Estimation of Thickness of a Low-Velocity Layer at the Surface of the Descending Oceanic Plate beneath the Northeastern Japan Arc by using Synthesized PS-wave

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Abstract: The upper seismic plane of the double-planed deep seismic zone beneath the Tohoku District, Japan, has recently been found to be located within a thin low-velocity layer at the surface of the descending Pacific plate. In order to estimate the thickness of the low-velocity layer, we synthesized PS-waves (converted from P to S at the upper boundary of the Pacific plate) for events in the double-planed deep seismic zone. We used a two-dimensional-raytracing program developed by Červený and Pšenčík (1981) and took the effect of a multiple reflection in the low-velocity layer into consideration. The comparison between the synthesized and the observed waveforms of PS-waves indicate that the low-velocity layer is as thin as about 5 km below 100 km depth. Seismicity data show that the shallower part of the low-velocity layer may be slightly thicker than 5 km.

1. Introduction

Investigation of the velocity structure in the upper mantle is important in order to clarify the cause of the deep seismic zone. Usually, deep and intermediate-depth earthquakes are thought to be located in the high-velocity slabs (*e.g.*, Utsu, 1967). Recently, some authors reported that there is a low velocity layer (LV layer) at the surface of the descending slab.

The first indication of the existence of such a LV layer was done by Okada (1979) using amplitude of ScSp-phase. However, his LV layer is considered to be located just outside of the descending oceanic plate. Nakanishi (1980) investigated the relation between the LV layer and the polarity of the ScSp-phase beneath the northeastern and the southwestern part of Honshu, Japan. He concluded that the LV layer has a sharp upper boundary and a transitional lower boundary. Hasegawa *et al.* (1978b) reported that the plate boundary estimated from ScSp-phase almost coincides with the upper seismic plane of the double-planed deep seismic zone beneath the Tohoku District (northeastern part of Honshu, Japan). These two reports indicate that the upper seismic plane is located very close to upper boundary of the LV layer at the surface of the descending plate beneath the Tohoku District.

The descending LV layer in which seismicity is active is found beneath Pamir-Hindu

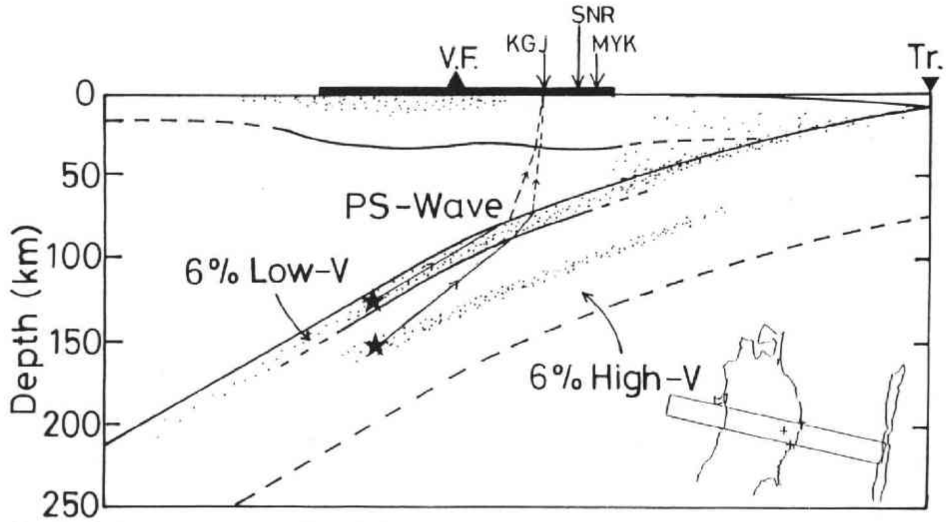

Fig. 1 Schematic representation of the upper mantle structure beneath the Tohoku District estimated by Matsuzawa *et al.* (1986b). Dotted areas indicate the region where seismicity is active. V.F. and Tr. represent the volcanic front and the trench axis, respectively. The locations of stations used in this study are shown by arrows (and crosses in the inserted map).

