THREE-DIMENSIONAL SEISMIC STRUCTURE OF THE CRUST AND UPPER MANTLE BENEATH THE OROGENIC BELTS IN SOUTHERN HOKKAIDO, JAPAN

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P-wave arrival time data, collected from 50 local earthquakes recorded at 11 stations located in and around the Hidaka range, near the junction between the Kuril and the Japan arcs, were inverted to construct the three-dimensional structure utilizing a method developed by AKI and LEE (1976). Hypocenter depths ranged from approximately 30 to 160 km. The present analysis reveals the lateral heterogeneities in the structure of the crust and upper mantle beneath the Hidaka district. A distinct and well-constrained low-velocity zone in the upper mantle appears beneath the western side of the Hidaka range, while the high-velocity zone is suggested to be present under the central and eastern side of the range. Both zones are parallel to the axis of the Hidaka range. These zonal structures extend to a depth of at least 90 km. Such a velocity contrast in the upper mantle may be explained by the effect of the collision between the Okhotsk Sea and the Eurasian plates. Although there is no clear lithosphere-aesthenosphere boundary beneath this orogenic belt, the regional velocity variations correlate well with the general geological features within the belt.

1. Introduction

As methods of interpretation in seismology have increased in mathematical sophistication, it has become evident that the usual assumption of only a few velocities and thicknesses is extremely restrictive and insufficient to account for the detail of observed seismic data. As more data become available, lateral variations of increasing complexity are found. AKI and LEE (1976) developed a sophisticated method for determining laterally varying velocity structure from seismic array recordings of local earthquakes and applied this method to the Bear Valley area of Central California in the United States. HORIE and AKI (1979) have recently applied it on a larger scale, to the three-dimensional seismic structure beneath the Kanto district at around the junction between the Philippine Sea and the Pacific plates. This technique has been extensively used in the investigation of upper mantle heterogeneity, for example, in an arc-trench area such as Japan (HIRAHARA, 1977, 1980).

In this paper the three-dimensional velocity structure of the axial zones under the Hidaka district in southern Hokkaido, Japan, with special reference to the orogenic belt, is investigated by using data from the Research Center for Earthquake Prediction (RCEP) of Hokkaido University (see Fig. 5) together with those of the three temporary stations consisting of cassette data recorders and 2.0 Hz geophones. We also used the data from the bulletins of the Japan Meteorological Agency (JMA). The orogenic belt in the Hidaka district is located near the junction of the Kuril and the Japan arcs. The seismological network coverage on the belt enables detailed sampling of the crust and upper mantle beneath the belt. Investigations of deeper structure have been less numerous, and the thickness of the high-velocity lid of the lithosphere is not known.

The aim of the present study is to determine the extent of regional velocity variations in the upper mantle beneath the orogenic belt. After describing the geological setting and tectonic evolution of the Hidaka province, the data set that was utilized, the variations of travel-time delays obtained by the time-term method, and their relationship to the crust and upper mantle velocity structure are discussed. Aki and Lee's method can help to put some constraints on the deeper structure beneath the orogenic belt, over which there has already been some controversy (DEN and HOTTA, 1973; DICKINSON, 1978 and so on).

2. Geological Setting and Tectonic Evolution

The Hidaka orogenic belt has undergone a long and extremely complex geological development. Dating and stratigraphic correlation between regions have been complicated by numerous orogenic episodes. On the whole, the axial orogenic zone can be divided into four major tectonic units from east to west: The Tokoro Belt, the Hidaka Belt (the Hidaka Metamorphic Belt), the Kamuikotan Belt (the Kamuikotan Structural Belt), and the Cretaceous-Tertiary Folded Belt (see Fig. 1). The paleogene deposits are well developed in the Cretaceous-Tertiary Folded Belt (MINATO *et al.*, 1965). The Kamuikotan Structural Belt mainly consists of basic effusive rocks activated in a geosynclinal stage ranging from Triassic through to latest Jurassic. This structural belt is regarded as a metamorphic belt related to the Hidaka Orogenic Movement proposed by the Hidaka Research Group (MINATO *et al.*, 1965). Following HASHIMOTO (1978), the Hidaka Metamorphic Belt is typified by a monoclinical structure which dips to the east and plunges southward. The metamorphic belt shows marked zonal structure. Namely, three major zones are distinguished; the Western, the Axial and the Eastern. The Western zone is made up of thrust-bounded amphibolite, metabasite and peridotite complex. The up-thrusting seems to occur during the late Miocene to Pliocene periods. The Axial zone is characterized by a high temperature metamorphism, and was formed during tectonic movement. The Eastern zone was formed at a later stage of tectonic movement.

