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P-wave Velocity Structure of Lithosphere-asthenosphere beneath the Western Northwest Pacific Basin Determined by an Ocean-bottom Seismometer Array Observation

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Abstract

A P-wave velocity-depth structure was studied in the western Northwest Pacific Basin by making a composite record section using six earthquakes which covered the epicentral distances from 6° to 18°. Assuming an uppermost mantle layer of 8.0 km/sec at a depth of 13 km, depths of the velocity boundaries of 8.3 km/sec and 8.4 km/sec were determined at about 61 km and 132 km respectively. The depth of the low-velocity layer which forms the shadow zone of P-wave refraction beyond 15° was estimated at about 149 km. This low-velocity layer could be regarded as the major transition from the lithosphere to the asthenosphere. The thickness of the lithosphere in the western Northwest Pacific Basin is thicker than formerly reported.

1. Introduction

This paper reports that the thickness of the lithosphere beneath the western Northwest Pacific Basin is thicker than formerly reported. In the previous papers (NAGUMO *et al.*, 1986a and b) we reported that the apparent velocities of the P-waves refracted from the lower-lithosphere are 8.4–8.6 km/sec and that the P-waves go into a shadow zone beyond the epicentral distance of 15°. The latter implies that the lithosphere is underlain by a low-velocity layer which could be the most significant transition from the lithosphere to the asthenosphere. In order to estimate the depth of this transition, that is, the thickness of the lithosphere, we need detailed travel-time data which cover the whole epicentral distances from the source region to the shadow zone. We made a composite record section covering the epicentral distance from 6° to 18° and estimated a velocity-depth structure by a

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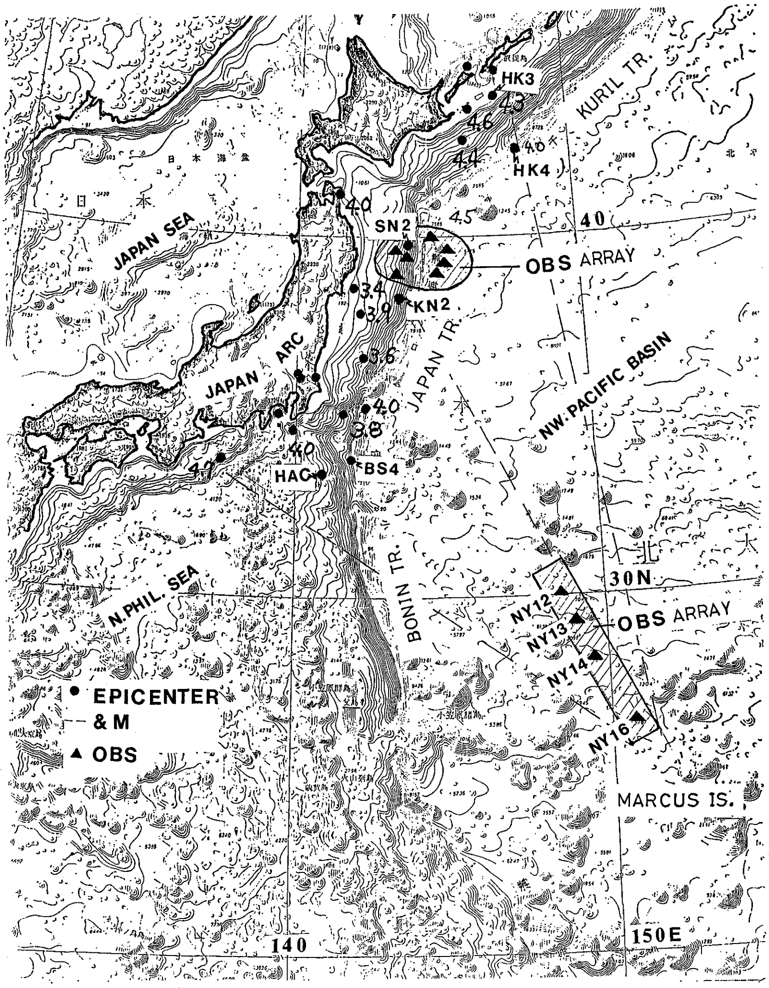

Fig. 1. Locations of OBS (Ocean-Bottom Seismometer) array (solid triangles) and epicenters of natural earthquakes (solid circles). The one array deployed far off Northeast Japan is for increasing the accuracy of the earthquake source data, and the other array deployed southeasterly towards Marcus Island is for covering long range observations. The record sections are made for 6 earthquakes HK3, HK4, SN2, KN2, BS4 and HAC.

Table 1. Data of the Ocean-Bottom Seismometer (OBS) stations.

Station number	Position		Water depth	Deployment		Recovery	
			m		H M		H M
NY12	30° 05.03′N	148° 27.27′E	6160	1981 JUN 11	20:05	JUL 11	03:45
NY13	29 09.64	149 03.05	6040	JUN 12	06:12	JUL 14	02:35
NY14	28 17.42	149 34.61	6100	JUN 12	14:23	JUL 12	02:54
NY16	26 26.07	150 24.10	5760	JUN 13	08:19	JUL 12	18:35

