Anomalously Early First Arrivals to the J-Array from Teleseismic Events

Hiroko Sugioka, Yoshio Fukao, and Shin'ichi Sakai

Earthquake Research Institute, The University of Tokyo
Bunkyo-ku, Tokyo 113, Japan

An analysis of hundreds of seismograms from a seismic array (J-Array) in the Japanese archipelago shows travel-time anomalies on the order of 4–10 s for P-waves from teleseismic events in some source-receiver geometries. The source-receiver pairs exhibiting such large anomalies are 1) Philippine to Indonesia–Kanto, Japan, 2) Kamchatka–Southern Hokkaido to Tohoku, Japan, and 3) Taiwan to Ryukyu–Shikoku to Kinki, Japan. A simple ray-path consideration indicates that the rays responsible for these anomalies are those incident to the base of the Izu–Bonin slab from below at near critical angles and those propagating laterally within the Kurile slab and the Ryukyu slab along their strikes. The travel times of the first arrivals are calculated for slab structural models using the method of Sakai (1992), which traces progressive expansion of a wavefront away from the source based on the Huygens' principle. This method has an advantage over conventional ray tracing methods in its robustness against the complexity of a laterally heterogeneous velocity structure. A slab model is specified by its thickness and velocity contrast with the surrounding mantle. Numerical experiments show that the effect of slab thickness is different from the effect of velocity contrast, and that their effects are separable. With an important assumption that the IASP91 model approximates the surrounding mantle, the observed negative anomalies can be consistently explained by a 50-km thick slab with a velocity contrast of 5% for the Pacific plate and by a 30-km thick slab with a velocity contrast of 5% for the Philippine Sea plate, although the observed records exhibit negative anomalies that are unable to be explained by the mere introduction of a high-velocity slab.

1. Introduction

Various P-wave tomographic studies have revealed the first order feature of subducted slabs of lithosphere along the Circum Pacific (Kamiya *et al.*, 1988, 1989; Suetsugu, 1989; Spakman *et al.*, 1989; Zhou and Clayton, 1990; van der Hilst *et al.*, 1991, 1993; Fukao *et al.*, 1992; Zhou, 1990; van der Hilst and Spakman, 1987, 1989). In particular, the upper surface of the subducting slab beneath northern Honshu and Izu has been carefully investigated using the ScS-to-P converted phase (Okada, 1977; Nakanishi *et al.*, 1981), the P-to-S or S-to-P converted phase (Matsuzawa *et al.*, 1986, 1989; Matsuzawa, 1989; Iidaka *et al.*, 1990), and the critically reflected P-phase (Fukao *et al.*, 1978). The upper surface of the subducted slab of the Philippine Sea plate has been studied using the ScS-to-P converted phase (Nakanishi and Horie, 1980) and mantle-

trapped P- and S-waves (Fukao *et al.*, 1983; Hori *et al.*, 1985; Oda *et al.*, 1990).

On the other hand, little is known about the lower surface of subducted slabs and, therefore, the thickness of the slab. The tomographic image of the slab under northern Honshu has been obtained in the finest resolution by a group from Tohoku University (Zhao *et al.*, 1990, 1992; Zhao and Hasegawa, 1993), which indicated a slab thickness of 80–90 km. A possible problem associated with this type of regional three-dimensional (3-D) tomography is a preferential propagation of first arrival rays within the slab without any outward refraction. The inverted structure near the base of the slab may therefore tend to remain close to the initial model (Kamiya, personal communication, 1995). Addition of the data of first arrival rays near-critically incident to the slab from below would increase the uniqueness of the solution. This study was motivated

Received October 5, 1995; Accepted January 10, 1996

* To whom correspondence should be addressed.

ed by our finding of such unique rays in the array records made on the Japanese archipelago.

Another type of first-arrival ray that may carry information about the slab thickness are those rays propagating within a slab over long distances along a trench that grazes the lowermost part of the slab. Ansell and Gubbins (1986), Gubbins and Snieder (1991) and Smith *et al.* (1994) reported negative anomalies of more than 10s for first-arrival rays travelling over a distance of 2,000 km along the Tonga–Kermadec Trench. They attributed such large anomalies to the preferential propagation of the first-arrival rays within the high-velocity slab. With the hope that such rays could provide us with information about the lowermost part of the slab, we searched for a similar phase in the array records in Japan.

Analyses of the above phases of slab origin do not allow a conventional one-dimensional treatment for the Earth's structure. To calculate the travel times of high-frequency seismic waves in a three-dimensional structure, ray tracing methods are most commonly used. These are classified into the shooting method (Jacob, 1970; Cervený and Ravindra, 1971; Cervený *et al.*, 1977; Kim and Cormier, 1990) and the bending method (Wesson, 1971; Julian and Gubbins, 1977; Koketsu, 1991). An alternative approach is to calculate progressive expansion of the wavefront based on the Huygens' principle, a method developed by Sakai (1992). This method is advantageous over ray tracing methods in its robustness against the complexity of three-dimensional structures and in its computational speed of dealing with a large number of observations over a wide area. We take full advantage of this method to analyze the slab-related phases.

