Double-planed deep seismic zone and upper-mantle structure in the Northeastern Japan Arc

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Summary. The ScSp wave converted from the ScS wave at the boundary between the descending lithospheric slab and the mantle above it was clearly observed from a nearby deep earthquake with magnitude 7.7 at some stations of the seismic network of Tohoku University which covers the Tohoku District, the northeastern part of Honshu, Japan. By applying the three-dimensional seismic-ray tracing method, the location of this boundary was determined from the difference in arrival time between the ScS and ScSp waves. The result shows that the upper boundary of the descending slab lies exactly on the upper plane of the double-planed deep seismic zone found in the Northeastern Japan Arc.

There is an additional evidence that the boundary is located on the upper plane of the double-planed deep seismic zone. The hypocentre distribution of intermediate-depth earthquakes located by the small-scale seismic-array observation is extremely different from that obtained by the relatively large-scale seismic network. The discrepancy in the distribution of hypocentres of the same earthquake independently located is well explained by the inclined lithospheric slab model derived from the difference in arrival time between the ScS and ScSp waves.

The earthquakes with reverse faulting or with down-dip compressional stresses occur at the upper boundary of the descending slab. Within the descending slab, the earthquakes with down-dip extensional stresses also occur in a very narrow zone from 30 to 40 km below the dipping boundary in the depth range from 50 to about 200 km, and these shocks form the lower plane of the double-planed deep seismic zone.

1 Introduction

It is widely known that there exists a deep seismic zone dipping downwards from the vicinity of an oceanic trench beneath an island arc and that deep and intermediate-depth earthquakes occur only within this dipping zone and not elsewhere. This deep seismic activity is related to the so-called descending high-Q, high-velocity slab beneath island arcs. In previous investigations on the shallow and intermediate-depth microearthquakes in the
Northeastern Japan Arc (Umino & Hasegawa 1975; Hasegawa, Umino & Takagi 1977), it has been revealed that the deep seismic zone in this region is distinctly separated into two planes, which are almost parallel to each other. The distance between the two planes is about 30 ~ 40 km. This prominent structure of the deep seismic zone has been found on the basis of the data from the seismic network of Tohoku University.

The relationship between the deep seismic zone and the high-velocity lithospheric slab which descends into the mantle has been discussed by many workers, and it has been considered that deep and intermediate-depth earthquakes occur within the descending slab. In order to investigate the subduction process beneath island arcs, it is necessary to have a better understanding of where deep and intermediate earthquakes are actually occurring. Mitronovas & Isacks (1971) reported that the upper boundary of the descending slab beneath the Tonga ridge was a velocity discontinuity and coincided approximately with the inclined seismic zone. They interpreted the unusual secondary phase between P and S arrivals observed at the stations in Tonga as the P to S-wave conversion at the boundary between the downgoing slab and the mantle above it. Okada (1971) noticed a pronounced longitudinal phase about 8 s before the normal ScS phase from several of the nearby deep earthquakes recorded at seismic stations in Hokkaido, northern Japan. He explained this phase in terms of the ScS to ScSp conversion at the boundary between the dipping high-velocity slab and the land-side low-velocity mantle, and located the conversion plane from the difference in arrival time between the two phases. The conversion plane, that is, the upper boundary of the descending lithospheric slab, was located near the upper boundary of the deep seismic zone. He also found the same phase at seismic stations in the Kanto District, the central part of Honshu, and obtained the same result as that in Hokkaido (Okada 1974). The location of the boundary between the descending slab and the surrounding mantle was also estimated in the Kurile Arc by Shimamura (1973). He interpreted the secondary phase between P and S arrivals observed at a station in Hokkaido as the P wave which is reflected and converted from the S wave at the boundary between the descending slab and the surrounding mantle. Based on this interpretation, he determined the location of the boundary, and revised the hypocentres of the earthquakes. The result shows that the revised hypocentres are located at the upper part of the descending slab about 10 km below the upper boundary of the slab.

