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Seismic structure of the northeastern Japan convergent margin : A synthesis

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Abstract. Many studies recently made on the basis of seismic observations have revealed a detailed structure of the crust and the upper mantle beneath the northeastern Japan arc and its relationship to seismic and volcanic activity. Spatial distributions of the depths to the Conrad and the Moho discontinuities, estimated from shallow earthquake data and seismic explosion data, show that both discontinuities are deep in the middle of the land area and shallow toward the coastlines of the Japan Sea and the Pacific Ocean. The *Pn* velocity has a lateral variation; it is as low as ~7.5 km/s beneath the land area, while that beneath the Japan Sea and the Pacific Ocean is 8.0-8.2 km/s. It changes abruptly at the transition zones, which are located along the coastlines. Precise structure and location of the subducted Pacific plate beneath the land area is inferred from converted or reflected seismic waves at the top or bottom of the plate. The Pacific plate is composed of a thin (~5 km) low-velocity upper layer and a thick high-velocity lower layer, its total thickness being 80-90 km. The upper plane seismicity of the double seismic zone is confined to the thin low-velocity upper layer, which probably corresponds to the subducted former oceanic crust. The lower plane seismicity lies at the middle of the high-velocity lower layer, and the lower half of the plate below it is incapable of generating earthquakes. The shallower portion of the upper surface of the plate beneath the Pacific Ocean, along which major seismicity with low-angle thrust faultings is actually occurring, is also located by seismic observations on land and in the sea. The Pacific plate subducts at an extremely low angle of ~5° for the first ~25-km depth, and then the dip steepens rather abruptly to ~30°. Normal-fault type events at the top of the plate have not been detected in the portion where the downward- bending is the largest, but have been ditected near the trench axis, where it is rather small. Tomographic inversions for seismic velocity structure clearly delineate the inclined high-velocity Pacific plate with a thickness of 80-90 km and low-velocity zones in the crust and the mantle wedge beneath active volcanoes. Seismic attenuation tomography also shows similar zones of low-Q value beneath active volcanoes, although its spatial resolution is much lower. The low-velocity zones with 2-6% velocity lows are continuously distributed from the upper crust just beneath active volcanoes to a depth of 100-150 km in the mantle wedge, and are approximately parallel to the dip of the underlying Pacific plate. These low-velocity zones probably reflect the pathway of magma ascent from a depth in the mantle wedge to the Earth's surface, corresponding to a portion of the subduction-induced secondary mantle wedge flow.

Introduction

The northeastern Japan arc is one of the most typical subduction zones in the world, and is an excellent region in which to study the subduction process. In this arc, the oceanic Pacific plate subducts downward into the mantle at a convergence rate of about 10 cm/yr and at an angle of about 30°. Many shallow earthquakes occur beneath the Pacific Ocean along the upper boundary of the Pacific plate or within it associated with its subduction beneath Tohoku, northeastern Japan. Intermediate-depth and deep earthquakes are generated within the subducted Pacific plate to a depth of ~670 km. Beneath the land area, shallow earthquakes also occur in the upper crust of the continental plate, and sometimes at the base of the crust or at the top of the mantle wedge. Active volcanoes are distributed on the land area, mainly along the volcanic front, which runs parallel to the trench axis.

High seismic activity and spatially dense seismic networks covering this region have facilitated the progress of understanding of the seismic structure in this subduction zone. Many studies on seismic activity and on seismic velocity and attenuation structure of the crust and upper mantle have been conducted in this region using data acquired through natural earthquake observations based on the dense seismic networks or explosion seismic observations. These studies have revealed a precise structure of the descending Pacific plate, the crust and the mantle wedge above it, and its relationship to seismic and volcanic activity. In the present manuscript, we describe an outline of these studies, especially those recently made through microearthquake observations, and present a qualitative model of the subduction structure of this arc based on these observations.

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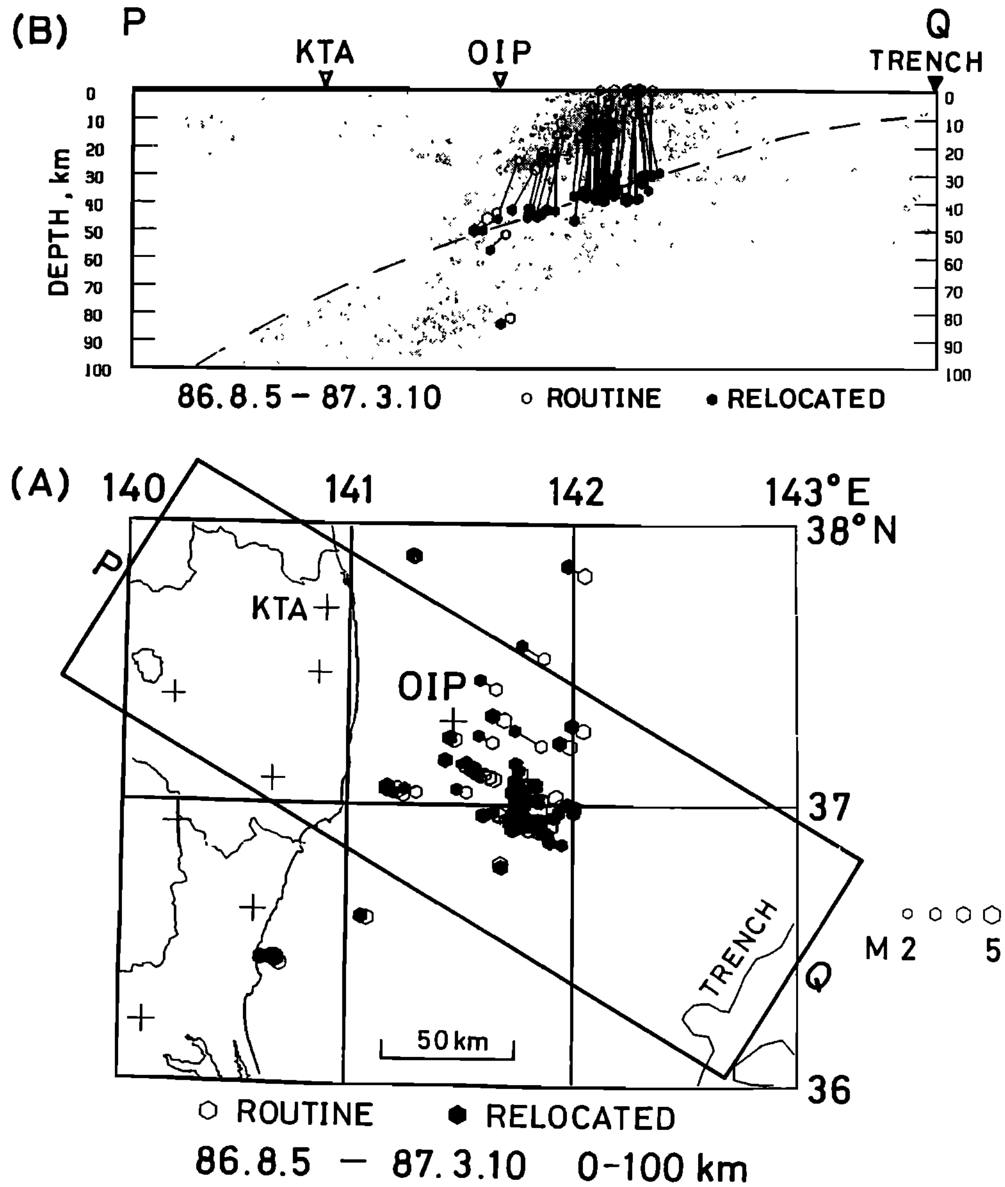


Figure 1. (a) Epicenter distribution and (b) vertical cross section of earthquakes detected at a hydrophone station (OIP) 40 km off the coast of Fukushima Prefecture for the period August 1986 to March 1987. Hypocenters located by the seismic network of Tohoku University on land (open circles) are relocated by adding P wave arrival time data at the hydrophone station and are shown by solid circles. Background seismicity by the land seismic network is also shown by small dots on the vertical cross section. Events in the rectangular area in Figure 1a are plotted on the vertical cross section in Figure 1b. Bold line on the top and broken line in the vertical section denote the land area and the estimated upper boundary of the descending Pacific plate, respectively [after *Hasegawa et al.*, 1987].

Seismic Activity

Most of shallow earthquakes in the northeastern Japan arc occur beneath the Pacific Ocean between the Japan trench and the Pacific coast, which can be partly seen from cross-arc vertical cross sections of seismicity shown by small dots in Figures 1 and 2. These shallow events are located mainly along the boundary (main thrust zone) between the subducting Pacific plate and the overlying continental plate. Historically, many destructive earthquakes with low-angle thrust faultings, such as the 1968 M 7.9 Tokachi-oki Earthquake and 1978 M 7.4 Miyagi-oki Earthquake, also have taken place along this main thrust zone beneath the Pacific Ocean. The shallow earthquakes beneath the Pacific Ocean shown in Figures 1 and 2, however, are considerably scattered in depth direction, and one cannot determine from these figures that they are actually generated along the plate boundary (or in its vicinity). This is due to the difficulty of accurate hypocenter determinations of earthquakes in this region beneath the Pacific Ocean. Since the seismic stations are limited to the land area far from the focal region, the accuracy of hypocenter locations, especially that of focal depths, is quite low.