Kush (Roeker, 1982) and Hokkaido (Miyamachi and Moriya, 1984) by using Aki and Lee's method. Hori *et al.* (1985) found a clear later phase in the seismograms of subcrustal earthquakes beneath the Kinki District, Japan. They interpreted this phase as a guided wave traveling along the LV layer at the surface of the descending Philippine-Sea plate.

Matsuzawa *et al.* (1986b) found PS-waves (converted from P to S at the upper boundary of the descending plate) and investigated the upper mantle velocity structure beneath the Tohoku District using PS-P time. They concluded that the upper seismic plane of the deep seismic zone is located in a thin LV layer (6% lower than in the overriding mantle) at the surface of the descending Pacific plate (Fig. 1). However, it is difficult to determine the thickness of the LV layer by using only PS-P time data because PS-P time is not so strongly depend on the thickness of LV layer. The waveform of the PS-phase must have some other information about the LV layer. It is also very important, of course, to ascertain whether or not these models can explain the observed waveform.

In this paper, we synthesize the waveform of PS-phase and estimate the thickness of the LV layer.

2. Synthesized PS-phase

Figure 2 shows an example of seismograms at station KGJ for an event in the upper seismic plane of the double-planed deep seismic zone. Seismometers used here are 1 Hz velocity type. PS-phase is clearly seen between P- and S-phases in horizontal components. We synthesized the PS-phase for various models and ascertain whether the

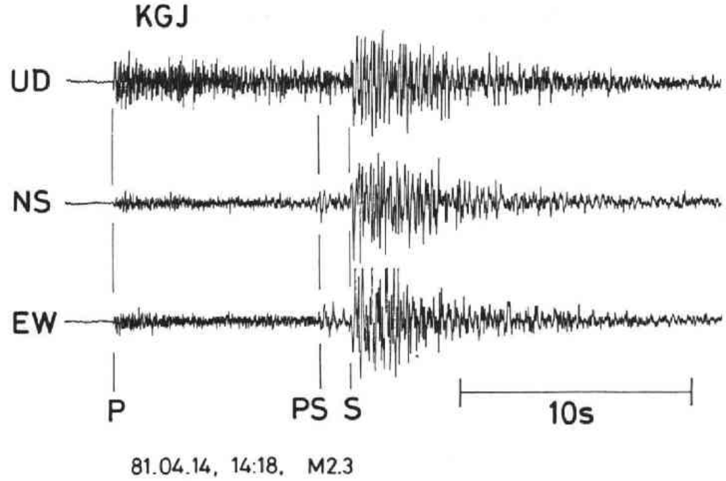

Fig. 2 An example of PS-phase observed at station KGJ. The PS-phase is clearly seen in the horizontal components.

synthetic waveform can explain the characteristics of the observed waveform.

At first, we calculated waveforms for an event in the upper seismic plane of the double-planed deep seismic zone. The velocity structure and the hypocenter location used for the synthesis of waveforms are shown in Fig. 3. The velocity structure above the descending slab is the same as the one used for the hypocenter determination at Tohoku University (Hasegawa *et al*., 1978a). In the slab, the LV layer and the HV layer have 6% lower and 6% higher velocity than the overriding mantle at the same depth, respectively, according to Matsuzawa *et al.* (1986b). The wave-theoretical calculation for such structure as shown in Fig. 3 is very difficult. We used the two-dimensional-raytracing program developed by Červený and Pšenčík (1981) and took account of the effect of a multiple reflection in the LV layer. In each wavelet, rays reflected many times in the LV layer, whose amplitudes are greater than one hundredth of the largest ray, are all included.

Another important factor which will affect the waveforms of PS phase is the focal mechanism of the event. PS-wave is hardly observed for events whose focal mechanism solutions could be determined, because the seismograms for these events are saturated at almost all the stations. We adopted here the down-dip compression type focal mechanism. This type of focal mechanism is predominant in the upper seismic plane at the depth of the hypocenter shown in Fig. 3 (Matsuzawa *et al*., 1986a).

