Structure of the Upper Mantle in the Island Arc-Systematic Errors in Focal Parameters and Inspections on the Suitability of Models-

Rep. Fac. Sci. Kagoshima Univ. (Earth Sci., Biol.) Nos. 5·6, p. 19–60, 1973

Structure of the Upper Mantle in the Island Arc — Systematic Errors in Focal Parameters and Inspections on the Suitability of Models —

By

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Investigations are made on systematic errors in focal parameters associated with the facts that the standard travel-time tables are different between JMA and ISC and that the existence of anomalous *S-P* is peculiar to JMA epicenter. For ISC parameters, *P* travel-time residuals are investigated on the assumption that they come from a slight deviation in velocity structure from the standard (J-B), the existence of station anomalies and errors in focal parameters.

P travel times from the events of shallow and intermediate focal depths in the Kurile region are analyzed by Herglotz-Wiechert method. These obtained models in addition to the models of HERRIN *et al.* and JEFFREYS are inspected on their suitabilities for the structure of the high-V zones near and in Japan.

I. Introduction

The studies on the structure of the upper mantle by body waves are made by many authors. Most of them are based on the assumption of lateral homogeneity in the structure. Then if there is some heterogeneity, an averaged structure may be estimated by taking it out of cosideration.

Taking heterogeneity in the mantle into consideration, KAILA *et al.* (1971) and TADA (1972) analyzed *P* travel times by the method very similar to GUTENBERG (1953). This method is applicable only for laterally homogeneous media. Then some proper section must be chosen for its application. Such a choice may, however, be hopeless in the existing situations of observatories.

Representative examples of studies on the inhomogeneous mantle are those by UTSU (1966, 1967), UTSU and OKADA (1968), OLIVER and ISACKS (1967) and *et cetera.* In these studies, travel times, waveforms and seismic intensities at the stations of different azimuths are compared respectively. In other words, an averaged structure is assumed and then deviations from this are investigated. Thus there probably remain some quantitative doubts in the results though they are satisfactory in qualitative sense. For instance, UTSU (1967) and ISHIDA (1970) estimated the ratio of *P* wave velocity in the high-V zone to that in the low-V zone, comparing travel-time residuals at various stations. In these estimations, the rough figures of the high-V zone and a standard velocity distribution are assumed. Then it is needed for the confirmation of the values to investigate these assumptions. When near-by events are used in the analyses, inspections of focal parameters are moreover necessary.

In this paper, a quantitative approach to the non-homogeneous mantle is introduced. First of all, the systematic errors in focal parameters are investigated by using travel time anomalies. A kind of investigation is made by the method similar to the joint epicenter method by DOUGLAS (1967). As a standard velocity distribution is needed in this method, observations at micro-earthquake observatories in Japan and JMA stations are analyzed by Herglotz-Wiechert method. In the last, inspections on the suitability of various models are made by two kinds of methods. One is an usual method by which the smoothed travel-time curves obtained from superposition for many events of nearly equal focal depths are compared with theoretical values. The other is a new one which is an application of the method for the analyses of travel-time residuals.

II. Systematic errors in focal parameters

It is necessary to make previous investigations of errors in focal parameters, such as the epicenter, the focal depth and the origin time, for the sake of speculating on the structure of the upper mantle by body waves.

For the earthquakes occurring near and in Japan, it is widely known that there are systematic differences in locations between JMA epicenters based only on the data from Japanese stations and ISC (or CGS) epicenters determined from a world-wide network. These differences are supposed to come mainly from anomalous structures in the upper mantle (UTSU (1967), ICHIKAWA (1969), ABE *et al.* (1971)). They are especially obvious for the earthquakes in the Kurile-Hokkaido region, the Sea of Okhotsk and the Izu-Bonin region and become over 100 km in the maximum (UTSU (1967, 1971)).

The epicenter which is based on the data only from one-sided observatories is affected by the anomalous structure, if any, between them and differs from that based on the data little affected by it. Systematic discrepancies in epicenter locations are, however, not always due to anomalous structures. They may result from another methods of calculations employed. Accordingly, in order to clarify the effects of anomalous structures, the effects arising from such causes must be removed.

2.1 Variations of epicenters due to another travel-time tables used

The focal parameters are fixed so that the observed arrival times of seismic waves are fitted best for one of the travel-time tables. Therefore, what type of tables is chosen has influence on the results.

Between JMA and ISC (or CGS), the travel-time tables for locations are different. ICHIKAWA and MOCHIZUKI (1971) prepared new travel-time tables, which were based on almost the same velocity distribution as J-B, and compared epicenters determined by JMA, CGS and them. They found that the averaged relations of the new epicenters to JMA ones were very similar to those of CGS ones to JMA ones though the new epicenters showed smaller differences from JMA ones than CGS ones did.

We shall now examine in this section how far an epicenter is relocated by using another table.

Let us the epicenter and the origin time be $(\lambda_0, \varphi_0, t_0)$, which are determined by using a certain table. λ_0 and φ_0 are the longitude and geocentric latitude of the epicenter, respectively. Then, the new parameters, $(\lambda_0+\delta\lambda, \varphi_0+\delta\varphi, t_0+\delta t)$, based on another table are calculated from the following equations:

$$\begin{aligned} &[\lambda'\lambda']\,\delta\lambda+[\lambda'\varphi']\,\delta\varphi+[\lambda']\,\delta t = [\lambda' v]\,, \\ &[\lambda'\varphi']\,\delta\lambda+[\varphi'\varphi']\,\delta\varphi+[\varphi']\,\delta t = [\varphi' v]\,, \\ &[\lambda']\,\delta\lambda+[\varphi']\,\delta\varphi+n\delta t = [v]\,, \end{aligned} \tag{2.1.1}$$

where

$$\lambda_i' = P_i(\partial T/\partial \Delta)_{\Delta=\Delta_{i0}}\,, \qquad \varphi_i' = Q_i(\partial T/\partial \Delta)_{\Delta=\Delta_{i0}}\,,$$

$$P_i = (\partial \Delta/\partial \lambda)_{\Delta=\Delta_{i0}}\,, \qquad Q_i = (\partial \Delta/\partial \varphi)_{\Delta=\Delta_{i0}}\,,$$

$$v_i = t_i - t_0 - T(\Delta_{i0})\,,$$

and n is the number of observations for the earthquake. $T(\Delta)$ is the standard travel time obtained from the new table, and i is the suffix for the i-th observatory.

Solving (2.1.1), we have

$$\delta\lambda = \frac{\begin{vmatrix} [\lambda' v] & [\lambda'\varphi'] & [\lambda'] \\ [\varphi' v] & [\varphi'\varphi'] & [\varphi'] \\ [v] & [\varphi'] & n \end{vmatrix}}{E}\,, \qquad \delta\varphi = \frac{\begin{vmatrix} [\lambda'\lambda'] & [\lambda' v] & [\lambda'] \\ [\lambda'\varphi'] & [\varphi' v] & [\varphi'] \\ [\lambda'] & [v] & n \end{vmatrix}}{E}\,,$$

$$\delta t = \frac{\begin{vmatrix} [\lambda'\lambda'] & [\lambda'\varphi'] & [\lambda' v] \\ [\lambda'\varphi'] & [\varphi'\varphi'] & [\varphi' v] \\ [\lambda'] & [\varphi'] & [v] \end{vmatrix}}{E}\,, \tag{2.1.2}$$

where

$$E = \begin{vmatrix} [\lambda'\lambda'] & [\lambda'\varphi'] & [\lambda'] \\ [\lambda'\varphi'] & [\varphi'\varphi'] & [\varphi'] \\ [\lambda'] & [\varphi'] & n \end{vmatrix}.$$

The solutions (2.1.2) represent that λ and φ are not independent with each other. From these solutions only, we are not able to know how the epicenter is relocated by exchanging the travel-time table. We now accept a straight line as a simplified travel time curve, and investigate the relation of the epicenter location with the gradient of the line.

Let V_0^{-1} be the gradient for the epicenter (λ_0, φ_0) and V^{-1} for $(\lambda_0+\delta\lambda, \varphi_0+\delta\varphi)$. At the j-th station, the epicenteral distance for the latter epicenter is put as Δ_j, then $\Delta_j=\Delta_{j0}+\delta\Delta_j$. Δ_{j0} is the epicentral distance for (λ_0, φ_0) and $\delta\Delta_j=P_j\delta\lambda+Q_j\delta\varphi$.

V_0^{-1} and V^{-1} are given by the method of least squares as

$$V_0^{-1}=X/Z\,,\qquad V^{-1}=(X+Y)/(Z+S+T)\,,$$

where

$$X=n[t\Delta_0]-[t][\Delta_0]\,,\qquad Z=n[\Delta_0^2]-([\Delta_0])^2\,,$$

$$Y=n[t\delta\Delta]-[t][\delta\Delta]\,,\qquad S=2n[\Delta_0\delta\Delta]-2[\Delta_0][\delta\Delta]\,,$$

$$T=n[(\delta\Delta)^2]-([\delta\Delta])^2\,.$$

Negelecting higher terms than the second power of $\delta\Delta_j$, we obtain

$$S-VY=\{(V/V_0)-1\}\,Z\,.$$

As $\delta\Delta_j=P_j\delta\lambda+Q_j\delta\varphi$, then

$$M\delta\lambda+N\delta\varphi=K\,, \tag{2.1.3}$$

provided that

$$M=n(2[P\Delta_0]-V[Pt])-(2[\Delta_0][P]-V[t][P])\,,$$

$$N=n(2[Q\Delta_0]-V[Qt])-(2[\Delta_0][Q]-V[t][Q])\,,$$

$$K=\{(V/V_0)-1\}\,(n[\Delta_0^2]-[\Delta_0]^2)\,.$$

In order to fix $\delta\lambda$ and $\delta\varphi$, we need another condition except (2.1.3). If we choose $[(\delta\Delta)^2]$=minimum as the condition, then we have

$$\begin{aligned}\delta\lambda&=-\frac{KN[PQ]-KM[Q^2]}{N^2[P^2]-2MN[PQ]+M^2[Q^2]}\,,\\ \delta\varphi&=\frac{KN[P^2]-KM[PQ]}{N^2[P^2]-2MN[PQ]+M^2[Q^2]}\,.\end{aligned} \tag{2.1.4}$$

Some examples of the relocated epicenters by (2.1.4) are shown in Figs. 1 and 2 for shallow earthquakes in the Kurile-Hokkaido region and the Shikoku-Kyushu region. For each earthquake, JMA epicenter is taken as the initial (λ_0, φ_0), which is indicated as a closed circle in the figures. The value of V recalculated for the epicenter at each end is also shown.

