clay (Chester et al., 2013; Kirkpatrick et al., 2015).

The incoming oceanic plate is characterized by the distinctive structural pattern of a horst and graben, which was first recognized from the seafloor topography by Ludwig et al. (1966). It is widely accepted that the horst and graben are formed by bend-related normal faulting of the oceanic plate toward the trench. Previous seismic studies have imaged a series of horst and graben structures beneath the frontal prism, and the throws of the normal faults become larger toward the trench axis and displacement continues to increase from the trench down the subduction zone (von Huene et al., 1994; Tsuru et al., 2000, 2002).

**DATA ACQUISITION AND PROCESSING**

The seismic reflection data used in this study were acquired during surveys in 2011 by R/V *Kairei*. The seismic system mounted on R/V *Kairei* consists of a 6000-m-long, 444 channel streamer cable as a receiver and a 7800 m³ tuned airgun array, which fired at 50 m intervals. The streamer cable was 20 m deep.

In order to examine structural characters in the rupture zone, we processed seismic data along five survey lines situated where the large coseismic slips...
are observed, from north to south, D19, D17, D11, TH03, and D02 (Fig. 1). The data were first processed following a conventional flow as described by Nakamura et al. (2014). The processing flow included minimum phase conversion, predictive deconvolution, common midpoint sorting, velocity analysis, normal moveout, stacking, and poststack time migration. Noise reduction by f-x projection (Soubra, 1995) and f-x prediction filtering (Canales, 1984) was applied to reduce the swell noise and noises from local earthquakes. Suppression of surface-related multiples and parabolic random transform were applied to reduce multiple reflections. Prestack time migration analysis was applied to produce a fine-tuned time-domain seismic image. The prestack time migration section was then used to establish a structural formation model for further prestack depth migration (PSDM).

To obtain precise depth images and P-wave velocity models, we applied PSDM. Details of the PSDM procedure were described in Nakamura et al. (2014). We applied a layer-cake approach for the PSDM velocity analysis, in which we determined the velocity downward from the shallowest unit. Once a satisfactory layer-cake model was achieved, the velocity model was improved by a horizon-based residual velocity analysis, which evaluates the moveout of the migrated common reflection point gathers by using horizon semblance. The velocity model was then refined by using grid-based traveltime tomography to handle the fine-scale velocity variation. In order to construct the deeper part of the model, we also referred to velocity models previously obtained by wide-angle seismic reflection and refraction studies (e.g., Miura et al., 2005).

**SEISMIC IMAGES AND DISCUSSION**

**Slope Sediments**

North of 38°N, the slope sediments from the deep-sea terrace to the middle slope have been well examined (e.g., von Huene et al., 1994; Tsuru et al., 2000, 2002; Kimura et al., 2012; Boston et al., 2017). Therefore, we briefly summarize structural characteristics of the slope sediments overlying the unconformity between the Cretaceous prism and Miocene slope deposit in light of those previous studies. In seismic images, the slope sediments are sedimentary units with a thickness of ~2 km above the strong reflector of the aforementioned unconformity (Fig. 2). In this study, details of the structure vary among the profiles, but the overall structural character described previously (von Huene et al., 1994; Tsuru et al., 2000, 2002; Kimura et al., 2012) is recognized to continue along the trench in our survey area. For example, the slope sediments can generally be divided into two main units (Fig. 2; cf. Kimura et al., 2012, Fig. 2 therein) that extend from the deep-sea terrace toward the trench and appear to be truncated at the lower slope. The upper and lower units are interpreted as Pliocene–Pleistocene sediments and Miocene sediments, respectively (Fig. 2). Boston et al. (2017) examined seismic sections along 15 profiles across the forearc of the central to northern Japan Trench, including the five profiles reprocessed in this study, and inferred that the sediments of the upper to middle slope transition show both extensional and contraction features. These results imply that accretion and contraction structures are common across the lower slope, whereas extensional structures are common across the middle and upper slope.