The aim of this paper is to report the slab-related first-arrival phases we found and to show how the method of Sakai (1992) is useful to analyze such phases. It is beyond the scope of this study to obtain a detailed upper-mantle model based on an extensive data set.

2. Data

The J-Array is a large-aperture short-period (1 Hz) seismic array across Japan which began recording in April 1991. (J-Array Group, The J-Array Program) It combines several regional seismic networks developed under a national programme of Japan for researching structures of the Earth's interior. With an aperture of 3,000 km in length

by 300 km in width, it covers the entirety of the Japanese archipelago (Fig. 1). With more than 300 high-quality short-period vertical-component seismographs, the J-Array is useful for analyzing detailed structures of the Earth's interior. The data used are selected from earthquakes that occurred during the period from April 1991 to November 1993. Table 1 lists the hypocentral parameters of 10 earthquakes used in the analysis.

Since we use the IASP91 travel-time table (Kennett and Engdahl, 1991) in the analysis, we relocate the events using this table whenever the first arrival time data are available from the bulletins of the International Seismological Center (ISC) or Earthquake Data Report (EDR). The sources of these 10 earthquakes can be grouped into three regions: Philippine–Indonesia, Kamchatka, and Ryukyu–Taiwan. The first arrivals in each group show mutually similar patterns in travel-time anomaly, although the degree of anomaly varies depending upon the source locality. In the following, we take an example earthquake from each group. Figure 2

Fig. 1. Geographic distribution of the J-Array stations. With a length of 3,000 km and a width of 300 km, we can regard this array as a huge telescope to look into the Earth's interior. The region names and some selected station names are provided.

Table 1. Earthquake data from the ISC catalogue.

Date			Time			Lat. (deg.)	Lon. (deg.)	Depth (km)	Region	M_w	M_s
year	m	day	h	min	s						
From Philippines to Kanto											
1991	05	21	11	00	21.9	-7.52	126.53	32.3	Banda Sea	6.1	6.2
1991	11	13	11	12	15.6	8.38	126.41	57.4	Mindanao, Philippines	6.1	6.4
1991	12	29	18	39	10.6	-4.42	132.76	47.0	West Irian, Java	6.0	5.9
1992	01	13	11	58	26.0	1.35	127.38	118.0	Halmahere, Indonesia	5.8	
1992	02	06	01	12	38.2	-5.71	103.13	35.0	Southern Sumatra	6.0	6.5
From Kamchatka to Hokkaido and Tohoku											
1992	03	02	12	29	39.6	52.92	159.89	38.6	E. coast of Kamchatka	6.5	6.8
1993	11	13	01	18	04.2	51.93	158.65	34.0	Near E. coast of Kamchatka	7.1	
From Ryukyu to Shikoku and Kinki											
1991	04	14	08	08	56.7	27.19	127.48	91.0	Ryukyu Islands	6.1	
1992	03	15	01	01	27.1	23.53	123.57	37.0	Southwestern Ryukyu	5.6	5.9
1992	04	19	18	32	20.9	23.90	121.60	26.0	Taiwan	5.8	6.3

Fig. 2. Geographical distribution of the earthquakes used in this study. Different symbols are used for different groups of earthquakes (see Table 1). Representative ray paths are shown by solid lines for the three example earthquakes.

illustrates the epicenters of the 10 earthquakes and the typical ray paths from the three example earthquakes to a J-Array station, NIK (see Fig. 1 for the location).

Table 2 compares the relocated hypocenters to the original ISC hypocenters. The relocated foci always tend to be deeper than the original ones. We also calculate the hypocenter of the 1991 Mindanao event using the laterally heterogeneous mantle model of Fukao *et al.* (1992). The result is added in Table 2. The focal depth in this case is very close to the

ISC depth and quite different from the IASP91 depth.

3. Travel-Time Anomalies

Figure 3 shows the seismic section of the 1991 Mindanao earthquake. The solid line indicates the travel-time curve calculated from the IASP91 model. The observed first arrivals are significantly early relative to the calculated arrival times in the Kanto region, and to a lesser extent in the Kyushu region (see Fig. 1 for the region names). The maximum negative anomaly is about 4 s at epicentral distances around 30°. Travel-time anomalies of this magnitude at such teleseismic distances are rather surprising, which motivated us to study the travel-time anomalies of other teleseismic events. Figures 4 and 5 show the records of the 1992 Kamchatka earthquake and the 1992 Ryukyu earthquake, respectively. For the Kamchatka event, early arrivals with anomalies up to -10 s are observed in the southern Hokkaido to Tohoku regions (Fig. 1), where the observed first arrivals are rich in high-frequency content. For the Ryukyu event, on the other hand, negative anomalies up to -8 s are found in the Shikoku to Kinki regions, and to a lesser extent in the northern Kanto to Tohoku regions, where again high-frequency component is dominant in the first-arrival phase.