It is evident from these figures that the more V is high the more an epicenter is relocated on the continental side. This means, by analogical inference, that the more the averaged $dT/d\Delta$ of the travel-time cruve is low the more the epicenter based on the curve is on the continental side. According to ICHIKAWA and MOCHIZUKI (1971), the new epicenters are always on the continental side of JMA epicenters. And, certainly, comparing travel-time curves for the same depth, we know that the averaged values of $dT/d\Delta$ for new ones are lower than those for W-S-M (WADATI-SAGISAKA-MASUDA) ones.

MITRONOVAS and ISACKS (1971) found for the earthquakes in the Tonga-Kermadec region that the epicenter moves systematically in a certain direction as the focal depth is artificially restricted to different values. This is expected by the analogy of the above considerations though the epicenter does not always move in the

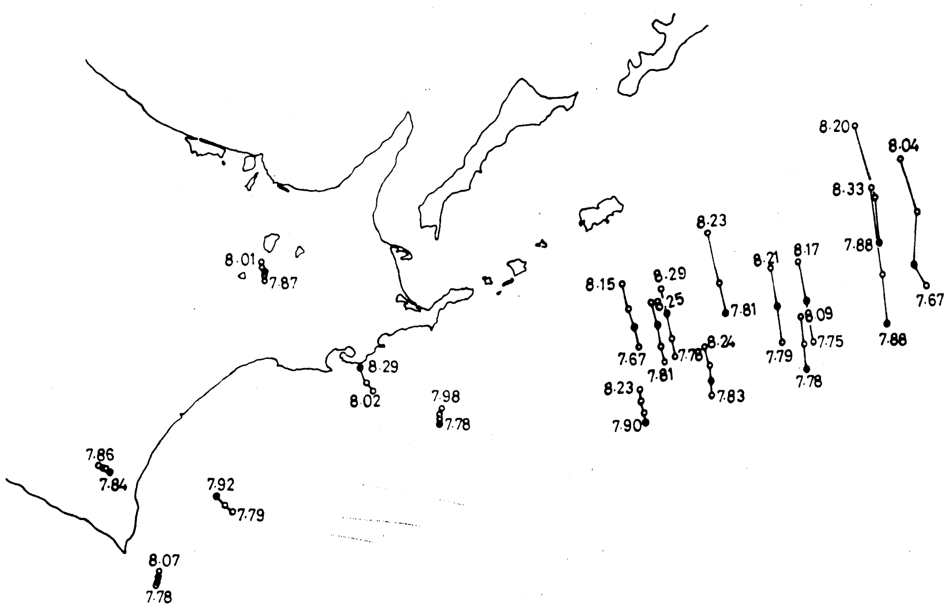

Fig. 1 Relocated epicenters for the events of shallow and intermediate focal depths in the Kurile-Hokkaido region. These epicenters are computed under the restriction of $\overline{V}$ to different values, where $\overline{V}$ is the inverse of the averaged gradient of the travel-time curve. $\overline{V}$'s for the epicenters on the outermost and innermost sides for each event are indicated by numerals in the figure. Closed marks are JMA epicenters.

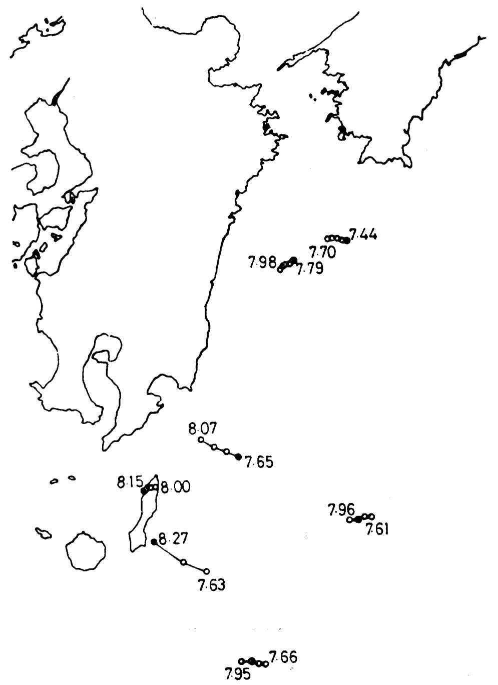

Fig. 2 Relocated epicenters for the events of shallow and intermediate focal depths in the Shikoku-Kyushu region.

direction of the continental side. The velocity distribution being fixed, then, the travel-time cruve of the lower value of the averaged $dT/d\Delta$ corresponds to the deeper hypocenter. This is why an epicenter moves in a certain direction as the focal depth is varied. Which direction the epicenter moves in may be a matter of situations of observatories.

Although it seems to be difficult to deny quantitatively the causes due to irregularities in the upper mantle if the results of ICHIKAWA and MOCHIZUKI (1971) are referred to, the differences in epicenter locations between JMA and ISC (or CGS) are qualitatively concordant with those due to the travel-time tables being different.

2.2 Systematic errors in JMA epicenter locations inferred from anomalous *S-P*'s

For the shallow earthquakes in the Kurile region, we often notice that the observed travel-time curve of *S-P* for an event is separated from that for another event though both events are determined by JMA as the same focal depth. Some examples are shown in Fig. 3 for very shallow events and those of 60 km depth. In the figure, *S-P*'s for the events on the west of the longitude of 148°E are indicated as closed marks and those on the east of that are as open marks. The focal parameters are tabulated with asterisk in Table 1. It is evident from the figure that *S-P*'s of open marks are later by ten seconds or over than those of closed marks at any distance within 1000 km or more.

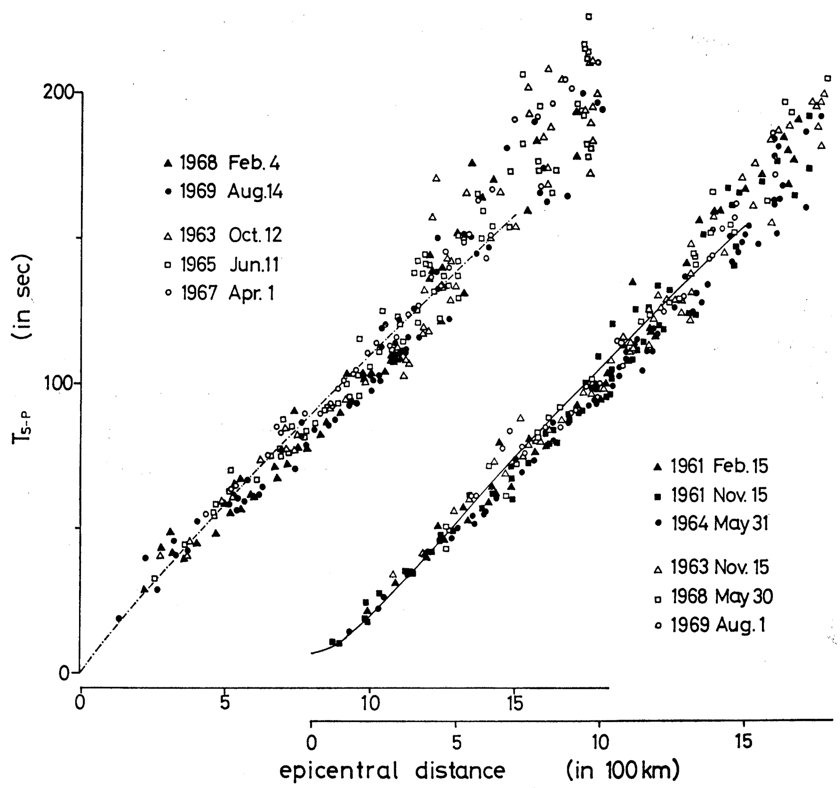

Fig. 3 Travel times of "*S-P*" based on JMA epicenter for the event in the Kurile-Hokkaido region. Left: for very shallow events with the theoretical curve (W-S-M) for surface focus. Right: for the events of focal depth of 60 km with the theoretical curve (W-S-M). The events on the east of 148°E are designated by open marks and the others are by closed marks.

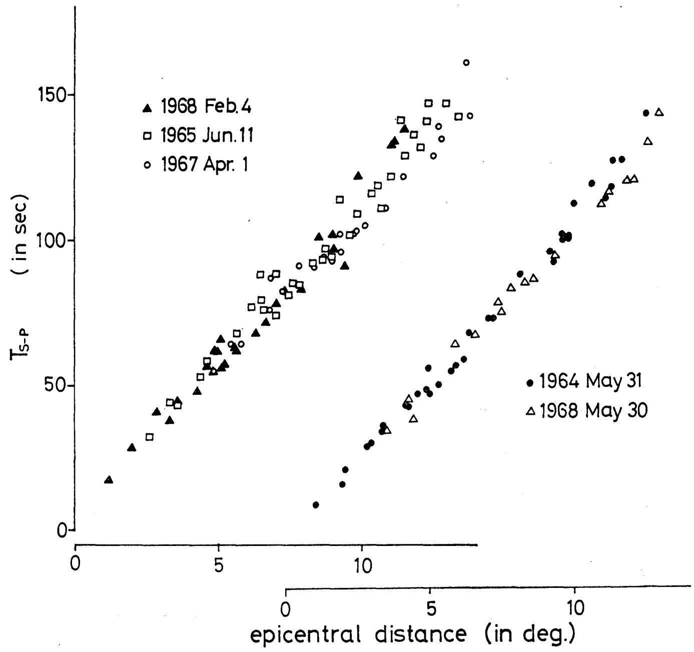

Fig. 4 Travel times of "*S-P*" based on ISC epicenter for the event in the Kurile-Hokkaido region. These events are all contained in Fig. 3.

Table 1. List of earthquakes with the arithmetic mean of *S-P* residuals and the epicentral distance to MAT. The events in Fig. 3 are denoted with asterisk. These focal parameters are given by JMA.