**Frontal Prism**

A characteristic structure observed in all five profiles is a wedge-shaped low-velocity block, the frontal prism, at the toe of the overriding plate bounded by a clear landward-dipping reflector (called the backstop interface) on its landward side (Figs. 3–7). The frontal prism, which is clearly imaged near the trench (to 20–30 km landward from the trench axis), has a seafloor width of 13.5–18.0 km (Table 1; Fig. 8). The P-wave velocity in the frontal prism varied from 2.0 km/s at the seafloor to 3.5 km/s at the bottom of the prism, and weak landward-dipping reflectors were imaged within the prism.

A recent result of IODP JFAST shows that a core sample at the toe of the frontal prism consists of pelagic sediment and the stratigraphic unit is well correlated with a core sample from incoming sediments to the Japan Trench at the western Pacific Deep Sea Drilling Project (DSDP) Site 436 (e.g., Rabinowitz et al., 2015). Although the JFAST site is located at the trenchward end of the frontal prism and the core is mainly sampled at the deeper part of the frontal prism, we believe that the results from JFAST are strong evidence in support of the idea that an accretionary process is a dominant process in forming the frontal prism. Assuming that the frontal prism has been developed since the Eocene, based on the age of the oldest pelagic sediments sampled at the JFAST site (e.g., Rabinowitz et al., 2015), we estimate that only ~2% of the input sediment contributes to form the frontal prism and the remaining ~98% of the input sediment has been subducted (i.e., an average of the volume of the frontal prism per 1 m along the trench is 4 × 10^7 m^3 (Table 1); a total input sediment volume since the Eocene per 1 m along the trench is 8 × 10^2 m/yr (plate convergent rate) × 5 × 10^7 yr (since Eocene) = 2 × 10^9 m^3].

To examine the along-trench variation of the frontal prism, we determined the volume and distribution of the frontal prism in six imaged sections, including the profile processed in Nakamura et al. (2014) (Fig. 9), and compared them with data from Tsuru et al. (2002), who processed 14 profiles in the Japan Trench subduction zone from 36°N to 41°N. Tsuru et al. (2002) showed that the wedge-shaped frontal prism is dominantly developed in the northern Japan Trench (north of 39.5°N), whereas the southern trench is characterized by an elongate low-velocity layer above the subducted oceanic crust with a thickness of 5–10 km. However, Tsuru et al. (2002, their figure 17 therein) could not resolve the transition between the prism and the elongate layer because the coverage of the profiles around 38°N is too sparse to resolve the structural transition. The structural information obtained in this study fills this gap, and shows that the wedge-shaped block is observed only north of 37.5°N; south of this latitude the structure is an elongate low-velocity layer. Thus, there is a sharp structural boundary in the toe of the overriding plate in the central Japan Trench at 37.5°N (Fig. 9).
Figure 3. Enlarged section of the landward-trench slope along D19. (Top) Prestack depth migration (PSDM) interval velocity model superimposed on the PSDM section. (Middle) PSDM section. (Bottom) Interpretative line drawing superimposed on the PSDM interval velocity model.
Figure 4. Enlarged section along D17. See Figure 3 caption for description.
Figure 5. Enlarged section along D11. See Figure 3 caption for description.
Figure 6. Enlarged section along TH03. See Figure 3 caption for description.
Figure 7. Enlarged section along D02. See Figure 3 caption for description.
The seafloor widths of the frontal prism determined in our study (Fig. 8) show an apparent inconsistency with those reported by Tsuru et al. (2002). Tsuru et al. (2002) estimated the seafloor width of the frontal prism in the northern Japan Trench (their profiles 1–5) to be 23–30 km, whereas the seafloor width in the profiles processed here ranged from 14 to 18 km, even though our profile D19 is in approximately the same location as their profile 5. Similarly, von Huene et al. (1994) estimated the wedge-shaped block to have a seafloor width of ~10 km in a profile at 39°40′N, whereas Tsuru et al. (2002) estimated the seafloor width of the frontal prism along profiles near that latitude to be 23.5–30 km. To examine the reason for these different seafloor widths, we compared how the low-velocity prism, or wedge-shaped low-velocity block, was defined among the three studies. von Huene et al. (1994) estimated the seafloor width of the wedge-shaped low-velocity block by measuring the width of the accreted sediment and excluding the width of what they called the tectonized block at the seaward tip of the continental block. In contrast, Tsuru et al. (2002) used the width of the deformed zone, consisting of the prism-shaped accretionary sediment and a reflective zone comprising a series of discontinuous landward-dipping reflectors in the continental block. In this study we excluded the reflective zone at the toe of the continental block when we measured the seafloor width of the frontal prism. These differences in definition are responsible for the apparent differences in the seafloor width of the wedge-shaped block.

