SIMULTANEOUS DETERMINATION OF P AND S VELOCITIES OF THE CRUST AND THE SUB-MOHO MANTLE WITH EARTHQUAKE HYPOCENTERS IN CENTRAL HONSHU, JAPAN

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P and S arrival times of crustal and subcrustal earthquakes, obtained from three-component wide-band seismographs of the 10-station telemetered network of Nagoya University, were inverted by a damped least squares method to obtain the P and S velocity structure of the crust and the sub-Moho mantle in the Chubu district, Honshu, Japan. The hypocentral parameters and the P and S station anomalies were determined simultaneously. The solutions showed satisfactory stability and resolution. The crust and the sub-Moho mantle were divided into four layers, based on results of refraction studies. The uppermost layer of 4 km thickness is the "sedimentary layer" which was not incorporated into the inversion. The second layer with a thickness of about 20 km is the "upper crust" for which P and S velocities of 6.1 and 3.6 km/sec were obtained with a V_P/V_S ratio of 1.69. This P velocity is consistent with the results of refraction studies. When this layer was divided into two parts at a depth of 12 km, the sums of squared residuals were reduced dramatically, suggesting a subdivision of the upper crust at this depth. The upper portion possesses P and S velocities of 6.0 and 3.55 km/sec and the lower portion, of 6.3 and 3.7 km/sec, respectively. Both portions are characterized by a V_P/V_S ratio of 1.69. The third layer of about 10 km thickness includes the "lower crust" there in, for which P and S velocities of 6.9–7.0 and 3.8–3.95 km/sec were obtained. The corresponding V_P/V_S ratio is 1.78–1.81, in marked contrast to the upper crustal V_P/V_S ratio. The fourth layer is the "sub-Moho mantle" for which we obtained P and S velocities of 7.7–7.8 and 4.4 km/sec that give a V_P/V_S ratio of 1.76–1.77, a value similar to that for the lower crust. The inverted sub-Moho P velocity is in agreement with the results of refraction studies. The Moho boundary probably lies at a depth of around 33 km. The boundary between the upper and lower crust is not well constrained by the data. The consistency of our P velocity structure with those derived from explosion seismic studies indicates the feasibility of the present method for obtaining an S velocity structure. Poisson's ratio changes significantly across the boundary between the upper and lower crust but not across the Moho.

1. Introduction

The damped least squares method for simultaneous determination of seismic velocities and hypocentral parameters has been developed by CROSSON (1976a, b), AKI and LEE (1976) and others. One advantage of this method for determining the velocity structure of the crust and the upper mantle is the capability of detecting a low velocity layer which is sometimes hidden in seismic refraction records. Another advantage is related to the fact that S waves are generated to a minor extent from a seismic explosion source. If one wishes to determine an S velocity structure, one must in general use natural earthquakes whose hypocenters cannot be determined very accurately. Up to now, however, the application of this method has been limited to the evaluation of P velocity structure (e.g., STEPPE and CROSSON, 1978; HORIE and SHIBUYA, 1979; SATO, 1979). The purpose of the present study is the simultaneous determination of P and S velocity structures from the arrival time data of natural earthquakes by using the damped least squares method. In this method the unknown parameters of P and S velocities are coupled with each other through hypocentral determination to give more reliable information on V_P/V_S ratio or Poisson's ratio than with the method based on the independent data sets of P and S arrival times. The V_P/V_S ratios obtained in this study have been in part discussed by UKAWA and FUKAO (1981) in conjunction with mineralogy of the crust and the upper mantle.

One of the problems associated with S velocity study is the difficulty of detecting accurate onsets of S arrivals. We used the data recorded by broad-band and wide-dynamic range seismographs that were telemetered to the Regional Center for Earthquake Prediction Observation, Nagoya University. The response of this seismograph, which is compared in the inset of Fig. 1 with that of a conventional short-period seismograph, is particularly suitable for detecting S arrivals (UKAWA and FUKAO, 1981). The locations of the ten seismic stations operated by the Regional Center are shown in Fig. 1. Since we used earthquakes occurring within the network, our calculation would give an average velocity model in that region.

The P velocity structures in this region have been elucidated from several refraction studies (e.g., AOKI *et al.*, 1972; IKAMI *et al.*, 1977; IKAMI, 1978). Their results can be summarized as follows:

- 1) The crust is divided into three parts, the "sedimentary," "granitic" and "basaltic" layers.
- 2) The "sedimentary layer" is about 4 km thick with a P velocity of about 5.5 km/sec. The lateral inhomogeneity of this layer is, however, rather considerable.
- 3) The "granitic layer" with a P velocity of about 6.0 km/sec is relatively thick, about 20 km, in the central region of the Chubu district but about 15 km thick in the south.

Fig. 1. Map showing the region investigated and the station locations of the telemetered network of the Regional Center of Nagoya University. The upper-right insertion shows a seismograph response of this network (solid line) in comparison with that of a conventional short-period seismograph (broken line).

4) The "basaltic layer" with a P velocity of about 6.8 km/sec is only about 6 km thick in the central region, but a thickness of about 15 km has been suggested for the southern region.