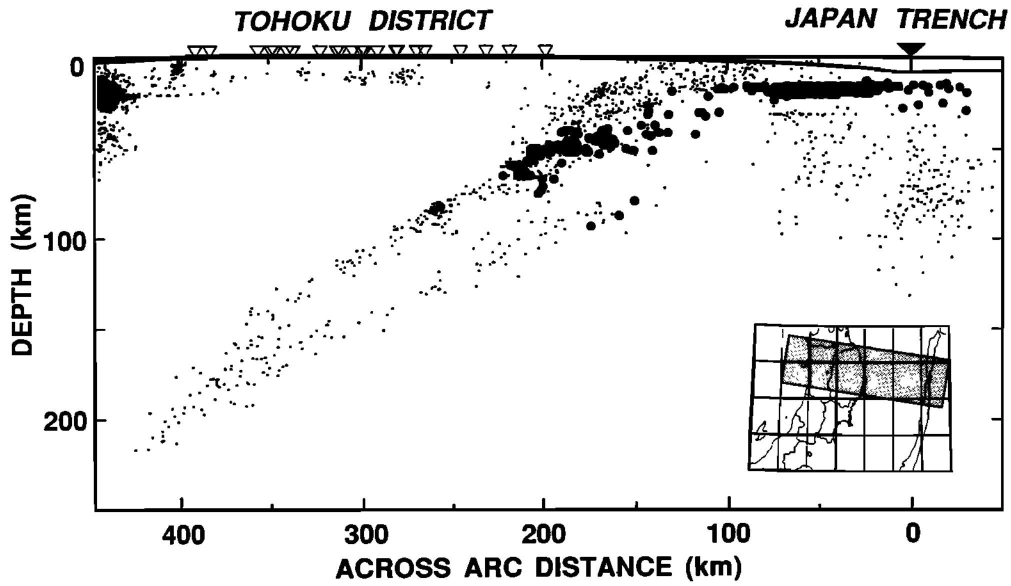


Figure 2. Cross-arc vertical cross section of earthquakes located by using the *sP* depth phases with small epicentral distances. Hypocenters within the rectangular area in the inset map are plotted by solid circles. Background seismicity by the land seismic network is also shown by small dots. Bold line, inverted solid triangle, and inverted open triangles represent the locations of the land area, the trench axis, and the seismic stations, respectively [after *Umino et al.*, 1992].

The direct way of improving the accuracy of focal depths is to set up seismic stations just above the focal region in the sea. *Hasegawa et al.* [1987] carried out a continuous seismic observation by using hydrophones at a platform located 40 km off the coast of Iwaki, Fukushima Prefecture. An addition of observed *P* arrival time data at the hydrophone station to those at the stations on land strikingly improves the accuracy of focal depths of shallow earthquakes in this region. Hypocenter distribution of earthquakes relocated by adding data at the hydrophone station is shown by solid circles in Figure 1, together with that (open circles) originally located by using data obtained only at a seismic network on land. Low-angle thrust faulting mechanisms of some of these events and the relocated hypocenter distribution gently dipping to the west-northwest indicate that these events have taken place along the upper boundary of the subducting Pacific plate. This observation suggests that many of the events scattered in focal depths beneath the Pacific Ocean in Figures 1 and 2 are actually located along the plate boundary.

Yoshii [1979] relocated earthquakes off Sanriku by using *pP* depth phase reported by the International Seismological Center. In his hypocenter relocation, Yoshii corrected the travel time within the water layer for the depth phase. The result shows that the shallow earthquakes off Sanriku form a thin seismic zone gently dipping to the west from the Japan trench and that focal mechanisms of earthquakes in the deeper portion are characterized by the low-angle thrust faulting. This observation indicates that the inclined thin seismic zone delineates the plate boundary between the subducting Pacific plate and the overlying continental plate. Earthquakes in the shallower portion near the trench axis have normal faulting mechanisms, suggesting that these events are occurring at the top of the Pacific plate caused by the bending of the subducting plate.

Recently, *Umino et al.* [1992] found that *sP* depth phase can be clearly observed even at small epicentral distances of about 200 km. This phase is expected to exist from some standard Earth model [e.g., *Kennett*, 1991; *Kennett and Engdahl*, 1991], although actual observations at such small epicentral distances have not been reported. Observations of this phase enable us to estimate accurately focal depths of smaller magnitude (~3) events because teleseismic waveform data are not required for this estimation. *Umino et al.* [1992] relocated earthquakes that occurred off Sanriku and off Fukushima by using the *sP* phases observed at the seismic network of Tohoku University, which covers the land area of the northeastern Japan arc. Arrival time difference of direct *P* and *sP* phases depends not only on focal depths but also on the seismic velocity structure, especially the structure above earthquake foci. Thus absolute errors of estimated focal depths rely on how accurate the velocity structure is derived. In their estimation, they adopted as the velocity structure beneath the sea, the velocity model obtained by an airgun-ocean bottom seismograph (OBS) seismic profiling across the Japan trench [*Suyehiro et al.*, 1990].

Earthquakes off Sanriku relocated by using the *sP* phases at small epicentral distances are shown in Figure 2 by solid circles on a vertical cross section which is nearly perpendicular to the trench axis. Hypocenters determined by the land network alone are also shown by small dots in the figure. In this case again, many shallow events off Sanriku scattered in focal depths (small dots) concentrate in a thin plane inclined gently to the west after the hypocenter relocation. If this plane is the boundary between the two converging plates or very close to it, the present result shows that the Pacific plate beneath this region subducts at an extremely low angle of ~5° for the first descent to ~25-km depth and then dips at a steeper angle of 25-30° at depths deeper than ~30 km. Figure 2 suggests that there is no aseismic belt beneath the inner wall of the trench axis, as pointed out by OBS observations [*Hirata et al.*, 1983, 1985; *Byrne et al.*, 1988]. Errors of estimated epicenters, particularly in the EW direction, by the land network, depend on the accuracy of observed *S-P* times and again on the velocity structure model. Thus it is difficult to know from the land network data alone whether or not the aseismic belt, caused by unconsolidated and semiconsolidated sediments in the plate interface, actually exists.

The shallow seismicity beneath the Pacific Ocean merges continuously with the deep seismic zone which dips beneath northeastern Japan at an angle of 25-30° and extends to 670-km depth. The deep seismic zone beneath the land area (in the depth range 70-150 km) is composed of two thin planes which are almost parallel to each other and 30-40 km apart. The focal mechanisms of the events in the upper and lower seismic planes, except right under the volcanic front, are characterized by downdip compression and downdip extension, respectively [*Umino and Hasegawa*, 1975; *Hasegawa et al.*, 1978a]. Beneath the volcanic front, the dominant focal mechanism in the upper plane is not the downdip compression type but is a normal fault type. Furthermore, N-S tensional normal fault type mechanisms, instead of downdip extension, appear in the lower plane beneath the volcanic front [*Matsuzawa et al.*, 1986a].

The upper and lower seismic planes forming the double seismic zone merge together at a depth of 150 km or so. Since the hypocenter determinations by the routine procedures use a horizontally homogeneous velocity structure model, the existence of the inclined high-velocity Pacific plate would affect calculated hypocenter locations for intermediate-depth and deep events, particularly at greater depths. *McLaren and Frohlich* [1985 Figure 8] pointed out by a simple model study using three-dimensional (3-D) ray tracings that the dipping high-velocity plate causes separate parallel event groups to appear to merge together. This suggests that the present case beneath the northeastern Japan arc is also produced artificially because of the source-station geometry. However, this effect is not very strong for our seismic network having a cross-arc distance of about 200 km. The systematic shift of hypocenters at a depth of 150 km or so, estimated by using 3-D ray tracings, is several kilometers, which allows us to discriminate separate parallel event groups that are 30-40 km apart. Moreover, events below a 150-km depth do not have the two focal mechanism types opposite to each other but are characterized only by downdip compression, in contrast to events at depths of 70-150 km [*Umino et al.*, 1984]. These observations indicate that earthquakes at depths deeper than ~150 km form a single plane.

The double-planed deep seismic zone is also found beneath the Kanto District (central Honshu) and beneath Hokkaido [*Tsumura*, 1973; *Suzuki and Motoya*, 1981]. The depths to the upper and lower seismic planes beneath Hokkaido, Tohoku, and Kanto are shown in Figure 3 by contour lines [*Hasegawa et al.*, 1985]. At the junction between the Kurile and the northeastern Japan arcs and that between the northeastern Japan and the Izu-Bonin arcs, the deep seismic zone is contorted but is still double- planed at least in the upper 100- to 150-km depth range. However, the stress state in the double seismic zone beneath Hokkaido is slightly different from that beneath Tohoku and Kanto. The upper seismic plane is not downdip compression but has various types of focal mechanisms, while downdip extensional stress is predominant in the lower plane, as in the case of Tohoku and Kanto [*Suzuki et al.*, 1983; *Umino et al.*, 1984]. The extension of the deep seismic zone deeper than about 150 km forms a single plane both in Hokkaido and Tohoku. Stress states again are different from each other. Downdip extensional stress is predominant beneath Hokkaido, whereas downdip compression is beneath Tohoku. The deepest portion of the deep seismic zone (deeper than about 300 km) is characterized by downdip compression both in Hokkaido and Tohoku [e.g., *Isacks and Molner*, 1971].

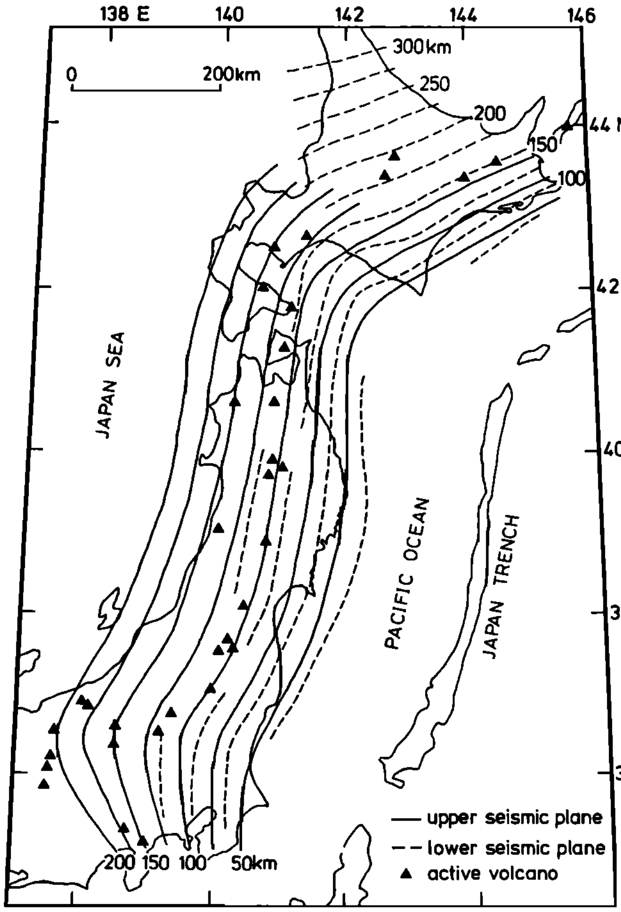


Figure 3. Geometry of the double-planed deep seismic zone beneath Kanto, Tohoku, and Hokkaido. Depths to the upper and lower seismic planes are shown by solid and dashed contours, respectively. Solid triangles denote the locations of active volcanoes [after *Hasegawa et al.*, 1985].