Origin Time (GMT)		Epicenter		Depth	M	$(\overline{S\text{-}P})_{res}$	Δ_{MAT}
*1961 Feb. 15	$10^h 45^m 14.9^s$	147°56′E	43°16′N	60^{km}	6.3	$-15.4^s \pm 4.0^s$	1114^{km}
*1961 Nov. 15	7 17 9.9	145 34	42 39	60	6.9	−17.2 ±4.5	926
1962 Jul. 17	17 20 22.3	145 10	42 38	60	5.9	−18.0 ±3.7	903
*1963 Oct. 12	11 26 58.1	148 54	43 53	0	6.3	−13.2 ±3.8	1220
1963 Nov. 10	17 17 44.2	149 04	43 36	40	6.0	− 9.2 ±1.8	1212
*1963 Nov. 15	21 6 34.4	149 04	43 32	60	6.2	−11.4 ±4.6	1206
1964 May 2	16 11 0.4	150 24	44 14	0	6.1	− 9.8 ±3.2	1340
*1964 May 31	0 40 35.7	147 14	43 16	60	6.7	−16.0 ±4.7	1072
1964 Jun. 23	1 26 34.9	146 28	42 59	80		−11.2 ±3.1	1006
1964 Jul. 24	13 25 21.6	153 53	45 49	80		− 3.6 ±2.0	1667
1964 Jul. 24	17 2 55.3	153 13	45 24	0	6.0	− 5.2 ±2.7	1598
*1965 Jun. 11	3 33 44.2	148 48	43 39	0	6.4	−11.2 ±4.2	1198
1966 Jun. 4	23 48 29.8	152 01	44 57	20	6.0	− 4.5 ±2.1	1491
*1967 Apr. 1	12 23 42.7	150 50	44 15	0	5.8	− 7.6 ±2.8	1369
*1968 Feb. 4	11 0 48.3	147 05	42 40	10	5.8		
1968 May 16	10 39 1.3	142 51	41 25	40	7.5	− 8.3 ±4.4	674
*1968 May 30	5 23 51.2	149 59	43 39	60	5.8	−10.9 ±5.0	1275
1968 Jun. 8	5 29 46.6	147 05	43 08	40	5.7	−10.7 ±3.8	1054
1968 Jun. 13	0 42 12.4	143 01	39 17	40	5.2	−12.1 ±4.9	521
1968 Jul. 12	0 44 33.6	143 29	39 34	40	6.4	− 8.3 ±6.0	572
1968 Aug. 7	8 0 13.2	144 58	42 58	80		−13.4 ±1.4	919
1968 Sep. 15	10 50 10.7	143 31	40 46	20	5.8	−11.7 ±6.1	658
*1969 Aug. 1	23 43 45.5	151 04	44 46	60	5.8		
*1969 Aug. 14	14 19 1.2	147 15	42 54	0	6.2		
1970 Feb. 2	17 22 5.2	147 45	42 50	60	5.4	−17.5 ±2.6	1075
1970 Feb. 2	17 49 48.7	147 48	42 58	80		−12.2 ±1.5	1087
1971 Dec. 2	17 18 32.8	152 06	43 44	0	6.5	− 7.4 ±2.2	1425

Travel-time curves of *S-P* for both groups with closed and open marks are on the whole parallel with each other within the range of epicentral distance. These features of anomalous *S-P*'s are characteristic for JMA epicenter, and do not appear for ISC epicenter (Fig. 4).

To clarify the characteristics of anomalous *S-P*'s, the arithmetic means of differences between the observed and theoretical *S-P*'s (based on W-S-M) are calculated for the part of *S-P*'s of which the relation to *P* arrivals is almost linear. They are listed in Table 1 with focal parameters by JMA. They are plotted in Fig. 5 as a function of epicentral distance to MAT. Strictly, the distance to the center of gravity of the observational network must be used. It is the reason why the epicentral distance to MAT is used in stead of it that MAT may be in the neibourhood of the gravity center of Japanese network for the events in the Kurile region. Fig. 5 demonstrates that the more epicentral distance to MAT, except less than about 700 km, increases the more the absolute value of the arithmetic mean of *S-P* residuals decreases. Namely, the observed *S-P*'s are more fitted on the theoretical curve as the epicenter locates more distant from the network. The characteristics in Fig. 5 may be expected when the travel times are abruptly delayed at a certain distance from the network or when *S-P* increases gradually with epicentral distance. If the irregularities in the upper mantle are main causes of anomalous *S-P*'s, their effects may appear in not only travel-time curves of *P* and/or *S* but the values of $\bar{V}_P/\bar{V}_S$ which are obtained from the relations between *S-P*'s and *P* arrivals. In the Kurile region, in spite of such expectations, it is difficult to classify the values of $\bar{V}_P/\bar{V}_S$ according to the situation of the epicenter (KAKUTA (1969)) and "the regional anomaly of travel-time curve" is not evident. Consequently, we cannot help to deny that anomalous *S-P*'s are due to such irregularities. It is deduced from the above investigations that the systematic

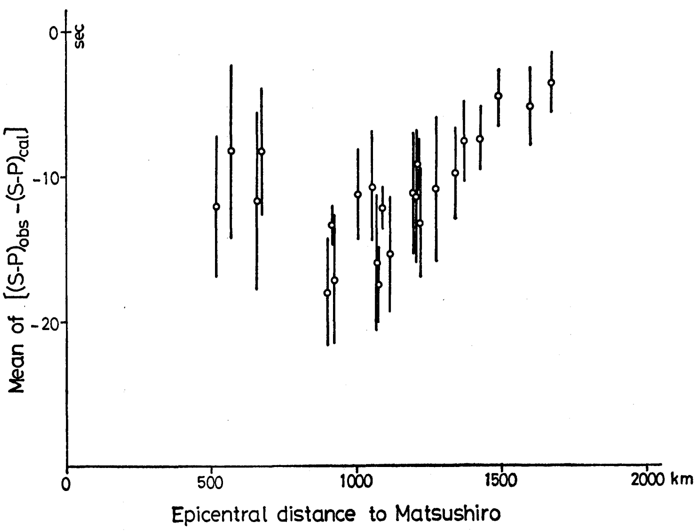

Fig. 5 The arithmetic mean of *S-P* residuals with its standard deviation based on JMA epicenter for the event of shallow or intermediate focal depth in the Kurile-Hokkaido region. This is plotted as a function of the epicentral distance to MAT.

errors in JMA epicenters, at least for the events in the Kurile region, contain the effects due to something but the irregularities in the upper mantle.

According to Seismological Section, JMA (1963), focal parameters are determined by GEIGER's method from P and S arrivals with appropriate weights. If there is no question in the procedure of location, a hypocenter is located to its own situation in accordance with the procedure. In such a case, anomalous S-P's would not be expected. The existence of them, therefore, suggests that something is improper in the procedure by JMA.

It is only thing we think possible that improper weights are applied to P and S arrivals, because the programing may hardly be in error. In location by JMA, weights of P and S arrivals are varied with the range of epicentral distance (SEINO, personal communication). The ratios of weights between P and S are about 2.2, 2.6

Table 2. Weights for P and S arrivals used by JMA for the determination of earthquake parameters (personal communication from SEINO (1972)).

P arrival		S arrival	
Epicentral distance	Weight	Epicentral distance	Weight
$0° \leq \Delta \leq 2°$	1	$0° \leq \Delta \leq 2°$	$1/\sqrt{5}$
$2° < \Delta \leq 6°35'$	$1/\sqrt{3}$	$2° < \Delta$	$1/2\sqrt{5}$
$6°35' < \Delta$	$1/2\sqrt{5}$		

and 1.0 in the range of epicentral distance less than 2°, from 2° to 6°35′ and more than 6°35′, respectively (Table 2). Namely, the weight ratio of S to P increases suddenly to more than twice when the epicentral distance to the station is over about 700 km.

On the other hand, travel-time residuals of S for shallow events in the Kurile region are negative at most of JMA stations in the range of epicentral distance less a thousand and some hundreds kilometers (see KAKUTA (1963, 1968) and HISAMOTO (1965)).

Considering these two things, that is, variations of weight ratio of S to P with the range of epicentral distance and negative residuals of S travel times, we easily recognize why mean residuals of S-P decrease with increasing epicentral distance to the gravity center of JMA network except less than 700 km. The reason is the following: S waves are more seriously accounted in location of a relatively distant event while a near-by event is located mainly from P data because of most observatories being at the epicentral distance less than 700 km. As a consequence, residuals of S travel times are small for a relatively distant event but large for a near-by one.

ABE *et al.* (1971) stated for the Eturup Earthquake of 1963 and its aftershocks that JMA epicenters are located at the distance of 100 km or more in the south or south-southwest direction of ISC or CGS epicenters. UTSU (1967, 1971) being referred to, the relative positions of JMA and CGS epicenters change gradually in a counterclockwise direction as the distance to the event from Japanese network decreases. That is, JMA epicenters lie in the southeast direction of CGS ones when the events are near the east coast of Hokkaido. Anomalous *S-P*'s are quite concordant with this, because the separation of *S-P* between the events of the open and closed marks in Fig. 3 disappears if the event of the open mark is relocated to more distant situation from Japanese network by some tens or about 100 km. In other words, the event of the open mark must be relocated in a northeasterly direction relative to the epicenter determined by JMA for the sake of anomalous *S-P*'s disappearing. If JMA epicenters are always located on the oceanic side of ISC (or CGS) epicenters and they are in the southeast of ISC epicenters for the events near the east coast of Hokkaido, then they become to lie in the south or southwest direction of ISC epicenters for the events far off the east coast of Hokkaido.

In a summary, anomalous *S-P*'s arise because the weight ratio between *P* and *S* arrivals is not kept constant in all ranges of epicentral distance. This is the main cause which brings systematic errors in JMA epicenters.

2.3 Residuals of *P* travel times based on ISC focal parameters

The residuals of travel times based on ISC (or CGS) epicenters are generally larger than those based on JMA epicenters. Though on the whole the geographic distributions of these residuals are systematic and consistent with the irregularities in the upper mantle, the ratios of the residual to the standard travel time are so high at the stations near the epicenter that they are hardly explained only by the irregularities. From this it is expected that there are also systematic errors in ISC epicenters. UTSU (1967) suggested on this subject that CGS epicenters deviate from the true epicenters by the influence of the lateral variations in the upper mantle. In this chapter, however, it is not our problem to estimate how much ISC epicenters deviate from the true ones but to inspect whether there are contradictions in ISC epicenters. For such inspections, we investigate the residuals of *P* travel times, taking account of anomalous structures in the upper mantle.

The structure of the upper mantle in an island arc is simplified as shown in Fig. 6. This is a vertical section perpendicular to the arc. The seismic waves emitted by the events in the Kurile region come to Japanese stations along the paths nearly parallel

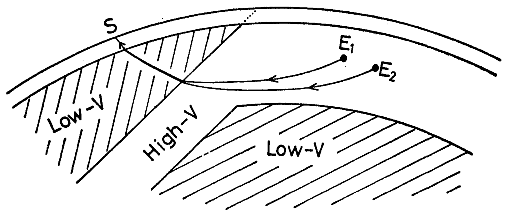

Fig. 6 A simplified anomalous structure in the upper mantle in a section perpendicular to an island arc. The cross section of the Japanese islands in the direction along the axis of the Kurile arc is approximated to this.

to the Kurile arc and perpendicular to the Izu-Mariana arc. Then we are able to treat the problem as a two-dimensional problem.

The velocity distributions are assumed as

$$V_h(r)=V_c(r)\{1+(\delta V/V_c)_h\} \quad \text{in the high-V zone,}$$
$$V_l(r)=V_c(r)\{1-(\delta V/V_c)_l\} \quad \text{in the low-V zone,}$$

where $V_c(r)$ is the standard velocity distribution. $(\delta V/V_c)$'s are far smaller constants than the unity and $(\delta V/V_c)_h+(\delta V/V_c)_l>0$.