In the Northeastern Japan Arc, the double-planed structure of the deep seismic zone was ascertained from an accurate determination of hypocentres, and such a pronounced structure of the deep seismic zone has not been detected in other regions hitherto. In order to understand the process of plate motion beneath the arc, we must know the geometrical relationship between the double-planed deep seismic zone and the descending lithospheric slab in more detail. In this paper, the location of the upper boundary of the descending slab is estimated by using the arrival time of the forerunning phase of the ScS wave observed at some stations of the seismic network of Tohoku University, and also by using the travel-time anomaly of intermediate-depth earthquakes observed at the small-scale seismic array situated at the central part of our seismic network.

2 ScSp wave generated by a large deep earthquake

A large deep earthquake took place in the western part of the Japan Sea on 1975 June 29. The magnitude of this event is 7.7 and the hypocentre is located at 38° 39'N, 130° 24'E and 600 km deep by the Japan Meteorological Agency. The location of this event is illustrated in Fig. 1 together with the stations of the seismic network of Tohoku University. The network was then composed of nine stations (15 stations at present). The locations and
Figure 1. Locations of stations of the seismic network of Tohoku University (solid circles) and the epicentre of a deep earthquake which occurred on 1975 June 29 (double circle). The broken line shows the axis of the Japan Trench. Seismic-array configuration at the Kitakami Seismological Observatory is illustrated at the bottom of the right-hand side of the figure.

the station codes are listed in Table 1. The station numbers indicated in the table correspond to those in the figure. Each station has the three-component velocity-sensitive seismometers with a natural frequency of 1 Hz. From each site, the amplified signals are transmitted by telephone telemetry to the central station at Sendai, which is numbered 5 in the figure. All the stations in the network detected very clearly the ScS wave generated by this deep shock about 12 min after the P arrival, and at several of the stations the forerunning phase of the ScS wave was also observed very distinctly.

Fig. 2 shows these phases recorded at each station. At the central station at Sendai, the data collected with telephone telemetry are recorded on the trigger system of 22-track FM magnetic tape recorders which start to work at the instant of quakes as well as on the monitor system of a 14-channel pen recorder with paper speed 2 mm/s. The records given in Fig. 2 are those reproduced from the magnetic tape recorders. The ScS phase is clearly seen on the horizontal components at each station as designated by open circles. In addition to this phase, a pronounced forerunning phase of the ScS wave is also seen on the vertical component of several stations denoted by solid circles. This phase is easily observable at stations HSK, MYK, KGJ and KWT which are located in the eastern part of the Tohoku District. At western stations OGA, FUT, HOJ and ATM, this phase is not observed. This phase cannot be seen at station AOB which is located in the eastern part. This is perhaps partly because of the small magnification of the instrument of this station, at which the magnification is about 20 dB less than at the other stations. The observed arrival times of the two phases at each station together with the locations of stations, epicentral distances and azimuths are presented in Table 1.
Figure 2. Tracings of seismograms showing the ScS phase (open circles) and the forerunning phase of the ScS phase (solid circles). The upper two traces and the lower trace at each station are the horizontal and vertical components, respectively. The magnification of the instrument in the horizontal components is about a half of that in the vertical component.
Table 1. List of stations and the observed arrival time of ScS and ScSp waves.