From the viewpoint of tectonics, these zonal orogenic belts are located in the

Fig. 1. Major geologic map of the Hidaka region and its neighborhood in the southern extremity of the Okhotsk plate. Zone A is the Tokoro Belt, B the Hidaka Belt, C the Kamuikotan Belt, and D the Cretaceous-Tertiary Folded Belt. Hatched *a* and dotted *b* narrow zones are the Hidaka Metamorphic Belt and the Kamuikotan Structural Belt, respectively. Initial sampled rectangular blocks are shown.

southern extremity of the Okhotsk Sea plate as proposed by DEN and HOTTA (1973). During the Early Cenozoic period, the Okhotsk Sea plate was sutured against the western terrain and then became attached to the Eurasian plate prior to the opening of the Sea of Japan. Prior to the collision, the Cretaceous and early Cenozoic orogene was an east-facing arc-trench system on the Eurasian mainland (OKADA, 1974).

Recent seismic data relevant to the interpretation of the tectonics of the Hidaka Orogenic Belt have provided good location data for events down to magnitude 2. Data are also available for focal mechanism solutions. Compiled solutions from shallow focus earthquakes beneath the Hidaka district indicate the dominance of the E-W or NE-SW compressional axes; for example, the 1970 Hidaka earthquake of magnitude 6.7, which occurred beneath the eastern flank of the Hidaka range, showed a thrust-type solution (SASATANI *et al.*, 1976). In contrast to this result, the Hidaka earthquake of magnitude 5.0, which occurred under the western flank

of the range, showed a normal fault-type solution, though located near the 1970 event (TAKANAMI *et al.*, 1978). Both events were accompanied by many aftershocks, which were somewhat unusual because such events accompanied by many aftershocks have seldom occurred before. Directions of null vectors of their focal mechanisms were parallel to the strike of the axis of the Hidaka range.

In the Kamuikotan Structural Belt and its vicinity, shallow earthquakes have appeared predominantly to be strike-slip faultings; for example, the 1977 Misono earthquake of magnitude 5.8 showed left lateral strike-slip faulting (TAKANAMI, 1978).

ONO (1980) investigated active faults around the Hidaka range by means of air-photographs and proposed that recent seismic activity is related to movement of small blocks according to gravity force. DEN and HOTTA (1973) tried to explain such shallow earthquakes as a product of the remaining movements of the crust along the Hidaka range which is a weak zone created by the Mesozoic tectonic process.

3. Data

The main data used in this study are the P-wave arrival times of local earthquakes given by RCEP, for the 4-year period July, 1976 through December, 1979. The seismological stations of RCEP were installed and operated by the research group at Hokkaido University. The station distribution is fairly uniform, although there are some gaps. All stations are equipped with one vertical and two horizontal components, short-period, 1 Hz geophones. Signals are telemetered to a central recording site at Hokkaido University in Sapporo, Hokkaido and recorded on a high-density digital data recorder (PCM data recorder) as well as on a multichannel analog pen recorder. The arrival times of P waves are read from the multichannel pen recorder with a paper speed of 0.5 cm/sec. In addition to the data by RCEP, we utilized the P-arrival times obtained by three temporary seismological stations as well as the data in the seismological bulletins of JMA. Portable slow-running cassette tape recorders at the temporary stations have been used successfully for the purpose of filling gaps in the permanent seismic network. The recording method using a cassette tape recorder is direct-recording, which has the advantage of maximum tape economy for a given bandwidth of unattended operation for long periods of time; for instance, the recorder can be operated continuously for 2 weeks on a cassette tape at a tape speed of 0.1 mm/sec and the timing accuracy of P arrivals is 0.1 sec or better.