If we assume there are no errors in ISC focal parameters, the observed travel time is

$$T_{\rm obs}=T_{\rm high}\{1-(\delta V/V_c)_h\}+T_{\rm low}\{1+(\delta V/V_c)_l\}+\gamma_s\,. \tag{2.3.1}$$

$T_{\rm high}$ and $T_{\rm low}$ are the standard travel times for the ray paths in the high-V and low-V zones, respectively. γ_s which is a constant peculiar to the station is concerned with the irregularities in the immediate neibourhood of the station.

The standard travel time is put as $T_{\rm cal}$, then

$$T_{\rm cal}=T_{\rm high}+T_{\rm low}-\alpha_s\,.$$

α_s represents the effect of refraction of the wave due to the anomalous structure in the upper mantle. So long as the structure is in Fig. 6, $\alpha_s>0$.

Substituting this into (2.3.1) and eliminating $T_{\rm high}$,

$$T_{\rm o-c}=-(\delta V/V_c)_h\,T_{\rm cal}+\gamma\,, \tag{2.3.2}$$

where $T_{\rm o-c}=T_{\rm obs}-T_{\rm cal}$, and

$$\gamma=\gamma_s+\{(\delta V/V_c)_h+(\delta V/V_c)_l\}\,T_{\rm low}+\alpha_s\{1-(\delta V/V_c)_h\}\,.$$

γ is usually called the station correction when $(\delta V/V_c)=0$. By the analogy, γ is hereafter called so. Though the negative γ is not always denied, γ is usually positive because all terms except the first on the right-hand side are positive. In the strict sense, γ is a function of epicentral distance. However, γ is regarded as a constant in a certain interval of epicentral distance, especially in the interval of $dT/d\Delta$ being nearly constant, as the increasing rate of $T_{\rm cal}$ is far larger than those of $T_{\rm low}$ and α_s in such a structure as that in Fig. 6.

It is also ascertained from observations that the relation (2.3.2) holds. For some Japanese stations, the values of $T_{\rm o-c}/T_{\rm cal}$ are computed for the events of which azimuths from each station range from 40° to 60°. They are shown in percent as a function of $T_{\rm cal}$ in Fig. 7. These data are from the Bulletin of the International Seismlogical Center. The figure demonstrates that for each station $T_{\rm o-c}$'s are well approximated by a relation (2.3.2) in the range of $T_{\rm cal}$ less than 210 or 240 seconds. Similar results are seen in KEBEASY (1969) though $T_{\rm o-c}$'s are plotted as a function of epicentral distance in his figures.

$(\delta V/V_c)_h$ and γ are obtained from observations by the method of least squares for each station. They are tabulated in Table 3 with their standard deviations. If the structure of the Kurile arc is uniform along the axis of the arc, $(\delta V/V_c)_h$'s must be

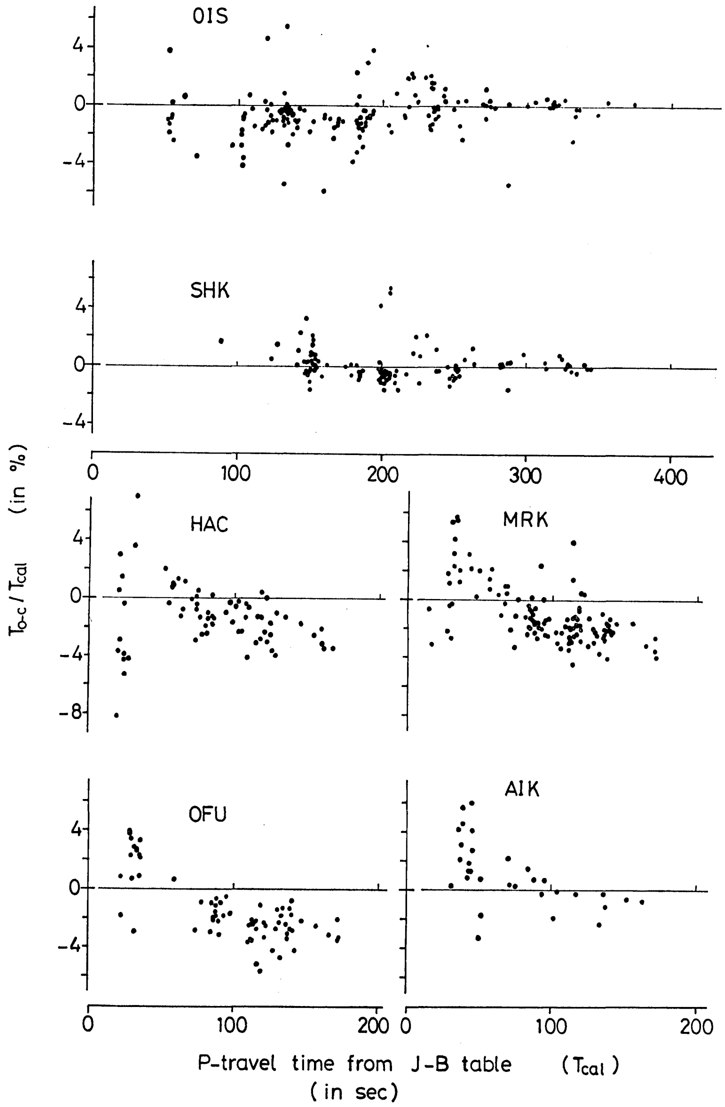

Fig. 7(a)

Fig. 7 *P* travel-time residuals for the events of shallow and intermediate depths in the Kurile-Hokkaido region. They are shown in T_{o-c}/T_{cal} as a function of T_{cal}, where T_{cal} is the theoretical travel time. These values are picked up in the bulletins of ISC.

nearly the same value for all stations. However, contrary to our expectations, they vary from 0.7 to 4.5% and it is difficult to assign a proper value of $(\delta V/V_c)_h$ satisfying the data of all stations. Moreover, the values of $(\delta V/V_c)_h$ seem to be concerned with the geographical situations of stations. That is, they are large at the nearer stations from epicenters, such as HAC, MRK and OFU, and small at the more distant stations, such as ABU and OIS. Among the stations of nearly equal epicentral distances, the values are larger for the stations on the more northern side, such as MAT and AIK, than for TSK and DDR.

These systematic variations of $(\delta V/V_c)_h$ are probably caused by errors in ISC focal parameters or by $(\delta V/V_c)_h$ being not constant.

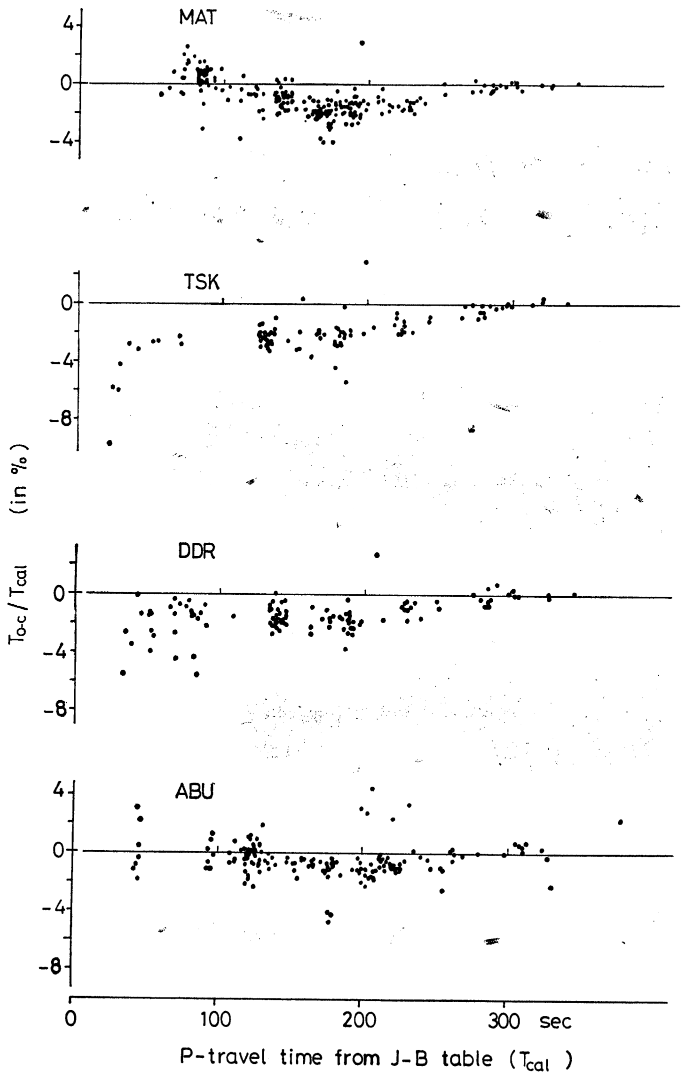

Fig. 7(b)

Table 3. Deviation ratio and station correction with their standard deviations calculated from the travel-time residuals at each station by least-squares technique.

	$(\delta V/V_c)_h$	γ
HAC	4.5±0.4%	2.9±0.3sec
MRK	4.1±0.2	2.4±0.2
OFU	4.2±0.2	2.2±0.2
AKI	1.4±0.6	1.5±0.3
AIK	3.1±0.3	3.0±0.3
TSK	2.1±0.1	−0.2±0.2
DDR	1.7±0.1	0.2±0.1
MAT	3.4±0.2	3.1±0.2
ABU	1.1±0.2	0.7±0.2
OIS	0.7±0.2	−0.2±0.2
SHK	2.6±0.4	4.2±0.7

2.4 Travel-time residuals due to errors in focal parameters

When there are errors in focal parameters though they are not so large, the travel-time residual for the i-th event is expressed at the i-th station by

$$(T_{\text{o-c}})_{ij} = -(\delta V/V_c)_h\,(T_{\text{cal}})_{ij} + \{1-(\delta V/V_c)_h\}\,[(\partial T/\partial\Delta)_{ij}\,\{P_{ij}\delta\lambda_i + Q_{ij}\delta\varphi_i\} + (\partial T/\partial h)_{ij}\delta h_i] + \delta T_i + \gamma_j\,, \tag{2.4.1}$$

where $\delta\lambda_i$, $\delta\varphi_i$, δh_i and δT_i are the errors in the longitude, latitude, focal depth and origin time of the i-th event, respectively, and P_{ij} and Q_{ij} are the same quantities as those in section 2.1.

If N events are observed at M stations, we have MN sets of observational equations. Such treatments of travel-time residuals were already performed by DOUGLAS (1967) as the joint epicenter method. His treatments, however, lacks the first term on the right-hand in (2.4.1).

