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Relation between seismic velocity structure of subducting oceanic crust and interplate micro-seismicity

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Abstract

I present a seismic velocity structure model to explain strong seismic reflections from the aseismic part of the plate boundary recently reported from an airgun-OBS survey in the northern Japan trench subduction zone. Although no P wave velocity anomaly models could explain the reflection amplitudes and the first arrival travel times consistently, it is possible to simulate significant amplitudes observed in relative small offset ranges using a model with a low S wave velocity in the subducting oceanic crust. The upper oceanic crust with the low S wave velocity anomaly may reduce the shear strength of the plate boundary thrust, as fault damaged zones with a considerably low S wave velocity make faults weak. The result of this study indicates the importance of clarifying the S wave structure in the vicinity of the plate interface to understand the frictional propeties of subduction megathrust seismogenic zone. Not only conventional seismic surveys but also seismic observations in deep boreholes are required to do this.

Key words: OBS seismic experiment, S wave velocity structure, fault strength, interplate seismicity

1. Introduction

The landward slope of the Japan trench is a zone of intense interplate seismicity, including both large thrust events with M>7 recurring with intervals of less than a hundred years and microearthquakes occurring stationarily. Such interplate seismicity is known to have considerable spatial heterogeneities (e.g. Kawakatsu and Seno, 1983) and the spatial extent of rupture areas of large earthquakes (e.g. Yamanaka and Kikuchi, 2002) and also the spatial distribution of small events are considered to reflect the heterogeneous structure of the interplate coupling (e.g. Umino *et al.,* 1990).

It is natural to relate spatial variations of the seismic velocity structure around the plate boundary to the patterns of interplate earthquake activities, and to seek parameters controlling earthquake generation along the subduction megathrust. Based on such motivations, numerous seismic surveys have been conducted for more than twenty years in this area (e.g. Suyehiro and Nishizawa, 1994). Especially in the northern part of the region, the off-Sanriku area, detailed seismic imaging experiments have been attempted in the last decade, because shallow interplate seismicity has been well studied through recent ocean bottom seismographic (OBS) observations (Nishizawa *et al.,* 1990; Hino *et al.,* 1996; Matsuzawa *et al.,* 1999). Figure 1 shows the locations of recent seismic surveys performed in the off-Sanriku area by Takahashi *et al.* (2000), Tsuru *et al.* (2000) and Fujie *et al.* (2002).

Tsuru *et al.* (2000) performed multi-channel seismic reflection profiling along a line in the dip direction of the subduction zone and showed fine images down to the seismogenic part of the plate boundary (>10 km in depth). Their result showed that there is a prism-shaped layer of low P wave velocity materials beneath the inner trench wall. The spatial extent of the layer corresponds to that of the aseismic region pointed out by OBS seismicity studies (e.g. Hirata *et al.,* 1985), suggesting that the seismic structure of the overriding plate is one of the important factors controlling interplate seismicity.

The line 'ERI96L2' shown in Fig. 1 is the survey

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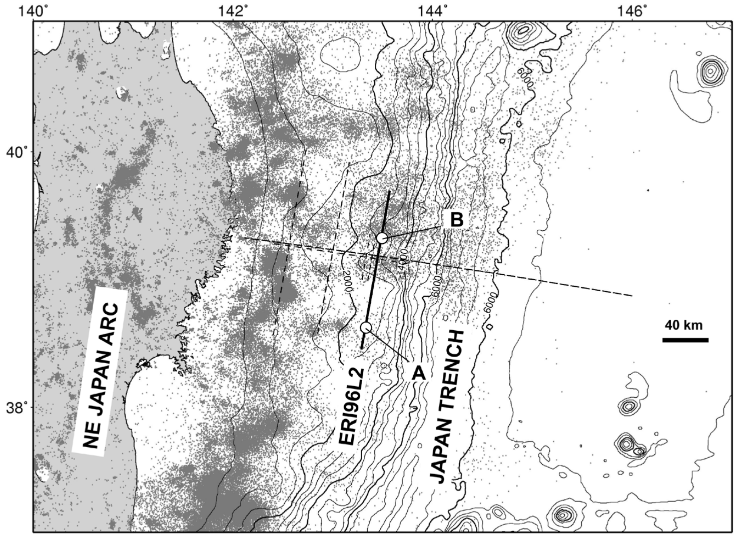