**Figure 8.** Schematic diagram of the geometry of the frontal prism, modified from Tsuru et al. (2002). Frontal prism dimensions (distances A and B and thickness) in each profile are given in Table 1.

**Figure 9.** Map showing the size and distribution of low-velocity sedimentary units along the Japan Trench, modified from Tsuru et al. (2002). Black bars indicate the thickness of the low-velocity wedge-shaped and elongate units as mapped by Tsuru et al. (2002), and the areas of the frontal prism determined by this study are in light orange. Coseismic slips (colored squares and contours) of the 2011 Tohoku-oki earthquake are inferred from tsunami waveform data (Satake et al., 2013). The light blue diamonds and the purple squares indicate the locations of ocean-bottom pressure gauges and global positioning system wave gauges that were used with coastal tide or wave gauges for the tsunami waveform inversion (Satake et al., 2013). The yellow rectangle outlines the inferred fault plane of the 1896 Sanriku tsunami earthquake (Tanioka and Satake, 1996).

**TABLE 1. FRONTAL PRISM DIMENSIONS**

<table>
<thead>
<tr>
<th>Profile</th>
<th>Distance A (km)</th>
<th>Distance B (km)</th>
<th>Thickness (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>D19</td>
<td>13.5</td>
<td>25</td>
<td>3</td>
</tr>
<tr>
<td>D17</td>
<td>16.5</td>
<td>24</td>
<td>3.5</td>
</tr>
<tr>
<td>D11</td>
<td>15</td>
<td>24</td>
<td>3</td>
</tr>
<tr>
<td>TH01</td>
<td>16</td>
<td>24</td>
<td>3</td>
</tr>
<tr>
<td>D02</td>
<td>18</td>
<td>29</td>
<td>4</td>
</tr>
</tbody>
</table>

Notes: Distance A is the seafloor width projected on the horizontal plane. Distance B is the plate boundary width projected on the horizontal plane. Thickness is the maximum thickness of the prism measured perpendicular to the plate boundary (see text Fig. 8).
We next examined the distribution of the frontal prism in relation to the source faults of tsunami earthquakes. As described in the Introduction, large coseismic slips with lower levels of high-frequency seismic energy are observed close to the trench. For example, the 1896 Sanriku earthquake, whose estimated source fault was close to the trench (Tanioka and Sataka, 1996) (Fig. 10), can be considered a typical tsunami earthquake. The cause of tsunami earthquakes is still under debate. Previous geophysical and geological studies have suggested that the physical and chemical properties of sedimentary units in the trench around the plate boundary, such as a very low dynamic frictional coefficient (Ujiie et al., 2013; Fulton et al., 2013) and uplift of the seafloor caused by the deformation of the frontal prism (Tanioka and Seno, 2001), primarily control the fault slip behavior of a tsunami earthquake. Large coseismic slips with tsunami earthquake slip characteristics seem to occur in the area where the frontal prism is well developed (Fig. 9). Therefore, the development of a frontal prism that consists of low-velocity pelagic sediments may affect the cause of a tsunami earthquake, at least along the Japan Trench.

Tsunamigenic fault slip along a splay fault in the Nankai seismogenic zone has been discussed (e.g., Park et al., 2002). The splay fault imaged in the Nankai Trough is a clear landward-dipping reflector within a low velocity accretionary wedge and is terminated at a plate interface. Due to the structural characters of the backstop interface, which resembles the splay fault in the Nankai Trough, we speculate that the backstop interface could act as a splay fault where a coseismic slip propagates. Note that a time-lapse bathymetry survey along TH03 (Fig. 6) (Fujiwara et al., 2011) that images coseismic seafloor displacements during the 2011 Tohoku-oki earthquake does not show any significant coseismic displacement at the seafloor where the backstop interface is imaged. However, this does not imply that the backstop interface is no longer seismically active.