We selected 50 events that have well-determined hypocenters with good coverage by a large number of stations over the Hidaka district. The important criteria used for selection were the widely distributed hypocenters within the area of the network and within a depth range of 120 km, based on preliminary hypocenter calculations.

Fifty local earthquakes with magnitudes greater than 2.0 were selected from

Table 1. List of hypocenters for earthquakes used in this study.

No.	Date	Coordinate			<i>M</i>
		Lat. (°N)	Long. (°E)	Depth (km)	
1	Aug. 11, 1977	42.96	142.05	105.5	2.6
2	July 1, 1978	42.86	142.06	134.7	2.1
3	Nov. 10, 1978	42.90	142.08	159.4	3.0
4	Apr. 29, 1978	43.00	142.47	153.3	2.4
5	Jan. 31, 1979	42.82	142.54	40.3	3.2
6	Dec. 20, 1978	42.86	142.46	106.5	4.7
7	Aug. 3, 1977	42.81	142.93	121.2	2.8
8	Nov. 14, 1977	42.83	142.89	83.1	?
9	Mar. 18, 1979	43.01	142.62	112.7	3.4
10	Feb. 19, 1979	42.91	143.40	121.6	4.1
11	Aug. 31, 1979	42.87	143.25	45.4	2.9
12	Nov. 18, 1976	42.68	142.24	142.9	2.0
13	May 12, 1978	42.70	142.29	86.8	2.2
14	Mar. 13, 1979	43.00	142.62	112.7	3.4
15	June 24, 1977	42.65	142.68	75.1	3.0
16	May 21, 1978	42.70	142.76	54.8	2.6
17	Aug. 28, 1978	42.67	143.07	129.2	2.4
18	Apr. 28, 1979	42.72	143.23	77.3	2.9
19	Dec. 23, 1976	42.47	142.09	42.9	3.5
20	June 29, 1978	42.50	142.02	109.9	2.1
21	Nov. 17, 1979	42.49	142.10	146.4	3.0
22	Oct. 11, 1977	42.50	142.59	102.2	3.3
23	Dec. 15, 1978	42.44	142.53	49.5	2.3
24	Apr. 1, 1979	42.52	142.52	71.1	2.4
25	Feb. 11, 1979	42.49	143.12	45.5	3.3
26	Dec. 7, 1977	42.55	142.95	121.6	2.9
27	Apr. 18, 1979	42.50	142.84	76.4	3.3
28	May 2, 1978	42.55	143.34	83.5	?
29	Aug. 28, 1979	42.50	143.24	48.2	2.3
30	Aug. 29, 1979	42.57	143.33	113.7	2.0
31	Oct. 27, 1979	42.30	142.00	126.9	3.3
32	May 19, 1979	42.04	142.44	61.8	2.9
33	July 2, 1977	42.29	142.21	90.1	2.9
34	July 29, 1978	42.26	142.32	57.4	1.8
35	July 3, 1978	42.30	142.68	47.5	2.6
36	Nov. 21, 1978	42.28	142.73	125.4	3.2
37	Oct. 28, 1977	42.35	143.01	49.7	3.2
38	May 10, 1979	42.30	143.02	62.9	4.1
39	July 15, 1979	42.28	143.15	105.8	3.9
40	Sep. 14, 1978	42.09	142.20	49.1	2.2
41	Oct. 26, 1978	42.02	142.08	72.8	3.1
42	Nov. 11, 1978	42.04	142.01	113.7	2.0
43	Sep. 8, 1976	42.00	142.46	105.7	2.5
44	Dec. 11, 1976	42.02	142.56	70.9	5.0
45	May 2, 1978	42.06	142.48	40.1	3.1
46	June 12, 1977	42.10	142.83	49.9	2.6
47	Feb. 23, 1979	42.08	142.77	51.9	3.4
48	Oct. 25, 1978	42.05	143.01	91.1	3.3
49	May 15, 1977	42.10	143.24	66.4	3.6
50	Sep. 2, 1978	42.17	143.36	42.2	1.7

The hypocenter parameters were determined by the RCEP.