Fig. 1. Location map of seismic surveys conducted in the northern Japan trench area in the last decade. Dashed lines are profiles of experiments by Takahashi *et al.* (2000) and Tsuru *et al.* (2000). Thick solid line (ERI96L2) is the location of the airgun-OBS profiling by Fujie *et al.* (2002) showing the correlation between seismic reflective intensity and microseismicity along the plate boundary. Open circles are locations of the OBS stations at the aseismic region (site-A) and at the seismic region, whose records are shown in Figs. 2a and 2b, respectively. Grey dots are epicenters of shallow (focal depths < 60 km) earthquakes located by Tohoku University from 1992 to 2002. Water depths are in meters and isobath interval is 500 m.

line of the airgun-OBS wide-angle seismic experiment performed by Fujie *et al.* (2002). They found that the amplitudes of the seismic reflection signals from plate boundary varied considerably in space, and strong reflections tended to be observed in the region without background microseismicity along the plate boundary, suggesting another significant correlation between seismic structure and interplate seismicity. They interpreted the cause of the high reflection intensity at the plate boundary to be the existence of a thin layer with a very low seismic velocity. They discussed that the thin interplate layer might be aqueous fluid or hydrated materials, which reduce mechanical coupling and thus make the plate boundary aseismic.

The correlation between the seismic activity and the seismic velocity structure along the thrust fault of the off-Sanriku area provides a very important clue to understanding the frictional properties of natural seismogenic faults, little of which have been clarified. However, the thin low seismic velocity layer along the fault may be only one of the possible solutions explaining the large reflection amplitudes and there may be alternative models explaining intense seismic reflections from the aseismic part of the plate boundary interface. In this paper, I re-examine the OBS records showing the seismic reflections from the plate boundary obtained in the off-Sarniku experiments to seek an alternative model.

2. Appearances of wide reflection arrivals from subducting oceanic crust in OBS records

Let me begin with a review of the characteristics of the airgun-OBS seismic data on which the wide angle reflection arrivals from the plate boundary appear. Figure 2 shows two typical airgun-OBS profiling records that Fujie *et al.* (2002) used for their analyses. These two records are for the vertical component and were obtained along the same profile (ERI96L2 in Fig. 1). One is from the OBS located above the plate boundary without microseismicity (site-A, in Fig. 1) and the other is from that located at the region of active seismicity (site-B).

These records manifest evident differences in appearances of the later arrivals in the distance range of 10–30 km. In this range, we see a clear branch of later arrivals (PB) at 5–5.5 sec in reduced travel times in the site-A record (Fig. 2a). Fujie *et al.* (2002) showed that this branch is the reflection phase from the plate boundary located at a depth of about 14 km from their travel time inversion analyses. Meanwhile, the reflection phase PB is not apparent as