For all the events listed in Table 1, the observed arrival times are read visually and the residuals from the calculated arrival times are plotted on maps. Figures 6(a), 7(a), and 8(a) show such plots for the

Table 2. Comparison of the hypocentral determinations

	Date			Time			Lat. (deg.)	Lon. (deg.)	Depth (km)
	year	m	day	h	min	s			
For Banda Sea earthquake									
ISC	1991	05	21	11	00	21.9	-7.52	126.53	32.3
IASP91	1991	05	21	11	00	24.3	-7.50	126.55	36.3
For Mindanao earthquake									
ISC	1991	11	13	11	12	15.6	8.38	126.41	57.4
IASP91	1991	11	13	11	12	19.7	8.42	126.44	79.2
FOIN*	1991	11	13	11	12	18.4	8.43	126.34	56.8
For E. coast of Kamchatka earthquake									
ISC	1992	03	02	12	29	39.9	52.92	159.89	38.6
IASP91	1992	03	02	12	29	45.7	52.85	159.86	81.9
For near E. coast of Kamchatka earthquake									
ISC	1993	11	13	01	18	04.2	51.93	158.65	34.0
IASP91	1993	11	13	01	18	08.3	51.94	158.59	55.8

* Hypocenters based on the model of Fukao *et al.* (1992).

Fig. 3. Seismograms of the Mindanao earthquake of 13 Nov. 1991. The theoretical travel-time curve for the IASP91 model is indicated by the solid line. The broken line indicates the theoretical travel-time curve for the spherically averaged version of the laterally heterogeneous model of Fukao *et al.* (1992). The first arrivals read visually are marked by dots. Station codes are attached to selected traces.

Fig. 4. Seismograms of the Kamchatka earthquake of 2 Mar. 1992. See Fig. 3 for other explanations.

example earthquakes. The negative anomalies as explained above are clearly recognized. For the Kamchatka event, the high-frequency phase with negative anomalies as large as 10 s has a ray path from Kamchatka to Japan along the Kurile arc. For the Ryukyu event, the ray path runs from Ryukyu

Fig. 5. Seismograms of the Ryukyu earthquake of 15 Mar. 1992. See Fig. 3 for other explanations.

Fig. 6. Travel-time anomalies for the Mindanao earthquake. The residuals are calculated relative to the IASP91 model. (a) Observed travel-time anomalies; (b), (c), and (d) calculated travel-time anomalies. The values in brackets are the thickness and velocity contrast of the subducted slab of the Pacific plate. Those of the slab of the Philippine Sea plate are 30 km and 5%.

Fig. 7. Travel-time anomalies for the Kamchatka earthquake. See Fig. 7 for other explanations.

Fig. 8. Travel-time anomalies for the Ryukyu earthquake. The residuals are calculated relative to the IASP91 model. (a) Observed travel-time anomalies; (b), (c), and (d) calculated travel-time anomalies. The values in brackets are the thickness and velocity contrast of the subducted slab of the Philippine Sea plate. Those of the slab of the Pacific plate are 50 km and 5%.

Fig. 9. Observed travel-time anomalies plotted against the incident angle to the deep seismic plane. The rays from the Mindanao earthquake are incident to the slab from below.

to Japan along the Ryukyu arc. The arrivals of this high-frequency phase are often followed by the later arrivals of a lower-frequency phase with larger amplitudes around the expected first arrival times from the IASP91 table. These characteristics are remarkably similar to those of the first arrival phase reaching New Zealand laterally through the subducting slab beneath the Tonga–Kermadec arc (Ansell and Gubbins, 1986; Gubbins and Snieder, 1991; Smith *et al.*, 1994), suggesting a common mechanism for seismic wave propagation. We, therefore, interpret the observed high-frequency phase as the wave propagating either within the slab of the Pacific plate along the Kurile Trench or within the slab of the Philippine Sea plate along the Ryukyu Trench. The lower-frequency later phase may be interpreted as one arriving along a normal path. These initial and later phases cannot be interpreted as those refracted above and at the 410 km discontinuity in the normal mantle, since the initial phase refracted below the discontinuity follows the general trend of the IASP91 curve.

The early arrivals from the Mindanao event are dissimilar to those from the Kamchatka or Ryukyu event in that they are not particularly rich in high-frequency component (Fig. 3). The early arriving rays are incident to the base of the Izu–Bonin slab from below. Figure 9 plots the observed travel-time anomalies against the incident angle of a ray to the

Fig. 10. Contour lines of the deep seismic plane which is modeled as the upper surface of the subducted slab. The contour lines are drawn by referring to the hypocentral data given by the ISC.

deep seismic plane. The incident angle is measured from the perpendicular to the deep seismic plane. Figure 10 shows the deep seismic plane simulated by contouring the hypocentral distribution of deep and intermediate-depth earthquakes, referring to the ISC catalogue, which will be modelled later as the upper surface of a subducted slab.