In addition to these seismic activities, shallow earthquakes occur beneath the land area; relatively high seismicity is distributed along the volcanic front or the central mountainous range (Ou backbone range). Shallow seismicity is also relatively high along or off the coastline of the Japan Sea. Most of these shallow earthquakes are confined to the so-called "granitic layer" [*Takagi et al.*, 1977]. A recent seismicity study reveals that the cutoff depth for the shallow seismicity does not correspond to the Conrad depth (~18 km) but is slightly shallower (~15 km) [*Hasegawa et al.*, 1991a]. The focal-depth cutoff of these shallow events is caused by a brittle to ductile transition or a stick-slip to stable-sliding transition due to increasing temperature with depth [e.g., *Brace and Byerlee*, 1970; *Meissner and Strehlau*, 1982; *Sibson*, 1982; *Tse and Rice*, 1986], and the upper 15 km of the crust forms a brittle seismogenic zone. The shallow depth of

~15 km of the transition zone is due to a large geothermal gradient beneath this volcanic arc [*Shimamoto*, 1991]. The lower portion of the crust and the mantle wedge are governed by creep or flow.

However, anomalously deep (22-47 km) microearthquakes, well below the base of the brittle seismogenic zone, also occur at the base of the crust or at the top of the mantle wedge beneath the land area, although their occurrence is rather infrequent (~1% of shallow earthquakes in the land area) [*Hasegawa et al.*, 1991a; *Yamamoto and Hasegawa*, 1990]. In total, 153 deep events have been found so far at 12 locations in northeastern Japan, and they are located beneath active volcanoes or around *P* wave low-velocity zones in the uppermost mantle. All 153 events have extremely low predominant frequencies (1.5-3.5 Hz), both for *P* and *S* waves. The close locations of these anomalous events to active volcanoes and their extremely low predominant frequencies of *P* and *S* waves suggest that the occurrence of these events is directly related to deep-seated magmatic activity in this depth range.

Crustal Structure

The *P* wave velocity structure of the crust has been investigated by several workers [*Hashizume et al.*, 1968; *Research Group for Explosion Seismology* (RGES), 1968, 1977; *Yoshii and Asano*, 1972; *Asano et al.*, 1979; *Okada et al.*, 1978, 1979], based on seismic explosion data along the Oga-Kesennuma profile, which traverses the central part of Tohoku. *Yoshii and Asano* [1972] analyzed the seismic explosion data along the Oga-Kesennuma profile by using the time term method. The crustal structure along the profile that they obtained is shown in Figure 4. An important result that they revealed is a lateral variation of *Pn* velocity. The *Pn* velocity beneath the land area is as low as about 7.5 km/s, whereas that beneath the Pacific Ocean and beneath the Japan Sea is 8.0-8.2 km/s, a value usually seen in many regions of the world. The lateral change in *Pn* velocity is confirmed through seismic explosion experiments subsequently carried out beneath the land area and beneath the surrounding seas. *Okada et al.* [1979] and *Asano et al.* [1979] found a sharp lateral change in *Pn* velocity across a transition zone near the Pacific coast. The location of this *Pn* velocity boundary is consistent with the aseismic front proposed by *Yoshii* [1975], which is originally estimated from a characteristic spatial distribution of seismicity. A sharp *Pn* velocity change from 7.5 km/s beneath the land area to about 8.0 km/s beneath the Japan Sea is also confirmed by *Okada et al.* [1978].

Horiuchi et al. [1982a] developed a method to investigate the crustal structure by using local earthquake data. In their method, the lateral depth variation of the boundary layer is expressed by a power series in latitude and longitude. Unknown parameters are coefficients of the power series, hypocenter parameters, and station corrections, and these parameters are estimated by inverting data of travel time anomalies from local earthquakes. Applying this method, *Horiuchi et al.* [1982a, b] estimated the depth distributions of the Conrad and Moho discontinuities beneath the central part of northeastern Japan. Both the Conrad and Moho discontinuities are deep in the middle of the land area and become shallower toward both the coastlines of the Japan Sea and the Pacific Ocean.

The Conrad depth in the western part of the Oga-Kesennuma profile estimated by *Horiuchi et al.* [1982a] coincides with that derived from the time term analysis of the seismic explosion data [*Yoshii and Asano*, 1972]. However, the Conrad depth in the eastern part is about 6 km deeper than that from the explosion data. *Horiuchi et al.* [1982b] showed that observations of first *Pg* arrivals with epicentral distances of about 80 km at a seismic array, located close to the eastern part of the profile, can be explained by the velocity model with the thick granitic layer, which they derived from local earthquake data. Their estimation of the Moho depth in the central part of the profile is again slightly different from, and is about 5 km deeper than that derived by the explosion seismic data.

Zhao et al. [1990a, b] systematically investigated a spatial variation of the *Pn* velocity beneath the northeastern Japan

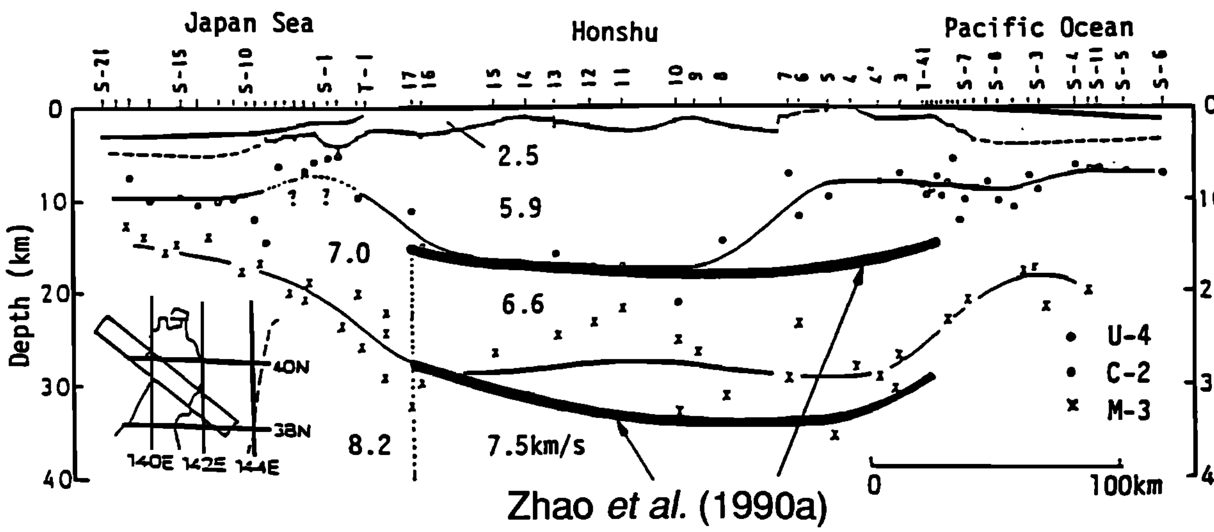


Figure 4. Crustal structure along the Oga-Kesennuma profile estimated from explosion seismic observations [*Yoshii and Asano*, 1972]. Location of the profile is shown in the inset map. Numerals denote the *P* wave velocity at each location. Depth distribution of the Conrad and the Moho discontinuities estimated by using shallow earthquakes [*Zhao et al.*, 1990a] is shown by bold lines.

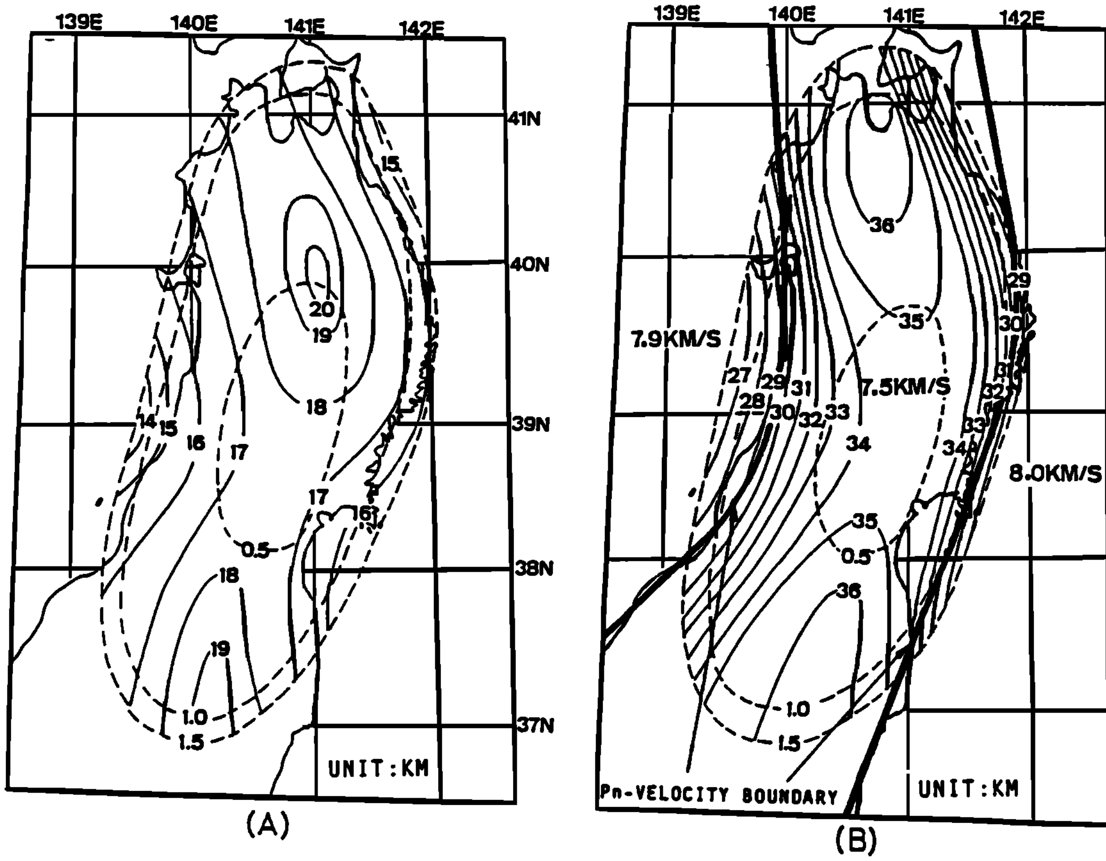


Figure 5. Maps showing depth distribution of (a) the Conrad and (b) the Moho discontinuities. Locations of the *Pn* velocity boundaries are shown by bold lines. The areas where the standard deviation of inverted depths is smaller than 0.5 km, 1.0 km, and 1.5 km are represented by three closed curves with dashed lines [after *Zhao et al.*, 1990a].