The synthetic waveforms are computed for three velocity models with no LV layer (Model I), 5 km-thick (Model II) and 10 km-thick LV layer (Model III); and they are shown in Fig. 3. S-phase and converted phases related with Moho are neglected here. The PS-phases calculated for Model I (top traces) have very small amplitudes, while those for Model II (middle) and for Model III (bottom) have large amplitudes. Characteristics of the observed waveform of PS-phase are (i) its large amplitude (in some cases

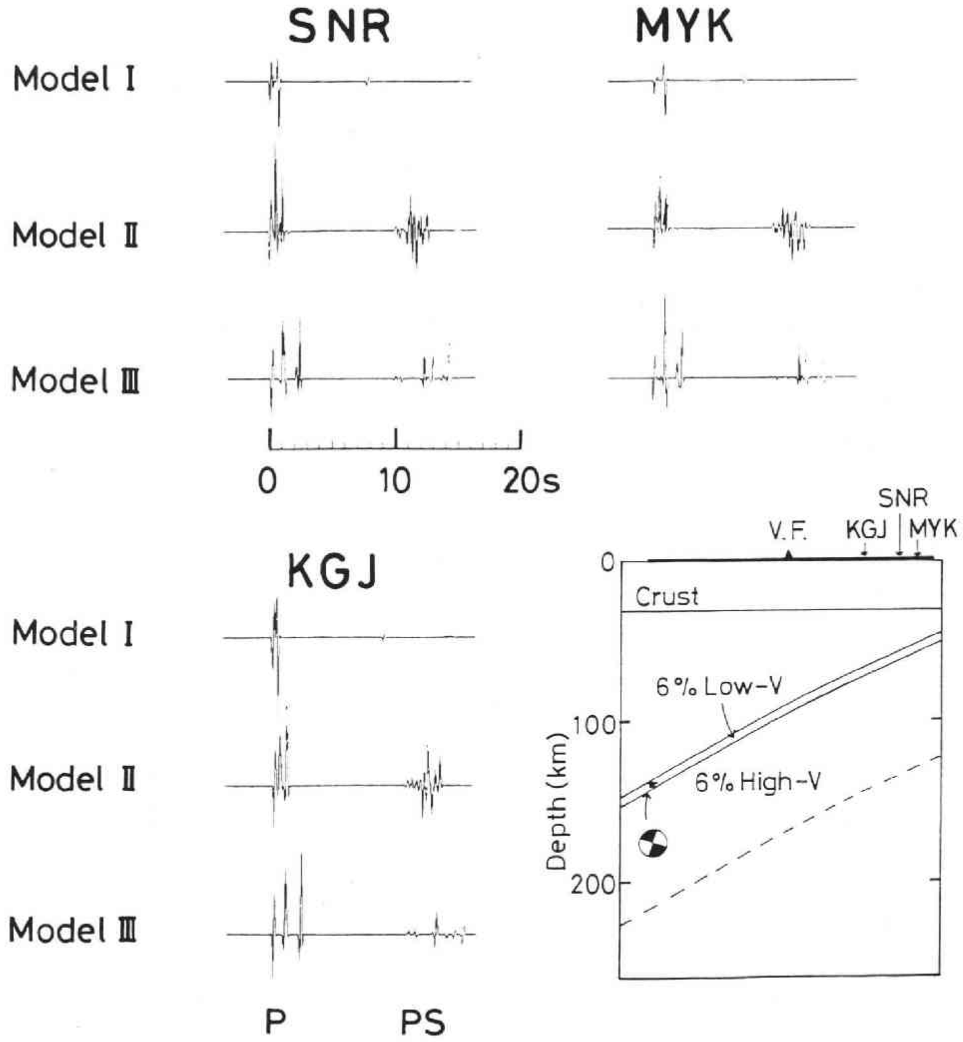

Fig. 3 Synthesized waveforms of horizontal component for an event in the upper seismic plane. The hypocenter location and the focal mechanism of the event are also shown. The focal mechanism is projected onto the vertical plane, and shaded quadrants indicate the compressional P-waves. Models I, II and III corespond to velocity models with no LV layer, 5 km-thick LV layer and 10 km-thick LV layer, respectively.

PS-phases are as large as a half of the P-wave), (ii) its dull onset and (iii) its spindly shape (Matsuzawa *et al.*, 1986b ; see also Fig. 2). We can see the synthesized PS-phases for Model II have waveforms similar to the observed ones.