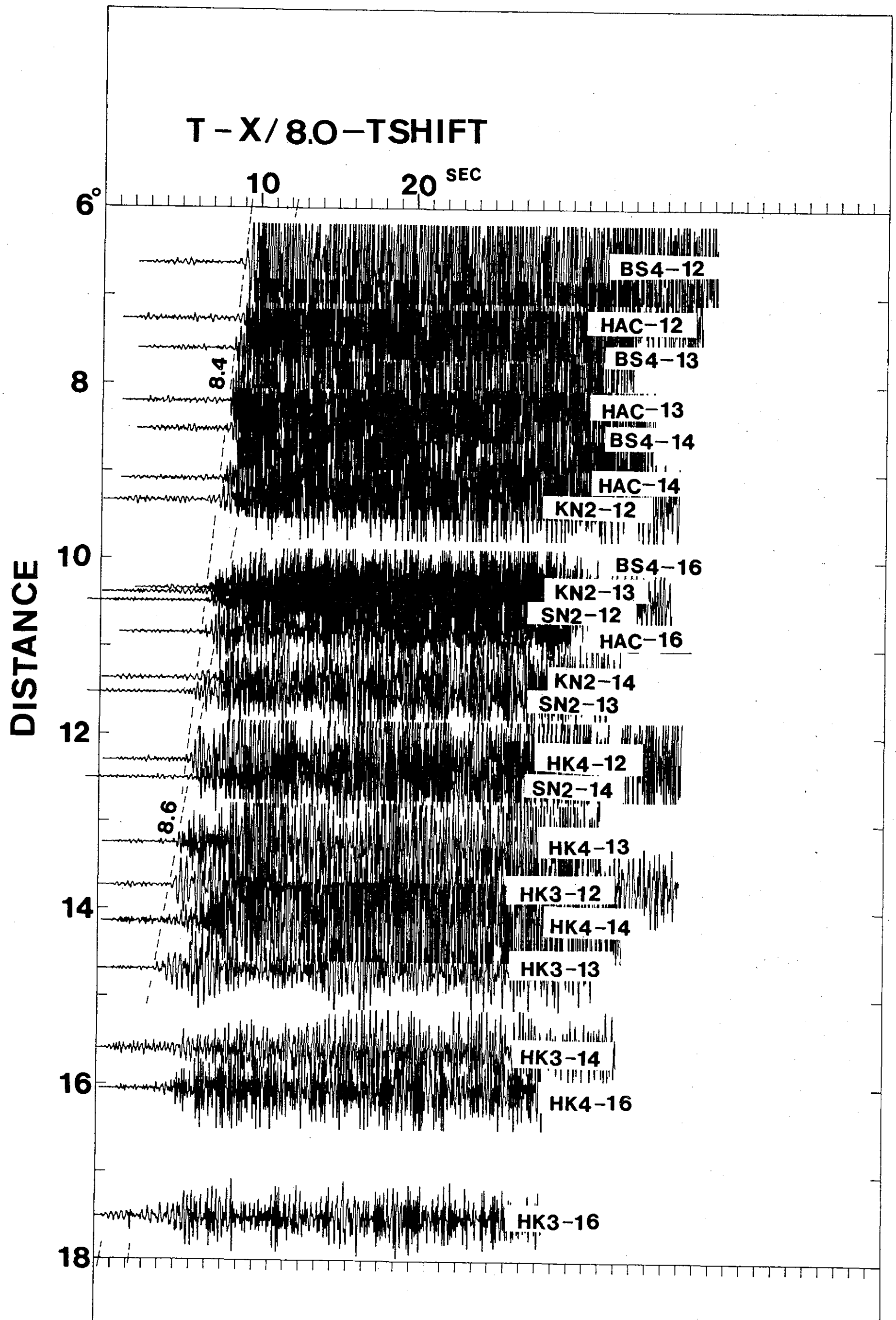

Fig. 2. Composite record section for the earthquakes HK3, HK4, SN2, KN2, BS4 and HAC. The travel times are reduced by a velocity 8.0 km/sec. The unit for the epicentral distance is degree. TSHIFT: an operational constant for the delay starting time. TSHIFT=−7.86 sec on this section. Each trace is labeled at the right margin by the earthquake name-station number. For example, BS4-12 indicates the record of the earthquake BS4 (see Table 2) at the station NY12 (see Table 1).

method of ray-tracing.

2. Data and results

A composite record section was made covering the range from 6° to 18° (Fig. 2) using six earthquakes whose source data are shown in Fig. 1 and Table 1. These earthquakes were selected from among the many earthquakes recorded during the observation period so as to exclude too much overlapping of records. The travel-times in Fig. 2 are reduced by a velocity of 8.0 km/sec. In order to adjust the errors of the origin times, we used the base line fitting and calibrated them by the array deployed near the source region as described in the previous paper (NAGUMO *et al.*, 1986b). In Fig. 2, the inter-lacing of the record sections are indicated by the notation of the earthquake name and the station number. The station data are shown in Table 2. We did not make the distance correction for the focal depth, because a part of this is automatically included in the base line fitting as a time correction and also because the accuracy of the focal depth which is reported by the seismological bulletines in this region is not always sufficient. In order to get the intercept time, we need the exact origin time. Fortunately, an earthquake of M 4.5 (9 July 1981, 10 h 28 m; SN2 in Fig. 1) occurred within the ocean-bottom seismometer (OBS) array which was deployed far-off NE Japan (KASAHARA *et al.*, 1982). The array could determine with high accuracy the location of the hypocenter and the origin time (NAGUMO *et al.*, 1984). The earthquake is one of those which occurred along the interface between the subducting oceanic plate and overlying landward mass. The origin time determined by the ocean-bottom seismometer array differs from those reported by the Japan Meteorological Agency (JMA) and the International Seismological Center (ISC) by 2.66, and 3.86 seconds respectively (Table 3). The data of the hypocenter determination by the ocean-bottom seismometer array are presented in the Appendix. We

Table 2. Data of the earthquakes (after the Seismological Bulletin of the Japan Meteorological Agency).