Table 2. List of stations used in this study.

No.	Station code	Coordinate		
		Lat. (°N)	Long. (°E)	Altitude (m)
1	HIC (Hidaka)	42.89	142.46	330
2	MSN (Misono)	42.42	142.54	100
3	KMU (Kamikineusu)	42.24	142.97	185
4	ERM (Erimo)	42.02	143.16	40
5	MYR (Moyori)	42.29	143.28	90
6	IWN (Iwanai)	42.67	143.05	235
7	URA (Urakawa)	42.16	142.78	30
8	OBI (Obihiro)	42.92	143.22	39
9	MUKW (Mukawa)	42.61	141.95	50
10	KUSN (Kushinai)	43.07	142.67	480
11	MOMY (Momijiyama)	42.93	142.10	350

No. 1–No. 6 are routine stations of the Research Center for Earthquake Prediction (RCEP). No. 7 and No. 8 are those of the Japan Meteorological Agency (JMA). No. 9–No. 11 are temporary stations operated for this study.

1,024 that were gathered according to the criteria mentioned above. A total of 325 first P-arrival times were available for inversion. Hypocenters of events that are used in this study are listed in Table 1. The coordinates of seismological stations are shown in Table 2.

4. Method of Analysis

4.1 An outline of the method

The modeling technique developed by AKI and LEE (1976) is used to estimate the three-dimensional velocity structure beneath the Hidaka district. Briefly, the velocity perturbations relating to an initial model are calculated by using the following equation;

$$\tau = G\chi + \epsilon, \quad (1)$$

where τ is a vector containing travel time residuals computed from the starting model, G is a matrix which is calculated from the derivatives of travel time with respect to the X , Y , and Z coordinates and travel time spent within each block for the initial unperturbed model, χ is a column vector consisting of correction terms for coordinates and origin times of all the sources and for slowness parameters of the medium, and ϵ is an error vector.

The least squares solution of (1) satisfies the normal equation;

$$\tilde{G}G\chi = G\tau, \quad (2)$$

where $\tilde{G}$ is the transpose of G . If we have sufficient data, we can expect the matrix $\tilde{G}G$ to be nonsingular. In practice, the least squares solution of (2), however, becomes unstable because the matrix $\tilde{G}G$ contains very small eigenvalues if the effects

of any two different parameters are indistinguishable from the observations. Damped least squares (LEVENBERG, 1944) offer an alternative, and can be used to approximate the generalized inverse solution (LANCZOS, 1961). The damped least squares solution to (2) is given by

$$\chi = (\tilde{G}G + \theta)^{-1} \tilde{G}\tau, \quad (3)$$

where χ is an approximate solution of (1), θ a diagonal matrix with positive elements θ_i (damping parameters). The resolution matrix is given by

$$R = (\tilde{G}G + \theta)^{-1} \tilde{G}G, \quad (4)$$

and the covariance matrix for the error in solution due to the random error in data is given by

$$D = \sigma^2 (\tilde{G}G + \theta)^{-1} \tilde{R}, \quad (5)$$

where σ^2 is an estimate of variance of errors in the data. The i -th diagonal element of weight matrix θ is given by $\theta_i = \sigma^2 / \sigma_i^2$, where σ_i is the root mean square level of the solution χ_i . For example, if the standard error of the time measurement is 0.5 sec and the standard level of the fractional slowness fluctuation is 0.05, the corresponding diagonal element of θ_i is 100 sec². If the expected mean square levels of the correction terms Δx , Δy , Δz , and ΔT to the source parameters are σ_x^2 , σ_y^2 , σ_z^2 , and σ_T^2 , respectively, the corresponding diagonal elements of θ are chosen as $\theta_x = \sigma^2 / \sigma_x^2$, $\theta_y = \sigma^2 / \sigma_y^2$, $\theta_z = \sigma^2 / \sigma_z^2$, $\theta_T = \sigma^2 / \sigma_T^2$. Following the experiments of HORIE and AKI (1979), $\theta_i = (10 \text{ sec})^2$, $\theta_x = \theta_y = (0.1 \text{ sec/km})^2$, $\theta_z = (0.06 \text{ sec/km})^2$ and $\theta_T = (0.6 \text{ sec/sec})^2$ were initially assumed in this study.