<table>
<thead>
<tr>
<th>No.</th>
<th>Station</th>
<th>Latitude (deg)</th>
<th>Longitude (deg)</th>
<th>Distance (km)</th>
<th>Azimuth (deg)</th>
<th>ScS phase (min)</th>
<th>Forerunning phase (min)</th>
<th>Lead time (s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>HSK</td>
<td>40.412 N</td>
<td>141.583 E</td>
<td>978</td>
<td>74.9</td>
<td>51</td>
<td>24.1</td>
<td>14.4</td>
</tr>
<tr>
<td>2</td>
<td>MYK</td>
<td>39.590 N</td>
<td>141.982 E</td>
<td>1004</td>
<td>80.4</td>
<td>24.8</td>
<td>15.1</td>
<td>7.2</td>
</tr>
<tr>
<td>3</td>
<td>KGJ</td>
<td>39.388 N</td>
<td>141.565 E</td>
<td>967</td>
<td>81.6</td>
<td>24.1</td>
<td>15.1</td>
<td>9.0</td>
</tr>
<tr>
<td>4</td>
<td>KWT</td>
<td>38.753 N</td>
<td>140.760 E</td>
<td>899</td>
<td>86.0</td>
<td>23.7</td>
<td>15.1</td>
<td>9.0</td>
</tr>
<tr>
<td>5</td>
<td>AOB</td>
<td>38.248 N</td>
<td>140.847 E</td>
<td>910</td>
<td>89.6</td>
<td>23.7</td>
<td>15.1</td>
<td>9.0</td>
</tr>
<tr>
<td>6</td>
<td>FUT</td>
<td>40.147 N</td>
<td>140.217 E</td>
<td>859</td>
<td>75.7</td>
<td>23.4</td>
<td>15.1</td>
<td>9.0</td>
</tr>
<tr>
<td>7</td>
<td>OGA</td>
<td>39.905 N</td>
<td>139.777 E</td>
<td>819</td>
<td>77.2</td>
<td>22.7</td>
<td>15.1</td>
<td>9.0</td>
</tr>
<tr>
<td>8</td>
<td>HOJ</td>
<td>39.338 N</td>
<td>140.173 E</td>
<td>848</td>
<td>81.8</td>
<td>22.7</td>
<td>15.1</td>
<td>9.0</td>
</tr>
<tr>
<td>9</td>
<td>ATM</td>
<td>38.565 N</td>
<td>139.662 E</td>
<td>804</td>
<td>87.8</td>
<td>22.4</td>
<td>15.1</td>
<td>9.0</td>
</tr>
</tbody>
</table>

Fig. 3 shows the travel times of the ScS wave and the forerunning wave plotted against the epicentral distance. The data from the stations of the Japan Meteorological Agency are indicated by the small circles. The large circles denote the arrival times of the ScS phase and the large squares represent the forerunning phase of the ScS wave listed in the table. This forerunning phase is also observed at a station of the Japan Meteorological Agency, located at Ishinomaki, northeast to AOB, which is also included in Fig. 3. Travel times of ScS wave shown by broken curves are taken from the Jeffreys-Bullen table and Randall's table (Randall 1971) for the hypocentre determined by the Japan Meteorological Agency. The focal depth determined is too deep to explain the observed data, and the depth is estimated to be about 50 km shallower than that determined by the Japan Meteorological Agency. The J-B travel-time curve in the case of focal depth 550 km is illustrated in the figure by the solid line. If we take this fact into consideration, the data from our seismic network (large circles) fit considerably well to the expected curve in comparison with those from the
The difference in arrival time between the ScS phase and the forerunning phase of ScS phase is plotted in Fig. 4 as a function of the depth to the deep seismic zone right under each station as was done by Okada (1974). The double circles are the results obtained from this work in the Tohoku District and open circles are those of Okada (1971, 1974) in the Hokkaido and Kanto Districts. The travel-time interval increases uniformly with the increase in depth to the deep seismic zone. Okada (1971, 1974) interpreted this forerunning phase as the P wave converted from ScS wave at the boundary between the land-side low-Q, low-velocity zone and the descending high-Q, high-velocity slab, and he termed this phase the ScSp wave. Assuming an inclined plate model with a seismic wave velocity 8 per cent higher than the surrounding mantle and with a thickness of 80 km, he calculated the travel-time interval between the two phases for several kinds of dip angles of the plate. The results are plotted in Fig. 4 with the broken lines. In this calculation, he also assumed the vertical incidence of the ScS wave beneath the descending slab, and the upper boundary of the slab is assumed to be located 15 km shallower than the deep seismic zone. The data observed almost coincide with the curve with a dip angle of 30°.