Earthq. name	Epicenter		Focal depth	Origin time					Magnitude
			km	1981		H	M	S	M
HK3	43° 49′N	147° 25′E	00	JUN	29	14	50	26.2	4.3
HK4	42 24	147 58	10	JUL	05	19	22	48.9	4.0
SN2	39 53	143 50	00	JUL	09	10	28	45.2	4.5
KN2	38 32	143 36	00	JUL	01	13	51	18.2	4.8
BS4	33 47	141 58	30	JUN	29	14	34	11.5	4.5
HAC	33 34	140 57	70	JUL	01	04	58	21.2	

Table 3. A comparison of the origin time of the earthquake SN2.

	Origin time (JST)			Epicenter		Focal depth	Magnitude M
				Latitude	Longitude		
RIHYPO8*1	1981 JUL 09	$10^{H}\ 28^{M}$	47.86^{s}	N39°53.67′	E143°40.18′	11.7^{km}	5.4
JMA*2			45.2	39.53°	143.50°	0.	4.5
ISC*3			44.	39.90°	143.69°	13.	4.5
PEK*3			47.0	39. 9°	143. 5°	—	Mb=4.5
NEIS*3			47.3	39. 8°	143. 6°	33.	Mb=4.6
MOS*3			50.2	40.27°	143.46°	33.	Mb=4.8

*1 from NAGUMO *et al.* (1984).
*2 from the Seismological Bulletin of the Japan Meteorlogical Agency, July, 1981.
*3 from Bulletin of the International Seismological Centre, July, 1981.

apopted the origin time determined by the ocean-bottom seismometer array as the most reliable for estimating the intercept time of the refraction travel-times. The origin times reported by JMA and ISC give large intercept times and require an unreal thick crust so as to satisfy the cross-over distance of the refractions from the Moho and the lower-lithosphere.

The first arrivals of P-waves are of sufficient amplitude in the distance range from 6° to 10°, where the apparent (surface) velocity is about 8.4 km/sec. The apparent velocity increaces a little in the range from 12° to 15°, becoming about 8.6 km/sec (NAGUMO *et al.*, 1986a). Beyond 15°, the amplitudes of the first arrivals decrease and present a shadow zone (NAGUMO *et al.*, 1986b). There is a slight decrease of the amplitude in the range from 10° to 11°, which may indicate the presence of a weak zone. However, we do not see any significant off-set of travel-times. Therefore, the first arrivals which are seen in the distance range from 6° to 10° seem to continue to 12° with the eye-fitting apparent velocity 8.35 km/sec. The change of the apparent velocity from 8.35 km/sec ($6°<X<10°$) to 8.6 km/sec ($12°<X<15°$) could be regarded as either a gradual change or a sharp bend of the travel-time curves. The large amplitudes seen in the later phases of P-arrivals may include cusp generated by a sharp velocity boundary. However their identification is obscure, because they are contaminated by the arrivals of High-Frequency Pn-phases (OUCHI *et al.*, 1983). In the shadow zone, we can see an off-set of travel-times whose amount is about 1.7–2.0 seconds. However, because of the emergent nature of their first arrivals in the shadow zone, the determination of their apparent velocity is uncertain.

The travel-time curves which can be read on the Fig. 2 are

$$T=X/8.35+4.96$$
$$T=X/8.58+9.49\ , \qquad (1)$$

where T: travel-time (sec), X: epicentral distance (km), the constants: intercept times. These travel-time curves are the expression for the source of a focal depth 10 km, and for the receiver at a depth 6.0 km (sea-floor) respectively. The apparent surface velocity of a refracted wave is the horizontal velocity along the sea-floor. The intercept times shown above are important data which constrain the depth of the velocity boundary.

3. Estimation of the velocity-depth structure

As a starting model for further calculation, we preliminarily computed the depths of the 8.35 km/sec (apparent) and 8.58 km/sec (apparent) layers by assuming a flat-layer constant velocity model from the intercept times of the formula (1). Since the crustal structures are different for the source and receiver regions, we assumed a model

Tabble 4. A starting model (a constant-velocity, flat-layered model) for the P-wave velocity structure beneath the western Northwest Pacific Basin derived from the travel-time formula (1). The different crustal structures are assumed in the source region and receiver region.

Source region			Receiver region			
Depth from the surface	Thickness	P-velocity	Depth from the surface	Thickness	P-velocity	P-velocity Sphericity corrected
km	km	km/s	km	km	km/s	km/s
(10) *1		(6.5)	(6.0) *3		(4.5) *4	(4.5)
	(10)			(6.0)		
(20)		(6.5)	(12.0)		(8.0)	(8.0)
(20) *2		(8.0)				
	36.1			44.1		
56.1		(8.0)	56.1			(8.0)
56.1		8.35**	56.1		8.35**	8.29
	65.7			65.7		
121.8		8.35**	121.8		8.35**	
121.8		8.58**	121.8		8.58**	8.43

*1 The focal depth of 10 km from the earth's surface is assumed to correspond to the earthquake SN2 (see Appendix). This depth also corresponds to the interface between the subducting oceanic plate and overlying land-mass.

*2 The depths of the Moho are assumed to be 20 km from the earth's surface at the source region and 12 km from the earth's surface at the receiving region.

*3 the water depth in the receiving region is assumed to be 6 km.

*4 The ocean crust is assumed to be single layer which is equivalent for computing travel-times through the layer 3, layer 2 and the sediment layer.