arc by using travel time data of shallow earthquakes. They found that the *Pn* velocity of 7.5 km/s beneath the land area sharply changes to 7.9 km/s across a transition zone near the Japan Sea coast and then gradually increases up to 8.0-8.2 km/s with the distance from the coast. They also found an abrupt *Pn* velocity change across a transition zone near the Pacific coast. Taking into account this regional variation of the *Pn* velocity, *Zhao et al.* [1990a] extended the method of *Horiuchi et al.* [1982a] and estimated the *P* wave velocity structure of the crust in the whole area of northeastern Japan. In their inversion scheme, coefficients of the power series which express the locations of the *Pn* velocity boundaries are also taken as unknown parameters. Subsequently, *Zhao et al.* [1992a] estimated the *P* wave velocity structure of the crust covering the whole area of the Japanese Islands by using the same method.

Figure 5 shows the depth distributions of the Conrad and the Moho discontinuities (solid contours in the figure) and the locations of *P*n velocity boundaries estimated by inverting travel time data of *P* and *S* wave first arrivals and later arrivals of *Pg* and *P** waves from shallow earthquakes [*Zhao et al.*, 1990a]. Estimated *Pn* velocity boundaries are located along the Pacific coastline and along the Japan Sea coastline, which is consistent with the result derived from the explosion seismic observations. The Conrad and the Moho depths have spatial distributions similar to each other. They are deep beneath the land area and shallow progressively toward the Japan Sea coast and toward the Pacific coast. The depth distributions of the Conrad and the Moho shown in Figure 5 are almost consistent with the result of *Horiuchi et al.* [1982a, b], which covers a smaller area of 200 $\times$ 200 km^2 of the central part of northeastern Japan. The estimated Conrad and Moho depths along the Oga-Kesennuma profile are drawn by bold lines on the crustal structure obtained from the explosion seismic observations [*Yoshii and Asano*, 1972] in Figure 4. As in the case of Horiuchi et al., the Conrad depth in the eastern part and the Moho depth in the central part of the profile are different from those obtained by explosion seismic observations.

Takahashi [1982] found an *S*-to-*P* mode-converted phase at the Moho in seismograms at a station along the Pacific coast from earthquakes beneath the station. A ray path of this phase is schematically shown on a vertical cross section in Figure 7a. The depth to the Moho beneath the station estimated from arrival time data of this phase ranges from 25 to 32 km. Subsequently, *Matsuzawa et al.* [1988] more systematically surveyed seismograms of earthquakes that occurred in the deep seismic zone. They estimated the depth distribution of the Moho beneath the eastern part of Tohoku by expressing spatial distribution of the Moho depth by a power series in latitude and longitude similarly to *Horiuchi et al.* [1982a] and by inverting travel time data of *S*-to-*P* converted waves observed at 15 seismic stations from intermediate-depth earthquakes. Estimated location of the Moho is shown on an

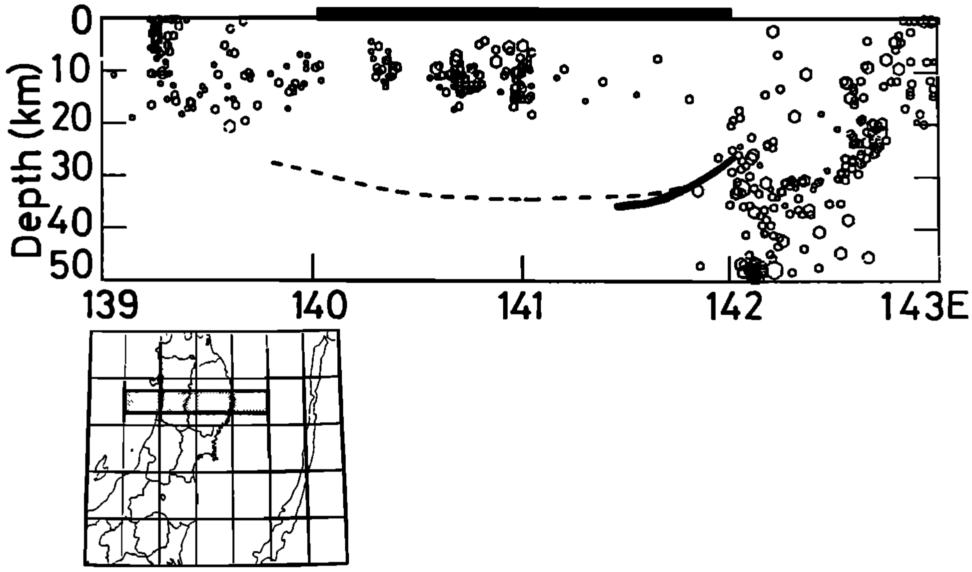


Figure 6. Depth distribution of the Moho discontinuity along an EW profile in the central part of Tohoku. Location of the profile is shown in the inset map. Estimated depth of the Moho by *Zhao et al.* [1990a] is shown by a dashed line, and that by using the *SP* converted wave [*Matsuzawa et al.*, 1988] is shown by a bold line. Earthquakes located along the profile are also plotted.

EW vertical cross section in Figure 6 along with that estimated by *Zhao et al.* [1990a]. Both are in good agreement with each other, suggesting the adequacy of the estimated depth distributions of the Conrad and Moho boundaries.

Plate Interface Morphology

A powerful method for estimating a 3-D seismic velocity structure by using an inversion technique has been proposed by *Aki and Lee* [1976] and *Aki et al.* [1977]. Since those pioneering works of seismic tomography, many researchers have investigated the 3-D seismic velocity structure on both local and regional scales in various regions of the world. Spatial resolution of estimated tomographic images by this method is limited to an average spacing between observation stations (and to the distribution of earthquakes). An effective way to estimate a more detailed structure of the subducted slab is the use of reflected waves [e.g., *Snoke et al.*, 1974; *Mizoue*, 1977; *Fukao et al.*, 1978; *Hurukawa and Hirahara*, 1980; *Stefani et al.*, 1982; *Obara and Sato*, 1988; *Obara*, 1989; *Umino and Hasegawa*, 1993] or converted waves [e.g., *Okada*, 1971, 1979; *Snoke et al.*, 1977; *Hasegawa et al.*, 1978b; *Nakanishi*, 1980; *Nakanishi et al.*, 1981; *Chiu et al.*, 1985; *Matsuzawa et al.*, 1986b, 1990; *Iidaka et al.*, 1990] at the top or bottom of the subducted slab, because the reflected and converted waves are directly influenced by the velocity structure near the upper or lower surface of the slab, in contrast to waves propagating directly through the slab, which are influenced by average properties along their ray paths.

Hasegawa et al. [1978b] first investigated a geometrical relationship between the double-planed deep seismic zone and the subducted oceanic plate beneath the northeastern Japan arc. They located the upper boundary of the subducted Pacific plate below about 100-km depth by analyzing observed travel times of *ScS*-to-*P* converted waves (*ScSp*: a ray path of the *ScSp* wave, along with that of the *ScS* wave, is schematically shown in Figure 7b) at the boundary and from an analysis of *P*-wave travel time anomalies observed at a small-scale array from intermediate-depth earthquakes. The upper boundary of the subducted plate estimated by *Hasegawa et al.* [1978b] lies just above the upper seismic plane of the double seismic zone, as shown by a broken line and a hatched zone on a cross-arc vertical cross section of seismicity in Figure 8.

Detailed seismic velocity structure near the upper boundary of the subducted plate has been investigated by analyzing observed amplitude and polarity of the *ScSp* phase [*Okada*, 1979; *Nakanishi*, 1980]. *Okada* [1979] and *Nakanishi* [1980] pointed out the existence of an extremely thin low-velocity layer just above the upper surface of the high-velocity subducted plate in order to explain observed large amplitudes of the *ScSp* phase. Further, *Nakanishi* [1980] showed that observed polarity of the *ScSp* phase requires that the low-velocity layer has a sharp velocity boundary on the top and a transitional velocity boundary on the bottom.

Detection of *PS* phase mode- converted from *P* to *S* at the upper boundary of the subducted plate (Figure 7c) was made by *Matsuzawa et al.* [1986b] beneath northeastern Japan. They analyzed travel time data of the *PS* waves from earthquakes in the upper seismic plane and those in the lower plane. The observed difference in *PS*-*P* time between the upper seismic plane and the lower one leads to the conclusion that the subducted Pacific plate is composed of a thin (thickness less than 10 km) low-velocity upper layer and a thick high-velocity lower layer. The upper and lower layers have 6% lower and 6% higher velocities than the overlying mantle wedge, respectively. The top of the thin low-velocity layer, which is shown by a thick line in Figure 8b, is located just above the upper seismic plane of the double seismic zone. Earthquakes in the upper seismic plane, at least in the depth range 60-150 km, occur within the thin low-velocity upper layer.