In Fig. 9, the negative travel-time anomaly increases as the incident angle increases, although the scatter is large. The largest negative anomaly of about 4 s is observed near an incident angle of 70°. We note that the incident angle of 70° corresponds to a critical angle for refraction across a velocity contrast of 6%. A velocity contrast of this magnitude appears to be reasonable for being between the subducting slab and surrounding mantle (e.g., Zhao and Hasegawa, 1993). Figure 9 therefore indicates that the rays responsible for the large negative anomalies are those near-critically refracted into the subducted slab from below to propagate within the slab over long distances. The relevant ray paths are schematically illustrated in Fig. 11(a). Figure 11(b) illustrates the ray paths of the high-frequency waves laterally propagating through the slab from Kamchatka to Japan and from Ryukyu to Japan. Next, we will make a more quantitative analysis based on the hypothesis developed above.

4. Wavefront Propagation Method

There are several methods to calculate travel

Fig. 11. Schematic illustration of the propagation mechanism for the first-arrival phases: (a) from Philippines to Kanto, Japan, and (b) from Kamchatka to Hokkaido and Tohoku, Japan, or from Ryukyu to Shikoku and Kinki, Japan. Broken line indicates the ray path based on a ray theory for a laterally homogeneous Earth.

times for a laterally heterogeneous Earth model, most typical of which are two-point ray tracing methods such as the bending method (Wesson, 1971; Julian and Gubbins, 1977) and shooting method (Červený and Ravindra, 1971; Červený *et al.*, 1977; Jacob, 1970). Vidale (1988, 1990) proposed a different approach; a finite difference scheme to calculate the times of passage of a wavefront through grids at each of which the velocity is given. Sakai (1992) also used the concept of wavefront propagation to calculate travel times but applied the Huygens' principle to trace its progressive expansion away from the source. We adopt, in this paper, the wavefront propagation method of Sakai (1992) because of its ease of dealing with complex three-dimensional structures. A conventional ray tracing approach, in general, requires a more time-consuming and more involved scheme to find the

Fig. 12. Iterative construction of a progressively expanding wavefront (Sakai, 1992). (a) Construction based on the Huygens' principle. A two-dimensional wavefront is approximated by a chain of straight lines. See the text for details. (b) Some detail of the procedure for wavefront construction. Add an extra point if the adjacent new points are too widely spaced, as in W. Add extra points if the curvature of the wavefront is too large, as in S. Discard the points, as in L, unless they are confirmed to be on the wavefront of the first-arrival phase.

first arrivals among other later arrivals if the structure is more complex. In the wavefront propagation method, on the other hand, no special care has to be taken into account for complexity of the three-dimensional structures: head waves are accurately treated and shadow zones are filled with diffracted waves.

Figure 12(a) outlines the essence of this method

in a two-dimensional case. A curved wavefront is approximated by a set of straight lines. Two examples of such straight lines are AP and BP with their junction at point P. Draw two lines perpendicular to AP and BP with their common foot at P and take their bisector PQ with its length in proportion to the velocity at P. Point P on the previous wavefront is now replaced with a new point, Q. When all the points on the previous wavefront are replaced with new points, the envelope of these new points defines the next wavefront.

Figure 12(b) illustrates some detail of the procedure. If the adjacent new points are too widely separated from each other, fill this gap by adding an extra point between them, taking into account the curvature of the wavefront (W in Fig. 12(b)). In practice we take such an extra point whenever the gap exceeds 80% of the distance PQ . If a part of the wavefront has too large of a curvature, we add extra points to it to prevent excessive spacing of new points for the next wavefront (S in Fig. 12(b)). Let a new point and its nearest several new points be Q_0 and Q_n ($n=1, 2, \dots$), and their corresponding older points be P_0 and P_n , respectively. We can regard Q_0 for the first arrival if the interval P_nQ_0 is greater than P_nQ_n for any n . If not, discard Q_0 from the subsequent calculation (L in Fig. 12(b)). Repeat the above process until the whole space is swept by the wavefront. No particular care has to be taken into account in this scheme when a wavefront crosses a velocity discontinuity.