** Observational data

( ) assumed

Table 5. (Model A) A model for the P-wave velocity structure beneath the western Northwest Pacific Basin (spherical earth model). The source side: The focal depth is 10 km from the surface beneath a trench slope. Ocean side: The average depth of the receiving stations (ocean floor) is six kilometers from the surface.
(Model B) A reference model for estimating the depth of the low-velocity layer which forms the shadow zone.
( ) Assumed crustal structure.
(( )) non uniqueness.

Model A						Model B		
Source side			OBS side			OBS side		
Depth from the surface km	V_p km/s	Gradient 1/s	Depth from the surface km	V_p km/s	Gradient 1/s	Depth from the surface km	V_p km/s	Gradient 1/s
			(6.0)	(1.50)				
			(6.0)	(2.00)				
					(0.01)			
			(6.7)	(2.07)		the same as right		
			(6.7)	(5.00)				
					(0.50)			
			(8.2)	(5.75)				
			(8.2)	(6.70)				
(10.0)	(6.50)				(0.01)			
		(0.02)	(13.1)	(6.75)				
			(13.1)	(8.00)				
					0.0000			
(20.0)	(6.70)							
(20.0)	(8.00)							
		0.00	the same as right					
61.1	8.00							
61.1	8.25							
		0.0005						
131.8	8.29							
131.8	8.39					131.8	8.39	
		0.0005						0.0005
149.1	8.40		the same as right					
149.1	((8.24))							
		0.0005				(Without low-velocity layer)		
((183.1))	8.26							
((183.1))	8.41							
		0.0005						

for the crust as shown in Table 4. In this model the source is located at a depth of 10 km from the earth's surface corresponding to the focal depth of the earthquake SN2. This depth also corresponds to the interface between the subducting oceanic plate and the overlying land-mass. Since we do not have data for the refraction from the Moho discontinuity on the record section of Fig. 2, we assumed the

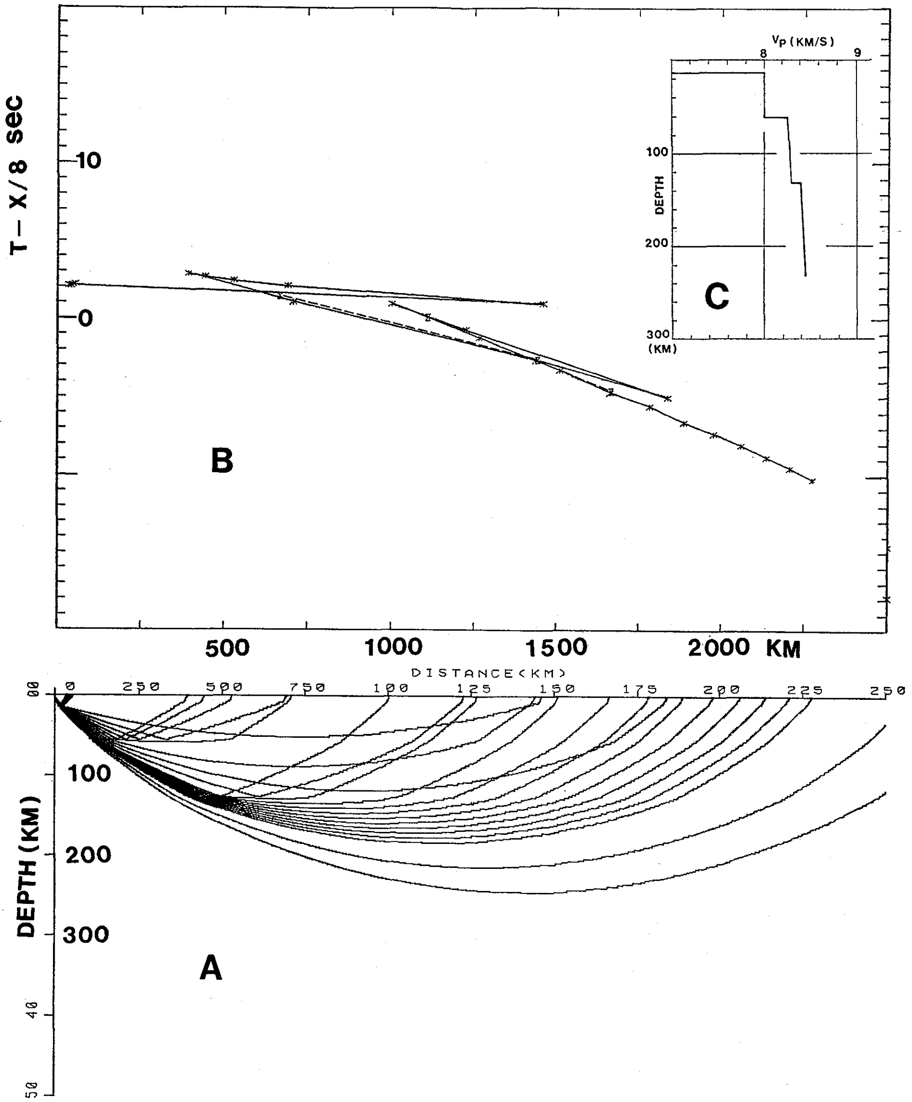

Fig. 3. Ray-path diagram (A), travel-times (B) and the model (C), Table 5-Model B, which has no low-velocity layer.

velocity and the depth of the Moho beneath the source region as 8.0 km/sec and 20 km from the surface respectively. As for the receiving stations, we adopted an average oceanic crust model based on the data in this region (LUDWIG *et al.*, 1966) with a water depth 6.0 km. Then, from the intercept times the depths of 8.35 km/sec and 8.58 km/sec layers were obtained as 56.1 km and 122 km from the earth's surface respectively. In this calculation the depths of the source and the station were handled separately.