The thickness of the low-velocity upper layer of the subducted Pacific plate was estimated by *Matsuzawa et al.* [1987]. They calculated synthetic seismograms of the *PS*

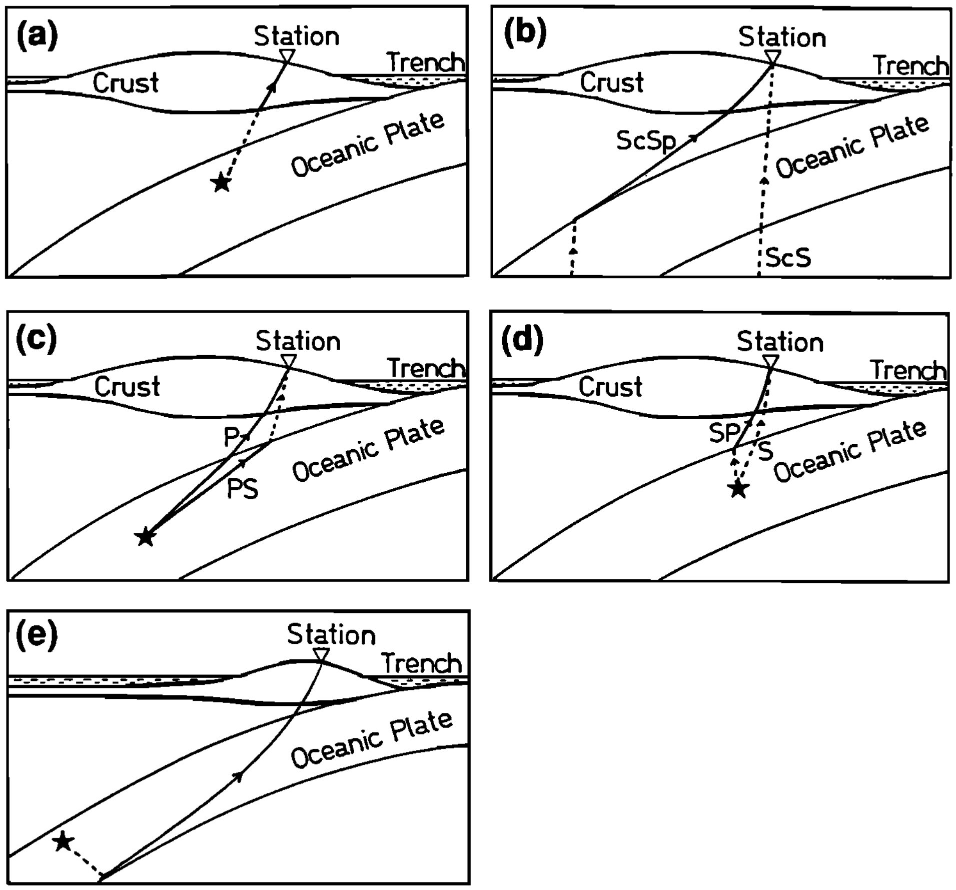


Figure 7. Schematic illustration of ray paths of (a) S to P converted wave at the Moho, (b) ScS to P converted wave ($ScSp$), (c) P to S converted wave (PS), (d) S to P converted wave (SP) at the top of the plate, and (e) S to P converted and reflected wave at the bottom of the plate. Ray paths of nonconverted or direct waves are also shown. Solid and broken lines denote P and S waves, respectively. Stars and inverted triangles indicate locations of source and station.

converted waves from events in the upper and lower seismic planes by using a two-dimensional (2-D) ray tracing algorithm developed by *Cerveny and Psencík* [1981]. Comparison of synthesized with observed wave forms of the PS phase indicates that the low-velocity upper layer is as thin as ~5 km. They also showed that the events accurately located in the upper seismic plane beneath northeastern Japan actually form a thin inclined seismic zone with a thickness of ~5 km at least below about 100-km depth. The upper plane seismicity in the shallower part is somewhat thicker than 5 km, which suggests that the low-velocity upper layer in the upper 100-km depth is slightly thicker than 5 km.

The S-to-P converted phase at the upper boundary of the subducted plate, as schematically shown in Figure 7d, was also detected by *Matsuzawa et al.* [1990] in seismograms of intermediate-depth earthquakes. They found that this phase can be observed at almost all stations in northeastern Japan. This enables them to estimate the depth distribution of the upper plate boundary in a much wider region than the case of the P-to-S converted wave. The depth distribution of the upper plate boundary in a beneath northeastern Japan using the inversion technique is nearly consistent with that of the upper seismic plane of the double seismic zone (Figure 3). The estimated location of the upper plate boundary is shown on a cross-arc vertical cross section of seismicity (Figure 8a), along with that estimated from the $ScSp$ phase and that from the travel time anomaly at a seismic array. The upper plate boundary is located just above the upper seismic plane, which is in good agreement with the results obtained from the analyses of $ScSp$ waves, PS converted waves, and seismic array observations.

Recently, *Umino and Hasegawa* [1993] detected a reflected and S-to-P mode-converted wave at the lower boundary of the subducted Pacific plate (Figure 7e) in seismograms of both intermediate-depth and deep earthquakes. Figure 9 is an example of seismograms showing clear later arrivals of the reflected SP wave at the bottom of the plate. The identification of these later arrivals as the reflected SP wave was made as follows. This unusual phase is most clearly defined on vertical component seismograms. Its directions of wave approach to observation stations are nearly the same as those of the direct P wave, and apparent velocities at the stations are the same as or slightly larger than those of the direct P wave. It has approximately the same frequency content as those of the direct P wave. The arrival time difference between this phase and the direct P wave ranges from 9 to 13 s, depending on the epicentral distance. It gradually decreases with increasing epicentral distance. This phase can be detected at many stations of northeastern Japan from many events with focal depths ranging from 250 to 550

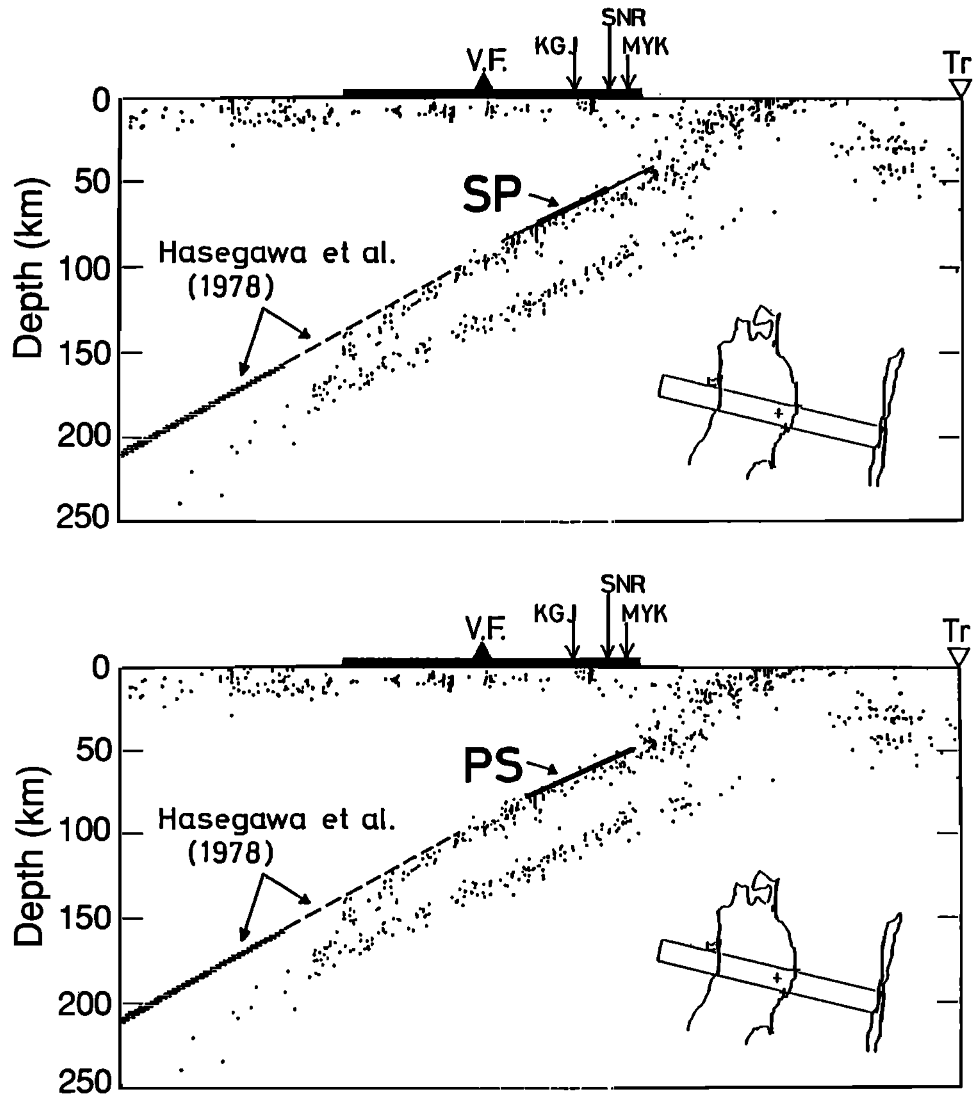


Figure 8. Estimated locations of the upper boundary of the subducted plate projected onto a cross-arc vertical cross section. Hatched line, dashed line, and solid line in the upper figure, and solid line in the lower denote the locations estimated from the *ScSp* wave, travel time anomaly at a seismic array [*Hasegawa et al.*, 1978b], the *SP* wave [*Matsuzawa et al.*, 1990], and the *PS* wave [*Matsuzawa et al.*, 1986b], respectively. Events located in the area indicated in the inset map are also plotted on the vertical section. Hypocenters are from the routine location procedures by the land seismic network, and so focal depths of shallow events beneath the Pacific Ocean are systematically biased (see Figures 1 and 2). The land area, the volcanic front, and the trench axis are shown on the top of each figure by a bold line, a solid triangle, and an open inverted triangle, respectively.

km. The above features show that this phase is generated by a reflection or conversion at a sharp velocity discontinuity near earthquake foci, then propagates through the subducted plate for a large part of its ray path, and finally reaches the observation stations at the surface as *P* waves.