5) The Moho discontinuity lies at a depth of 30–33 km. The sub-Moho P velocity is about 7.8 km/sec.

These features will be used as a guide in the following analysis.

The region under investigation is tectonically characterized as an incipient subduction zone of the Philippine Sea plate. YAMAZAKI and OOIDA (1982) found a subcrustal seismic zone, inclined at a dip of 15° toward the north from the Nankai trough, interpretable as the descending slab of the Philippine Sea plate. This situation makes somewhat difficult the estimation of a real velocity structure by using such a laterally homogeneous layered model as assumed in the present study. Although we made an effort to minimize the effect of lateral heterogeneity, our analysis may still be influenced by it. Our final model should be regarded in this sense as a best-fitting one-dimensional model for the real three-dimensional structure.

2. Method

In the present study the observations are P and S arrival times. The unknowns are P and S velocities, hypocentral parameters and station anomalies.

Usually we have already reasonably good solutions for the unknowns. Taking these solutions as starting values of the unknowns, we obtain such quasi-linear equations relating the observations to the unknowns as

$$\Delta T_{ij}^P = \sum_{k=1}^4 \left(\frac{\partial T_{ij}^P}{\partial x_{ki}} \right) \Delta x_{ki} + \sum_{l=1}^L \left(\frac{\partial T_{ij}^P}{\partial V_l^P} \right) \Delta V_l^P + \Delta \tau_j^P \quad (1)$$

for P waves and

$$\Delta T_{ij}^S = \sum_{k=1}^4 \left(\frac{\partial T_{ij}^S}{\partial x_{ki}} \right) \Delta x_{ki} + \sum_{l=1}^L \left(\frac{\partial T_{ij}^S}{\partial V_l^S} \right) \Delta V_l^S + \Delta \tau_j^S$$

for S waves, where

$$\begin{aligned} \Delta x_{ki} &= x_{ki} - x_{ki}^{(0)} \\ \Delta T_{ij}^P &= T_{ij}^P - T_{ij}^{P(0)}, & \Delta T_{ij}^S &= T_{ij}^S - T_{ij}^{S(0)} \\ \Delta V_l^P &= V_l^P - V_l^{P(0)}, & \Delta V_l^S &= V_l^S - V_l^{S(0)} \\ \Delta \tau_j^P &= \tau_j^P - \tau_j^{P(0)}, & \Delta \tau_j^S &= \tau_j^S - \tau_j^{S(0)} \\ i &= 1, \dots, I \quad \text{and} \quad j = 1, \dots, J. \end{aligned}$$

T_{ij}^P and T_{ij}^S are P and S travel times from the i th event to the j th station. V_l^P and V_l^S ($l=1, \dots, L$) are P and S velocities of the l th layer. x_{ki} ($k=1, \dots, 4$) is a hypocentral parameter of the i th event. τ_j^P and τ_j^S are P and S station anomalies of the j th station. The parameters with suffix (0) denote initial values. I , J and L are the total numbers of the earthquakes, the stations and the layers with unknown velocity parameters. Due to the characteristics of Eqs. (1), we must fix one of the station anomalies (CROSSON, 1976a). The P station anomaly of INU was fixed at 0 sec.

Equations (1) can be written in a matrix form,

$$A \Delta X = \Delta T, \quad (2)$$

where A is a coefficient matrix, ΔX a solution vector and ΔT an observational data vector composed of travel time residuals. We attempt to solve Eq. (2) using the damped least squares method. Refer to CROSSON (1976a), AKI and LEE (1976), and HORIE and SHIBUYA (1979) for the details of this method. Before solving this equation we must standardize the data and solution vectors, ΔT and ΔX , into non-dimensional vectors, $\Delta T'$ and $\Delta X'$, using weight matrices N and W , respectively.

$$\begin{aligned} \Delta T' &= N \Delta T \\ \Delta X' &= W \Delta X. \end{aligned} \quad (3)$$

N and W are the diagonal matrices whose diagonal elements are the reciprocals of the standard deviations of the observations and those of the maximum allowable errors of the solutions respectively (MATSU'URA, 1977). The standard deviations of the observations were estimated to be $\sigma=0.2$ sec for P and $\sigma=0.35$ sec for S on the basis of the mean residuals of P and S arrival times. This es-

timate of σ should be of the same magnitude as the combined uncertainty associated with the reading of arrival times and with the assumption involved in the analysis to insure the consistency of the problem (WIGGINS, 1972). It is rather difficult to estimate the maximum allowable errors of the solutions. After some trial and error, we found that the maximum allowable error for a parameter α can be empirically estimated from

$$\varepsilon_\alpha = \sqrt{\frac{n_0}{n_\alpha}} \left(\sigma / \sqrt{(\partial T / \partial \alpha)^2} \right), \quad (4)$$