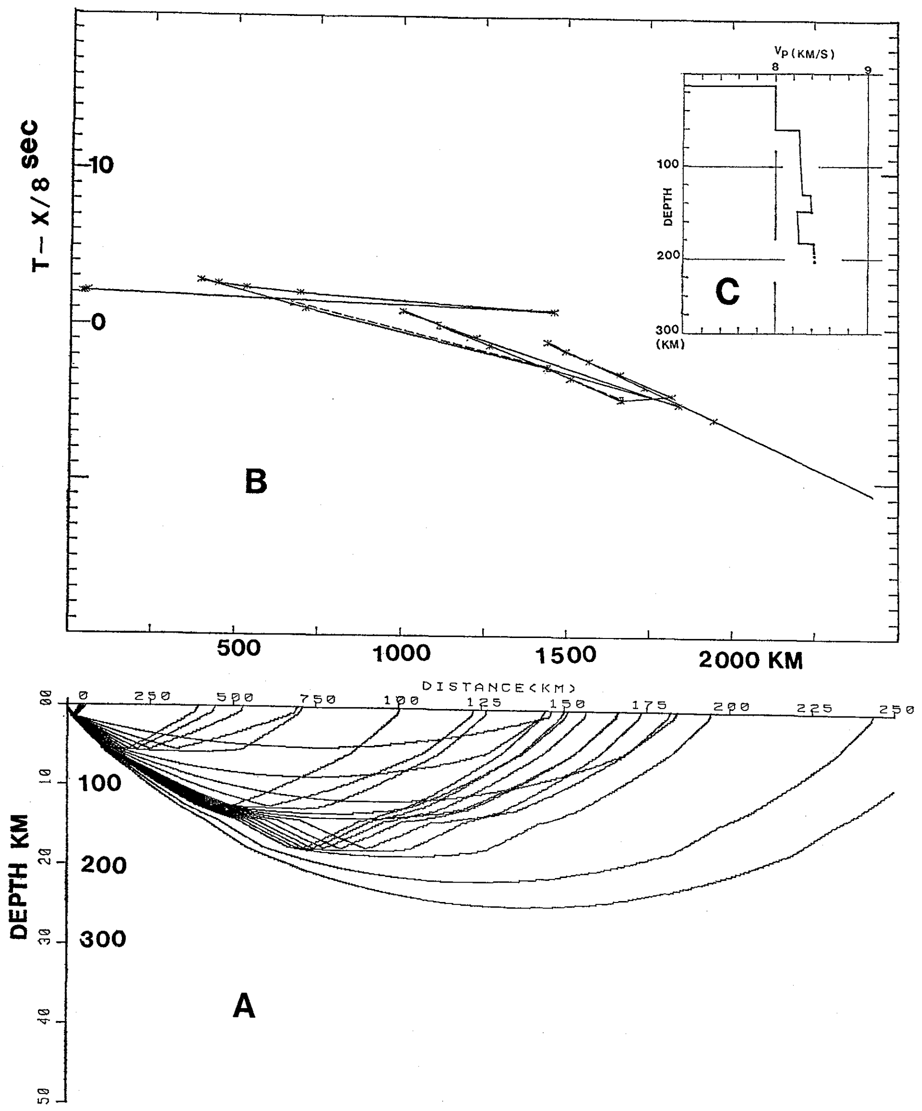

Fig. 4. Ray-path diagram (A), travel-times (B), and the model (C), Table 5-Model A in which a low-velocity layer underlies the lower-lithosphere.

Then the curvature effect of the earth's surface was taken into consideration by the formula $v_{app}/v_t = r_0/r_t$ (MEREU, 1967), where r_0: radius of the earth; r_t: distance from the center of the earth to the turning point; v_{app}: apparent velocity along the sea-floor; v_t: true velocity at the turning point. The true velocities which correspond to the apparent velocities of 8.35 km/sec and 8.58 km/sec were obtained as 8.29 km/sec at the depth 61 km and 8.43 km/sec at the depth 132 km respectively. The results are shown in Table 4.