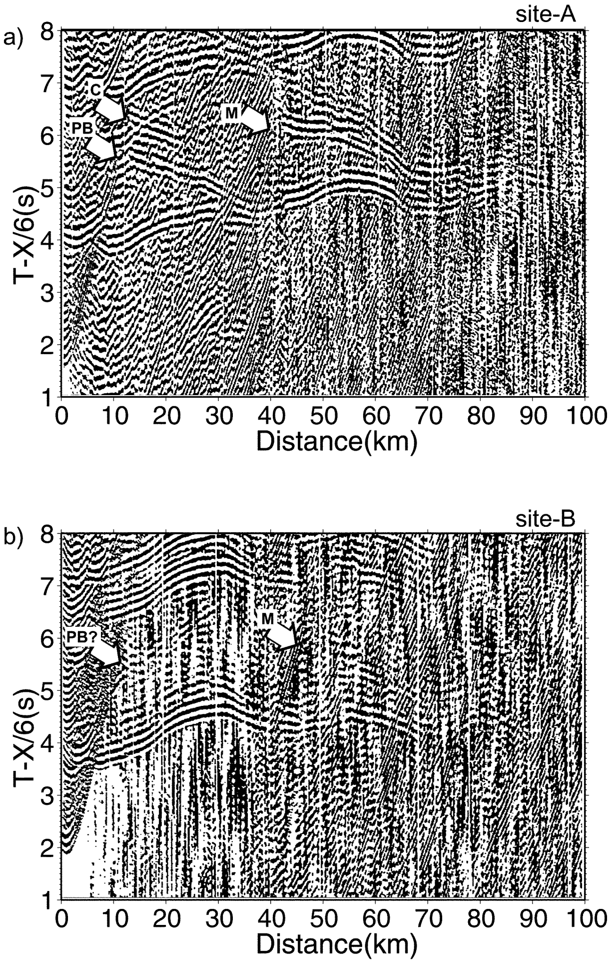

Fig. 2. Typical record sections (vertical component) obtained by the airgun-OBS surveys along the line ERI96L2. Each trace is band-pass filtered (5–15 Hz) and only positive polarity parts are plotted as lines. a) Records obtained at site-A located in the region of inactive interplate seismicity. Three coherent later arrivals are clearly observed; PB: reflection from the plate boundary, C: reflection from the reflector within the upper oceanic crust and M: reflection from the Moho of the oceanic crust. b) Records at site-B in the region of active seismicity.

the evident later arrivals on the record at site-B (Fig. 2 b) in the corresponding offset and travel time ranges. Such observed differences in appearance of the reflections from the plate boundary form basis of the discussion by Fujie *et al.* (2002) claiming that there is the distinctive relation between reflectivity and seismicity along the subduction plate boundary.

In the record sections at sites A and B, later arrivals can be seen at 45–80 km and 5.5–6 sec in the offset and the travel time, respectively (denoted as 'M' in Figs. 4 a and b). These arrivals are explained as reflections from the bottom of the subducting oceanic crust, the oceanic Moho, assuming the Vp structure model presented by Fujie *et al.* (2002). The later phase M seems to be observed more clearly at site-A than at site-B, but this might be due to the low S/N ratio of the site-B records, which are more degraded by multiples of direct water arrivals from the previous airgun shots than those of site-A. Because the amplitude ratio of the first arrivals to the M arrivals is not significantly different, I did not think the difference in the appearance of the M phase between the sites A and B is substantial.

Examining the record sections shown in Fig. 2 a, the PB reflection arrivals seem to be accompanied by other later arrivals, delayed about 0.5 sec in the offset range of 15–25 km, as indicated by 'C' in Fig. 2 a. Although the apparent velocity of this later phase is almost the same as the later phase PB, the phase C seems not to be multiples of the PB, because no other arrivals including the first arrivals accompany the later arrivals with about 0.5 sec separation. Moreover, the offset ranges where the PB and C phases appear are different from each other. I interpret the later phase C as wide angle reflections from a reflector within the igneous oceanic crust. The seismic reflection profile by Tsuru *et al.* (2000) showed that there is a sedimentary layer covering the top of the subducting igneous oceanic crust, but the thickness of the sedimentary layer estimated at a depth of over 10 km is too small to account for the travel time difference between the later phases PB and C observed on our OBS record sections. Assuming P wave velocity (Vp) values generally obtained for the upper oceanic crustal layer (4–6 km/s, e.g. Spudich and Orcutt, 1980), the observed travel time difference corresponds to a depth of around 2 km to the reflector from the top of the plate boundary.

It is notable that neither the later phase PB nor the C phase is apparent in the records obtained at site-B located at the seismic active region. This led me to the idea that not only the reflection strength from the plate boundary but also that from the reflector within the upper oceanic crust has a correlation with interplate seismic activity. If the topmost layer of the oceanic crust, whose top and bottom are defined by the reflectors corresponding to the PB and C arrivals, changes in seismic structure within the layer, affecting observed amplitudes of both these two arrivals, I think some variations of the seismic velocity structure of the upper oceanic crust may have relation with the seismicity of the plate boundary.

3. A model explaining the strong reflections from the oceanic crust

Hereafter, the layer between the reflectors corresponding to the PB and C reflection arrivals is referred as 'target layer'.

As I explained earlier, this layer composes the upper part of the igneous oceanic crust, but not the sedimentary layer. As the thickness of the layer should be almost 2 km to explain the travel time difference between the PB and C phases, this layer may correspond to oceanic layer 2, the upper crustal layer of the standard oceanic crustal structure model (e.g. Spudich and Orcutt, 1980).