Candidates for the velocity discontinuity are the upper and lower surfaces of the subducted plate and some velocity boundary within the plate such as the olivine-spinel phase transition zone. Two models have been proposed for the location and the configuration of the olivine-spinel phase boundary: one in which the location of the phase boundary, determined by equilibrium thermodynamics, is distorted upwards within the cold subducted plate [*Turcotte and Schubert*, 1971; *Schubert et al.*, 1975], and the other in which the boundary of the phase transition within the plate is distorted downwards due to a metastable olivine phase, which persists to a depth of ~600 km [*Sung and Burns*, 1976; *Liu*, 1983]. In any case, however, the phase transition zone cannot explain the observed feature that the seismic later phase can be detected even for events well above the estimated transition zone. Observed values of arrival time difference between the direct *P* wave and the later phase also cannot be explained by the phase transition zone model. Similarly, an *S*-to-*P* converted wave at the upper surface of the plate and a reflected *P* wave at the lower surface cannot explain the observed values of arrival time difference. *Umino and Hasegawa* [1993] showed on the basis of 2-D ray-tracing calculations that the arrival time of a reflected and *S*-to-*P* converted wave at the

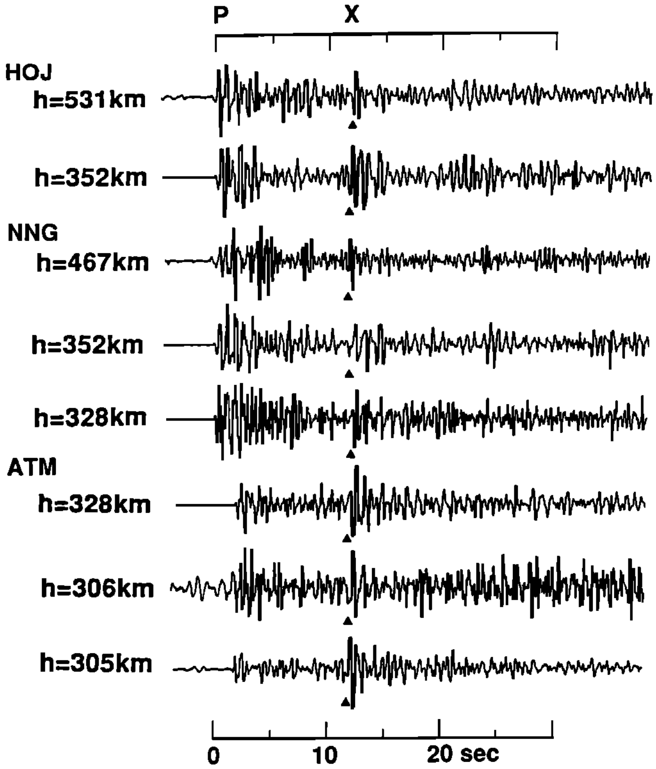


Figure 9. Examples of seismograms showing clear *SP* reflected waves at the bottom of the plate. Seismograms are at three stations (HOJ, NNG, and ATM) from deep events with focal depths (*h*) ranging from 305 to 531 km [after *Umino and Hasegawa*, 1993].

bottom of the plate coincides with that observed for wide ranges of focal depths of 250-550 km and of epicentral distances of 250-600 km.

Other types of converted or reflected waves at the bottom of the plate have not been detected yet, although reflected *S* waves, *S*-to-*P* converted waves, and *P*-to-*S* converted waves at the top surface of the plate have been successively reported [e.g., *Matsuzawa et al.*, 1986b, 1990; *Obara and Sato*, 1988; *Iidaka et al.*, 1990; *Okada*, 1971; *Hasegawa et al.*, 1978b; *Nakanishi*, 1980; *Snoke et al.*, 1977; *Tsumura et al.*, 1993]. This is considered to be partly because of a geometrical relationship between seismic stations, subducted plates, and earthquake foci within the plates. For example, *S*-to-*P* or *P*-to-*S* converted waves at the bottom of the plate from events within the plate, even if they actually exist, cannot reach the observation stations, which are not distributed homogeneously in space but usually on the land area of subduction zones. Similarly, *ScS*-to-*P* converted waves at the bottom of the plate, in the same depth range as the case of the upper surface conversion, cannot reach the stations on land but reach their trenchward side in the sea. Reflected waves at the bottom are the only case to reach the stations on land.

Umino and Hasegawa [1993] also detected the *S*-to-*P* converted wave at the upper boundary of the plate (Figure 7d), which was originally detected by *Matsuzawa et al.* [1990], in seismograms from earthquakes with the same depth range. Detection of the *SP* converted wave at the top and *SP* reflected wave at the bottom indicates that both the top and bottom of the subducted plate have sharp velocity boundaries. It also allows them to reliably estimate the thickness of the plate and to know where within the plate earthquakes are actually occurring. The result shows that the subducted Pacific plate is 80-90 km thick and that earthquakes in the depth range 250-550 km are occurring within the plate 20-30 km below the top of the plate.

Three-Dimensional Seismic Velocity and Attenuation Structures

Reflected or converted waves provide information on seismic velocity structures near the reflection or conversion planes. However, the generation of these waves requires the existence of a sharp velocity change at some boundary, and actually, the detection of these waves is not very frequent. Estimation of the seismic velocity structure by analyzing these phases is thus limited. We need other ways for surveying a much wider area, including the portions away from sharp velocity boundaries. The *P* and *S* wave travel time tomography developed by *Aki and Lee* [1976] and *Aki et al.* [1977] is a very effective method for estimating the 3-D seismic velocity structure in regions covered by dense seismic networks.

Studies of 3-D seismic velocity structure of subduction zones done most extensively on both regional and local scales are those of the Japanese Islands, because dense high-gain seismic networks have been installed by several institutions in this region. On a regional scale, *Hirahara* [1977] and *Hirahara and Mikumo* [1980] estimated the 3-D *P* wave velocity structure of the crust and upper mantle beneath the Japanese Islands by inverting travel time data of teleseismic and local earthquakes. They pointed out the existence of a *P*-wave high-velocity zone corresponding to the subducted Pacific plate. The inclined high-velocity Pacific plate has also been detected by the *P* wave tomographic studies of *Kamiya et al.* [1989] and *Zhou and Clayton* [1990]. They discussed the problem of further penetration of the Pacific plate into the lower mantle based on the tomographic images estimated to a depth of 1200 km, although the conclusions of the two studies opposed each other.

P wave tomographic images of the mantle to a depth of 1200 km beneath northwest Pacific island arcs, estimated recently by *van der Hilst et al.* [1991] by using a more realistic background Earth model and depth phase data as well as direct phase data, have yielded higher-quality images of the high-velocity Pacific plate. They have also shown that the subducted Pacific plate beneath the northeastern Japan arc does not penetrate into the lower mantle but is deflected at the boundary between the upper and lower mantle. Their tomographic images showing the slab deflection at the 670-km discontinuity agree well with the result of *Fukao et al.* [1992], who have also obtained high-quality images of the mantle to a depth of 1200 km beneath the Japanese Islands by solving simultaneously for updates of one-dimensional (1-D) background Earth models and for spatial variations in *P* wave velocity.

Many tomographic studies on a local scale also have been made in several regions beneath the Japanese Islands by inverting arrival time data from local earthquakes [e.g., *Miyamachi and Moriya*, 1984; *Nakanishi*, 1985; *Horie and Aki*, 1982; *Hasemi et al.*, 1984; *Obara et al.*, 1986; *Ishida and Hasemi*, 1988; *Hirahara et al.*, 1989]. These studies have also confirmed the existence of the inclined high-velocity Pacific plate. Moreover, some of these studies have detected the existence of low-velocity bodies in the crust and/or in the mantle wedge beneath the active volcanoes. *Hasemi et al.* [1984] obtained a detailed 3-D *P* wave velocity structure beneath the central part of northeastern Japan to a depth of 200 km, and have revealed low-velocity zones continuously distributed in the crust and in the mantle wedge beneath the active volcanoes. A tomographic study subsequently made by *Obara et al.* [1986] for 3-D *P* and *S* wave velocity structures covering a wider area of northeastern Japan confirms the low-velocity zones extensively existing in the crust and in the mantle wedge beneath the active volcanoes.

More distinct *P* and *S* wave tomographic images of the crust and the upper mantle beneath northeastern Japan recently obtained by *Zhao et al.* [1992b] and *Zhao and Hasegawa* [1993] have updated the works by *Hasemi et al.* [1984] and *Obara et al.* [1986] by improving the resolution. They developed a tomographic method that can deal with a general velocity model with complex velocity discontinuities in the modeling space and with 3-D velocity variations in each layer bounded by the velocity discontinuities. The estimated 3-D *P* wave velocity structure of the crust and the upper mantle is shown on three vertical cross sections which are nearly perpendicular to the trench axis (Figure 10a, 10b, and 10c).

The inclined high-velocity Pacific plate is clearly delineated in all the three vertical cross sections in Figure 10. The thickness of the high-velocity plate is estimated to be 80-90 km, which is in good agreement with the estimation from the reflected *SP* waves at the bottom of the plate [*Umino and Hasegawa*, 1993]. The bottom of the subducted plate as imaged by the tomographic inversion (Figure 10) has a sharp velocity contrast, which again agrees with the existence of the reflected seismic waves at the bottom. A recent work of seismic tomography of the Japan subduction zone by *Zhao et al.* [this issue] used local, regional, and teleseismic events simultaneously, and thus has a higher resolution for areas around the bottom of the plate due to the addition of many more seismic rays passing through those areas. This work also delineates the high-velocity Pacific plate with a thickness of 80-90 km and with sharp velocity boundaries on the top and bottom, although the lower boundary in some areas is rather gradational. Earthquakes forming the double-planed deep seismic zone beneath northeastern Japan [*Umino and Hasegawa*, 1975; *Hasegawa et al.*, 1978a] are located in the upper half of the high-velocity subducted Pacific plate. Low-velocity zones are distributed continuously from the crust to the mantle wedge beneath the active volcanoes. They are inclined to the west in the mantle wedge and are nearly parallel to the dip of the high-velocity subducting Pacific plate. This is seen from all three vertical cross sections shown in Figure 10.