We have now obtained the first arrival times in the whole space, including those at the Earth's surface. The computation time, therefore, does not depend on how large the number of stations is; the advantage of the wavefront propagation method is to analyze the data of a large-scale, high-density array such as the J-Array. Vidale (1988) proposed a finite difference approach based on the concept of wavefront propagation, which is also robust against the complexity of structures. Sakai (1992) compared the computational accuracy of his method to that of Vidale (1988). For this purpose, he adopted the same two-dimensional velocity model as the one used by van Trier and Symes (1991). Figure 13(a) shows this model, in which velocity increases linearly with depth and the source is placed at the surface. Also illustrated are the grid spacing in the finite difference approach and the successive wavefront spacing in the wavefront propagation method. Note that the wavefront spacing is taken to be more than

Fig. 13. An accuracy test of the wavefront propagation method (Sakai, 1992). (a) Velocity model (van Trier and Symes, 1991) used in the accuracy test. Velocity increases linearly with depth. Source is located at the surface. Travel time is calculated at a depth of 150 km and its residual from the exact solution is regarded as the computational error. The real grid interval and wavefront interval are both one-tenth of those illustrated in this figure. Note that the wavefront interval is more than 4 times as large as the grid interval. (b) Comparison of the computational accuracy among the finite difference methods of Vidale (1988), broken line; van Trier and Symes (1991), solid line, and the wavefront propagation method (Sakai, 1992), dots.

four times that of the grid spacing.

Figure 13(b) plots the travel-time difference at a depth of 150 km between the numerical and analytical solutions as a function of horizontal distance. As can be seen, the accuracy of the method of Vidale (1990) depends on the direction of wave propaga-

focus

Fig. 14. Equi-travel time lines of the first-arrival phase from the 1991 Mindanao earthquake, that are contoured along the Honshu and Izu-Bonin trenches. The line interval is 4 s. The upper surface of the high-velocity slab is illustrated in Fig. 10. The slab is 100-km thick and 5% faster than the surrounding mantle.

tion. This point was improved by van Trier and Symes (1991), but in their method computational error accumulates as the propagation distance increases. In the wavefront propagation method, on the other hand, accuracy does not depend on the direction of wave propagation and the increase in computational error is minimal as distance increases. For the grid spacing and the wavefront propagation method employed in this test, the wavefront propagation method appears to be superior to the finite difference approach in terms of both accuracy and computational time.

Sakai (1992) further extended his method to the three-dimensional case, where the intersection of the wavefront with the Earth's surface at each step defines an equi-travel time line. Figure 14 shows an example of such calculation. The time step is taken to be 4 s, for which the precision of travel-time

calculation is about 0.4 s. A slab model is inserted into the IASP91 spherical Earth model. The upper surface of the model slab coincides with the deep seismic plane (Fig. 10) and the slab is 5% faster than the surrounding mantle. The thickness is taken to be 100 km for the Kurile-Honshu-Izu-Bonin-Mariana arc and 30 km for the Ryukyu arc. Distortion of the equi-travel time lines is observed clearly along the strike of the model slab. The spatial distribution of travel-time anomaly can be obtained by comparing such equi-travel time diagrams with and without a slab model.

5. Slab Model

Figures 15–17 show the calculated travel-time anomaly maps of the Mindanao, Kamchatka, and Ryukyu earthquakes, respectively. To obtain these maps we assumed a velocity contrast of 5% between the slab and the surrounding mantle. The velocity structure of the surrounding mantle is given by IASP91. The subducted portions of the Pacific plate and Philippine Sea plate are assumed to be 50- and 30-km thick, respectively. These thicknesses are assumed to continue to the Pacific and Philippine Sea plates left on the Earth's surface, yet this assumption has little effect on the results of the analysis. As can be seen from these maps, the slab effect spreads over an extensive area in a more complex way than one might imagine. The effect extends not only to the Japanese islands but also to the ocean bottom. The effect on the Japanese islands is particularly large, for the Mindanao event, in the Kanto region. It is large for the Kamchatka event in the southern Hokkaido to northern Tohoku regions. For the Ryukyu event, the effect is most outstanding in the Shikoku to Kinki regions but also appears in the Tohoku region.

Figures 6(b), 7(b), and 8(b) plot the calculated anomaly values at the J-Array stations in comparison with the observed values (Figs. 6(a), 7(a), and 8(a)). The agreements between the observations and calculations are good in view of the simplicity of the upper-mantle model. The positive anomaly that exceeds the computational error (≤ 0.4 s) does not appear on any of the synthetic maps, since only high-velocity slab was introduced into the spherical model. To examine how the travel-time anomaly is affected by the slab thickness and velocity contrast, we make model calculations using different values of slab thickness and velocity contrast. Figure 6(c) shows the case of a velocity contrast of 5% faster

and a 100-km thick slab of the Pacific plate for the Mindanao event. As compared to the case of the 50-km thick slab (Fig. 6(b)), the negative anomaly zone extends further north, beyond the Kanto region. Note that little change occurs in the maximum intensity when the slab thickness is increased from 50 to 100 km. Figure 6(d) shows another case at 10% faster and a 50-km thick slab, where the intensity of anomaly becomes much larger than in the case of a velocity increase of 5%. There is an additional effect of broadening the negative anomaly zone. Therefore, the effect of doubling the slab thickness is quite different from the effect of doubling the velocity contrast.