Reflective Zone

According to von Huene et al. (1994), the tectonized block consists of a series of landward-dipping reflectors between the frontal prism and the continental block. The tectonized block is interpreted as a fragmented and detached part at the seaward tip of the continental framework. The deformed zone, as defined by Tsuru et al. (2000, 2002), consists of the frontal prism and the tectonized block of von Huene et al. (1994). In this study we refer to the tectonized block as the reflective zone owing to its seismic characteristics.

In section D19, a clear landward-dipping reflector is imaged between CDP (common depth point) 11700 and 10100 at 6–8 km depth, and landward of this reflector, a series of discontinuous landward-dipping reflectors with a thickness of ~5 km is imaged subparallel to the subducting oceanic basement (Fig. 3). In section D17, a strong landward-dipping reflector imaged from CDP 17000 to CDP 14400 at 9–11 km depth terminates at the subducted oceanic basement (Fig. 4). Discontinuous reflectors subparallel to the oceanic basement, rather than clear landward-dipping reflectors, are dominantly observed in the reflective zone in sections D11 (Fig. 5; CDP 15000–18000), TH03 (Fig. 6; CDP 6800–10000), and D02 (Fig. 7; CDP 11000–13000).

Ranero and von Huene (2000) illustrated a reflective subduction channel in the Costa Rica subduction zone extending 50 km landward from the trench axis to 10 km depth beneath the overriding plate. Bangs et al. (2015) also reported this reflective subduction channel farther southeast along the Costa Rica subduction zone; in a volume of three-dimensional seismic images they showed...
amplitude variations indicating variable porosity with depth. On the basis of the results of drilling, seismic, and bathymetric surveys in the Central America and Japanese subduction zones, von Huene et al. (2004) proposed a generic model of subduction erosion, in which a zone of hydrofracturing caused by high pore pressure fragments at the base of the overriding plate and facilitates removal of material in the subduction channel. Kimura et al. (2012) quantitatively estimated the distribution of high pore fluid pressure in the Japan Trench by using a thermal model and considering the dehydration kinetics of opal-A to quartz and the transformation of smectite to illite; their results suggest that dehydration of underthrust sediments takes place at a distance of 40–80 km from the trench and peaks at 50–60 km landward of the trench axis. The reflective zone identified in this study 30–80 km landward from the trench occurs within the dehydration range of underthrust sediments estimated by Kimura et al. (2012), and the newly processed seismic sections of this study show that the reflective zone at the base of the overriding plate is not a thin layer but is locally as much as ~5 km thick. This result suggests that high pore fluid pressure can extend to ~5 km above the subducted igneous oceanic crust.

**Subducted Oceanic Crust**

Previous seismic reflection studies in the Japan Trench focused mainly on seismic reflective structures along the plate boundary interface beneath the continental slope (von Huene et al., 1994; Tsuru et al., 2000, 2002). The supporting seismic images did not image the plate interface below a depth of 15 km. In this study we image the deeper part of the interface with a larger airgun array, which allowed reception of signals from the plate boundary interface to a depth of ~25 km (Fig. 2).

The plate boundary was not imaged continuously along the profiles, but the two northern profiles (D19 and D17) show clearer images of the plate boundary than the more southern profiles. Along both D19 and D17 it is possible to trace a landward-dipping reflector, interpreted as the plate boundary interface, down to a depth of 25 km; the plate boundary reflector is clearly imaged at intervals of ~10 km (Fig. 2; D19, CDP 11700–24500 and D17, CDP 3200–14400). This spacing appears similar to the spacing of the horst and graben structures observed on the incoming plate and on the subducted oceanic crust beneath the frontal prism and reflective zone (Fig. 10). Tsuru et al. (2000) showed that the horst and graben structures become larger landward from the outer rise of the trench. At 100 km seaward from the trench, the normal fault throw is ~100 m, whereas along the subducted oceanic crust at 30 km landward from the trench, it is ~800 m; the throw is even larger (~2 km) 50 km landward from the trench (Figs. 3–7).