Synthesized PS-phases for Model III in Fig. 3 have long durations owing to a multiple reflection in the LV layer. This model predicts the duration of about 5 seconds or more. Duration-frequency diagrams for the observed PS-phases are shown in Fig. 4. When the S-phase arrives before the end of coda of the PS-phase, we do not measure the duration of the PS-phase. Observed PS-phases whose durations are longer than 5

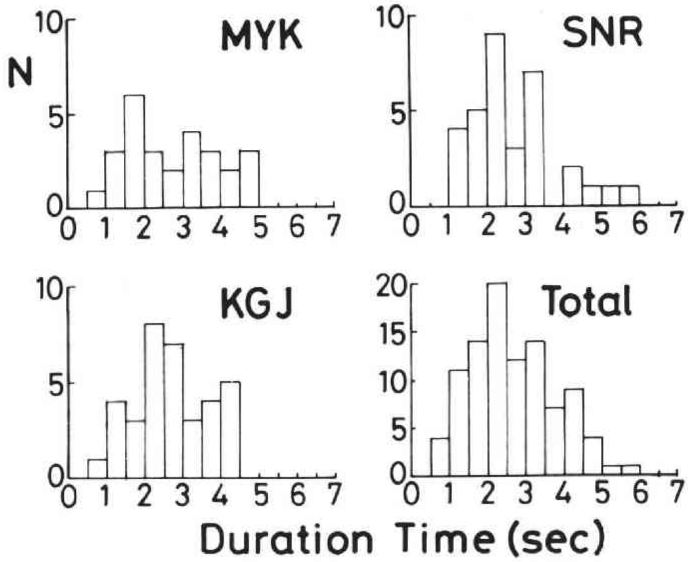

Fig. 4 Duration-frequency diagrams for observed PS-phases.

seconds are very few. Therefore, Model III cannot explain the observed waveforms.

As discussed above, 5 km-thick LV layer model (Model II) can fully explain the observed waveform of PS-phase for the event in the upper seismic plane of the double-planed deep seismic zone. Model I predicts the smaller amplitudes and the shorter durations of PS-phases than observations. However, the amplitude and the duration of a seismic wave are affected by the structure in the crust. Thus, we still cannot reject Model I from the amplitudes and durations of the PS-phases for the upper plane events only. Next, we compare the synthesized PS-phases for the lower plane event with those for the upper plane event to check whether we can reject Model I. Figure 5 shows an example of synthesized waveforms for a upper plane event (upper traces) and for a lower plane event (lower traces). Synthetic waveforms are computed for the two velocity models with no LV layer (Model I) and 5 km-thick LV layer (Model II) and are shown in the figure. The hypocenter and the focal mechanism for the lower seismic plane event are also shown in the figure. The conditions for the upper plane event are the same as the ones in Fig. 3.

For Model I, synthesized PS-phases for the lower plane event have larger amplitudes than those for the upper plane event. Although the absolute amplitudes of the phases are affected by the structure in the crust, the effects of the structure will be almost the same on the waves from the upper plane events and on those from the lower plane events. If Model I is correct, PS-phases will be detected mainly in seismograms for the events in the lower seismic plane. However, the SP-phase was detected in only 15% of the seismograms for lower plane events at stations KGJ, SNR and MYK. On the other hand, we could detect the PS-phase in 50% of the seismograms for upper plane events. We suppose lower detectability of PS-phase for lower plane events is due to low