Fig. 4 and the predominance of the longitudinal component of the particle motion indicate that the phase is the ScSp wave converted from the ScS wave at the upper boundary of the descending slab.
3 Upper boundary of descending lithospheric slab

The interpretation of the forerunning phase of the ScS wave in terms of the ScS to ScSp conversion at the boundary between the descending high-velocity slab and the mantle above implies that the boundary is a velocity discontinuity and its location approximately coincides with the deep seismic zone. In order to obtain the geometrical relationship between the double-planed deep seismic zone found in this region and the descending slab in more detail, we use a three-dimensional seismic-ray tracing method developed by Jacob (1970).

The 80-km thick plate model with P and S wave velocities 6 per cent higher than the surrounding mantle is assumed. The conclusions to be drawn here are not altered very much, even if the thickness of the plate is changed within a reasonable range. The velocity structure by Randall (1971) for the S wave is adopted as a standard earth model. In the upper 400 km of the mantle and crust, the velocity model used for our seismic network is to be assumed because the hypocentres of microearthquakes have been determined by using this velocity structure. The velocity model has been reported in a previous paper (Hasegawa et al. 1978). The hypocentre of the deep shock is that determined by the Japan Meteorological Agency, although its focal depth must be a little deeper than the actual value. This does not affect the general conclusions to be presented in this work.

Applying the three-dimensional ray-tracing technique to this model, the travel time difference between ScS and ScSp waves for each station is calculated. At the boundary between the descending slab and the surrounding mantle, which is a velocity discontinuity, Snell’s law is used to calculate the directions of rays of converted waves. The observed arrival-time differences at three stations HSK, MYK and KGJ only are adopted for the comparison with the calculated values, because the ScSp phase at these stations is observed more clearly than that at the other stations. The shape of the descending slab is assumed to be a simple flat plate with constant thickness. The location of the upper boundary of the descending slab is determined so that the calculated travel-time intervals between the two phases may simultaneously agree with those observed at the three stations within ±0.2 s. The results are shown in Fig. 5(a). The figure exhibits the focal depth distribution of the microearthquakes located in the region from 39 to 40° N in the Tohoku District during the period from 1975 April to December. The shocks are plotted on the vertical section in the E–W direction. The triangle and the reverse triangle represent the positions of the volcanic front and the Japan Trench, and the thick horizontal line in the upper part of the figure denotes the land area. As we have already noted (Hasegawa et al. 1978), the deep seismic zone is distinctly separated into two planes. The calculated range of the position of the upper boundary of the descending slab is illustrated by a hatched zone. The position of the boundary coincides with the upper plane of the double-planed deep seismic zone.

If the plane of conversion is within the range denoted by the hatched zone in Fig. 5(a), the angle between the incident ScS wave and the normal of conversion plane is from 35.3 to 36.5°, the latter value being the critical angle. A theoretical calculation of the S to P transmission (e.g. Ewing, Jardetzky & Press 1957) suggests that, in this range of incident angles, the transmission coefficient is relatively large for the probable contrast in the seismic parameters. Thus the converted ScSp wave may represent an efficient conversion process at the boundary between the descending slab and the mantle above. The observed amplitude ratio of the ScSp wave to the ScS wave at each station is the order of 0.2 which coincides roughly with theory.