where n_0 and n_α are the numbers of the data theoretically and actually available respectively for determining parameter α . For example $n_0^P = n_0^S = I$ for τ_j^P and τ_j^S , $n_0^P = n_0^S = I \times J$ for V_i^P and V_i^S and $n_0 = 2J$ for x_{ki} . $\sqrt{(\partial T / \partial \alpha)^2}$ represents a root mean square of the $I \times J$ individual $\partial T / \partial \alpha$. σ is 0.2 sec for P and 0.35 sec for S. To calculate the term in the parentheses of Eq. (4) for a hypocentral parameter x_{ki} , the contribution from P observations and that from S observations were averaged with relative weights n_α^P and n_α^S . Equation (4) may be understood as describing how the observational error propagates to the estimate of α . The use of this formula lead us almost always to the solutions with satisfactory resolution and stability. If we put

$$A' = NAW^{-1}, \quad (5)$$

then Eq. (2) can be written as

$$A' \Delta X' = \Delta T'. \quad (6)$$

The damped least squares method gives the following solution for Eq. (6)

$$\Delta X' = H \Delta T', \quad H = (A'^t A' + \theta^2 I)^{-1} A'^t, \quad (7)$$

where θ is a weighting coefficient. Since there is a trade-off between resolution and stability, we must search for an optimum value of θ . The optimum value of θ^2 is 1.0 in our case as the data and solution vectors have been already standardized as in Eq. (3) (AKI and LEE, 1976; HORIE and SHIBUYA, 1979). The resolution and covariance matrices are found from

$$R = H A', \quad C = H H^t \quad (8)$$

(CROSSON, 1976a).

We inverted two sets of arrival time data separately. They consist of the data of shallow crustal earthquakes (Data-1) and those of deeper crustal and sub-crustal earthquakes (Data-2). Since we restricted the epicentral distances within 100 km for each data set, the results of our inversion would be little affected by a structure at depths substantially below the earthquake foci. For this reason we first inverted Data-1 to determine upper crustal velocities. We then fixed the upper crustal velocities and used Data-2 to obtain lower crustal and sub-Moho velocities. This procedure reduces the number of unknown velocity parameters in one inversion and is expected to provide a more stable solution. The con-

sistency of this procedure may be checked against the two sets of station anomalies obtained from the two data sets.

One of the problems of the damped least squares method is a criterion for the convergence of the iterative solutions. We made in general 4–5 iterations for an inversion. The iterative solutions were visually extrapolated to the 8–10th iteration (see, for example, Figs. 5 and 7). Our criterion for the convergence is to judge whether or not each element of the extrapolated solutions is still within the standard error bounds obtained the final run of the iteration.

3. Data

The P and S arrival times were read to 0.01 sec on the vertical and transverse seismograms respectively. The transverse seismograms were obtained by rotating the original two horizontal seismograms. As described in the last section, the arrival time data were grouped into two, Data-1 and Data-2, according to the focal depths of the earthquakes. The hypocentral parameters are listed in Table 1, and were determined by the Regional Center of Nagoya University. In both the data sets the number of available data for each earthquake is more than 8 and

Fig. 2. Distribution of the initial hypocenters (closed circles) of Data-1 in comparison with the relocated hypocenters (open circles). Left, epicentral map; right, vertical cross-section of hypocenters. Stations are indicated by crosses (left) and by triangles (right), respectively.

usually 12 to 15 (see Table 1). Data-1 is composed of 35 upper crustal earthquakes with focal depths of less than 20 km. The hypocenters routinely computed are shown in Fig. 2 in comparison with the relocated hypocenters. The epicentral distances are less than 100 km so that the data are primarily related to the upper crustal portion. Data-2 consists of 46 subcrustal earthquakes in a depth range from 20 to 60 km. The routinely computed hypocenters and the relocated ones are shown in Fig. 3. The hypocentral distribution indicates a plane-like seismic zone gently dipping from south to north. Since the data are