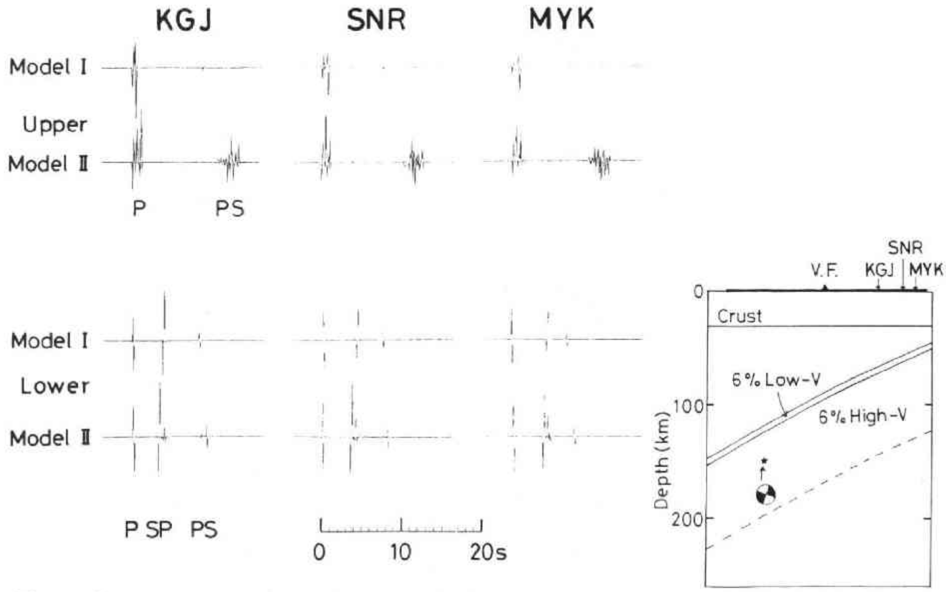

Fig. 5 Synthesized waveforms for events in the upper (upper traces) and the lower seismic plane (lower traces). The clear phase between P- and PS-phase for the lower plane event is SP-phase converted from S- to P-wave at the plate boundary. The hypocenter location and the focal mechanism for the event in the lower seismic plane are also shown on the right-hand side of the figure. For further detail, see the caption of Fig. 3.

amplitude of the phase. Model II predicts the amplitudes of PS-phases for lower plane events are smaller than those for the upper plane events. Thus, the 5 km-thick LV layer model (Model II) can explain the difference in detectability of PS-phase between the upper plane events and the lower plane events, while no LV layer model (Model I) cannnot explain the difference.

Matsuzawa *et al.* (1986b) reported that the observed PS-P time for the upper seismic plane event is 1~2 seconds longer than that for the lower plane event. The PS-P times calculated for Model I are nearly the same in both cases, while those for Model II are clearly different between the events in the upper and the lower seismic plane. The difference in PS-P time is another strong evidence for the existence of the LV layer.

3. Discussion

In the preceding section, we found that the 5 km-thick LV layer model is most probable. If the upper plane events occur in such a thin LV layer, the thickness of the upper plane seismicity must be less than 5 km. Figure 6 shows the vertical cross-section of the seismicity, projected onto a plane perpedicular to the trench axis, in the area indicated in the inserted map. In the figure, we show only the events whose standard errors of focal depths are less than 3 km located using more than nine stations. Most of the upper seismic plane events between the western seashore and the volcanic front are within the two solid curves whose distance is 5 km. On the other hand, between the

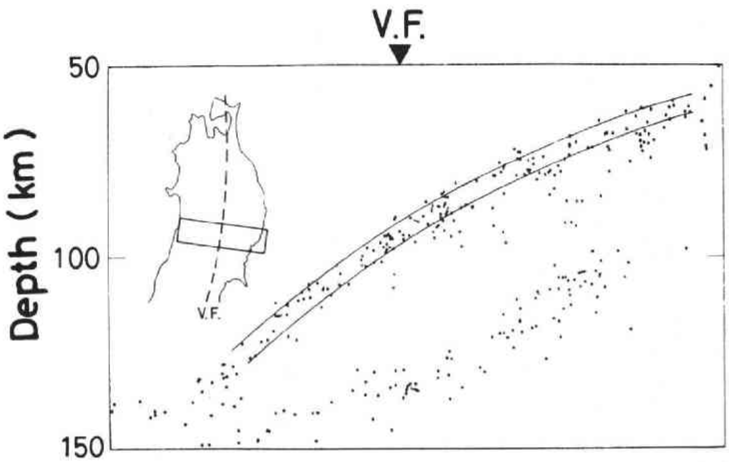

Fig. 6 Vertical cross-section of the seismicity of the deep seismic zone beneath the land area of the Tohoku District for the period from January, 1977 to July, 1985. Dots indicate the events whose standard errors of focal depths are less than 3 km and determined by using more than nine stations. V.F. denotes the volcanic front. Two solid curves whose interval is 5 km indicate the probable location of the LV layer.

volcanic front and the eastern seashore, the upper seismic plane seems to be thicker than 5 km and thinner than 10 km taking the errors into consideration. However, most of the data used in Fig. 4 are for the events on the western side of the volcanic front. Therefore, the thickness of the LV layer estimated in the preceding section almost coincides with the thickness of the upper seismic plane between the western seashore and the volcanic front.