4.2 Application of the method

In applying the method described above for solving the inversion model problem, a slightly modified version of the computer program written by AKI and LEE (1976) was utilized. Following AKI and LEE (1976), we shall divide the earth under a seismological network into blocks and assign a parameter to each block describing the perturbations of seismic wave slowness in the block. We have selected an area of 150 by 150 km in the Hidaka district for our application of the inversion method (see Fig. 1). We used a Cartesian coordinate system with the Z axis vertically downward. The coordinate origin is placed at 42° 00.0'N, 143° 24.0'E and at sea level, while the Y axis is chosen parallel to the meridian line. We then considered the volume of earth under the station network and divided it into equal rectangular blocks with sides parallel to the X, Y, and Z axes. The side lengths of each block in the X, Y, and Z directions are 25, 30, and 30 km, respectively. P-wave velocity of 6.6 km/sec for all blocks in the first layer and 8.0 km/sec for all the deeper blocks was also initially assumed. In order to smooth the effect of block configuration on slowness perturbation in each layer, we averaged the result for two configurations as shown in Fig. 2.

$$P1 = (V1 + V2 + V7 + V8)/4, \quad P1' = (V1' + V2' + V7' + V8')/4,$$

$$P2 = (V2 + V3 + V8 + V9)/4, \quad P2' = (V2' + V3' + V8' + V9')/4,$$

$$P20 = (V23 + V24 + V29 + V30)/4, \quad P20' = (V23' + V24' + V29' + V30')/4.$$

Fig. 2. Schematic representation of the computational scheme for smoothing the configurations of the slowness perturbations. $V1, V2, \dots, V30$ represent the slowness perturbations at the blocked elements in a layer. $V1', V2', \dots, V30'$ also represent the slowness perturbations in the elements re-blocked after shifting the origin of coordinates horizontally 12.5 km east and 15 km south.

5. Results

The use of the inversion method developed by AKI and LEE (1976) provides a direct way of determining slowness perturbations within a number of rectangular blocks.

In this section we present results from a three-dimensional inversion of the local P-wave travel time data following the method described in the preceding section. The slowness perturbations in each layer are shown in Fig. 3 as a final model. In this figure, areas with negative slowness perturbations are marked 'H' which are equivalent to higher velocities, and areas with positive slowness perturbations are marked 'L' which represent lower velocities. For example, since the average velocity in the first layer is assumed as 6.6 km/sec, 6% 'L' corresponds to 6.2 km/sec. Unfortunately, because of the small number of stations, information regarding the surface blocks which correspond to the crust is little. Although the resolution is not very high and the standard errors are high, the pattern of the velocity perturbation shows a good correlation with local average station residuals of P-arrival times from distant nuclear explosion sources (Table 3). The local average station residuals are mainly dependent on the crustal structure; they may reflect crustal thickness and velocity variations. The uppermost mantle (layer 2) between depths of

Fig. 3a. Result of three-dimensional inversion. The numbers show percentage variation of P-wave slowness from an average slowness for crust layer 1. The letters L and H refer to low- and high-velocity anomalies, respectively.

Fig. 3b. Result of three-dimensional inversion for layer 2. See Fig. 3a for explanation of symbols.

Fig. 3c. Result of three-dimensional inversion for layer 3. See Fig. 3a for explanation of symbols.

Fig. 3d. Result of three-dimensional inversion for layer 4. See Fig. 3a for explanation of symbols. As the slowness perturbations in this layer are very small, we drew the contour lines more closely than those on the above figures (Fig. 3a-Fig. 3c).

Table 3. Regional variation of travel-time residuals of P waves from underground nuclear explosions ($M > 5.5$).