The assumed difference in velocity between the descending slab and the surrounding mantle would be reasonable. Utsu (1967) estimated the difference in velocity between the two portions to be about 6 per cent for both P and S waves in the vicinity of Japan by
Figure 5. Focal-depth distribution of microearthquakes in the central part of the Tohoku District projected on the vertical section in the E–W direction (solid circles) and the expected range of the boundary between the descending slab and the mantle above obtained for several plate models (hatched zones). (a) Plate model with $P$ and $S$ wave velocities 6 per cent higher in the surrounding mantle, (b) and (c) are those with 4 and 8 per cent, respectively.
studying the travel times of deep earthquakes. Kanamori (1968, 1970) concluded the velocity difference for P waves to be 4 ~ 5 per cent by investigating the travel-time residuals from longshot to the Japanese stations. Tada (1972) determined the velocity distribution for P waves within the downgoing slab beneath the island arcs of Japan by applying the method developed by Kaila (1969), and estimated the velocity difference to be 5 per cent at the depth of 200 km. Recently, Yamamizu (1973) and Hamada (1973) investigated the upper-mantle structure beneath the islands arcs of Japan by means of the three-dimensional ray-tracing method, and they obtained the velocity difference for P waves between the descending slab and the mantle above it to be 6 and 7 per cent, respectively.

Two other models with different velocity contrasts are also adopted in order to confirm the conclusion presented here. One is the plate model with P and S wave velocities 4 per cent higher than the surrounding mantle, and the other is that with 8 per cent. The seismic-ray tracing technique is also applied to these models, and the results are shown in Fig. 5(b) and (c). Hatched zones in the figures are the obtained range of the upper boundary of the descending slab. Fig. 5(b) and (c) correspond to the models with velocity differences of 4 and 8 per cent, respectively. The locations of the boundary in both models coincide exactly with the upper plane of the double-planed deep seismic zone in a similar way to that in the model with 6 per cent velocity difference.

The dotted line in Fig. 5(a) represents the range of the upper boundary of the descending slab obtained with the different velocity structures for P and S waves in the upper 400 km of the mantle and crust. The P-wave velocity structure reported by Herrin et al. (1968) and the S wave structure by Randall (1971) are assumed in the upper 400 km in the mantle instead of those used for our seismic network. In the crust, the velocity structure of Ichikawa & Mochizuki (1971) is adopted. The same calculation is performed by assuming the velocity difference to be 6 per cent. In this case, however, we cannot find the range in which the calculated travel-time intervals between ScS and ScSp waves account simultaneously for the observed data at the three stations within ±0.2 s. Therefore, the permissible range of ±0.25 s is adopted instead of ±0.2 s. The boundary lies on the upper plane of the double-planed deep seismic zone even for this velocity model.

Strictly speaking, the locations of hypocentres will be affected by the lateral variation of seismic-wave velocity in the upper mantle. Since the hypocentres of the microearthquakes shown in Fig. 5 are determined by neglecting this lateral variation, the hypocentres may differ systematically from their real positions. We relocated the hypocentres of these microearthquakes, taking the lateral variation of the velocity structure into consideration. In this calculation, the 80-km thick plate model with velocities 6 per cent higher than the surrounding mantle is used. The shape and location of the descending high-velocity slab have already been determined in this section as illustrated in Fig. 5(a), and this laterally heterogeneous structure is assumed for the calculation. The travel-time difference from the source of the first approximation to each recording station between the laterally heterogeneous and homogeneous models is calculated by using the three-dimensional ray-tracing method. By adding this value to the observed arrival time at each station, the hypocentre is redetermined for the laterally-homogeneous velocity structure. In the next step, the travel-time difference from the relocated source to each station is calculated again, and the hypocentre is redetermined in the same manner as that in the first step. The procedure is iterated until the solution converges to a certain position. The revised focal-depth distribution indicates that the position of the upper plane of the double-planed deep seismic zone is almost unchanged, and that the dip of the lower plane becomes slightly less steep than before. Thus the boundary between the descending slab and the mantle above coincides exactly with the upper plane of the double-planed deep seismic zone in the Northeastern Japan Arc.
A small-scale seismic-array observation has been carried out since 1970 at the Kitakami Seismological Observatory. This seismic array is located in the eastern part of the Tohoku District by 13 stations (Fig. 1), each containing short-period seismometers. The hypocentres of the microearthquakes located by our seismic network are also located independently by this seismic array, and the hypocentre distribution of intermediate-depth earthquakes determined by this array observation is extremely different from that of our seismic network (Yamamoto & Kohno 1975).