The normal equations containing unknowns of $(4N+M+1)$ are expressed by a matrix as

$$U\vec{X} = \vec{S} \tag{2.4.2}$$

We solve (2.4.2) by the method of conjugate gradients. The solutions of the first approximation are found by the following. The ratio of velocity deviation, $(\delta V/V_c)_h$, and the station corrections are calculated at first under the assumption of no error in focal parameters. Then, putting them into (2.4.1) as the known quantities, we compute $\delta\lambda_i$, $\delta\varphi_i$, δh_i and δT_i.

If there is no error in focal parameters, the normal equation is immediately derived from (2.3.2). In a matrix expression, we have

$$A\vec{Y} = \vec{C} \tag{2.4.3}$$

where the components of the matrix, A, and vectors, $\vec{Y}$ and $\vec{C}$, are represented as $a(s, t)$, $y(s)$ and $c(s)$, respectively. Then,

$$a(1,1) = \sum_{i=1}^{N}\sum_{j=1}^{M}(T_{\text{cal}})_{ij}^2\,,\qquad a(1,k+1) = \sum_{i=1}^{N}(T_{\text{cal}})_{ik}\,,$$

$$a(k+1,1) = \sum_{i=1}^{N}(T_{\text{cal}})_{ik}/n_k\,,\quad a(s,t) = \delta_{st}\qquad (\text{for } s,t \geq 2)\,,$$

$$y(1) = -(\delta V/V_c)_h\,,\qquad y(k+1) = \gamma_k\,,$$

$$c(1) = \sum_{i=1}^{N}\sum_{j=1}^{M}(T_{\text{cal}})_{ij}\,(T_{\text{o-c}})_{ij}\,,$$

$$c(k+1) = \sum_{i=1}^{N}(T_{\text{o-c}})/n_k\,,\qquad (k = 1, 2, 3, \cdots, M)\,,$$

where n_k is the number of the observed events at the k-th station and δ_{st} is the Kronecker-delta.

The inverse matrix of A is obtained without difficulty. By expanding A on the last row or column, we have the recurrence formula

$$|A_k| = |A_{k-1}| - \alpha(1, k)\, \alpha(k, 1)\,.$$

Then the determinant of A is

$$\det(A) = \alpha(1, 1) - \sum_{m=2}^{M+1} \alpha(1, m)\, \alpha(m, 1)\,.$$

If the inverse matrix of A is written as B, its components, $\beta(s, t)$, are

$$\beta(1, 1) = 1/\det(A), \qquad \beta(1, k) = -\alpha(1, k)/\det(A)\,,$$
$$\beta(k, 1) = -\alpha(k, 1)/\det(A)\,,$$
$$\beta(s, t) = \delta_{st} - \beta(s, 1)\, \alpha(1, t) \qquad (\text{for} \quad s, t \geq 2)\,.$$

Then, the solutions of the first approximation are

$$(\delta V/V_c)_{h0} = -\sum_{m=1}^{M+1} \beta(1, m)\, c(m)\,, \qquad \gamma_{j0} = \sum_{m=1}^{M+1} \beta(j+1, m)\, c(m)\,.$$

Putting these solutions into (2.4.1), we have

$$\delta(T_{\text{o-c}})_{ij} = (\partial T/\partial \Delta)_{ij} \left\{P_{ij} \delta\lambda'_{i0} + Q_{ij}\, \delta\varphi'_{i0}\right\} + (\partial T/\partial h)_{ij}\, \delta h'_{i0} + \delta T_{i0}\,, \tag{2.4.4}$$

where $\delta(T_{\text{o-c}})_{ij} = (T_{\text{o-c}})_{ij} + (\delta V/V_c)_{h0}\,(T_{\text{cal}})_{ij} - \gamma_{j0}$, and $\delta\lambda'_{i0}$, $\delta\varphi'_{i0}$ and $\delta h'_{i0}$ correspond to the solutions of the first approximation, $\delta\lambda_{i0}$, $\delta\varphi_{i0}$ and δh_{i0}, multiplied by $\{1-(\delta V/V_c)_{h0}\}$, respectively.

The normal equations for (2.4.4) are set up for every event. They are expessed by a matrix

$$E_i \vec{Z}_i = \vec{W}_i \tag{2.4.5}$$

In order to determine the value of $(\delta V/V_c)_h$ in the Kurile region, we must choose the interval of T_{cal} so that the seismic ray comes only once through the high-V zone. We assign it from 60 sec to 210 sec, taking Fig. 7 into consideration, and treat T_{cal}'s only in this interval as the observations. For some sets of events above ten which occurred in an extent of about 2° in each dimensions and of which observations were not less than ten, (2.4.2) is solved according to the above process.

If percent is the unit of $(\delta V/V_c)_h$ and kilometer is that of $\delta\lambda$, $\delta\varphi$ and δh, $(\partial T/\partial h)$ is far lower than the coefficients of another unknowns except δh in (2.4.1). As a result, anomalously high δh's frequently occur in the above analyses. To avoid such anomalous δh's, we must exclude δh_i from (2.4.1). It is possible if the focal depth is confirmed by another methods, such as using of *pP-P*, *et cetera*. Then (2.4.2) is rewritten to the simultaneous linear equations of $(3N+M+1)$. Such procedure does not produce no effective changes in another parameters from the former case because $(\partial T/\partial h)$'s are very low.

(2.4.2) or its modification is also solved by alternate iterations of (2.4.3) and (2.4.5) if corrected residuals of T_{o-c}'s are all converged to zero.

III. Structure of the upper mantle along the axis of the Kurile arc analyzed by Herglotz-Wiechert method

In the last section, we tried to determine $(\delta V/V_c)_h$, γ's and errors in focal parameters simultaneously by the method of least squares, assuming that the residual of P travel time is the sum of their effects. In spite of our expectations, however, $(\delta V/V_c)_h$'s vary from region to region and errors in focal parameters of some events are so serious that we cannot help to consider the analyses unsuccessful. Such unsuccess may be concerned with some reasons. We consider in this chapter that it comes from the wrong assumption of $(\delta V/V_c)_h$ being constant and intend to construct a model of the structure of the upper mantle which satisfies the condition of $(\delta V/V_c)_h=0$.

It is generally difficult to prepare a quantitatively satisfactory model including the anomalous structure by using the existing observatories and natural earthquakes. To escape from such difficulties, we choose such a vertical section that the assumption of the layered earth is permitted. For the shallow events in the Kurile region, the greater parts of ray paths to Japanese stations are nearly parallel to the axis of the arc and are restricted within narrow limits in the direction perpendicular to the axis. If the structure is laterally homogeneous in the direction along the axis, this is just the case that the use of Herglotz-Wiechert method is admitted.

3.1 Data for the analyses

For the purpose of the analyses by Herglotz-Wiechert method, it is necessary to know the gradients of a travel-time curve for the surface focus as a function of epicentral distance. As a matter of fact, we are obliged to use the travel-time curves for arbitrary focal depths because there are very few such favorable conditions. Accordingly, the epicentral distances where the gradients are observed must be reduced to the values for the surface focus.

There are some kinds of methods by which we obtain the gradients from observations though the gradients are not the strict values but the approximated values. We now introduce some of them.

a) Difference in travel time between two stations

The difference in travel time divided by the difference in epicentral distance between two stations is an approximated gradient of a travel-time curve at the distance in the middle of two stations.

If Δ' is the apparent epicentral distance, the apparent gradient is

$$(\partial T/\partial \Delta)_{\mathrm{ap}} \fallingdotseq (T_j-T_k)/(\Delta'_j-\Delta'_k)\,. \tag{3.1.1}$$

On the other hand, its true value is

$$(\partial T/\partial \Delta)_{tr} \fallingdotseq (T_j - T_k)/(\Delta_j - \Delta_k)\,.$$

Designating the error in epicentral distance as $\delta\Delta$, where $\delta\Delta = \Delta' - \Delta$, we have from the above equations

$$(\partial T/\partial \Delta)_{tr} \fallingdotseq \kappa_{jk}(\partial T/\partial \Delta)_{ap}\,. \tag{3.1.2}$$

If $|\delta\Delta_j - \delta\Delta_k| \ll |\Delta_j - \Delta_k|$,

$$\kappa_{jk} = 1 + (\delta\Delta_j - \delta\Delta_k)/(\Delta_j - \Delta_k)\,.$$

Now, $\delta\Delta_i = P_i\delta\lambda + Q_i\delta\varphi$, where $P_i = -\cos\varphi_e \sin Az_i$ and $Q_i = -\cos Az_i$, then

$$\delta\Delta_j - \delta\Delta_k = -\left\{(\sin Az_j - \sin Az_k)\cos\varphi_e\,\delta\lambda + (\cos Az_j - \cos Az_k)\,\delta\varphi\right\}.$$

φ_e is the geocentric latitude of the epicenter and Az_i is the azimuth to the i-th station from the epicenter. Therefore if $Az_j = Az_k$, $\kappa_{jk} = 1$ and then $(\partial T/\partial\Delta)_{tr} = (\partial T/\partial\Delta)_{ap}$.

The apparent gradients are computed from the differences in travel time between two stations of which azimuths are equal in the bulletins of ISC. They are plotted in Fig. 8 as a function of epicentral distance reduced to the value for surface focus. The reductions are made under the assumption of the structure by ICHIKAWA and MOCHIZUKI (1971). In Fig. 8, the smoothed curve is derived by the method of summary values (ARNOLD (1968)). In this smoothing, the data from only *iP* phases, which are designated as open circles, are weighted by 5 times as much as those containing *eP* phases, which are represented in cross marks. This is a decreasing function of epicentral distance so that we can solve the problem immediately from this curve by Herglotz-Wiechert method.