Table 1a. List of earthquakes in Data-1

No.	Date				Time		Lat. (°N)	Long. (°E)	Depth (km)	Magni- tude	Number of data	
	Y	M	D	(hr)	(min)	(sec)					P	S
1	78	4	2	21	8	49.60	35.732	137.130	8.7	3.8	7	7
2	78	5	6	3	16	27.26	35.180	137.777	9.0	2.0	6	6
3	78	5	2	15	29	53.54	35.871	137.321	11.2	3.8	6	6
4	78	5	23	21	9	40.27	35.271	137.872	7.6	2.2	6	6
5	78	6	19	23	15	15.00	34.951	138.201	14.9	3.1	4	4
6	78	6	23	11	14	7.00	35.941	137.734	6.7	4.0	6	6
7	78	6	27	5	9	57.00	34.891	137.070	8.4	3.7	6	6
8	78	6	30	12	40	14.39	34.892	137.899	3.3	2.6	7	6
9	78	7	3	13	10	41.50	35.879	137.369	4.1	3.6	6	6
10	78	7	22	8	34	48.91	35.072	137.995	12.9	2.8	4	4
11	78	8	1	10	15	13.20	35.760	137.104	11.3	4.5	7	7
12	78	8	4	13	23	25.00	35.840	137.091	4.7	3.0	7	6
13	78	8	15	8	8	3.63	34.926	136.811	12.5	3.7	5	5
14	78	8	17	1	20	7.02	35.207	138.211	5.3	3.8	4	4
15	78	8	18	20	32	34.85	35.539	137.015	6.3	3.6	9	9
16	78	9	29	10	54	33.23	35.328	136.902	11.8	2.6	7	6
17	78	10	9	14	21	53.49	35.781	137.530	10.5	3.7	6	6
18	78	10	10	3	13	47.67	35.685	137.624	8.2	1.9	7	7
19	78	10	10	3	10	28.22	35.310	136.519	9.2	4.1	5	4
20	78	10	10	4	2	25.92	35.313	136.491	10.5	4.0	5	5
21	78	10	20	8	2	16.13	35.182	136.516	7.0	2.6	5	5
22	78	10	27	0	54	4.55	35.414	137.568	14.4	2.1	8	8
23	78	11	06	15	21	11.55	35.033	136.881	7.2	3.0	7	7
24	78	11	11	20	35	17.83	35.694	136.903	5.9	*	6	5
25	78	12	05	23	26	30.63	36.003	137.347	7.4	2.8	5	3
26	78	12	08	23	58	57.61	35.310	137.166	14.9	3.1	9	6
27	78	12	18	10	10	43.66	35.518	136.747	13.2	3.7	7	7
28	78	12	29	21	56	47.86	35.627	137.045	4.1	2.7	7	6
29	79	1	3	8	31	39.94	35.469	137.702	15.1	2.3	8	8
30	79	1	10	1	22	53.82	35.835	136.804	8.5	2.8	5	4
31	79	1	12	2	6	45.54	35.734	137.520	9.4	3.2	6	6
32	79	1	16	11	27	4.49	35.410	136.967	11.3	2.6	8	7
33	79	1	23	21	19	25.77	35.004	136.809	16.3	3.1	7	7
34	79	2	2	3	10	28.22	35.730	137.095	12.8	3.0	7	6
35	79	2	4	4	44	50.63	34.809	137.112	6.3	3.3	7	7

Y, year; M, month; D, day

Table 1b. List of earthquakes in Data-2.

No.	Date				Time		Lat. (°N)	Long. (°E)	Depth (km)	Magni- tude	Number of data	
	Y	M	D	(hr)	(min)	(sec)					P	S
1	78	4	2	9	2	42.12	35.120	137.459	38.7	3.6	7	7
2	78	4	4	3	48	26.82	35.322	137.433	46.8	3.5	10	8
3	78	4	10	20	1	54.13	35.110	137.907	37.3	2.0	5	7
4	78	4	12	14	28	31.78	35.251	137.368	47.1	1.7	8	8
5	78	4	22	18	40	57.50	35.099	137.075	43.0	2.0	7	7
6	78	5	18	3	55	45.45	35.036	136.790	36.3	2.2	6	5
7	78	5	18	20	06	10.61	34.700	137.608	27.0	3.0	4	4
8	78	5	25	1	41	19.98	35.209	136.496	35.2	1.8	4	4
9	78	6	1	17	42	30.54	34.951	137.382	35.3	2.8	5	5
10	78	6	7	5	31	49.32	34.876	137.439	40.2	2.4	5	5
11	78	6	14	0	32	47.77	34.988	137.283	36.7	3.3	6	6
12	78	6	24	21	14	3.27	35.251	137.098	41.9	2.0	6	6
13	78	7	7	23	49	23.86	35.109	137.795	41.2	2.4	6	6
14	78	8	2	15	37	44.72	35.107	136.903	41.8	2.4	7	7
15	78	8	7	1	26	43.01	34.667	137.222	29.5	2.5	4	4
16	78	8	11	9	28	34.10	35.125	136.971	40.9	3.5	7	7
17	78	8	11	12	0	58.28	35.288	136.960	40.0	2.6	6	6
18	78	8	11	20	22	44.74	34.926	137.645	35.7	4.2	5	5
19	78	8	21	16	01	41.88	35.090	137.321	39.5	*	6	6
20	78	8	27	0	51	0.87	35.327	137.357	41.2	*	7	7
21	78	8	29	13	20	6.11	34.819	137.155	33.5	2.0	4	4
22	78	9	5	23	51	20.72	35.104	137.291	39.1	2.1	6	5
23	78	9	21	12	30	1.28	34.998	137.306	39.6	*	6	6
24	78	9	28	16	4	22.90	35.269	137.294	45.8	2.3	9	6
25	78	10	14	21	25	26.44	35.008	137.369	41.0	2.0	5	5
26	78	10	14	23	28	37.29	35.174	136.817	35.9	2.1	7	7
27	78	10	15	9	27	54.56	35.127	137.299	40.0	2.5	6	6
28	78	10	19	10	46	20.94	35.045	137.312	41.9	2.0	6	6
29	78	10	20	16	0	14.75	34.862	136.456	54.3	3.9	6	6
30	78	10	21	2	59	13.86	35.223	137.223	39.9	2.2	6	6
31	78	10	22	13	07	16.77	35.163	137.315	40.0	2.4	7	7
32	78	10	23	13	32	10.57	35.416	137.115	45.4	2.1	8	8
33	78	10	23	13	57	26.07	34.994	137.625	32.2	2.8	7	6
34	78	10	26	23	37	35.28	34.831	136.799	36.6	2.3	4	4
35	78	11	03	3	31	55.99	34.912	138.041	26.6	2.2	4	4
36	78	11	10	23	49	39.73	35.278	137.309	44.4	2.2	9	8
37	78	11	11	9	13	2.10	35.135	137.555	40.2	3.1	7	7
38	78	11	16	6	15	18.77	35.116	137.346	42.7	2.3	7	7
39	78	11	18	14	31	59.84	35.387	136.900	43.1	2.2	7	7
40	78	11	23	2	50	4.72	34.883	137.284	32.4	2.2	5	5
41	78	11	24	5	21	19.07	34.880	136.956	38.5	2.6	5	5
42	78	11	25	13	48	59.11	35.476	137.433	57.9	2.8	8	10
43	78	11	27	14	42	21.80	35.311	137.427	44.0	2.3	8	7
44	78	11	30	7	35	15.32	34.771	137.683	24.2	2.5	4	4
45	79	1	15	11	26	40.10	35.232	137.303	42.8	3.5	9	9
46	79	1	24	10	3	2.10	34.878	137.295	38.3	3.0	5	5