In this section, we examine how the amplitudes of the wide angle reflections PB (from the top of the target layer) and C (from the bottom) are affected by changing the seismic velocity within the target layer by calculating synthetic seismograms assuming four different velocity structural models. The waveform calculations were performed using an asymptotic ray theoretical code (Červený and Pšenčík, 1983), so only major arrivals are included in the seismograms obtained; the calculated record sections shown below contain refraction arrivals and reflections from oceanic crustal layers (PB, C, and M phases) and direct water waves, reflections from the acoustic basement, and multiple reflections are not included. Therefore, the discussion hereafter is based on rather qualitative comparisons of appearances of calculated and observed record sections, and I did not make quantitative amplitude analyses.

All of the trial structure models are 1-D velocity models, in which seismic velocity depends only on depth. Figure 3 shows P and S wave velocities (Vp and Vs) as a function of depth. From the travel time analyses by Fujie *et al.* (2002), the seismic velocity structure of the overriding plate is well determined and its variations along the profile direction are known to be small. Therefore, I made a 1-D velocity structure model by simplifying their 2-D model and used it for all of the calculations. Vp/Vs ratio is assumed to be 1.73 for all the layers in the overriding plate, except for the surface sedimentary layer in which Vp/Vs is assumed to be 3.0.

First, I adopted the Vp model of the Pacific oceanic crust obtained at the outer slope of the northern Japan trench (Nishizawa and Suyehiro, 1990) as the structure of the subducting oceanic crust. In this model (model-1) the upper layer of the igneous oceanic crust (layer 2) has an average Vp of about 5 km/s with a steep vertical increase. Vp at top of the layer (4.2 km/s) is smaller than that of the lowermost part of the overriding plate (6 km/s), the subducting oceanic crust becomes a significant low-velocity layer (Fig. 3), and this makes remarkable shadow and jump

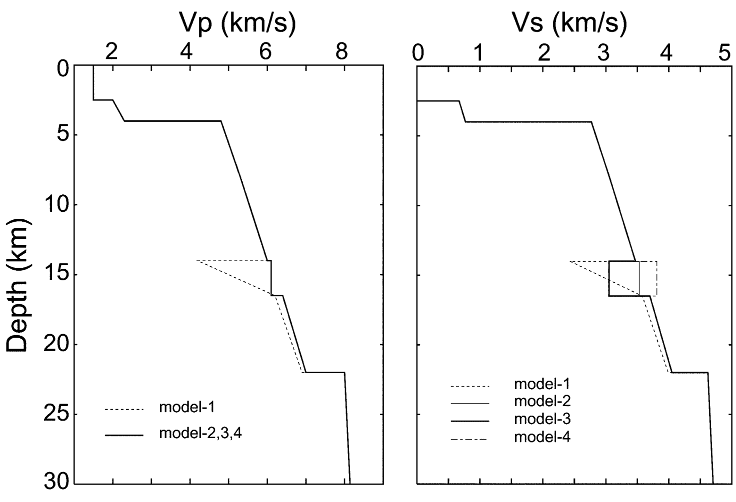

Fig. 3. Seismic velocity structure models for calculating synthetic seismograms shown in Fig. 4. For all of the models, the same Vp and Vs are assumed in the overriding plate underlain by the plate boundary at a depth of 14 km. See the text for detailed descriptions of the four models.

in the travel time curve of the first arrivals as seen in the calculated record section (Fig. 4 a). In the calculation, the Vp/Vs is also assumed to be 1.73 for the oceanic crust. Although large amplitudes of the PB arrivals can be expected, no significant shadows and gaps of the first arrivals are observed (Fig. 2), therefore, the model-1 cannot be a possible structural model for the subducting oceanic crust in the landward slope area. This suggests that the subducting oceanic crust changes velocity structure as it goes beneath the land plate.

Vp in the oceanic crust may exceed that of the overriding plate due to the seismic velocity change in the oceanic crust, and the resulting positive velocity jumps may cause reflection signals at the plate boundary. However, the velocity jump cannot be large enough to explain the large amplitude observed at the site-A (Fig. 2 a). Any models with averaged Vp larger than about 6.8 km/s in the oceanic crust cannot explain the first arrival travel time in offsets over 50 km, where the refractions from the oceanic crust become the first arrivals. Fig. 4b shows synthetic seismograms calculated for model-2 (Fig. 3), in which the velocity jumps at the PB and C reflectors are as large as possible under such constraints on the Vp in the oceanic crust. Also in this case, we assumed a Vp/Vs of 1.73 in the oceanic crust. The calculated amplitudes of reflected arrivals are too small to be observed evidently on the airgun-OBS record section. However, the observed record section at site-B (Fig. 2 b), without large amplitude PB and C phases, seems to resemble the synthetics calculated from model-2. Although the phase PB is not evident in offsets less than 30 km in the site-B records, it seems to be visible enough in larger offsets, and this variation of appearances seems to be presented in Fig. 4 b. The resemblance of the site-B record sections to the synthetics for model-2 suggests that the model is a good approximation of the seismic structure in the region with active microseismicity along the plate boundary.