The station corrections are not taken into consideration in deriving (3.1.2). If we consider these,

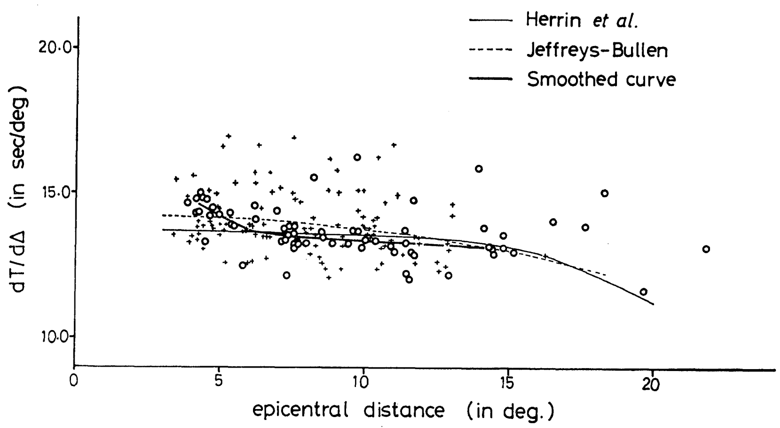

Fig. 8 The gradient of a travel-time curve calculated from the difference in travel time between two stations of which azimuths are equal in the bulletins of ISC. The value computed from only *iP* phases is indicated by open circle. If it contains *eP* phase, it is indicated by cross. It is plotted at the epicentral distance in the middle of two stations after the reduction to the values for surface focus. The smoothed curve is calculated by the method of summary values (ARNORLD (1968)).

$$(\partial T/\partial \varDelta)_{ap} \doteqdot \kappa'_{jk}(\partial T/\partial \varDelta)_{tr}+\nu'_{jk}\,, \tag{3.1.3}$$

where

$$\kappa'_{jk} = 1-(\delta\varDelta_j-\delta\varDelta_k)/(\varDelta'_j-\varDelta'_k)\,,$$

$$\nu'_{jk} = (\gamma_j-\gamma_k)/(\varDelta'_j-\varDelta'_k)\,.$$

Then $(\partial T/\partial \varDelta)_{ap}$ is the upper limit of $(\partial T/\partial \varDelta)_{tr}$ for the event in the Kurile region, because $\gamma_j>\gamma_k$ if $\varDelta'_j>\varDelta'_k$.

b) Superposing travel times from the events of nearly equal focal depths

Only the stations of equal azimuth to be used not only is extremely inefficient, but also gives rise to such a trouble that the same pairs of stations tend to be used. For sufficient efficiency, it is desirable to use all observations. In this case, however, the errors in focal parameters have much influence on the differences in travel time between two stations and, as a result, the values of $\partial T/\partial \varDelta$ may be highly scattered.

To avoid such troubles, superposing travel times from the events of nearly equal focal depths and dividing them into some groups according to their epicentral distances, we compute the smoothed travel times by the method of summary values for each group. Then $\partial T/\partial \varDelta$ is obtained with its accuracy. If there are many events, the effects due to errors in focal parameters may be diminished by canceling each other. The difficulties arising from the existence of station anomalies may also be settled by using many observations for events in long ranges.

The computations are carried for each class of focal depths of 0, 10, 20, 30, 40, 50 and 60 km. The focal depths except those over 65 km are grouped into one of the classes, by counting 5 and higher fractions inclusive and disregarding the rest. The interval of epicentral distance where the method of summary values is applied is set

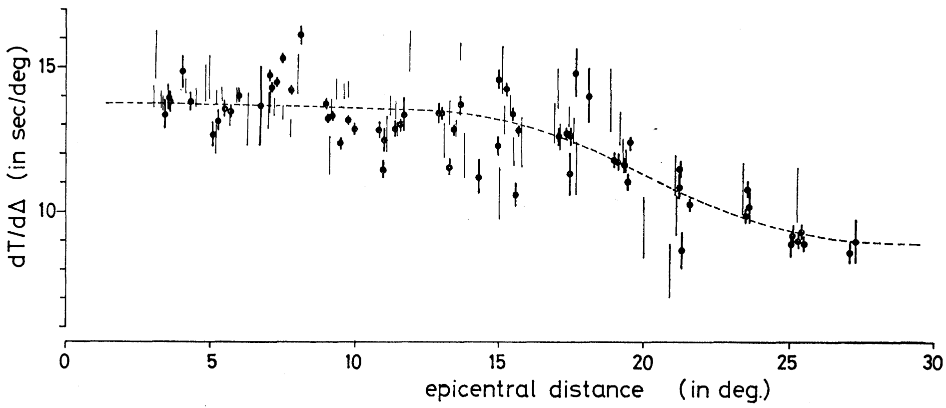

Fig. 9 The gradient of a travel-time curve calculated from the difference in travel time between the first and second summary points of the superposed travel times for the events of nearly equal focal depths. It is plotted in the middle of both points after the reduction to the values for surface focus. A thick bar centered with a closed circle is the gradient with its standard deviation from only the data of the micro-earthquake observatories and a thin bar corresponds to that from only the data of JMA stations. The theoretical curve of HERRIN *et al.* is indicated by a dashed line.

as 2°. The data of P travel times are obtained from the bulletins of micro-earthquake (or seismological) observatories, such as Urakawa, Dodaira, Inuyama, Abuyama, Tottori, Shiraki and Kochi, and JMA in addition to ISC.

$\partial T/\partial \Delta$ with its accuracy normalized to surface focus is shown in Fig. 9. A thick bar centered a closed circle on is computed from the data of the micro-earthquake observatories and a thin bar is from JMA stations. It seems that there are some maxima and minima in the curve of $\partial T/\partial \Delta$ as a function of epicentral distance. This is because of the distribution of the events being not uniform and means that there remain the effects of station anomalies. We take, therefore, this out of consideration and fit the curves decreasing with the increase of epicentral distance for these data.

c) Seismic array

The method of seismic array is very useful to determine most accurately $\partial T/\partial \Delta$.

According to OTSUKA (1966), the travel-time residual at the j-th station of the array is defined as

$$R_j = (T_j - T_0) - D_j \cos(\phi_j - \Phi)/V\,, \tag{3.1.4}$$

where D_j and ϕ_j are the distance and azimuth of the station from the origin of the coordinates in the array. T_j and T_0 are the arrival times at the station and origin. The apparent velocity V and azimuth Φ of seismic waves are determined so that the sum of squares of the residuals is minimum. If the structure is laterally homogeneous, we are able immediately to deal with the inverse of V as the data for the analyses. In the case of the non-homogeneous and/or anisotropic structure, however, we require corrections to V and Φ.

If we consider a station correction after the correction of altitude, the residual is

$$R_j = (T_j - \gamma_j - T_0) - D_j \cos(\phi_j - \Phi)/V\,. \tag{3.1.4'}$$

When no station corrections are taken account of,

$$R'_j = (T'_j - T'_0) - D_j \cos(\phi_j - \Phi')/V'\,. \tag{3.1.5}$$

Then, from the conditions that both sums of squares of residuals are minimum,

$$V = (P^2 + Q^2)^{-1/2}\,, \qquad \Phi = \tan^{-1}(P/Q)\,, \tag{3.1.6}$$

$$V' = (P'^2 + Q'^2)^{-1/2}\,, \qquad \Phi' = \tan^{-1}(P'/Q')\,, \tag{3.1.7}$$

where

$$P = \sum_{j=1}^{M} U_j (T_j - \gamma_j - T_0)/M = \sin \Phi/V\,,$$

$$P' = \sum_{j=1}^{M} U_j (T_j - T'_0)/M = \sin \Phi'/V'\,,$$

$$Q = \sum_{j=1}^{M} W_j (T_j - \gamma_j - T_0)/M = \cos \Phi/V\,,$$

$$Q' = \sum_{j=1}^{M} W_j(T_j - T_0')/M = \cos\Phi'/V' ,$$

$$T_0 = \sum_{j=1}^{M} (T_j - \gamma_j)/M , \quad T_0' = \sum_{j=1}^{M} T_j/M ,$$

$$U_j = ([YY]X_j - [XY]Y_j)/ME , \quad W_j = ([XX]Y_j - [XY]X_j)/ME ,$$
$$E = ([XX][YY] - [XY]^2)/M^2 ,$$
$$X_j = D_j \sin\phi_j , \quad Y_j = D_j \cos\phi_j .$$

If we put

$$\alpha = \sum_{j=1}^{M} U_j\left(\gamma_j - \sum_{k=1}^{M} \gamma_k/M\right)\Big/ M ,$$
$$\beta = \sum_{j=1}^{M} W_i\left(\gamma_j - \sum_{k=1}^{M} \gamma_k/M\right)\Big/ M , \tag{3.1.8}$$

we have

$$\alpha = P' - P , \quad \beta = Q' - Q .$$

V and Φ correspond to the corrected values of V' and Φ'.

Thus, from the above relations, the corrected value of V'^{-1} is

$$V^{-1} = (\alpha \cos\Phi' - \beta \sin\Phi')/\sin\Delta\Phi , \tag{3.1.9}$$

and the difference $\Delta V^{-1} = V'^{-1} - V^{-1}$ is given as

$$\Delta V^{-1} = [\alpha(\cos\Phi - \cos\Phi') - \beta(\sin\Phi - \sin\Phi')]/\sin\Delta\Phi , \tag{3.1.10}$$

where $\Delta\Phi = \Phi' - \Phi$. When $\Delta\Phi$ is small,

$$\Delta V^{-1} = \alpha \sin\Phi - \beta \cos\Phi .$$

In this case, we are able to regard ΔV^{-1} as a constant if both variations of α and β are not so serious and Φ is nearly the same.

Although we need nothing but to know only one of α and β or station corrections for the purpose of the analyses, we have no means to determine it without any assumption. Thus, in this chapter, we deal with V' of small $\Delta\Phi$, assuming that α and β are nearly zero.

The main data we analyze are the observations at micro-earthquake observatories. These observatories are Abuyama (ABU, TAT, TAN, KOB, TSU), Tottori (HKM, OYA, MZK, IZM, FNK), Wakayama (WKU, OIS, ISE, HBR, SRT, KNK, ARD, SCK, HDK, GZS, TDK, INN, KMT), Kochi (URS, WMY, IHR) and Shiraki (SHK). Their situations are shown in Fig. 10. The data of P travel times obtained from the bulletins published by the observatories and ISC.