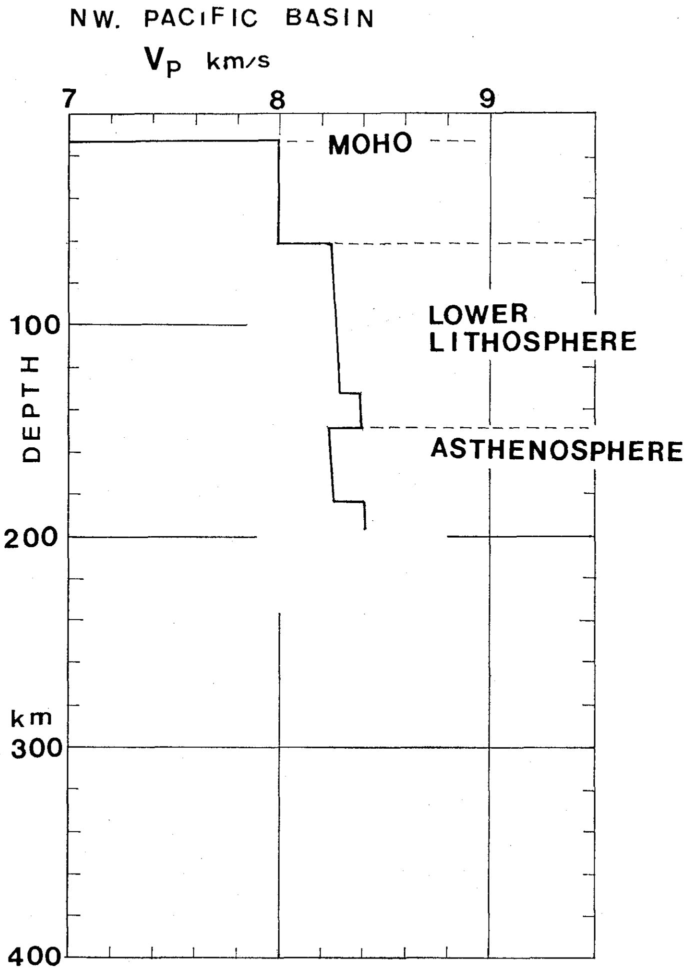

Fig. 5. P-wave velocity structure of the lithosphere-asthenosphere beneath the western Northwest Pacific Basin (spherical earth model).

The layers of the true velocities about 8.3 and 8.4 km/sec at depths about 61 km and 132 km from the earth's surface form the "lower-lithosphere" in the sense used by GREEN and HALES (1968). The lower-lithosphere is a zone of sub-Moho high-velocity layers which underlies the 8.0 km/sec velocity layer of the Moho-discontinuity. It is interesting to note that the "lower-lithosphere" presents a feature of double zoning which might have some important petrological significance for the genesis of the lower-lithosphere.

In order to estimate the depth of the low-velocity layer which forms the shadow zone beyond 15°, a ray-path diagram was prepared for the spherically layered structure using a program developed by

ODEGARD (1975). We handled the crustal structures separately for the source and the receiver region. We put the source at a depth of 10 km from the surface as determined by the OBS array and the receiver at a depth of 6.0 km (the average depth of the OBS upon the ocean floor). First, we computed the ray-paths and travel-times for the velocity model which has no low-velocity layer below the lower-lithosphere. This is shown in Table 5-Model B and the results in Fig. 3B. The sea-floor depth of 6 km from the surface was taken as the reference level of the presentation. A trial-and-error fitting was attempted for the refraction arrivals from the lower-lithosphere. The fitting is shown on Fig. 3A where the broken lines connecting the mark ⧖ are observations and the solid lines are calculations. Both the velocity gradient and the velocity itself were adjusted for finding the best fitting. The result shown on Table 5-Model A is one of the best fittings so far attempted. The gradient of the velocity is sensitive for the fitting. Getting a nice fitting of travel-times to the lower-lithosphere, we can see that the depths of the turning points of the rays which appear beyond the epicentral distance 15° are deeper than about 140 km from the sea-floor. This shows that the low-velocity layer begins to appear below this depth. Next, in order to obtain the thickness of the low-velocity layer below the "lower-lithosphere", we tried ray-tracing for several models for finding the best fit. The calculation shows that the fitting to the off-set of travel times beyond 15° also related to the location of the cusp of refraction/reflection from the base of low-velocity layer, that is, the fitting of travel-times also relates to the large amplitude of later phases in the distance range 12°–15°. Although there is no uniqueness about the value of the velocity of the low-velocity layer, it was found that the thickness of the low-velocity layer appears to be rather thin. An example of the best fit is shown in Fig. 4B and the structure is shown in Fig. 5 and Table 5-Model A. As regards the low-velocity layers, because of the insufficient data of the observation, the solution is not unique with respect to its velocity, its gradient and its thickness. However, the travel-time off-set of about 1.7–2.0 sec constrains the thickness of the low-velocity layer. Since the velocity step at the layer boundary generally causes large energy of reflections, the velocity steps seen in Fig. 5 should be modified so as to match with the amplitude pattern of the reflection/refraction from the lithosphere. Some of the velocity steps may be reduced to a smooth transition. Such features will be further studied by using synthetic seismograms.

4. Discussion

Major discontinuity

The P-wave shadow zone which appears beyond 15° seems to be indicative of a low-velocity layer which forms a transition from the lithosphere to the asthenosphere. In the seismic refraction studies, the velocity depth distribution is profiled by the distance section of travel-times. The deeper the depth of the objective boundary, the longer the distance of the observation-line. Therefore, when the distance increases, the lateral hterogeneity is likely to contaminate the depth profile. However, the composite record section involves in itself a process of averaging and can smooth out a small amount of local heterogeneity. In this sense, the record section presented in this study could be regarded as an average profile for a broad region shown in Fig. 1 in the western Northwest Pacific Basin. The P-wave first arrivals in the distance ranges from 6° to 15° cover a certain depth range of the lithosphere. Therefore, it will be quite certain that the profile which extends up to 18° could cover the depth range from the lithosphere to the asthenosphere. Even though there is a weak zone in the range between 10° and 12° as stated above, because of no visible off-set of travel-times, the shadow zone beyond 15° is the most significant discontinuity in the whole range from 6° to 18°. Thus, we will be able to conclude that the low-velocity layer indicated by the 15° shadow zone is the major transition from the lithosphere to the asthenosphere.

Thickness of the lithosphere

The thickness of the lithosphere defined above is about 143 km and is greater than what had been formerly reported in this region (YOSHII, 1975; ASADA and SHIMAMURA, 1976; FORTHYS, 1977). Though the depth of the shadow zone is dependent on the velocity gradient in the overlying layers, the distance range 15° is definite evidence that shows a thick lithosphere beneath the western Northwest Pacific Basin. As discussed in the previous paper (NAGUMO *et al.*, 1986b), the distance range of the P-wave shadow zone is the most direct data of the regional difference of the lithosphere thickness.