Fig. 6(a) shows the focal-depth distribution of the microearthquakes in the region from 39 to 40° N projected on the vertical section in the E–W direction. The thick horizontal
line in the upper part of the figure denotes the land area, and the triangle represents the location of the seismic array. The figure is exaggerated in the vertical direction, and the reduced scale in the horizontal direction is half that in the vertical direction. The shocks located by our seismic network are indicated by the solid circles. These hypocentres are those which have been revised by taking the laterally-heterogeneous structure into account as mentioned in the preceding section. The open circles in the figure denote the hypocentres of the shocks located by the seismic array (Yamamoto & Kohno 1975). The same earthquakes are connected by straight lines, respectively. Fig. 6(b) is the epicentre distribution of the shocks plotted in Fig. 6(a). The cross in the figure indicates the location of the array. The solid and open circles are the same as those in Fig. 6(a). The shocks with focal depths shallower than 50 km are omitted in these figures. It is shown that the hypocentres of the earthquakes which occurred westward to 141°E are determined by the seismic array more deeply and more closely to the array than that determined by our seismic network. The discrepancy between the hypocentres increases with increasing focal depth. This is due to the existence of the descending high-velocity slab. The seismic rays of intermediate-depth earthquakes to the array will be refracted and bent at the upper boundary of the descending high-velocity slab and, therefore, the apparent velocities and the directions of approach of P waves at the array will be altered. The hypocentres are calculated from the observed apparent velocities and directions of wave approach. Thus the hypocentres of the intermediate-depth earthquakes will be located by the seismic array at different positions from real ones. However, such effects are not taken into account for the hypocentre determination of the seismic-array observation. The lateral heterogeneity of the upper mantle is reflected in the discrepancy between the hypocentres of intermediate-depth earthquakes located by the small-scale seismic array and those by relatively large-scale seismic network. Consequently, the upper-mantle structure is estimated from the amount of this discrepancy.

The 80-km thick plate model with velocities for P and S waves 6 per cent higher than the surrounding mantle is assumed as described in the preceding section. The location of the dipping high-velocity plate has been determined in the preceding section. We calculate the apparent velocities and the directions of approach of P waves at the array for the shocks located by our seismic network (solid circles in Fig. 6) by applying the three-dimensional ray-tracing method to this model. The hypocentres of these shocks can be determined from the calculated apparent velocities and directions of wave approach by the same method as that used in the seismic array. The calculated hypocentres are plotted in Fig. 7 by open circles. Solid circles and other marks in the figures are the same as used in Fig. 6. The earthquake hypocentres calculated by assuming the laterally-heterogeneous upper mantle (open circles in Fig. 7) almost coincide with those observed by the seismic array at the Kitakami Seismological Observatory (open circles in Fig. 6). This means that the predicted model is reasonable.

We cannot explain the observed hypocentre distribution of the shocks located by the seismic array if the upper boundary of the descending high-velocity slab is located at a depth deeper than the upper plane of the double-planed deep seismic zone. This situation is illustrated in Fig. 8. Fig. 8(a) and (b) show the focal-depth distributions of the shocks calculated by assuming the laterally-heterogeneous upper mantle. The location of the upper boundary of the descending slab is assumed at depths 15 and 30 km deeper than the upper plane of the double-planed deep seismic zone and is illustrated by straight lines in the figures. The calculated locations of the intermediate-depth earthquakes in Fig. 8(a) and (b) do not agree with the observed ones (open circles in Fig. 6(a)).

The plate models with different velocity contrasts are also tested. Fig. 9(a) and (b) are focal-depth distributions of the shocks calculated for the plate models with velocity
differences of 4 and 8 per cent, respectively. The locations of the upper boundary of the descending slab are shown in Fig. 5(b) and (c). The hypocentres are calculated for these models. Comparing these calculated results with the observed one, we can see that the hypocentre distribution obtained from the seismic-array observation is also well explained by the plate models with these velocity contrasts. In both cases, the upper boundary of the inclined high-velocity lithospheric slab lies on the upper plane of the double-planed deep seismic zone.