KANAMORI (1967) had studied the structure of the upper mantle by using the the observations at the Wakayama Micro-earthquake Observatory. In 1965, this observatory has 12 sub-stations in a comparatively limited area, so that it seems to

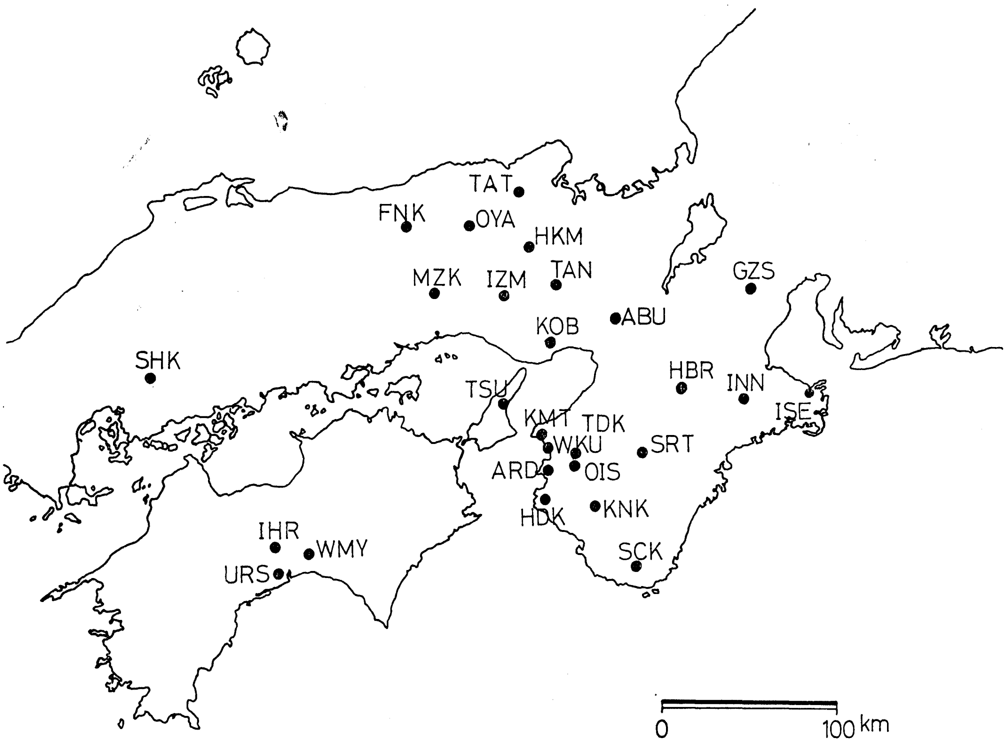

Fig. 10. Stations in the west of Japan. These are all used as stations of seismic arrays.

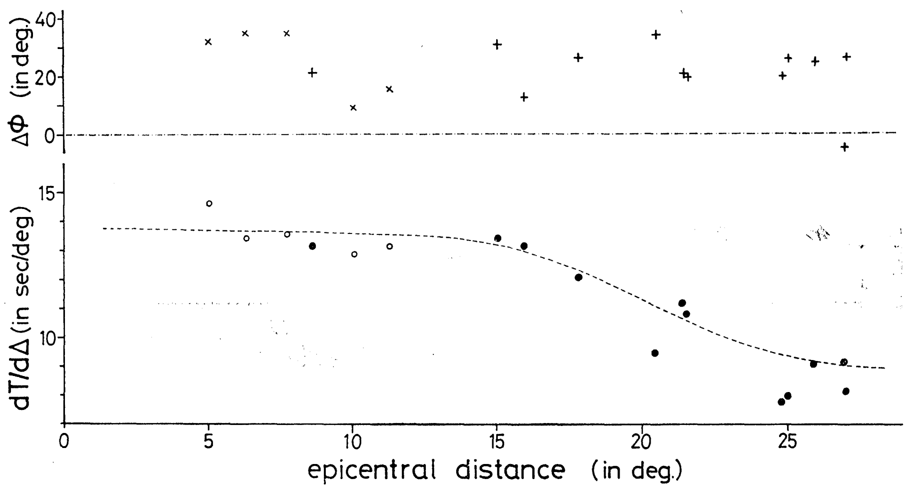

Fig. 11 Apparent slowness and azimuthal deviation of the incident wave at the seismic array composed of the Wakayama Micro-earthquake Observatory and ABU. These are indicated by closed circle and cross, respectively. Those at the array composed of the Tottori Micro-earthquake Observatory in addition to the above-mentioned stations are shown by open circle and x. The dashed curve is the theoretical curve of HERRIN *et al.*

be adequate for a seismic array. As a matter of fact, $\Delta\Phi$ from this array is serious though V'^{-1} does not so differ from the curve of HERRIN *et al.* (1968) (Fig. 11). The results are almost the same if the observations at the Tottori Micro-earthquake Observatory are combined with these. They are indicated in Fig. 11 as small open circles

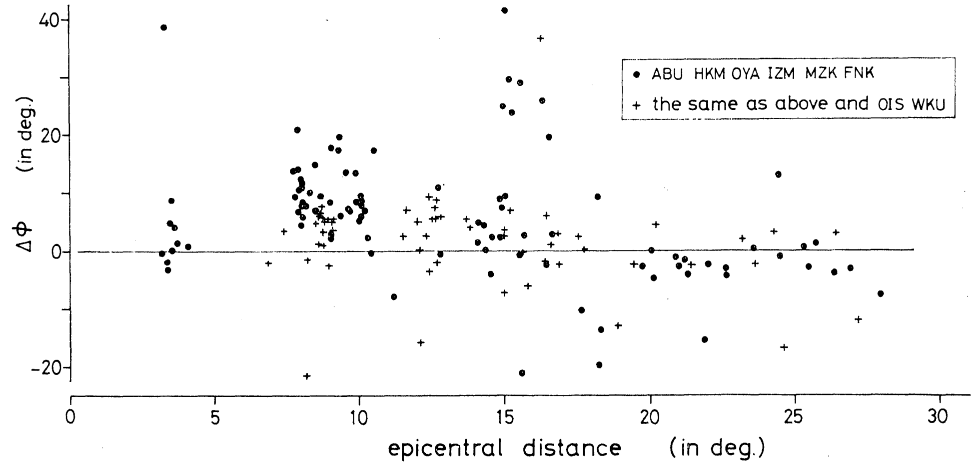

Fig. 12 Azimuthal deviations of incident waves at the seismic array composed of the Tottori Micro-earthquake Observatory and ABU. For the cases of including OIS and WKU, they are also plotted.

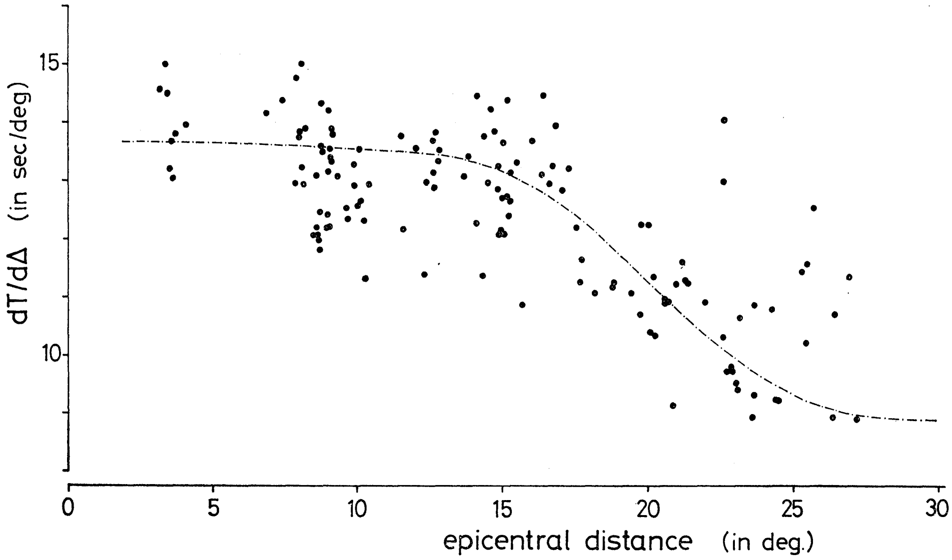

Fig. 13 Apparent slowness for which the absolute value of $\Delta\Phi$ is less than 10° at the seismic array of stations in Fig. 10 (those of the Wakayama Micro-earthquake Observatory are excluded except OIS and WKU). This is plotted at the reduced epicentral distance for surface focus. The curve is the theoretical one of HERRIN *et al.*

in the lower and as small x's in the upper. It seems that something anomalous may be concerned with this, so that it is better to lay aside this array in the beginning.

In Fig. 12, $\Delta\Phi$'s from the arrays composed of ABU, HKM, OYA, IZM, MZK and FNK and of WKU and OIS in addition to the former are shown as a function of epicentral distance. These are small as compared with the former case. We use the array composed of Abuyama, Tottori to which WKU and OIS are added for relatively near-by events and Kochi and Shiraki in addition to the former stations for distant events. V'^{-1}'s from the array of which the absolute values of $\Delta\Phi$ are less than 10° are shown with the curve of HERRIN *et al.* in Fig. 13. In this figure, the events near Matsushiro are also contained as the data at small epicentral distance are necessary for the analysis of the uppermost part of the mantle.

3.2 Epicentral distance where a ray passing through a low velocity layer begins to appear

An evidence of a low velocity layer in the upper mantle is the discontinuity in the relation between T_{S-P} and T_P (KAKUTA (1968)). The examples of the relations are adduced in Fig. 14 for the events of May 31 and July 24, 1965. These arrival times are those in the Seismological Bulletins published by JMA. The straight line for each event is corresponding to $\overline{V}_P/\overline{V}_S$ of 1.765, which is the averaged value for the shallow events in the Kurile region (KAKUTA (1969)). As seen in the figure, the maximum T_{S-P} which is fitted by the line is about 150 sec for the event of May 31 and it is about 180 sec for the event of July 24. The values of T_{S-P} over the maximum are fitted either by another lines of almost the same gradient but separated from the former or by another curves. For the events of shallow and intermediate focal depths in the

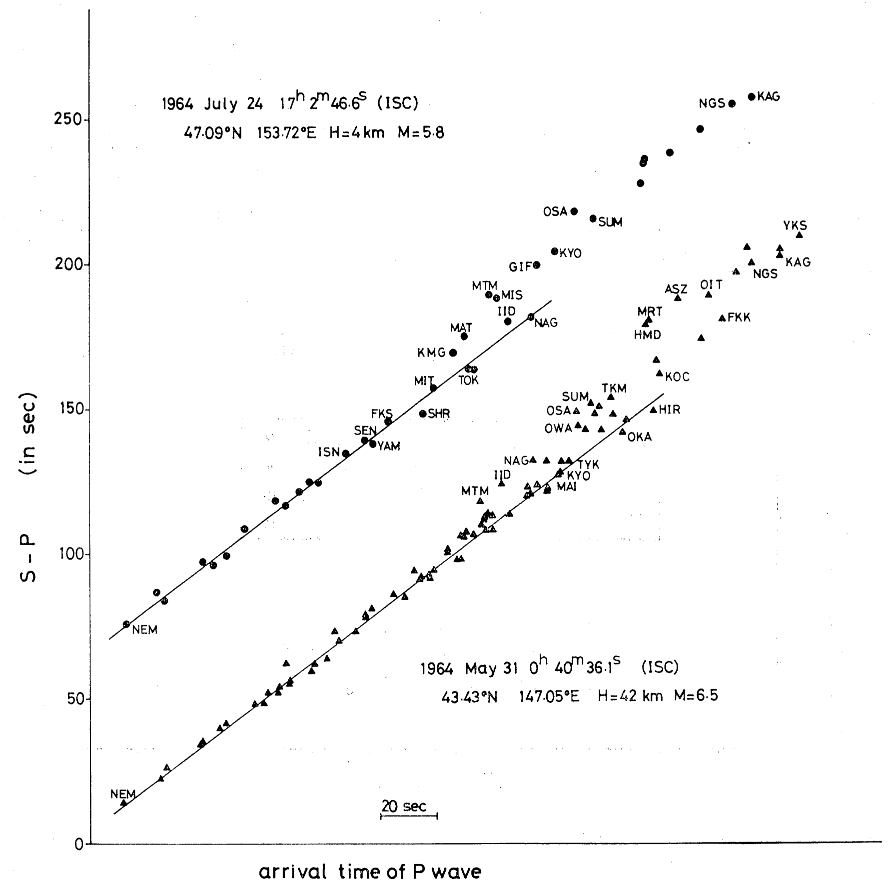

Fig. 14 Examples of the relation between *S-P* and *P* arrival for the event of shallow or intermediate focal depth in the Kurile region. The solid line corresponds to $\overline{V}_P/\overline{V}_S=1.765$.