It is expected that large events do not occur within such a thin LV layer. Figure 7 shows the epicentral distribution of intermediate-depth earthquakes beneath the Tohoku District determined by the seismic network of Tohoku University for the period from April, 1975 to July, 1985. Although microearthquake activity ($M \geq 2$) for the upper seismic plane is more active than that for the lower plane, large earthquakes with $M \geq 4.5$ did not occur in the upper plane on the western side of the volcanic front. We also searched the data file of JMA (Japan Meteorological Agency) for intermediate-depth earthquakes beneath the land area for the period from January, 1926 to March, 1975. No large events with $M \geq 5$ occurred in the upper seismic plane on the western side of the volcanic front for this period, while there occurred seven large events in the lower seismic plane. Although the hypocentral data in the early days are somewhat unreliable, relatively large events with $M \geq 5$ are not considered to be mislocated so badly. Large events with $M \geq 5$ occurred on the eastern side of the volcanic front even in the upper seismic plane. However, magnitudes for those events are not greater than 6.0, and probably some of these events are not intra-plate events but inter-plate ones (low-angle thrust type).

Stress drops of intermediate-depth earthquakes are reported to be almost 100bars (Mikumo, 1971). Using equations $\Delta\sigma = (7/16)\, Mo/r^3$ (Eshelby, 1957) and $\log Mo =$

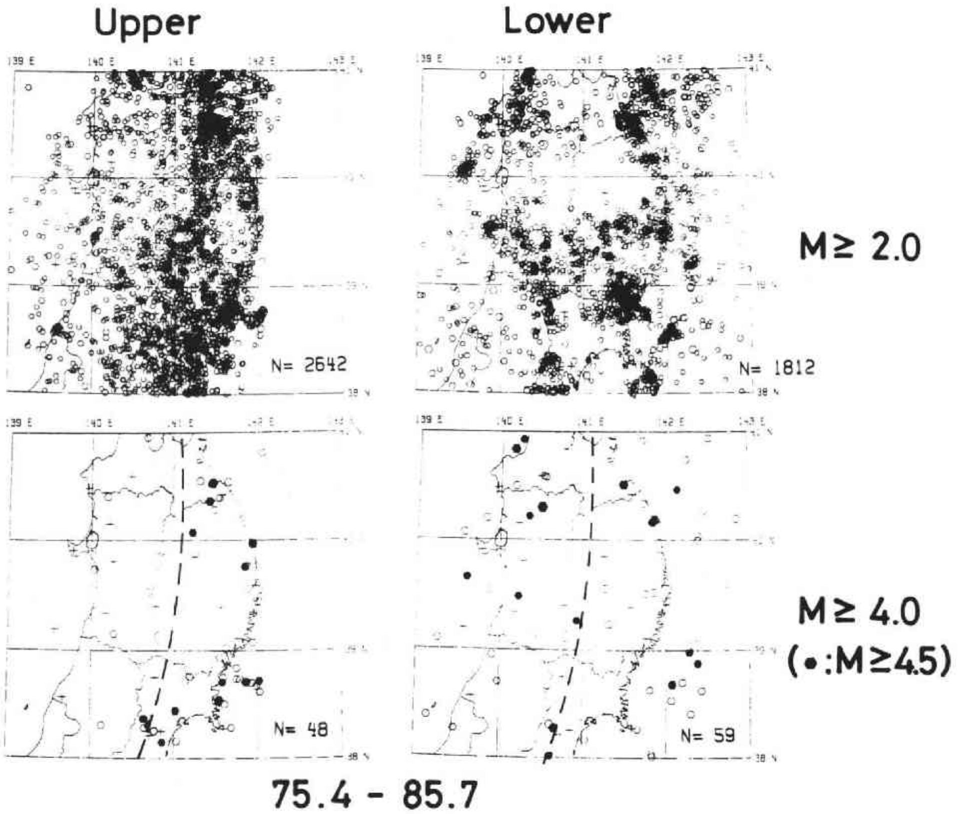

Fig. 7 Seismicity maps for $M \geq 2.0$ (top) and for $M \geq 4.0$ (bottom) events in the double-planed deep seismic zone beneath the Tohoku District. The upper and the lower seismic plane are shown on the left-hand and the right-hand side, respectively. In the bottom maps, solid circles indicate the events with $M \geq 4.5$ and dashed lines represent the volcanic front.