Station code	Source						Reference**
	(Southern Nevada)		(Novaya Zemlya)		(Eastern Kazakhstan)		
	M. res. (sec)	L. res.* (sec)	M. res. (sec)	L. res.* (sec)	M. res. (sec)	L. res.* (sec)	
ESH	+0.1	−0.2	+1.4	+0.6	+1.7	−0.1	Esan
HSS	+0.2	−0.1	+1.2	+0.4	+2.8	+1.0	Sapporo
HIC	+0.1	−0.2	+1.2	+0.5	+1.8	+0.0	W
MSN	+0.1	+0.8	+1.2	+0.4	+2.1	+0.3	W
KMU	+1.0	+0.7	+1.4	+0.6	+2.3	+0.5	W
ERM	+1.3	+1.0	+1.8	+1.0	+2.4	+0.6	W
MYR	−0.0	−0.3	+0.1	−0.7	+1.6	−0.2	E
IWN	−0.6	−0.9	−0.1	−0.9	+0.9	−0.9	E
AKK	−0.6	−0.9	−1.0	−1.9	+0.7	−1.1	Akkeshi

Locations and origin times of explosions reported by the ISC bulletins (1976.7-1977.12) are used. E and W indicate the east- and west-side stations of the Hidaka range, respectively.

* M. res. is mean residual of travel-times of P waves from Herrin's Travel Time Table (HERRIN, 1968). L. res. is M. res. minus averaged M. res. and is equivalent to local residuals.

** ESH (Esan), HSS (Sapporo), and AKK (Akkeshi) are other stations of the RCEP (see Fig. 5).

30 and 60 km is characterized by the northwest trending zones; there exists a low velocity zone of about +6% beneath the western orogenic belt outcropped at the surface. This suggests that the structural difference between the two orogenic belts extends into the uppermost mantle. The resolution in the uppermost mantle is much better than that for the crustal blocks. In contrast to the high resolution in the central area, the resolution deteriorate somewhat toward the peripheral region of the model because the number of ray-paths crossing each block decreases. The overall resolution and standard errors are best for this particular layer. In layer 3 (60-90 km), the zone of lower velocity seems to be shifted to the southwest. There is still evidence of higher velocities in most of the eastern blocks beneath the Hidaka range. In addition to this general trend of velocity perturbations, it is suggested that the velocity appears to increase to the east in this layer. The layer 4 (90-120 km) faintly resembles the overlying layers. As it appeared that the overall trends are small, we tried to contour the lines of equal slowness perturbations at narrow intervals of 0.2. There is little, if any, dissinction between the two geologic provinces observed at the surface. The deepest layer (layer 5, 120-150 km) bears little resemblance to the overlying layers. We neglected this layer in the present consideration because only a few blocks are sampled there. Because of the relatively sparse network geometry, addition of more layers or smaller blocks would degrade the resolution seriously. However, this analysis suggests that major lateral differences in structure extend to depths of at least 90 km.

6. Discussion and Conclusion

Three-dimensional inversion of P-wave travel time data illustrates that structures down to perhaps 90 km and deeper can be correlated with surface geology. This has an important implication in that major orogenic belts have effects that reach well into the deep structures. As mentioned before, the Kamuikotan Structural Belt, the westward-lying Hidaka uplift (Hidaka range), consists of early Cenozoic structures that may be related to the intervening suture.

This region is not only a site of large-scale geologic deformation, but it is also characterized by geophysical anomalies such as apparent low crustal velocities and negative Bouguer gravity anomalies; Hidaka is characterized by a strong velocity anomaly. This is consistent with the previous refraction results. The explosion seismic studies in Hokkaido were done by using the first arrival data from the explosion of 1968 in the sea off Cape Erimo, southern Hokkaido, and of 1969 in the sea off the Shakotan Peninsula, western Hokkaido, and at Teine Quarry near Sapporo (OKADA *et al.*, 1973). Because the quality of records was fairly poor at the stations, except at the stations northwest of the Ishikari-Yufutsu plain, western Hokkaido, no crustal structure in the southeastern half of the profile was obtained, although the data up to about 80 km from shot-points of the explosions in the sea off Cape Erimo suggested the existence of a 6 km/sec-layer. Such poor records

Fig. 4. Contour map of isolines of the Moho time-terms residuals. See Appendix for explanation of the residuals.

obtained in the profile can be explained by our results of three-dimensional inversion. Namely, a prominent low-velocity area in the west of the Hidaka range correlates clearly with the sites of poor explosion signals.