Figure 7(c) shows the case of velocity contrast at 5% faster and a 100-km thick slab of the Pacific plate for the Kamchatka event. The negative anomaly zone extends further south, beyond the Tohoku region, as compared to the case of the 50-km thick slab (Fig. 7(b)). Figure 7(d) shows another case with the velocity contrast 3% faster and a 50-km thick Pacific slab. The intensity of anomaly is small and the anomaly zone is narrow, relative to the case of the 5% faster slab. Figure 8(c) shows the case with the velocity contrast 5% faster and a 50-km thick slab of the Philippine Sea plate for the Ryukyu event. The negative anomaly extends further east, beyond the Shikoku to Kinki regions, as compared to the case of a 30-km thick slab (Fig. 8(b)). Figure 8(d) shows another case with the velocity contrast 10% faster and a 30-km thick Philippine Sea plate. The intensity of anomaly becomes much larger and the negative anomaly zone extends even more eastward than for the 50-km thick slab.

Figures 6–8 therefore demonstrate that an increase in slab thickness and an increase in velocity contrast produce different effects on the negative anomaly pattern, so they can be determined separately from the observations of travel-time anomaly. We have tested slab models with various combinations of velocity contrast (3, 5, or 10%) and thickness (30, 50, or 100 km) against the observed travel-time anomalies not only for the example earthquakes but for other earthquakes listed in Table 1. The best combination is obtained almost uniquely, 5% and 50 km for the subducted portion of the Pacific plate and, 5% and 30 km for the subducted portion of the Philippine Sea plate. Figures 15–17 (or 6(b), 7(b), and 8(b)) are the results for such best combinations. Note that the observations for the Mindanao event, related to the slab structure along the Izu–Honshu

Mindanao earthquake

Fig. 15. Contour map of travel-time anomaly for the 1991 Mindanao earthquake. The 50-km thick slab of the Pacific plate is 5% faster than surrounding mantle (IASP91) while the 30-km thick slab of the Philippine Sea plate is 5% faster.

Kamchatka earthquake

Fig. 16. Contour map of travel-time anomaly for the 1992 Kamchatka earthquake. See Fig. 15 for other explanations.

arc, and those for the Kamchatka event, related to the slab structure along the Kurile–Honshu arc, are explained by a single slab model of the Pacific plate.

The above slab model has been obtained under the following assumptions.

Ryukyu earthquake

Fig. 17. Contour map of travel-time anomaly for the 1992 Ryukyu earthquake. See Fig. 15 for other explanations.

1) The slab is one-layered, having a uniform velocity contrast with the surrounding mantle.

2) The surrounding mantle is laterally homogeneous, coinciding with the IASP91 model. Admittedly, the model is too simple to explain all of the important observations. For example, the model is unable to explain the positive anomalies in the Hokkaido region for the Mindanao event (Fig. 6(a)), positive anomalies on the Pacific side of southwest Japan for the Kamchatka event (Fig. 7(a)) and positive anomalies in the central part of Honshu for the Ryukyu event (Fig. 8(a)). These positive anomalies imply the presence of low-velocity zones relative to the IASP91 model, both above and below the subducting slab of the Pacific plate. The positive anomalies in the Hokkaido region (Fig. 6(a)) require a low-velocity zone below the slab, while those in the central part of Honshu (Fig. 8(a)) require a low-velocity zone above the slab for the IASP91 spherical Earth model.

In order to examine to what extent our conclusion depends on the reference model, we calculate theoretical travel times for the 1991 Mindanao event based on the spherically-averaged version of the laterally heterogeneous model of Fukao *et al.* (1992), with the relocation using the same model after correction for lateral heterogeneity (see Table 2). The calculated travel-time curve is shown by the broken line in Fig. 3, which does not coincide with the IASP91 curve (solid line). The model of Fukao

Table 3. Observed travel-time anomalies and slab models obtained.

	Kurile	Izu	Ryukyu	Honshu
Residual (s)	10	4	8	1–2
Slab thickness (km)	50	50	30	80–90
Velocity contrast (%)	5	5	5	3–6

* After Zhao and Hasegawa (1993).

et al. (1992) does not show positive anomalies in Hokkaido. This result suggests that if we use the model of Fukao *et al.* (1992) as a reference, a low-velocity zone below the slab is unnecessary and that the slab may be faster and/or thicker than the slab model obtained in this study. We emphasize that the conclusions obtained in this paper are valid only by referring to the IASP91 model.

5. Summary

We have detected, for the first time, a first-arrival phase near-critically incident to the base of the subducting slab, which shows a large travel-time anomaly of -4 s. This first-arrival phase is not particularly rich in high-frequency component. Even larger travel-time anomalies of -8 to -10 s have been observed for the first arrivals from earthquakes in Kamchatka and Ryukyu, respectively. The latter first-arrival phases, rich in high-frequency component, are those preferentially propagating through the high-velocity slab along the strike of the deep-sea trench, as first reported by Ansell and Gubbins (1986) on the first-arrival records in New Zealand for earthquakes occurring in Tonga.