The results of these simple waveform models indicate that neither positive nor negative Vp anomalies can explain the large amplitudes of reflections from the target layer and the travel times of the first arrivals consistently. However, by assuming the Vs anomaly, it is possible to explain both the reflection amplitudes and the first arrival travel

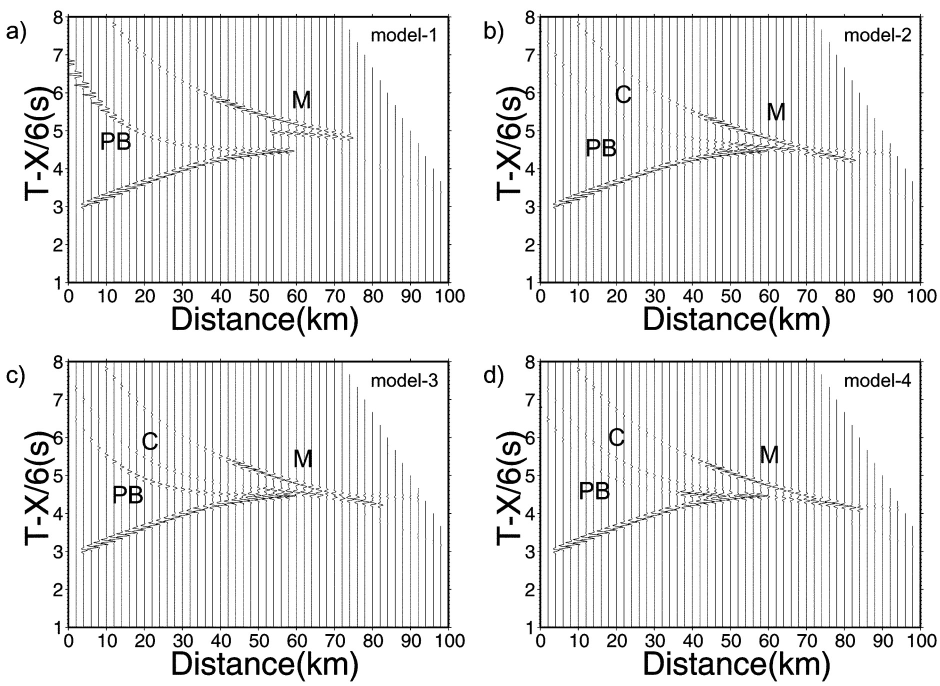

Fig. 4. Ray theoretical synthetic seismograms calculated from four different seismic structure models. Waveforms for model-1 to model-4 are shown in a) to d), respectively. Indexes denoting later arrivals are the same as in the observed record sections in Fig. 2. Direct water waves, reflections from the acoustic basement, and multiple reflections are not calculated.

times. Because the amplitudes of the PB and C phases are not only functions of Vp, but also of the Vs in the target layer, a large amplitude can be attributed to Vs anomalies in the target layer, while the Vs anomalies will not affect the travel times of the P wave first arrival. Next, I demonstrate how the amplitude of the PB and C phases varies by changing Vs in the target layer. In model-3 and model-4, the Vp structure is the same as in model-2, but I assumed a different Vp/Vs ratio for the target layer, 2.0 in model-3 and 1.6 in model-4 (Fig. 3).

Figure 4 c shows synthetic record sections for model-3 with large Vp/Vs in the target layer. Both the reflections from the top and the bottom of the target layer (PB and C, respectively) show significantly larger amplitudes than in the synthetics (Fig. 4 b) for model-2 (without Vs anomaly), especially in the offset range of 15–30 km as in the observed record section obtained at the site-A (Fig. 2 a). Meanwhile, for the amplitudes of the PB and C phases in these offsets of Fig. 4 d, the record sections calculated from model-4 are not large enough to explain the observed amplitudes. Therefore, it is strongly suggested that the intense wide angle reflections from the target layer are caused by the low Vs anomaly in the target layer, the upper oceanic crustal layer.