$1.5\,M+16.0$ (Aki, 1972), we obtain $r=3.5$ km for $M=6.0$ and $r=1.1$ km for $M=5.0$, where $\Delta\sigma$ is stress drop, Mo is seismic moment, r is source radius, M is magnitude. From this rough estimation, it is considered that events with $M<6.0$ are possible to occur within the 5 km-thick layer.

An exceptionally large intermediate-depth earthquake (M 6.6) occurred recently beneath the central part of Iwate (39.832°N, 141.764°E, 74.6 km) on January 9, 1987. Umino *et al.* (1987) reported that aftershocks of this event are divided spatially into two groups. The western group is located in the upper seismic plane and its dominant focal mechanism is down-dip compression type. On the other hand, the eastern group is located below the upper seismic plane and its characteristic focal mechanism is strike slip type. The major aftershock (M 5.8) belongs to the eastern group. The complicated aftershock distribution for this large event shows that this event is not an ordinary one but a very special case. A large event with $M>6.0$ like the present event cannot occur only within the thin upper seismic plane, but the inner portion of the oceanic plate must be also destroyed. Presumably a large event ($M\simeq 6.0$?) occurred at first in the thin LV layer, and triggered the rupture in the adjoining HV layer subsequently. Thus, the occurrence of the present event with M 6.6 is not inconsistent with our hypothesis that ordinary events in the upper seismic plane are located in a thin LV layer.

The apparent thickness distribution of the upper seismic plane is consistent with the fact that the largest event on the western side of the volcanic front is smaller than that on the eastern side. The eastern part of the LV layer may be slightly thicker than the western part.

One may think that earthquakes cannot occur in such a LV layer. However, earthquake occurrence in the LV layer is possible if the LV layer is brittle and has weak shear-strength. Generally, seismicities in the crust and in the lower seismic plane are inhomogeneous. We can easily find some aseismic regions there. On the other hand, as shown in Fig. 6 and Fig. 7, the seismicity in the upper seismic plane is almost homogeneous beneath the land area extending over several hundred kilometers, while its thickness is less than only 10 km. We cannot find conspicuous aseismic region in the upper seismic plane beneath the land area. Matsuzawa *et al.* (1986a) reported that the characteristic focal mechanism of the upper seismic plane beneath the volcanic front is normal fault type, while down-dip compression type events are dominant in other region of the upper seismic plane beneath the land area. It is impossible to explain the thin homogeneous seismicity in such inhomogeneous stress field by a simple model, and some mechanism of stress concentration or a thin weak zone should be introduced. This local weakness of shear strength is probably due to the difference in material. Of course, there is the possibility of an increase in pore pressure from dehydration of hydrous minerals or partial melt.

For further investigation, analyses of other phases in the seismograms and calculation of the stress distribution for a model including the LV layer are needed.

4. Conclusion

The thickness of the low-velocity layer at the surface of the descending Pacific plate is estimated by using the synthesized PS-waves (converted from P to S at the plate boundary) for events in the deep seismic zone beneath the Tohoku District, Japan. We used a two-dimensional-raytracing program developed by Červený and Pšenčík (1981) and took the effect of a multiple reflection within the low-velocity layer into consideration. The synthesized waveforms of PS-wave indicate that the low-velocity layer is as thin as about 5 km. The seismicity of the upper plane of the double-planed deep seismic zone shows that the shallower part of the upper seismic plane is apparently thicker than the deeper part, and the largest event in the shallower part is greater than that in the deeper part. The shallower part of the low-velocity layer may be a little thicker than the deeper part.

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