These first arrivals were analyzed for slab structure using the wave propagation method of Sakai (1992). The slab is characterized by its thickness and velocity contrast. The surrounding mantle is represented by IASP91. The tentative results of our analysis are summarized in Table 3. A velocity contrast of 5% appears to be reasonable for all the slabs investigated. The slab thickness of 50 km of the Pacific plate is apparently somewhat smaller than the slab thickness of 80–90 km obtained from tomographic inversions of the regional data by the group of Tohoku University (e.g., Zhao and Hasegawa, 1993). The value of slab thickness obtained in this study, however, may depend on the reference model, as discussed in the previous section. A more quantitative discussion must await a

simultaneous analysis of the positive and negative anomalies of the observed travel times.

An advantage of the wavefront propagation method is ease of obtaining travel-time anomaly maps as shown in Figs. 15–17. Since the computation is stable and computational speed does not strongly depend on the complexity of three-dimensional structures, we can generate maps like Figs. 15–17 for a large number of trial models. Such maps will be useful not only to analyze existing data but to search for critical places of observation to obtain new data. For example, Figs. 15–17 may be used to find an optimum location for deploying an array of ocean-bottom seismometers in a project of constraining the thickness of the Pacific plate oceanward of the trench.

We are grateful to the members of the J-Array Group for their efforts of maintaining the J-Array. We thank K. Hirahara of Kyoto University and T. Matsuzawa of Tohoku University for sending us their network data. We also thank F. Niu of the University of Tokyo for useful comments. This study was supported by a Grand-in-Aid for Scientific Research of the Ministry of Education, Science, Sports and Culture of Japan, the special project "Earth's Central Core."

REFERENCES

- Ansell, J. H. and D. Gubbins, Anomalous high-frequency wave propagation from the Tonga-Kermadec seismic zone to New Zealand, *Geophys. J. R. Astron. Soc.*, **85**, 93–106, 1986.
- Červený, V. and R. Ravindra, *Theory of Seismic Head Waves*, Univ. Toronto Press, Toronto, 1971.
- Červený, V., I. A. Molotov, and I. Pšenčík, *Ray Method in Seismology*, Univerzita Karlova, Praha, 214 pp., 1977.
- Fukao, Y., S. Hori, and M. Ukawa, A seismological constraint on the depth of basalt-eclogite transition in a subducting oceanic crust, *Nature*, **303**, 413–415, 1983.
- Fukao, Y., K. Kanjo, and I. Nakamura, Deep seismic zone as an upper-mantle reflector of body waves, *Nature*, **272**, 606–608, 1978.
- Fukao, Y., M. Obayashi, H. Inoue, and M. Nenbai, Subducting slabs stagnant in the mantle transition zone, *J. Geophys. Res.*, **97**, 4809–4822, 1992.
- Gubbins, D. and R. Snieder, Dispersion of P waves in subducted lithosphere: evidence for an eclogite layer, *J. Geophys. Res.*, **96**, 6321–6333, 1991.
- Hori, S., H. Inoue, Y. Fukao, and M. Ukawa, Seismic detection of the untransformed 'basaltic' oceanic crust subducting into the mantle, *Geophys. J. R. Astron. Soc.*, **83**, 169–197, 1985.
- Iidaka, T., M. Mizoue, I. Nakamura, T. Tsukuda, K. Sakai, M. Kobayashi, T. Haneda, and S. Hashimoto, The upper boundary of the Philippine Sea plate beneath the western Kanto region estimated from S–P-converted wave, *Tectonophysics*, **179**, 321–326, 1990.
- Jacob, K. H., Three-dimensional seismic ray tracing in a laterally heterogeneous spherical Earth, *J. Geophys. Res.*, **75**, 6675–6689, 1970.
- J-Array Group, The J-Array Program: System and present status, *J. Geomag. Geoelectr.*, **45**, 1265–1274, 1993.
- Julian, B. R. and D. Gubbins, Three-dimensional seismic ray tracing, *Geophys. J. R. Astron. Soc.*, **43**, 95–113, 1977.
- Kamiya, S., T. Miyatake, and K. Hirahara, How deep can we see the high velocity anomalies beneath the Japan Islands?, *Geophys. Res. Lett.*, **15**, 828–831, 1988.
- Kamiya, S., T. Miyatake, and K. Hirahara, Three-dimensional P-wave velocity structure beneath the Japanese Islands, *Bull. Earthq. Res. Inst., Univ. Tokyo*, **64**, 457–485, 1989.
- Kennett, B. L. N. and E. R. Engdahl, Travel times for global earthquake location and phase identification, *Geophys. J. Int.*, **105**, 429–465, 1991.
- Kim, W. and V. F. Cormier, Vicinity ray tracing: An alternative to dynamic ray tracing, *Geophys. J. Int.*, **103**, 639–655, 1990.
- Koketsu, K., Three-dimensional ray tracing in a subduction zone, *Zisin*, **44**, 165–176, 1991.
- Matsuzawa, T., Seismic-wave velocity structure of the subducting plate boundary estimated from later phases, *Zisin*, **42**, 525–536, 1989.
- Matsuzawa, T., T. Kono, A. Hasegawa, and A. Takagi, Subducting plate boundary beneath the northeastern Japan arc estimated from SP converted waves, *Tectonophysics*, **181**, 123–133, 1989.
- Matsuzawa, T., N. Umino, A. Hasegawa, and A. Takagi, Upper mantle velocity structure estimated from PS-converted wave beneath the north-eastern Japan Arc, *Geophys. J. R. Astron. Soc.*, **86**, 767–787, 1986.
- Nakanishi, I. and A. Horie, Anomalous distributions of seismic intensities due to the descending Philippine Sea plate beneath the southern Kanto district, Japan, *J. Phys. Earth*, **28**, 333–360, 1980.
- Nakanishi, I., K. Suyehiro, and T. Yokota, Regional variations of amplitudes of ScSp phases observed in the Japanese Islands, *Geophys. J. R. Astron. Soc.*, **67**, 615–634, 1981.
- Oda, H., T. Tanaka, and K. Seya, Subducting oceanic crust on the Philippine Sea plate in Southwest Japan, *Tectonophysics*, **172**, 175–189, 1990.
- Okada, H., Fine structure of upper mantle beneath Japanese island arcs as revealed from body wave