According to AKI (1977), the process which brought the serpentine to the surface might have weakened the crust causing the low velocity anomaly in Central California. Certainly, the Kamuikotan Structural Belt is dotted with many outcrops of serpentines of various sizes. MORIYA *et al.* (1980) obtained supporting results from quarry blast data; the lower crustal apparent velocities have been found in or just to the west of the Hidaka range. Also the seismic refraction measurements of the two-ship method were taken in the deep sea terrace off Hidaka (DEN and HOTTA, 1973; ASANO *et al.*, 1979). The refraction indicated a significant continental layer of 6.2 km/sec. It is possible that such a crustal structure with thick sedimentary layers lying over a layer of 6.2 km/sec must continue to the land as inferred from our inversion. Recently, the time-term method was employed to determine P_n velocity in and around Hokkaido using 17 earthquakes occurring in the crust and 57 stations in a range from 108.4 to 414.1 km, with a total of 240 observations being made (TAKANAMI, 1980). The P_n velocity over this area yielded a value of 7.4 ± 0.1 km/sec. The low P_n velocity of 7.4 km/sec recognized in the uppermost mantle in Hokkaido has been suggested for the northeastern part of Honshu and the northern Kanto to Chubu region (OKADA *et al.*, 1979). By using

Fig. 5. An example of seismograms recorded in the seismological network of the RCEP. All seismograms are vertical component records. Note that the epicentral distances and azimuths for the stations in the Hidaka region are almost the same.

several shallow events, SUZUKI (1978) also obtained a similar value for P_n velocity under the land side of the Aseismic Front as proposed by YOSHII (1975). Following Appendix, Moho time-term residuals are contoured in the map shown (Fig. 4). When we pay attention to Hidaka in particular, we can say that a northwest-southeast trend in the residuals contours are apparently observed in the same way as are the configurations of the slowness perturbations in the uppermost mantle in three-dimensional inversion. Along the Kamuikotan Structural Belt or the western foot of the Hidaka range, velocities in the second layer are lower than in the Hidaka range; for example, their slowness perturbations are about +6% and +4% for the former and latter regions, respectively. Namely, the lower negative residuals in the southwestern part of Hidaka can be identified with the lower velocities in the uppermost mantle and the higher negative values in the Hidaka range can be also identified with higher velocities in the uppermost mantle.

Fig. 6. Recalculated epicenter map using the data of the RCEP in due consideration of station adjustments inferred from travel-time residuals shown in Table 3. Hypocenters projected onto A-B and I-II vertical planes.

The existence of such strong lateral-inhomogeneity in the upper mantle beneath the Hidaka orogenic belt was also confirmed by the comparison of wave-shapes (Fig. 5). The seismic amplitudes of the events occurring in and around Hidaka do not decrease systematically with distance due to the strong inhomogeneity, and there is an remarkable difference in the distance dependence between the western and the eastern sides of the Hidaka orogenic belt. Therefore, this may be in agreement with the view that the western zone composed of low- Q and low- V materials runs in parallel with the eastern zone of high- Q and high- V materials. To be concrete, the eastern stations of IWN and MYR may be related to some buried mafic or ultra-mafic bodies whose appearances are indicated by local gravity highs. And the low-velocity anomaly zone in the uppermost mantle (30–90 km in depth) beneath the western stations of URA and ERM strikes northwest parallel to the structural grain of the Hidaka orogenic belt. Because of the limited extent of the seismological network, the vertical extent of this structure can only be constrained to lie somewhere between 90 and 120 km.