Comparison of the P-wave velocity structures

In Fig. 6 is shown a comparison of structures between the western Northwest Pacific Basin, the continental shield region and the continental tectonically active region. As discussed in the previous papers (NAGUMO *et al.*, 1986a, b and NAGUMO and KASAHARA, 1986), there are many varieties in the P-wave structures of the lithosphere-asthenosphere even within the similar geological provinces such as the continental shield regions and the continental tectonically active regions. Even though they are neither representative nor typical of these regions, we take the case of the Central U. S. (GREEN and HALES, 1968) and the western U. S. (MASSÉ *et al.*, 1972) as examples of the P-wave structure of the lithosphere-asthenosphere beneath these regions. These two cases were selected because we think that these structural features may show almost the mature and juvenile states of the evolution of the upper mantle (NAGUMO and KASAHARA, 1986).

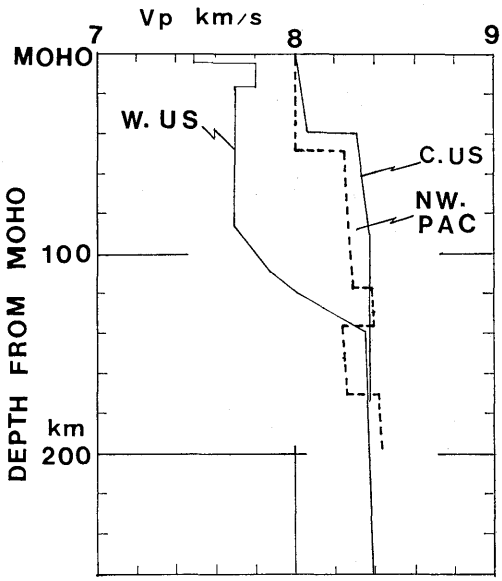

Fig. 6. A comparison of the P-wave velocity structures of the lithosphere-asthenosphere in the continent and the ocean. C.US: Central U. S. (GREEN and HALES, 1968); W.US: Western U. S. (MASSÉ *et al.*, 1972); NW.PAC: Northwest Pacific Basin (this study)

The structure of the lithosphere in the western Northwest Pacific Basin obtained in the preceding section resembles that of the Central U.S. with respect to the layered structure, but it is different in the low-velocity layer. In the Central North America (the Early Rise Model-1, GREEN and HALES, 1968), the lithosphere has a two-layered structure, namely, the uppermost mantle layer (about 40 km thick, P-velocity 8.0 km/sec) is underlain by a high-velocity layer of 8.4 km/sec at depth about 90 km from the surface, which is referred to as "lower-lithosphere". The lower-lithosphere extends to the depth of about 400 km without an intervening low-velocity layer. On the other hand, in the western Northwest Pacific Basin, the lithosphere is of a three-layered structure, namely, the uppermost mantle layer (about 48 km thick, P-velocity 8.0 km/sec) is underlain by the "lower-lithosphere" which is composed of a layer of 8.3 km/sec at the depth about 61 km and another high velocity layer of 8.4 km/sec at the depth of about 132 km. These two layers of 8.3 and 8.4 km/sec form the "lower-lithosphere" and are underlain by a low-velocity layer at the depth of about 149 km, which is the transition to the asthenosphere.

In the Wertern U.S., a tectonically active region, the development of the lithosphere is poor. As reviewed in the previous paper (NAGUMO *et al.*, 1986a) the uppermost mantle layer in the continental tectonically

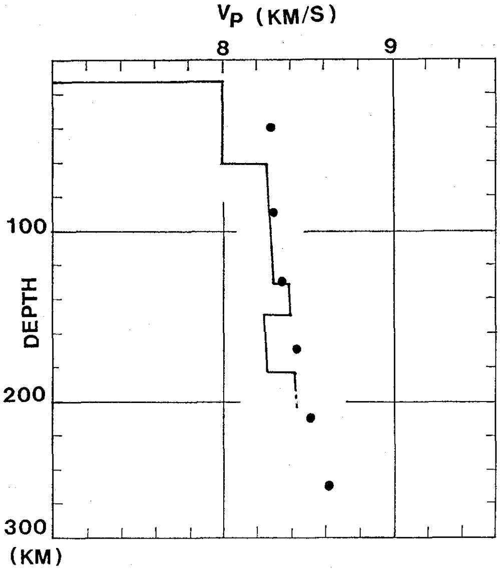

Fig. 7. A comparison of the mineralogically calculated velocities (solid circle: garnet-pyrolite model after HALES *et al.*, 1980) and the P-wave velocity structure of the lithosphere-asthenosphere beneath the western Northwest Pacfic Basin obtained from the P-wave travel-times.

active region, such as the Japanese island arc and the Western United States, are anomalous being characterized by the low-velocity layer in which the P-wave velocity is about 7.5–7.7 km/sec with or without a thin high velocity lid at the top of it. An example is the model in the Western United States (MASSÉ *et al.*, 1972) as shown in Fig. 6. The structure in the western Northwest Pacific Basin is quite different from such a structure.

From the viewpoint of evolution of upper mantle, the structure of the western Northwest Pacific Basin could be regarded as being similar to that of a continental shield region and is in accordance with the old age of the ocean floor (about 130–150 MYBP). However, the existence of the low-velocity layer and the velocity of 8.3 km/sec at the top of the lower-lithosphere may show that the evolution of the lithosphere in the western Northwest Pacific Basin is not yet fully matured. There may be many discussions about the maturity of the evolution of lithosphere as reviewed by SACKS and SNOKE (1984) and FUCHS and VINNIK (1982). The absence of a low-velocity layer within the lithosphere, however, could be regarded as an indication of a well matured state of the evolution.