- analyses, DSc thesis, Hokkaido Univ., 129 pp., 1977.
- Sakai, S., Travel time analysis for two-dimensionally and three-dimensionally heterogeneous structures: Wave-front propagation method, Ph. D. Thesis, University of Tokyo, Japan, 1992.
- Smith, G., D. Gubbins, and W. Mao, East P wave propagation in subducted Pacific lithosphere: Refraction from the plate, *J. Geophys. Res.*, **99**, 23787–23800, 1994.
- Spakman, W., S. Stein, R. D. van der Hilst, and R. Wortel, Resolution experiments for NW Pacific subduction zone tomography, *Geophys. Res. Lett.*, **16**, 1097–1100, 1989.
- Suetsugu, D., Lower mantle high velocity zone beneath the Kurile as inferred from P wave travel time and amplitude data, *J. Phys. Earth*, **37**, 265–295, 1989.
- van der Hilst, R. D. and W. Spakman, A tomographic image of the lesser Anilles subduction zone, *Eos. Trans. Am. Geophys. Union*, **68**, 1376, 1987 (abstract).
- van der Hilst, R. D. and W. Spakman, Importance of the reference model in linearized tomography and image of subduction below the Caribbean plate, *Geophys. Res. Lett.*, **16**, 1093–1096, 1989.
- van der Hilst, R. D., E. R. Engdahl, and W. Spakman, Tomographic inversion of P and pP data for aspherical mantle structure below the northwest Pacific region, *Geophys. J. Int.*, **115**, 254–302, 1993.
- van der Hilst, R. D., E. R. Engdahl, W. Spakman, and G. Nolet, Tomographic imaging of subducted lithosphere below northwest Pacific island arcs, *Nature*, **353**, 37–43, 1991.
- van Trier, J. and W. W. Symes, Upwind finite-difference calculation of travel times, *Geophysics*, **56**, 812–821, 1991.
- Vidale, J. E., Finite-difference calculation of travel times, *Bull. Seismol. Soc. Am.*, **78**, 2062–2076, 1988.
- Vidale, J. E., Finite-difference calculation in three dimensions, *Geophysics*, **55**, 521–526, 1990.
- Wesson, R. L., Travel-time inversion for laterally inhomogeneous crustal velocity models, *Bull. Seismol. Soc. Am.*, **61**, 729–746, 1971.
- Zhao, D. and A. Hasegawa, P wave tomographic imaging of the crust and upper mantle beneath the Japan Island, *J. Geophys. Res.*, **98**, 4333–4353, 1993.
- Zhao, D., A. Hasegawa, and S. Horiuchi, Tomographic imaging of P and S wave velocity structure beneath northeastern Japan, *J. Geophys. Res.*, **97**, 19909–19928, 1992.
- Zhao, D., S. Horiuchi, and A. Hasegawa, 3-D seismic velocity structure of the crust and the uppermost mantle in the northeastern Japan Arc, *Tectonophysics*, **181**, 135–149, 1990.
- Zhou, H. W., Mapping of P wave slab anomalies beneath the Tonga, Kermadec and New Hebrides arcs, *Phys. Earth Planet. Inter.*, **61**, 199–229, 1990.
- Zhou, H. W. and R. W. Clayton, P and S wave travel time inversions for subducting slab under the island arcs of the northwest Pacific, *J. Geophys. Res.*, **95**, 6829–6851, 1990.