Y, year; M, month; D, day

Fig. 3. Distribution of the initial hypocenters of Data-2 in comparison with the re-located hypocenters. Symbols are the same as in Fig. 2.

restricted to epicentral distances of less than 100 km, they are primarily concerned with the portion above the subcrustal seismic zone.

A comparison of Figs. 2 and 3 indicates that Data-1 and -2 are geographically rather complementary: subcrustal earthquakes occur in the area where there are relatively few upper crustal earthquakes and vice versa. Although we assumed a laterally homogeneous structure for these two complementary regions, the results from Data-1 and -2 are probably biased toward representing structures near the sampling regions for the two respective data sets.

4. Results

4.1 Upper crustal structure

We first inverted Data-1 to determine the velocity structure of the upper crust which is defined as a layer of about 20 km thickness with its top at a depth of 4 km. The surface "sedimentary" layer of 4 km thickness was not incorporated into the inversion for the reason described by CROSSON (1976b). The P velocity of this layer was fixed at 5.5 km/sec based on results of seismic explosion studies. The S velocity was rather arbitrarily assumed as 3.25 km/sec to give a V_P/V_S ratio similar to that of the upper crust (UKAWA and FUKAO, 1981). The inverted S station anomalies would depend on this choice of S velocity but otherwise it

little affects the results for the other parameters. The lower crustal velocities were fixed at 6.8 km/sec for P and 3.8 km/sec for S on the basis of the results of the seismic explosion study of AOKI *et al.* (1972) and the P* and S* study of UKAWA and FUKAO (1981). Small changes in these values little affect the results of the analysis.

Inversion was made with 5 structural models U-1 to -5. U-1 and -2 are one-layered upper crustal models and U-3 to -5, two-layered upper crustal models. Figure 4 shows the layer division. We tested in some models two sets of starting values for velocity parameters. We took $V_P=6.0$ km/sec and $V_S=3.57$ km/sec as the first set of starting values based on the results of AOKI *et al.* (1972) and

	km	U-1	U-2	U-3	U-4	U-5
V_P (km/sec)	0			<u>5.5</u>		
	10	6.11 ± 0.02 (0.971)	6.12 ± 0.02 (0.972)	6.06 ± 0.03 (0.973)	5.98 ± 0.03 (0.974)	5.97 ± 0.03 (0.973)
	20			6.35 ± 0.05 (0.987)	6.34 ± 0.04 (0.983)	6.33 ± 0.04 (0.983)
				<u>6.8</u>		
V_S (km/sec)	0			<u>3.25</u>		
	10	3.61 ± 0.01 (0.974)	3.62 ± 0.01 (0.975)	3.58 ± 0.02 (0.976)	3.55 ± 0.02 (0.977)	3.55 ± 0.02 (0.977)
	20			3.78 ± 0.03 (0.990)	3.74 ± 0.02 (0.988)	3.73 ± 0.02 (0.988)
				<u>3.8</u>		
V_P/V_S	0			<u>1.692</u>		
	10	1.691	1.690	1.694	1.686	1.683
	20			1.681	1.695	1.694
				<u>1.789</u>		
$\sum \sigma_P^2$ (sec ²)		2.960	2.883	2.899	2.051	2.237
$\sum \sigma_S^2$ (sec ²)		5.915	6.118	5.370	4.875	4.927

Fig. 4. Structural models for the upper crust and the results of inversion from Data-1. Values underlined are fixed throughout the inversion. Inverted velocities are indicated with their standard errors. Values in parentheses are the diagonal elements of the resolution matrices. Squared sums of final residuals of P and S arrival times are shown for each model.