**Figure 11.** (A) Enlarged view of the TH03 section between CDP (common depth point) 21245 and 13245. (B) Enlarged view of the D02 section between CDP 1601 and 9601. The reflector interpreted as the plate boundary is indicated by black dots; 2 km landward shallowing of the plate boundary is observed. The arrows and square brackets indicate the areas where landward shallowing of the plate boundary is observed.
among multiple reflections or noise, the shallowing plate boundary interface is observed in all three southern profiles (D11, TH03, and D02) at the same depth (18–22 km) and the same distance from the trench axis. If this shallowing is real, a possible interpretation is that there is a subducted seamount with a diameter of ~100 km at ~20 km depth. Moreover, the hypocenter (i.e., the initial break point of the fault) of the 2011 Tohoku-oki earthquake was at ~20 km depth in the middle of our survey area (Fig. 1, star).

The possible presence of a seamount around the epicenter is suggested by the results of a moment tensor inversion of regional broadband strong-motion waveforms of the Tohoku-oki earthquake (Kumagai et al., 2012). Kumagai et al. (2012) inferred that a strong localized asperity with an estimated radius of ~70 km ruptured around the epicenter, leading to a slip of ~50 m accompanied by a large stress drop (~40 MPa). Kumagai et al. (2012) suggested that one possible origin of the localized asperity would be a subducted seamount, but they indicated that more seismic data need to be acquired around the epicenter to conclude this. Since their study was published, however, no subducted seamount has been imaged.

**Depth-Varying Structural Features and Seismicity**

Although depth-varying physical, chemical, and geological properties in a subduction seismogenic zone have been examined (e.g., Hyndman and Wang, 1993; Lay and Bilek, 2007; Kimura et al., 2012), no depth-varying seismic slip behaviors were well documented before the 2004 Sumatra-Andaman earthquake. During the last decade, global and local broadband observations of great earthquakes, such as the 2004 Sumatra-Andaman earthquake, the 2010 Chile earthquake, and the 2011 Tohoku-oki earthquake, have revealed depth variations of seismic slip behaviors. In particular, Lay et al. (2012) described depth-varying seismic slip behavior along the megathrust of recent great earthquakes by examining globally observed broadband seismic data; as a result, they divided the megathrust zone from the trench to the downdip end of the seismogenic zone into four domains: domain A (trench to 15 km depth), tsunami earthquake region; domain B (15–35 km depth), large coseismic fault displacement region with modest short-period seismic wave radiation; domain C (35–55 km depth), region in which small patches produce bursts of short-period seismic wave during large ruptures or, sometimes, during repeated smaller ruptures; and domain D (at the lower part of domain C), region where low-frequency earthquakes, non-volcanic tremor, and slow-slip events are dominant in subduction zones of a young oceanic lithosphere. Here we describe the structural characteristics of domains A–C in the Japan Trench on the basis of the seismic images as well as seismic activity recorded by on-land and ocean-bottom seismic networks (Yamamoto et al., 2014).

Hypocenters of earthquakes observed by land and ocean-bottom seismic networks before and after the 2011 Tohoku-oki earthquake (Yamamoto et al., 2014) were plotted on the seismic reflection images of profiles JFD1 (Nakamura et al., 2014) and TH03 (Fig. 12). Then, by comparing seismic structures, seismicity, and slip domains, we characterized the slip domains A–C on these profiles. In domain A, a low-velocity frontal prism, a large subducted horst and graben structure, and a reflective zone above the oceanic crust were imaged. Background seismicity in this domain before the 2011 Tohoku earthquake (gray dots in Fig. 12) was very low, but aftershocks of the 2011 Tohoku-oki earthquake (black dots in Fig. 12) occurred in the mantle immediately below the Moho with mostly normal focal mechanisms (Obana et al., 2013). In domain B, subducted horst and graben structures were imaged. The background seismicity along the subducted oceanic crust seems to be low, but another seismicity study (Suzuki et al., 2012) shows that the seismicity along the plate interface extends as much as 40 km from the trench axis, which corresponds to the upper end of the domain B. Aftershocks were observed in the overriding plate, within and below the oceanic crust. The aftershock activity along the plate interface was very low. In domain C, background seismicity along the plate interface is high beginning around 25 km depth and continuing downdip. Aftershocks were distributed along the plate interface, in the mantle, and in the overriding plate.