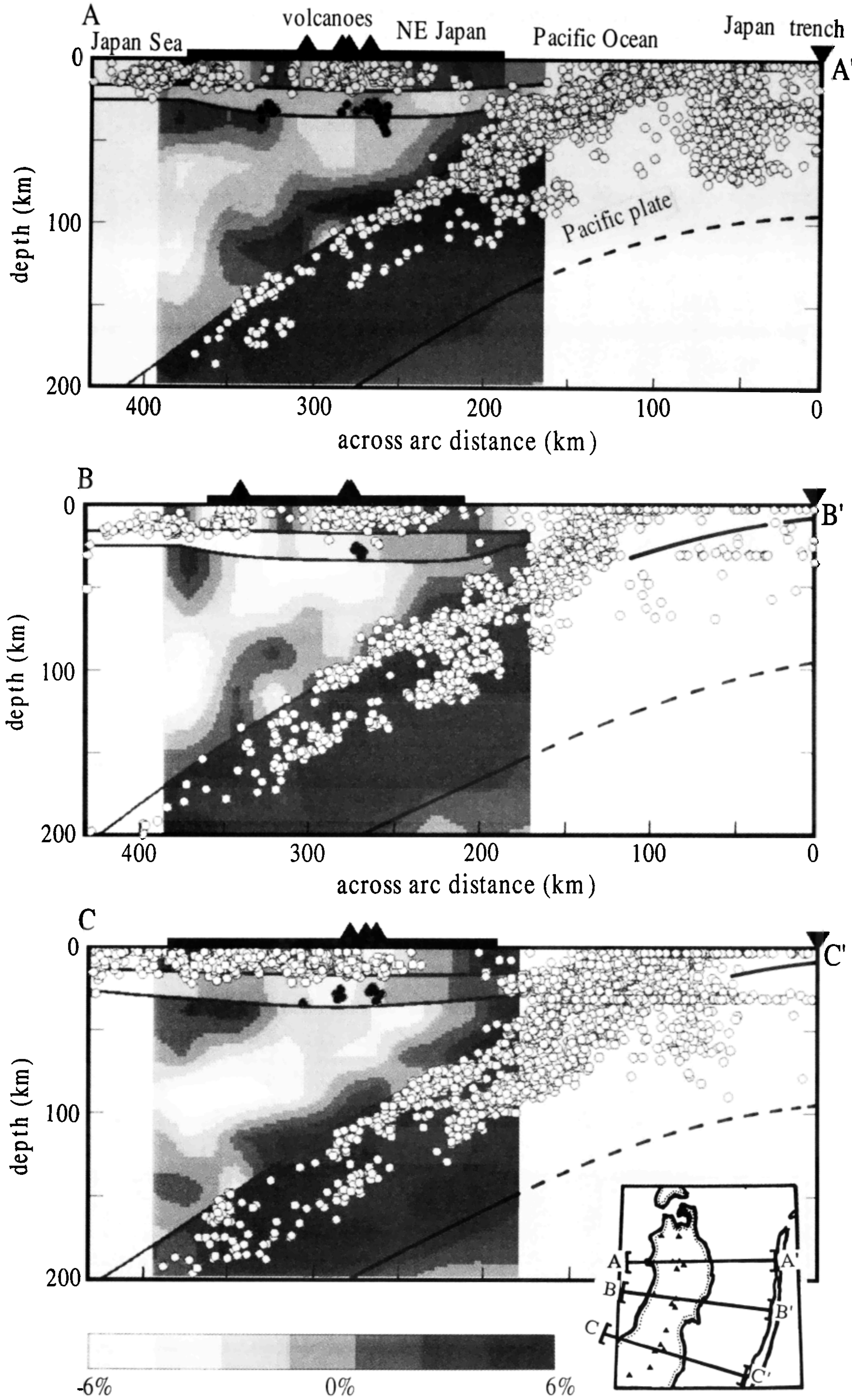
The seismic attenuation structures of the crust and the upper mantle have been estimated from observed data of seismic wave attenuation. Tomographic methods used in the travel time inversion also can be applied to seismic wave amplitude data or to amplitude spectrum data for estimating the 3-D attenuation structure of the medium. *Umino and Hasegawa* [1984] estimated the 3-D attenuation structure of the crust and the upper mantle beneath northeastern Japan by applying a block inversion technique to observed *S* to *P* wave spectral ratios. Their result shows low Q_s values in the crust beneath the active volcanoes, a low Q_s value in the mantle wedge on the back arc side of the volcanic front, an intermediate Q_s value in the mantle wedge on the forearc side of it, and a high Q_s value in the subducted Pacific plate. Figure 11 shows cross-arc vertical cross sections of the estimated attenuation structure. Regional variation of the Q_p value of the uppermost mantle beneath northeastern Japan, estimated by taking a twofold spectral ratio of station pair and source pair located on a straight line [*Matsuzawa et al.*, 1989], also shows a low Q_p value on the back arc side of the volcanic front and an intermediate value on the forearc side.

Matsumoto and Hasegawa [1989], based on the single scattering model, have estimated coda Q values in the crust and the uppermost mantle beneath northeastern Japan for several frequency bands from the time decay of observed coda wave amplitudes. They developed a method for estimating 2-D coda Q structure by inverting the obtained apparent values of coda Q between station and source. The estimated 2-D coda Q structure shows that the coda Q is low in the regions around the active volcanoes and near the coast of the Japan Sea, while it is high near the coast of the Pacific Ocean. A similar attenuation structure of the crust and the upper mantle has been estimated even from observed data of the seismic intensity [*Hashida and Shimazaki*, 1984, 1987]. *Hashida and Shimazaki* [1987] applied a 3-D block inversion method to observed seismic intensity data, and estimated the attenuation structure of the northeastern Japan arc. Subsequently, *Hashida* [1989] estimated the attenuation structure of the crust and the upper mantle beneath the Japanese Islands in the same way as *Hashida and Shimazaki* [1987]. They imaged low Q zones in the mantle wedge to a depth of ~90 km beneath the active volcanoes, and high Q zones corresponding to the subducted Pacific plate.

The estimated images of the low Q zones in the mantle wedge beneath the active volcanoes and the high Q zone corresponding to the Pacific plate are less clear than those of the *P* wave low-velocity zones and the high-velocity zone resolved by the travel time inversions. This is because of the lower spatial resolution of the attenuation tomographic studies. Considering the difference in spatial resolution, low (high) Q regions generally correspond to low (high) velocity regions.

Discussion

Figure 12 shows a schematic illustration of the cross-arc vertical cross-sectional model of the crust and the upper mantle beneath the central part of northeastern Japan, which is inferred from the seismic observations described in the previous sections. The oceanic Pacific plate subducts at an extremely low angle for its first descent to ~25-km depth. Normal-fault type events occur in the shallowest portion of the Pacific plate near the trench axis perhaps due to the bending of the subducting plate. The large Sanriku Earthquake of 1933 (*M* 8.1) occurred in this region just beneath the trench axis with a large-scale normal faulting extending over the entire thickness of the subducting Pacific plate [*Kanamori*, 1971]. Low-angle thrust-fault type events, directly related with the downgoing motion of the Pacific plate, occur along the boundary of the two converging plates down to a depth of ~50 km. The subducting Pacific plate dips steeper at an angle of 25-30° at depths deeper than ~30 km. The downward-bending of the plate at this position is much greater than that near the trench axis. Normal fault type events at the top of the plate (and thrust fault type events at the middle of it) are


A
volcanoes
Japan Sea
NE Japan
Pacific Ocean
Japan trench
A'
Pacific plate
depth (km)
0
100
200
400
300
200
100
0
across arc distance (km)
B
B'
depth (km)
0
100
200
400
300
200
100
0
across arc distance (km)
C
C'
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-6%
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6%
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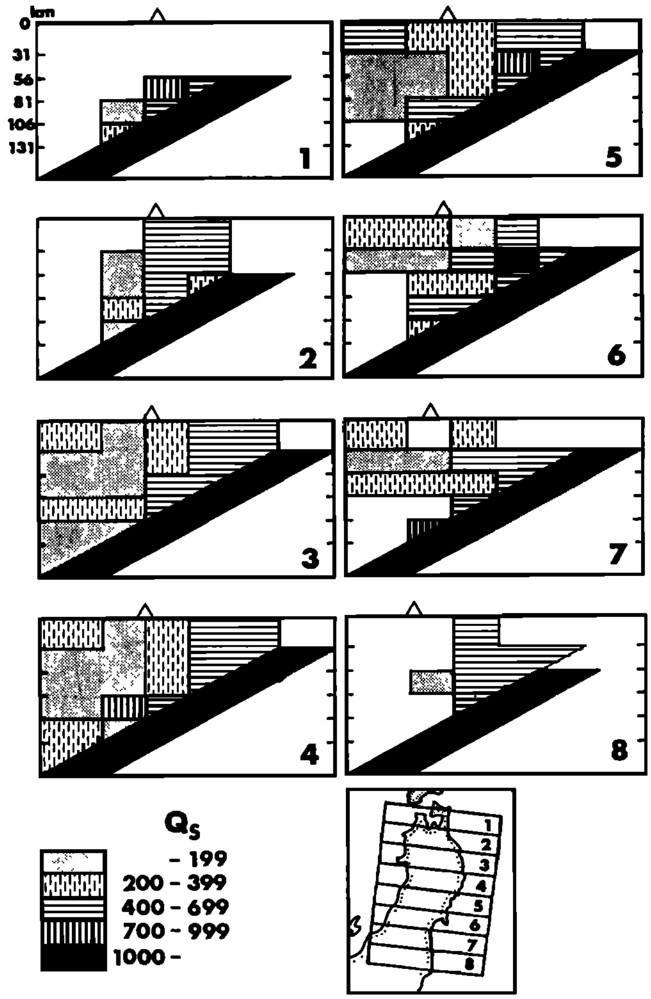

Figure 11. Cross-arc vertical cross-sectional view of the attenuation structure beneath northeastern Japan estimated from a block inversion by using spectral amplitude ratios of P and S waves from intermediate-depth earthquakes. Estimated Qs value in each block is shown by the scale at the bottom. Open triangles denote the location of the volcanic front [after *Umino and Hasegawa*, 1984].

expected to occur at this position from this steep downward-bending, which seems not to be the case. More precise study on seismicity and focal mechanisms for the events beneath the Pacific Ocean is required for correctly understanding the subduction process of the oceanic plate actually progressing in this arc.