UKAWA and FUKAO (1981). The second set of starting values consists of arbitrary fluctuations from the first set. Experiment confirmed that the results are little affected by which set of starting values are used.

Figure 5 shows the histories of the iterative solutions up to the 5th iteration for velocity parameters of model U-1 with the first set of starting values. The squared sums of P and S residuals, $\sum \sigma_P^2$ and $\sum \sigma_S^2$, are also shown. It is seen that $\sum \sigma_P^2$ and $\sum \sigma_S^2$ are dramatically reduced in the first two iterations, by the end of which the solutions almost entirely converge.

Figure 4 shows the inverted velocity structures to which the corresponding diagonal elements of resolution matrices are attached, all being greater than 0.97, indicating high resolution. The values of $\sum \sigma_P^2$ and $\sum \sigma_S^2$ are also presented as a measure of nearness to the real structure. The inverted P and S velocities are 6.1 and 3.6 km/sec with a V_P/V_S ratio of 1.69 for the one-layered models U-1 and -2. The corresponding velocities are 6.0–6.1 and 3.55–3.6 km/sec for the two-layered models U-3 to -5. The P velocities of 6.0–6.1 km/sec are in close agree-

Fig. 5. Histories of the iterative solutions for velocity parameters of model U-1. The corresponding V_P/V_S ratio and $\sum \sigma_P^2$, $\sum \sigma_S^2$ are also shown. The velocity parameters obtained at the final run are indicated with the standard error bounds.

ment with those obtained from refraction studies. For the second layer of the two-layered models we obtained P and S velocities of 6.3–6.4 and 3.7–3.8 km/sec. The first and second layers, thus have an almost identical V_P/V_S ratio of 1.68–1.69. This value is in good agreement with the one obtained from a Wadati diagram study made by UKAWA and FUKAO (1981). Among the three two-layered models, U-3 has a boundary at a depth of 14 km and U-4 and -5 at a 12 km depth. Note that $\sum \sigma_P^2$ and $\sum \sigma_S^2$ of U-4 and -5 are significantly smaller than those of the other models (see Fig. 4). It is believed to be a real feature that P and S velocities increase at a depth of about 12 km within the upper crust.

Our preferred models U-4 and -5 are different from each other in the bottom depth of the upper crust. Although U-4 has slightly smaller $\sum \sigma_P^2$ and $\sum \sigma_S^2$ than U-5, their differences are not significant. Similarly the one-layered models U-1 and -2 are different from each other in the bottom depth of the upper crust, yet there is little difference in $\sum \sigma_P^2$ and $\sum \sigma_S^2$. These results suggest poor resolution of Data-1 for locating the bottom depth of the upper crust.

4.2 Lower crustal and sub-Moho structure

We next inverted Data-2 to determine the velocity structure of the lower crust and the sub-Moho mantle. The upper crustal velocities were fixed by basing on the former results. Inversion was made with 4 structural models LM-1 to -4. Figure 6 shows the layer division. LM-1 and -2 are characterized by the one-layered upper crust and LM-3 and -4 by the two-layered upper crust. The upper crustal structures of LM-1, -3 and -4 are identical to those of U-1, -4 and -5, respectively. LM-2 is a modified version of LM-1 in which the lower crust is displaced upward by 2 km with respect to LM-1.

AOKI *et al.* (1972) suggested a relatively thin lower crust of 6 km thickness with its bottom at a depth of about 30 km in the central region of the Chubu district. We initially attempted to incorporate a 6 km-thick lower crust into the inversion but were unable to obtain a satisfactory stable solution due perhaps to the paucity of the hypocenters in such a thin layer (see Fig. 3). IKAMI (1978), on the other hand, suggested a lower crust of about 15 km thickness with its bottom at a depth of 33 km in the southern region. The thickness of the lower crust was fixed at 10 km in the present study, with which we were able to obtain a sufficient stability. It should be born in mind, however, that the real thickness may vary considerably from region to region because of the lateral undulation of the top boundary of the lower crust.

The starting values of velocity parameters were given on the basis of the results of the seismic explosion study of AOKI *et al.* (1972) and the P^* , S^* and P_n , S_n study of UKAWA and FUKAO (1981). More specifically the starting P and S velocities were set at 6.8 and 3.8 km/sec for the lower crustal layer and at 7.8 and 4.4 km/sec for the sub-Moho layer. Figure 7 shows the histories of the iterative solutions up to the 4th iteration for velocity parameters of model LM-1. Although the convergence is rapid, the standard errors of the solutions are large