In this section, we have confirmed from the different data the validity of the result that the boundary between the descending slab and the mantle above coincides with the upper plane of the double-planed deep seismic zone, which is obtained in the preceding section. In our previous investigation (Hasegawa et al. 1978), the composite focal-mechanism solution indicates that the earthquakes in the upper plane of the double-planed deep seismic zone are
Figure 8. Focal-depth distributions of microearthquakes calculated for the plate models in which the location of the plate boundary is different from that of Fig. 7. The plate boundary is assumed to be located at depths (a) 15 km and (b) 30 km deeper than the upper seismic plane.
Figure 9. (a) Focal-depth distribution of microearthquakes calculated for the plate model shown in Fig. 5(b), and (b) that for the plate model shown in Fig. 5(c).
characterized by reverse faulting, or some of them by down-dip compressional stresses and those in the lower plane by down-dip extensional stresses. From the result obtained here, two kinds of interpretation are possible as the mechanism generating the earthquakes in the upper plane. One is the thrusting motion on the boundary between the descending lithospheric slab and the mantle above it. If this is true, the focal mechanisms for the events in the upper plane should represent a thrust faulting dipping about 30° to the west. The other is a kind of elastic unbending of the descending lithospheric slab (Isacks & Barazangi 1977). The focal mechanisms should be characterized by down-dip compression in this case. More precise determination of focal mechanisms is essential to the correct understanding of the mechanism generating earthquakes within the descending slab.

5 Conclusions
The forerunning phase of the $ScS$ wave was clearly observed at several stations of the seismic network of Tohoku University for a large deep earthquake which occurred in the western portion of the Japan Sea on 1975 June 29. This phase can be interpreted as the $P$ wave converted from the $ScS$ wave at the boundary between the descending high-$Q$, high-velocity slab and the land-side low-$Q$, low-velocity mantle beneath the Northeastern Japan Arc. Inclined high-velocity plate models are assumed beneath the arc to account for the observed differences in arrival time between the two phases. A three-dimensional seismic-ray tracing method is applied to these models, and the location of the boundary is estimated in such a way that the calculated travel-time intervals between the two phases may agree with the observed ones. It is known from this calculation that the upper boundary of the descending slab coincides with the upper plane of the double-planed deep seismic zone.

The hypocentre distribution of the intermediate-depth microearthquakes which are located by the small-scale seismic-array observation is quite different from that by the relatively large-scale seismic network. This discrepancy is interpreted due to the existence of the inclined high-velocity slab beneath the arc. Applying the seismic-ray tracing technique to the same models mentioned above, the apparent velocity and the direction of approach of $P$ waves are calculated for the earthquakes located by the large-scale seismic network. The hypocentres are obtained from these values by the same method as that used in the seismic array. These hypocentres calculated by considering the laterally-heterogeneous upper-mantle structure are almost consistent with those derived from the seismic-array observation. This fact provides support for the conclusion obtained from the difference in arrival time between $ScS$ and $ScSp$ waves.

The composite focal-mechanism solutions derived for these two groups of earthquakes indicate that the earthquakes having occurred in the upper plane are characterized by reverse faulting, or some of them by down-dip compressional stresses, and those in the lower plane by down-dip extensional stresses. The obtained location of the upper boundary of the descending slab implies that the earthquakes in the upper plane of the double-planed deep seismic zone mostly occur at this boundary either by interaction between the descending slab and the mantle above or by elastic unbending of the descending slab. Within the descending slab, from 30 to 40 km below its upper boundary, the earthquakes with down-dip extensional stresses occur. These earthquakes form the lower plane of the double-planed deep seismic zone. Fig. 5(a) shows a typical vertical cross-section of the upper mantle perpendicular to the trench axis beneath the Northeastern Japan Arc.

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