The velocity 8.3 km/sec at a depth of about 61 km from the surface is in good agreement with the calculation for the garnet-pyrolite model by HALES *et al.*, (1980). In Fig. 7 is shown a comparison of their model calculation and the structure obtained in this study. The transition from 8.3 km/sec to 8.4 km/sec at the depth of about 132 km may correspond to another change of petrological composition. A mineralogical change of a higher content of garnet as suggested by HALES *et al.*, (1980), and LADLE STUDY GROUP (1983) might occur there. The possible existence of a weak zone which is seen in the distance range 10°–12° may be an indication of such a transition of mineralogical composition.

5. Conclusion

By preparing a composite travel-time record section in the distance range from 6° to 18°, we attempted to determine the P-wave velocity-depth structure beneath the western Northwest Pacific Basin. The main results are as follows. The uppermost mantle layer of P-velocity 8.0 km/sec is underlain by the "lower-lithosphere" which possesses a layered structure with a rapid velocity increase to 8.3 km/sec at the depth of 61 km and to 8.4 km/sec at the depth of 132 km. The depth of the low-velocity layer, which causes the shadow zone beyond the epicentral distance of 15°, is estimated as about 149 km. This low-velocity layer could be regarded as the most significant discontinuity which represents the transition from the lithosphere to the asthenosphere. The thickness of the lithosphere thus defined is thicker than that obtained by the surface wave studies. The P-wave velocity structure in the upper mantle beneath the westren Northwest Pacific Basin seems to be similar to that of the continental shield region in the Central U.S. except for the presence of an intervening low-velocity layer.

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Appendix Seismological data of the earthquake SN2 obtained by the ocean-bottom seismometer array.

Origin Time: 1981 JUL 09 10 H 28 M 47.86 S (JST)
Latitude: 39°53.67′N, Longitude: 143°48.18′E
Depth: 11.7 km, *M.*: 5.4

Station	P-time	S-time	S−P	(O−C)P	(O−C)S	Delta	Azimuth
KHQ 7	10^{H} 28^{M} 55.69^{s}	—	—	-0.07^{s}	—	40.53^{km}	195.77°
KHQ 9	58.72	—	—	0.41	—	64.42	157.96
KHQ 10	29 02.89	14.34	11.45	−0.36	−0.17	109.72	84.04
KHQ 1	02.94	14.44	11.50	−0.14	0.21	95.66	190.15
KHQ 4	06.98	22.48	15.50	−0.52	0.66	144.18	129.99
KHQ 2	08.33	—	—	1.1	—	143.36	100.40
KHQ 3	08.43	22.83	14.40	−0.26	−1.10	154.39	118.06

Station	Latitude	Longitude	Depth
KHQ 7	39°59.49′N	144°56.48′E	5.700^{m}
KHQ 2	39°38.81	145°19.13	5.365
KHQ 3	39°13.51	145°15.05	5.380
KHQ 4	39°02.99	144°57.63	5.530
KHQ 7	39°32.60	143°32.50	2.740
KHQ 9	39°21.21	143°57.35	4.815
KHQ 10	39°02.61	143°28.94	2.730

北西太平洋海盆西域リソスフィヤーアセノスフィヤのP波速度構造
——海底地震計群列観測による——

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前報に引続き，北西太平洋海盆にて得られた海底地震計長距離群列観測のデータから，同海域のP波速度構造を求めた．先づ震央距離 6°～18° にわたる走時記録断面を 6 ケの地震を用いて作成した(第 2 図)．震源における発震時の誤差の補正は走時曲線の基準となる直線をずらして合わせるという方法を用いた．幸い，1981 年 7 月 9 日 10 時 28 分，M=4.5 (JMA) の地震が起り，その震源データが震源付近に展開してあった群列によって精度よく求められた．それによってリソスフィヤからの屈折波の原点走時が決まり，各速度層の深度が決まった．見掛け速度 8.4 km/s, 8.6 km/s の深度がそれぞれ約 61 km, 132 km と求められた．15° 以遠の影領域 (Shadow zone) を作る低速度層の深度の推定は波線経路の断面を作ることによって行った．その結果，15° 以遠に入るべき波線の最深点から，低速度層の深度は約 149 km と推定された．勿論この深度は速度勾配に依存するので，更に詳しいことは波形振幅の距離による変化特性を調べねばならない．この深度は従来表面波の研究から求められている値より幾分大きめである．

今回求められた構造を大陸のシールド帯および活動的構造帯のそれらと比較してみると，北西太平洋海盆西域の構造はシールド帯のそれに近い．しかし，海盆下では下部リソスフィヤが更に2層構造

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をしていること，またその下位に低速度層を持っていることが大きく異っている．シールド帯ではリソスフィヤの底部が不明になるという一般的特性があるので，北西太平洋海盆下のリソスフィヤの底部が約 149 km と深く求められたということは，北西太平洋海盆西域下のリソスフィヤの進化がそれだけ進んでいることを表わしているものであろう．一方，活動的造構帯ではリソスフィヤに相当する深度の領域は厚い低速度層となっており，北西太平洋海盆西域下の構造とは全く異っている．

リソスフィヤの進化という観点からみると，シールド帯は進化の充分進んだ所，活動的造構帯は進化の初期の所とみなされるであろう．北西太平洋海盆西域は，その海底の年代 130–150 MYBP に対応して，進化が進んでいるものと見られよう．しかし，シールド帯の程度迄には未だ至っていないようである．