It is interesting to note that correlation of high-velocity zones and seismic zones can be explained by the presence of tectonic stress in a cold, elastic slab and its absence in a hot, ductile asthenosphere. Now, turning to the present zone, we see that a low-velocity zone is associated with a high seismicity (Fig. 6), in contrast to

Fig. 7. A schematic cross-section of the hypothetical plate boundary between the Eurasian and the Okhotsk Sea plates. KB, the Kamuikotan Belt; HB, the Hidaka Belt; TB, the Tokoro Belt; KS, the Kamuikotan Structural Belt; DB, the Diabase-Belt in the western part of HB; HM, the Hidaka Metamorphic Belt; LVZ, the low velocity zone.

what HIRAHARA (1977) found with this relation; deep-focus earthquakes indeed occur within a zone of high-velocity anomaly. The low-velocity zone may represent a weak spot in an otherwise rigid lithosphere where a concentration of stress occurs. Thus, the interpretation of velocity anomalies, in terms of seismicity, requires a careful assessment of various factors in the context of the physical conditions and material of the region. AKI (1979) has already pointed out such a similar relation of velocity structure and seismicity, in that the high seismic activity may be associated with either low- or high-velocity anomalies.

Recently, SHIMIZU and MAEDA (1980) proposed an inclined reflector dipping toward the northeast with an angle of about 40° beneath the Hidaka range by using data of the converted waves observed at a few of the stations in the RCEP network.

In summary, we have revealed that structural differences between two structural belts, namely, the Hidaka Metamorphic Belt and the Kamuikotan Structural Belt, extend to depths of 90 km and deeper. Velocity variations in deeper parts are correlative with superficial geologic and tectonic features. This indicates that major collisional episodes have been very long lasting (see Fig. 7). The lower velocities associated with western Hidaka are evident to a depth of at least 90 km and dip to the southwest showing a spatial correlation with the large boundary between the Okhotsk Sea and the Eurasian plates.

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APPENDIX

The time-term method in terms of structure, as summarized by several authors, e.g., BÅTH (1978), can be considered to be a "relative analysis"; i.e., under the assumption of no lateral velocity gradient, horizontal layering. And the time-terms (also called delay times or time intercepts) themselves are quite accurately determined for any given waves. Therefore, the mapping of time-terms would probably be more reliable than a mapping of thickness. At the least, it might be of interest to see time-terms mapped in the preparation of crustal thickness maps. From this point of view, such a structural analysis can hardly be considered to be more than a first approximation.

In practice, travel time between station i and source j cannot be defined ex-

actly. The observed travel time T_o will approximately be related to the solution values by

$$T_o = \tau_i + \tau_j + \Delta_{ij}/V_{P_n} + \epsilon_{ij}, \quad (\text{A.1})$$

where τ_i and τ_j are the Moho time-terms of the i -th and j -th sites, respectively, V_{P_n} is the Moho refractor velocity, Δ_{ij} is the distance between the i -th and j -th sites measured along the surface, and ϵ_{ij} is the observational error.

The general similarity between the Moho time-term contours and gravity anomaly contours has been also noted. We assume a relation between the Moho time-term and gravity anomaly of the form

$$\tau_i = a + b\Delta g, \quad (\text{A.2})$$

where τ_i is the Moho time-term in seconds, Δg is the Bouguer gravity anomaly in milligals, and a and b are constants. According to OKADA *et al.* (1979), the values of a and b are 3.23 and -0.0063 , respectively. In the above numerical calculations for P_n velocity in and around Hokkaido using the least squares method, such empirical time-terms were then subtracted from the observed travel times. Simultaneously, the focal depths of events in the crust were assumed to be zero, and differences of time-terms owing to their depths were then added to the observed travel times assuming that the velocities in the crust and uppermost mantle were 6.0 and 7.5 km/sec, respectively. Therefore, the time-terms obtained by the same procedure are equivalent to the residuals from empirical time-terms at the sites. We call them Moho time-term residuals. A contour of Moho time-terms residuals shown in Fig. 4 is roughly equivalent to an equi-deviation contour from the Moho time-terms calculated from (A.2) at sites. If the heavy anomalous values of the residuals exist in the contour map, we must doubt the application of the time-terms method. Conversely speaking, it is possible to examine the significance of the solution and the aptness of the method by means of the configuration of the contours. It is difficult to decide whether the assumptions such as lateral velocity gradient and horizontal layering exceed a tolerance limit or not. At least, the anomalous residuals do reflect the strong lateral-inhomogeneity. In this paper, we use the residuals map as a facility for the study of lateral-inhomogeneous structure.

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