4. Discussion

Waveform modeling of the airgun-OBS seismic profiling of this study shows that the amplitudes of the wide angle reflection signals from the top of the subducting oceanic crust are sensitive not only to the P wave but also to the S wave velocity in the oceanic crust. Strong reflection signals are expected if the Vp/Vs ratio is larger in the upper oceanic crustal layer than its surroundings. In this model, no thin low-velocity layers along the subduction thrust are necessary as Fujie *et al.* (2002) assumed to explain the large amplitudes of the reflection arrivals. They argued that the thin layer is not only the cause of the strong reflections, but also of the low seismicity along the plate boundary. Is the present alternative model concordant with the aseismicity along the interface ?

Let us assume a model in which a fault is supported by several asperities with large shear strength, and the asperities are surrounded by weak apertures. In such a case, the shear stress in the vicinity of the fault concentrated around the asperities, and such stress concentration can increase microseismicity around the asperities. Hino *et al.* (1996) showed that the microearthquakes in the off-Sarniku region show a cluster structure and suggested that the cluster distribution represents the locations of asperities along the plate boundary. If this is the case, the region of low seismicity can be interpreted having weak apertures. Yamamoto *et al.* (2002) indicated that the existence of zones of highly fractured rocks along a fault zone, damaged zones, effectively lowering the shear strength of the fault and that the apertures are filled with such materials.

The damaged zones are often recognized as layers of considerably low Vs along fault zones evidenced by the fault zone trapped wave observations (e.g. Li *et al.* 2000). The upper layer of the oceanic crust (layer 2) is composed of extrusive rocks such as pillow basalt (e.g. Fowler, 1990) having considerable porosity (e.g. Wilkens *et al.,* 1991) and may have mechanical properties similar to the fractured rocks of the fault damaged zone. Therefore, the oceanic layer-2 with the low Vs anomaly can play a similar role to the fault-damaged zone, and may reduce the effective strength along the plate boundaries, acting as fractured rocks along faults as predicted by Yamamoto *et al.* (2002). In other words, the apparent correlation between the reflective strength of seismic waves and the aseismicity of the plate boundary found in the off-Sanriku region is a manifestation of the correlation between the fault strength and S wave velocity in the vicinity of the fault.

The present study indicates that the low Vs anomaly in the oceanic crustal layer is an alternative to the thin interplate layer model presented by Fujie *et al.* (2002), as an explanation of the high reflection intensity at the aseismic plate boundary. Analyses of S arrivals are crucial for validating the existence of the Vs anomaly, however, it is very rare to observe evident S arrivals from horizontal component records of OBSs, because they are often severely degraded by PS conversions and multiples in the subbottom sedimentary layers of refracted and reflected P phases. In the case of the OBS data obtained by the 1996 experiment, it was only the refracted S waves from the overriding plate that we could identify on the horizontal component records and could not estimate the Vs structure in the oceanic crust using S wave records. Although polarity data of the P arrivals reflected from the plate boundary may provide us another important constraint on the seismic structure of the subducting oceanic crust, we did not exploit them due to the low S/N of the reflected arrivals.

Therefore, efforts to improve the quality of OBS experiment data, for example, using huge volume of airgun array, are necessary. However it is difficult to tell which model is preferable only from the OBS survey data analyses, because the surface-based seismic survey data do not have sufficient resolution to constrain the detailed structure in the vicinity of the plate interface at depth uniquely. Since how the seismic structure heterogeneity affects frictional properties and earthquake genesis is quite different in these two models, it is desirable to do structural studies with much more resolution in the fault zone. Deep borehole-based research is the only way to obtain high-resolution structure information around the fault zone. For example, if we want to apply an analysis of the fault zone trapped waves, a powerful tool to see fine seismic structures of the fault zone in the land area, we need seismic sensor arrays across the fault, which can only be realized by drilling into the subduction thrust.

5. Conclusion

I present a model to explain large amplitude reflected seismic waves from the aseismic part of the plate boundary other than the thin interplate layer model suggested by Fujie *et al.* (2002). The observed large amplitudes can be explained by a low S wave velocity anomaly in the upper layer of the subducting oceanic crust, even if no P wave velocity anomalies are present. The upper oceanic crustal layer with a low shear wave velocity, or with large Vp/Vs ratio, reduces the shear strength at the plate interface, which may cause low microseismicity along the interface.

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