		LM-1	LM-2	LM-3	LM-4
V _p (km/sec)	0				
			5.5		
	10			5.98	5.97
		6.11	6.11		6.33
	20			6.34	
					6.73 ± 0.13 (0.900)
30		7.04 ± 0.11 (0.934)	6.89 ± 0.12 (0.923)	7.00 ± 0.11 (0.933)	
	40	7.80 ± 0.10 (0.958)	7.69 ± 0.09 (0.920)	7.80 ± 0.10 (0.933)	7.57 ± 0.09 (0.926)
V _s (km/sec)	0				
			3.25		
	10			3.55	3.55
		3.61	3.61		
	20			3.74	3.73
					3.78 ± 0.07 (0.890)
30		3.95 ± 0.07 (0.931)	3.81 ± 0.07 (0.949)	3.91 ± 0.07 (0.958)	
	40	4.40 ± 0.05 (0.972)	4.38 ± 0.05 (0.965)	4.41 ± 0.05 (0.972)	4.29 ± 0.05 (0.948)
V _p /V _s	0				
			1.692		
	10			1.686	1.683
		1.693	1.693		
	20			1.695	1.694
					1.780
30		1.784	1.808	1.790	
	40	1.772	1.759	1.770	1.765
$\sum \sigma_p^2$ (sec ²)		1.780	1.754	1.788	1.833
$\sum \sigma_s^2$ (sec ²)		6.117	6.027	6.089	6.661

Fig. 6. Structural models for the lower crust and the sub-Moho mantle and the results of inversion from Data-2. See Fig. 4 for other explanations.

compared to those for Data-1 (see Figs. 4 and 5).

Figure 6 shows the inverted velocity structure and the corresponding diagonal elements of resolution matrices which are almost always greater than 0.90. The values of $\sum \sigma_p^2$ and $\sum \sigma_s^2$ are also shown. In models LM-1 and -3 the Moho is located at a depth of 34 km, for which P velocities of the lower crust and the

Fig. 7. Histories of the iterative solutions for velocity parameters of model LM-1. The corresponding V_p/V_s ratios and $\sum \sigma_p^2$, $\sum \sigma_s^2$ are also shown. The velocity parameters obtained at the final run are indicated with the standard error bounds.

sub-Moho mantle were obtained as 7.0 and 7.8 km/sec, respectively. In model LM-2 the Moho is located at a depth of 32 km and P velocities obtained are 6.9 and 7.7 km/sec. LM-4 has a Moho boundary at a depth of 30 km and produced P velocities of 6.7 and 7.6 km/sec. The inverted velocities thus little depend on whether the upper crust is one- or two-layered. They primarily depend on the boundary depths of the lower crustal layer. We note that LM-2 is such a version of LM-1 that the lower crustal layer is displaced upward by 2 km relative to LM-1. Experiments were made at this stage with other versions of LM-1 in order to examine the sensitivity of the boundary depths of the lower crustal layer to the inverted values of velocity parameters. The second version of LM-1 displaces only the top boundary of the lower crustal layer by 2 km upward. The third version displaces only the bottom boundary by 2 km upward. The second version generated a velocity structure very similar to the original model LM-1. The third version generated a structure similar to LM-2. It follows that the velocity differences among different models are due primarily to the difference in the bottom depth of the lower crustal layer and that the top depth is poorly constrained. This in turn suggests that the thickness of the lower crust is not well determined from the present data sets. Such a result has been expected since refraction studies revealed a considerable lateral undulation of the top boundary of the lower crust.

Among the four models shown in Fig. 6, the model with the Moho at a shallow depth of 30 km, LM-4, is a less reliable model judging from the largest $\sum \sigma_P^2$ and $\sum \sigma_S^2$ and the poorest resolutions of velocity parameters obtained. The remaining models have the Moho at a depth of 32–34 km with a sub-Moho P velocity of 7.7–7.8 km/sec in agreement with results of seismic explosion studies. The lower crustal portion has a P velocity of 6.9–7.0 km/sec, a value somewhat greater than values reported as P* velocities (AOKI *et al.*, 1972; IKAMI, 1978; UKAWA and FUKAO, 1981; MORI *et al.*, 1981). The reasons for this departure may be: a relatively high degree of coupling in velocity parameters between the lower crustal layer and the sub-Moho layer; fine structure in the vicinity of the Moho that cannot be resolved by the present data; lateral inhomogeneity which was not taken into account in the present analysis. The generated velocity structures give a V_P/V_S ratio of 1.76–1.77 for the sub-Moho mantle and a ratio of 1.78–1.81 for the lower crust, indicating no significant difference in the V_P/V_S ratio across the Moho.

4.3 Station anomalies

We obtained two sets of station anomalies from Data-1 and -2 in which the P station anomaly of INU was fixed at 0 sec. For each data set the solutions for different models agreed within their standard errors, indicating the independence of structural models and starting values. The diagonal elements of the resolu-

Fig. 8. Station anomalies for (a) model U-1 with Data-1 and for (b) model LM-1 with Data-2. The abscissa is P station anomaly and the ordinate, S station anomaly. The bars indicate standard errors. The regression lines of P and S station anomalies are indicated by straight line.

tion matrices are all greater than 0.8, suggesting high resolution. Figure 8 shows the P and S station anomalies for models U-1 and LM-1 that were derived from Data-1 and -2, respectively. It is to be noted that there is no significant difference between the two sets of station anomalies except for the P and S station anomalies at MIK and MZI and the P station anomaly at CHT. This approximate coincidence between the two sets of station anomalies indicates the consistency of our analysis as well as consistency between the two sets of data. The three exceptional stations are all located near the coastal region of the Philippine Sea where local heterogeneity may persist to greater depths than in the region further landward. Figure 8 also shows a strong correlation between the P and S station anomalies. A linear regression analysis gives a slope of about 1.35 for Data-1 and a slope of about 1.50 for Data-2. These values are somewhat smaller than the V_P/V_S ratio values obtained for the crust and the sub-Moho mantle.