**CONCLUSIONS**

We applied PSDM to seismic reflection data acquired in the rupture zone of the 2011 Tohoku-oki earthquake and examined depth-varying structural features along the megathrust fault. The resultant characteristic structures imaged by PSDM were a low-velocity frontal prism at the trenchward tip of the overriding plate, a reflective zone at the seaward end of the coherent continental framework above subducted oceanic crust, and subducted horst and graben structures that could be traced down to ~25 km depth.

The frontal prism had a seafloor width ranging from 13.5 to 18.0 km and was characterized by velocities (Vp) of 2.0–3.5 km/s. We considered the size and distribution of the frontal prism in this study together with data from a previous study and found that the frontal prism along the Japan Trench is well developed north of 37.5°N and extends to the northern end of the Japan Trench. The structure of the frontal prism transforms abruptly at 37.5°N to an elongate unit in the south. The association of the frontal prism and the large slip zone of the 2011 Tohoku-oki earthquake as well as the fault zone of the 1896 Sanriku earthquake indicates that tsunami earthquakes with large shallow slip have occurred where the frontal prism is well developed. A reflective zone, a zone with a thickness of ~5 km that consists of a series of discontinuous landward-dipping reflectors, was observed at the seaward end of the coherent continental framework, from 30 to 80 km landward of the frontal prism. We inferred that high pore fluid pressure caused by dehydration of the subducted sediment contributes to the formation of the reflective zone. The subducted oceanic crust was imaged from the trench axis to a depth of ~25 km. Clear horst and graben structures were imaged beneath the frontal prism and the reflective zone. These images show that the throws of the normal faults associated with the horst and graben structures are larger by as much as ~2 km beneath the reflective zone. Landward of the reflective zone, the deeper plate interface is imaged as a discontinuous series (at intervals of ~10 km) of landward-dipping reflectors. We interpreted these reflectors as the top of subducted horsts.
Figure 12. Earthquake hypocenters determined by on-land and ocean-bottom seismic networks (Yamamoto et al., 2014) superimposed on seismic images. (A) Profile JFD1 (Nakamura et al. 2014). (B) Profile TH03. Black and gray dots indicate hypocenters of earthquakes observed after and before the 2011 Tohoku earthquake, respectively. (C) Schematic diagram of the rupture domains of an interplate megathrust (after Lay et al., 2012). Domain A is the near-trench domain where tsunami earthquakes or aseismic deformation and stable sliding can occur. In domain B, the central megathrust domain, large slip can occur with minor short-period seismic radiation. Domain C, the downdip domain, is where moderate slip occurs with significant coherent short-period seismic radiation. LFE—low-frequency earthquake. Domain D is a transitional domain, mostly present in a young subducting plate and shallow megathrust dip, where slow slip events and seismic tremor can occur.
By considering seismic images and seismicity observed by both on-land and ocean-bottom seismograph networks, we identified the following depth-varying structural features: domain A, where tsunami earthquakes occur, is characterized by low levels of short-period seismic energy radiation; a well-developed low-velocity frontal prism and reflective zone, and low seismicity along the plate interface; in domain B, where large coseismic slips with low levels of short-period seismic energy are observed, subducted horst and graben structures are imaged and background seismicity along the plate interface is very low but may extend to the upper end of domain B; in domain C, which is deeper than 25 km, clear seismic images were not obtained, but landward-dipping background seismicity along the plate boundary was observed. Therefore, the plate interface at this depth is characterized by high seismicity.

ACKNOWLEDGMENTS

This study was supported by KAKENHI Grants-in-Aid for Specially Promoted Research, number JP26000002, and for Scientific Research (S), number JP15H06718, from the Japan Society for the Promotion of Science, and by the JAMSTEC Japan Agency for Marine-Earth Science and Technology research fund. We thank JGI, inc., for help with processing the seismic data by prestack depth migration. Comments by David Scholl, Roland von Huene, and Gaku Kimura greatly improved this manuscript.

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