Beneath the land area, the subducted Pacific plate is imaged by the tomographic studies as an inclined high-velocity and high Q zone with a thickness of 80-90 km. The uppermost portion of the plate with a thickness of 5 km or so, which is estimated as a low-velocity layer from the analysis of PS converted waves, cannot be detected by the tomographic studies using mainly direct P and S wave arrival time data due to their lower spatial resolutions. Earthquakes in the deep seismic zone form a double seismic zone in the depth range 70-150 km, which is located in the elastic upper half of the subducting plate; the lower half flows plastically rather than deforming by earthquake breakage. The upper plane seismicity is confined to the thin low-velocity upper layer of the plate, while the lower plane seismicity takes place at the middle of the thick high-velocity and high Q lower layer.

The thin low-velocity layer in the upper portion of the subducted Philippine Sea plate, similar to the present case, is found in the Kinki and Kanto districts, the central to southwestern part of Honshu, Japan, by analyzing guided waves traveling within the low-velocity layer [*Fukao et al.*, 1983; *Hori et al.*, 1985; *Hori*, 1990; *Oda et al.*, 1990]. Inclined low-velocity zones with high seismicities are also found in Pamir-Hindu Kush [*Roecker*, 1982] and in Hokkaido [*Takanami*, 1982; *Miyamachi and Moriya*, 1984]. All these low-velocity zones have been interpreted as the former oceanic crust subducted with the oceanic plate. The low-velocity layer beneath northeastern Japan attains a depth of ~150 km, which is much deeper than low-velocity layers detected in the other regions. Gabbro within the subducted oceanic crust will change to eclogite at a certain depth [e.g., *Ringwood and Green*, 1966]. The oceanic crust, after the phase transformation from gabbro to eclogite occurred, would not show a seismic low velocity compared with the overlying mantle wedge. These facts leave a serious problem to be solved. If we interprete the thin low-velocity layer beneath northeastern Japan as the former oceanic crust capping the subducted Pacific plate, the phase transformation would not occur at least to a depth of ~150 km, which does not agree with the result obtained so far from laboratory measurements. *Sacks* [1983] pointed out that the phase change rate is extremely slow if the temperature in the subducted oceanic crust is sufficiently low. Unfortunately, we do not know the exact temperature in such a depth, and thus the problem has been left unsolved.

The cause for formation of the double-planed deep seismic zone has not been revealed yet. Four models have been proposed as the possible cause for the double seismic zone: unbending of the subducted oceanic plate [*Isacks and Barazangi*, 1977; *Engdahl and Scholz*, 1977], phase changes within the plate [*Veith*, 1974], sagging of the plate in a less viscous asthenosphere [*Sleep*, 1979], and thermal stress within the plate [*Yang et al.*, 1977; *House and Jacob*, 1982; *Goto et al.*, 1983]. However, no single explanation, listed above, seems entirely satisfactory or to be universally applicable. For example, stresses generated by the unbending of the plate, one of the likeliest models of those proposed to

Figure 10. Cross-arc vertical cross sections of fractional P wave velocity perturbation (in percent) along the lines (a) AA', (b) BB', and (c) CC' in the inset map [*Zhao et al.*, 1992b]. Velocity perturbation is shown by the shading scale at the bottom. Circles are earthquakes located within a 60-km width along each line. Solid circles denote deep low-frequency microearthquakes. Hypocenters are from the routine location procedures, and so focal depths of shallow events beneath the Pacific Ocean are systematically biased. The land area, the active volcanoes, and the trench axis are shown by a bold line, solid triangles, and an inverted solid triangle at the top of each figure. Locations of the Conrad and the Moho discontinuities, and the top and bottom of the Pacific plate are also shown by bold lines.

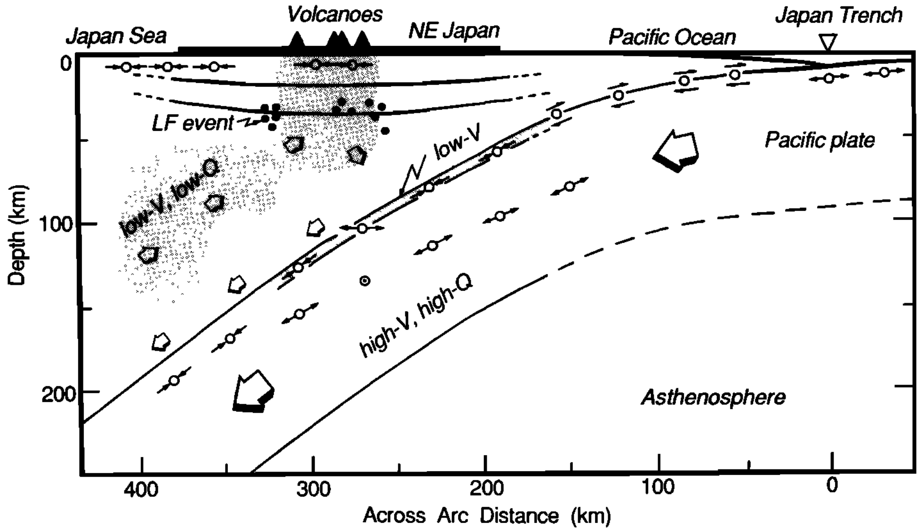

Figure 12. Schematic cross-arc vertical cross section of the crust and the upper mantle in the northeastern Japan convergent margin inferred from seismic observations. Open circles with arrows indicate dominant focal mechanisms of earthquakes. Solid circles show deep low-frequency microearthquakes occurring in or around the low-velocity zones beneath active volcanoes. Low-velocity zones in the crust and the mantle wedge and in the uppermost portion of the subducted Pacific plate are shown by shaded areas.

date, must surely be acting within the plate. The stress state actually observed on the double seismic zone, which is schematically shown in Figure 12, is a little more complicated than that simply expected from the unbending model. *Matsuzawa et al.* [1986a] attributed the local change in dominant focal mechanism of the double seismic zone just beneath the volcanic front to the local downward bending of the subducted plate at that depth. Even at the junction between the Kurile and the northeastern Japan arcs and that between the northeastern Japan and Izu-Bonin arcs, where the subducting plate is considerably bent in parallel to the arcuation of the trench axis, the double seismic zone is still distributed continuously from both sides of the arcs in the upper 100-150 km depths (Figure 3). Although the deep seismic zone beneath Hokkaido is also double-planed in nearly the same depth range, the upper seismic plane is not downdip compression but has various types of focal mechanisms. *Hasegawa and Takagi* [1987] pointed out that the difference in stress state between Tohoku and Hokkaido is caused by the difference in the intensity of slab pull force, superimposed on the stress system generating the double seismic zone.

The double seismic zones, similar to the present one, have been found in several other subduction zones. Among these, the upper seismic plane of the double seismic zone beneath the Shumagin Islands, Alaska [*Reyners and Coles*, 1982], and that beneath the central North Island, New Zealand [*Reyners*, 1980], predominantly have downdip extensional mechanisms. Moreover, the double seismic zone beneath the southern North Island of New Zealand has downdip extensional stresses both for the upper and lower seismic planes [*Robinson*, 1986]. In any case, these observations suggest that several separate stress systems are acting on the descending plates, and that the difference in their relative importances, as a result of their superposition, will cause diversity in the patterns of seismicities and earthquake-generating stresses within the subducted plates.

In the mantle wedge, low-velocity and low Q zones are distributed in parallel to the dip of the subducting plate. A mechanically induced secondary convection will be generated in the overlying mantle wedge by the subduction of the oceanic plate [e.g., *McKenzie*, 1969; *Sleep and Toksöz*, 1973; *Toksöz and Bird*, 1977]. The existence of the inclined low-velocity, low Q zones suggests that they are images of the ascending flow of hot mantle material from depth, that is, a portion of the secondary subduction-induced convection (small open arrows in Figure 12). Decompression melting within the ascending flow produces low seismic velocities and high seismic attenuations.

Magma within the ascending flow finally reaches the top of the mantle and there it stagnates temporarily as mantle diapir. Anomalously deep low-frequency microearthquakes (solid circles in Figure 12) are located at the base of the volcanoes and/or around the low-velocity zones [*Hasegawa et al.*, 1991b]. They are probably generated by the magmatic activity of mantle diapir at the top of the mantle. Magma further migrates upward into the crust, suffering fractional crystallization and chemical reaction with the surrounding materials. This upward- migrating magma in the crust again produces a low velocity and a high attenuation around it. In fact, evidence for the temporary storage of magma at midcrustal levels has been found in or around the low-velocity zones as distinct S wave reflectors, which are considered to be the tops of thin magma bodies existing in the midcrust [*Mizoue*, 1980; *Mizoue et al.*, 1982; *Horiuchi et al.*, 1988; *Iwase et al.*, 1989; *Hori and Hasegawa*, 1991; *Hasegawa et al.*, 1991b]. Repeated discharges of magma, that finally reached to the Earth's surface, form the arc volcanoes (solid triangles in Figure 12).

The upper ~15 km of the crust forms a brittle seismogenic zone. The lower portion of the crust and the mantle wedge are governed by creep or flow, being weak and incapable of supporting much stress. Horizontal compressional stress caused by the convergence between the subducting plate and the overlying continental plate is thus supported mostly by the upper ~15 km of the crust, the relatively strong seismogenic zone, resulting in the occurrence of shallow thrust fault type earthquakes within it. Relatively high seismicity along the volcanic front is perhaps due to the local elevation of the base of the brittle seismogenic zone beneath such volcanic regions with higher temperatures. Stress concentration will arise in or around the places, such as the volcanic regions, where the thickness of the relatively strong seismogenic zone is locally thin.

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