4.4 Hypocenters

Epicenters, focal depths and origin times were determined with average standard errors of ± 0.6 km, 1.8 km and 0.11 sec for Data-1 and ± 0.8 km, 1.4 km and 0.11 sec for Data-2 and within these uncertainties the relocated hypocenters were independent of structural models. Figure 2 compares these relocated hypocenters to those routinely determined by the Regional Center of Nagoya University, which were used as the starting values of inversion. For the subcrustal earthquakes (Fig. 4), there is some tendency that the relocated focal depths become progressively shallower than the original ones toward the north, although their differences are in general less than 5 km.

Table 2 shows examples for the elements of a resolution matrix relating among hypocentral parameters. The examples are the case for event No. 1 of Data-1 with model U-1 and the case for event No. 1 of Data-2 with model LM-1. As may be seen, the diagonal elements of the resolution matrices for hypocentral

Table 2. Elements of resolution matrices for hypocenter parameters.

(a) Event No. 1 of Data-1 with model U-1

	<i>X</i>	<i>Y</i>	<i>Z</i>	<i>t</i>
<i>X</i>	0.946	0.004	0.015	-0.011
<i>Y</i>	0.004	0.927	-0.021	0.032
<i>Z</i>	0.015	-0.021	0.406	0.180
<i>t</i>	-0.011	0.032	0.180	0.884

(b) Event No. 1 of Data-2 with model LM-1

	<i>X</i>	<i>Y</i>	<i>Z</i>	<i>t</i>
<i>X</i>	0.934	0.007	-0.001	-0.003
<i>Y</i>	0.007	0.936	0.012	-0.016
<i>Z</i>	-0.001	0.012	0.837	0.115
<i>t</i>	-0.003	-0.016	0.115	0.859

parameters are in general greater than 0.8 with a noticeable exception for the depth parameter. The diagonal elements for the depth parameter are often smaller than 0.8 and sometimes less than 0.5. The non-diagonal elements relating the depth parameter to the origin time parameter are relatively large, suggesting some trade-off in the determination of these parameters.

5. Discussion and Conclusions

Table 3 summarizes the result of our analysis for the velocity structure in the Chubu district, Honshu. The main features are as follows.

(1) The upper crust is characterized by a V_P/V_S ratio of 1.69, and possesses an internal boundary at a depth of about 12 km. This depth is well constrained by the data. A P velocity of 6.0 km/sec obtained for the upper portion is in good agreement with results of seismic explosion studies. A two layered feature of the upper crust has been also reported by IKAMI *et al.* (1977) in western Chubu to Kinki districts.

(2) The boundary between the upper and lower crust is not well constrained. Such a poor constraint may be indicative of a relatively large lateral undulation of this boundary, which has been suggested from the results of two seismic explosion studies of AOKI *et al.* (1972) and IKAMI (1978). We obtained a P velocity of 6.9–7.0 km/sec for the lower crust which is somewhat greater than the reported P* velocities of 6.7–6.8 km/sec (AOKI *et al.*, 1972; UKAWA and FUKAO, 1981). The inverted velocities, however, change by amounts as large as 0.1 km/sec when the bottom of this layer is displaced by only 2 km whereas no significant changes occur in $\sum \sigma_P^2$ and $\sum \sigma_S^2$. The P and S velocities are rather unstable in this sense. A more stable parameter against small displacement of the lower crustal layer is the V_P/V_S ratio for which a value of 1.78 to 1.81 was obtained in agreement with that derived from P* and S* velocities. This V_P/V_S ratio is in marked contrast to that of the upper crust (UKAWA and FUKAO, 1981).

Table 3. Inverted velocity structure: summary.

Depth (km)	V_P (km/sec)	V_S (km/sec)	V_P/V_S
0	5.5*	3.25*	1.69*
4	6.0	3.55	1.69
12	6.3	3.7	1.69
20–24	6.9–7.0	3.8–3.9	1.78–1.81
32–34	7.7–7.8	4.4	1.76–1.77
—			

* The values of the surface layer were fixed throughout the inversion.

(3) The data suggest that the boundary between the crust and the mantle is located at a depth around 33 km, in agreement with results of seismic explosion studies (e.g., IKAMI, 1978). There is little significant difference in the V_P/V_S ratio across this boundary. A sub-Moho P velocity of 7.7–7.8 km/sec is consistent with results of seismic explosion studies.

Thus, the P velocity structure generated through the inversion is consistent with results of independent refraction studies, though the agreement may be marginal for the lower crust. This in turn suggests the reliability of the corresponding S velocity and V_P/V_S ratio structure. It also shows the feasibility of the present method to determine S velocities of the crust and the upper mantle.

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