Kurile region, most of the maxima stated above are from 150 sec to 180 sec. This range of T_{S-P} corresponds to that of epicentral distance from 14° to 17° and that of T_P from 200 sec to 240 sec. It is already pointed out that an end of a continuous curve of T_P as a function of T_{cal} is in the range from 210 sec to 240 sec (Fig. 7). The variations of the end point may be concerned with the variations of focal depth, because the epicentral distance of the end point decreases as the focal depth increases as long as the focus is more shallow than the upper surface of the low velocity layer.

OLIVER and ISACKS (1967) discussed the figures of high-Q zones in the Tonga-Kermadec region by investigating whether the high-frequency phases exist or not. UTSU and OKADA (1968) demonstrated that the anomalous structure obtained from the studies of seismic intensities or travel times also agrees well with the variations of waveform between stations in Hokkaido. Both studies prove that the variation of waveform is an indicator suggesting whether the ray has passed through the low-Q zone, in other words, low-V zone. Then, if we observe seismic waves arriving from an azimuth at a certain station and investigate the relations of the frequencies of the seismic waves to epicentral distances, we are able to know where the ray reflected with a critical angle at the upper surface of the low velocity layer appears.

Dividing an arbitrary time interval of a seismogram by the number of crests or troughs in the interval, we have a quantity which is characteristic of the interval of the seismogram. It is named "the apparent mean period".

For the events in the Kamchatka-Kurile and Aleutian regions, the values of the apparent mean period for the initial parts of the seismograms at KMU are plotted in Fig. 15 as a function of epicentral distance. As the seismographs of low and high sensitivities are set up at KMU, the apparent mean period is computed for each of them. They are indicated in Fig. 15 by the cross for the low sensitivity and the open circle for the high sensitivity. Two small open circles and x's correspond to those for the events in the neibourhood of Hachijo-jima Island.

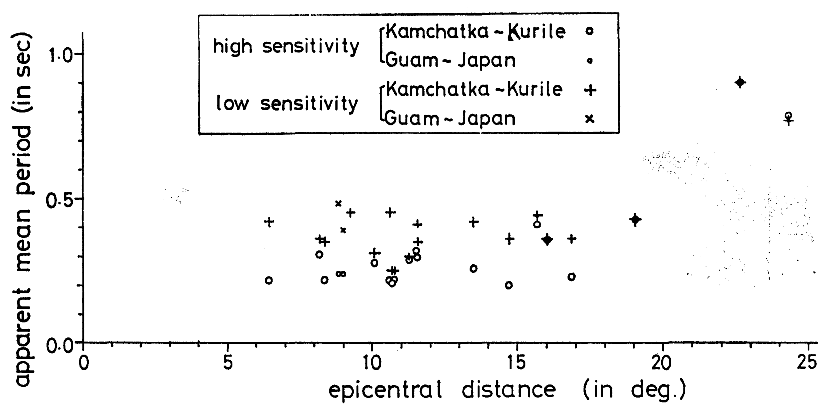

Fig. 15 Apparent mean periods calculated from the seismograms at KMU for the events of shallow and intermediate focal depths in the Kamchatka-Kurile and Aleutian regions. For reference, those for two events near Hachijo-jima Island are added to. They are calculated for each of seismograms of high and low sensitivities.

The appearances of seismograms are extremely different between the events in the Kamchatka-Kurile and Aleutian regions (Figs. 16 and 17). They are also seen in Fig. 15. No high-frequency waves in the seismograms for the events in the Aleutian region is probably due to the low velocity layer. Among the events in the Kamchatka-Kurile region, in the seismograms of the high sensitivity, the values of the apparent mean period for the events of $\Delta > 15°$ except that of December 14, 1967 are about twice of those for the events of $\Delta < 15°$. Although it is not evident whether it is related either with the attenuation proportional to the increasing distance or with the low velocity layer, it is certain that the seismograms of $\Delta < 15°$ are not affected by the low velocity layer, because the values of the apparent mean period for these are all the same as those for the events near Hachijo-jima Island.

In the seismic array composed of the Tottori Micro-earthquake Observatory and some stations, $\Delta\Phi$ for the event of shallow or intermediate depth in the Kurile region changes its sign at about $\Delta = 18°$ (Fig. 12). This is comprehended as the following: for the ray of $\Delta > 18°$, the direction of incidence is turned to the inner side by the down-going slab in the Kurile arc because the ray comes across the low-V zone under the slab if it travels in the azimuth to the array. On the other hand, for the ray of $\Delta < 18°$, which does not come to the low-V zone even if it travels in the azimuth to the array, it is turned to the outer side by the inclined Mohorovicic discontinuity or the high-V zone in the Izu-Mariana arc. Thus, in the neibourhood of $\Delta = 18°$ is the end point of the ray which is not affected by the low velocity layer.

3.3 Structure of the upper mantle

When there are discontinuities which cause triplications or cusps in the travel-time curve, the curve of $dT/d\Delta$ as a function of Δ is divided into portions corresponding to the discontinuities. Then, each portions is solved by Herglotz-Wiechert method after taking off the effects of the known structure.

The first of such discontinuities is the Mohorovicic discontinuity. The crustal structure in the northern part of the southwest Japan is analyzed by HASHIZUME *et al.* (1966) and SASAKI *et al.* (1970). In these analyses, the thickness of the crust is about 30 km if the velocity of P wave over 7.4 km/sec is peculiar to the mantle. Along the inner side of the trench in the Kurile region, it varies from 15 km to 30 km (KOSMINSKAYA *et al.* (1969), TUYEZOV (1971)). Most researches by artificial explosions in Japan demonstrate that the thickness of the crust is about 30 km (MATUZAWA *et al.* (1959), MIKUMO *et al.* (1961), HASHIZUME *et al.* (1968), AOKI *et al.* (1972)). Therefore we take the averaged thickness as 30 km. The assumed crustal structure is very similar to that of HASHIZUME *et al.* (1968).

The structure solved from the smoothed curve in Fig. 8 is designated as K-2 type. K-2-B is the model obtained by extrapolating the curve to $\Delta = 18°$. The depth of the upper surface of the low velocity layer is 170 km for K-2-A and it is 195 km for K-2-B (Fig. 18).

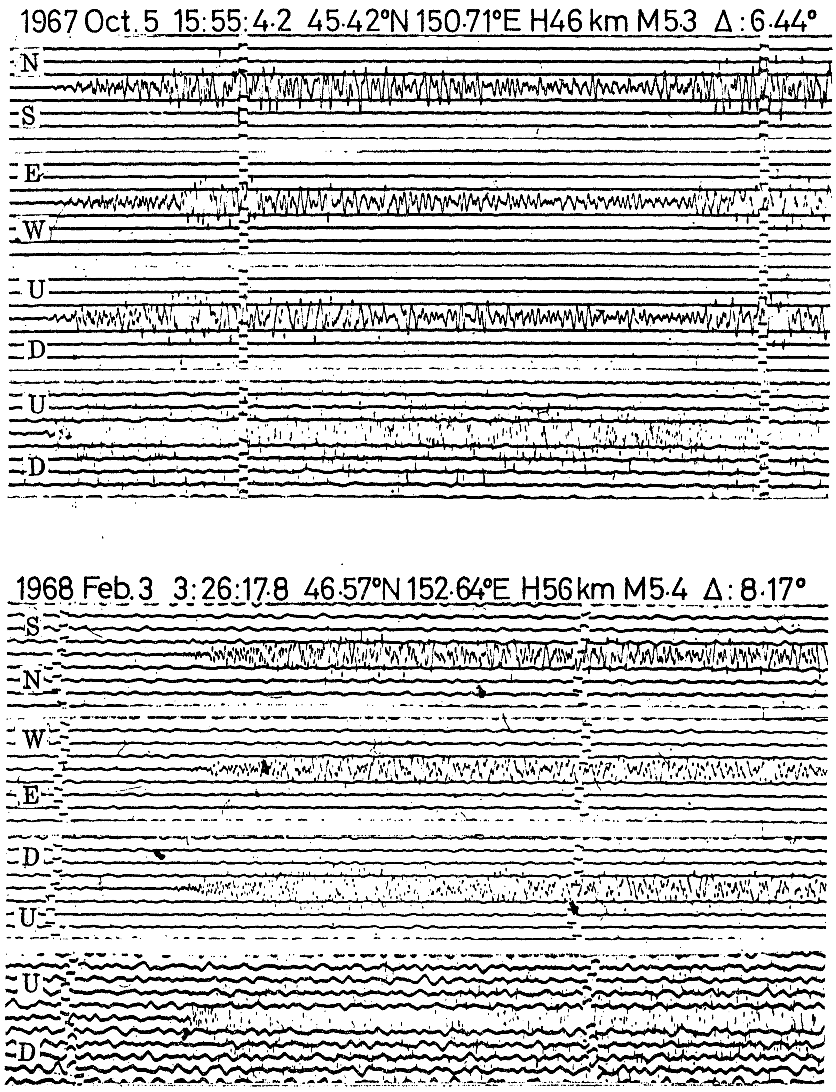

Fig. 16(a)

Fig. 16 Examples of seismograms at KMU for the events of shallow and intermediate focal depths in the Kamchatka-Kurile region. Focal parameters are those by ISC.

For the data from the superposed travel-time curves and seismic arrays, we construct the smoothing curves by the method of summary values, rejecting the scattered data and applying appropriate weights. In these treatments, the data of $10° \leq |\varDelta\Phi| < 20°$ and some of those from the Wakayama Micro-earthquake Observatory are also added with light weights. As the smoothing curves obtained by such procedures are not always decreasing functions of $\varDelta$, each of them must be revised so that the curve decreases with increasing $\varDelta$.

The inclined Mohorovicic discontinuity produces the systematic differences in the values of V and Φ. For the array composed of eight stations, that is, ABU, HKM, OYA, IZM, MZK, FNK, WKU and OIS, $\varDelta\Phi$'s tend to be positive for the events in

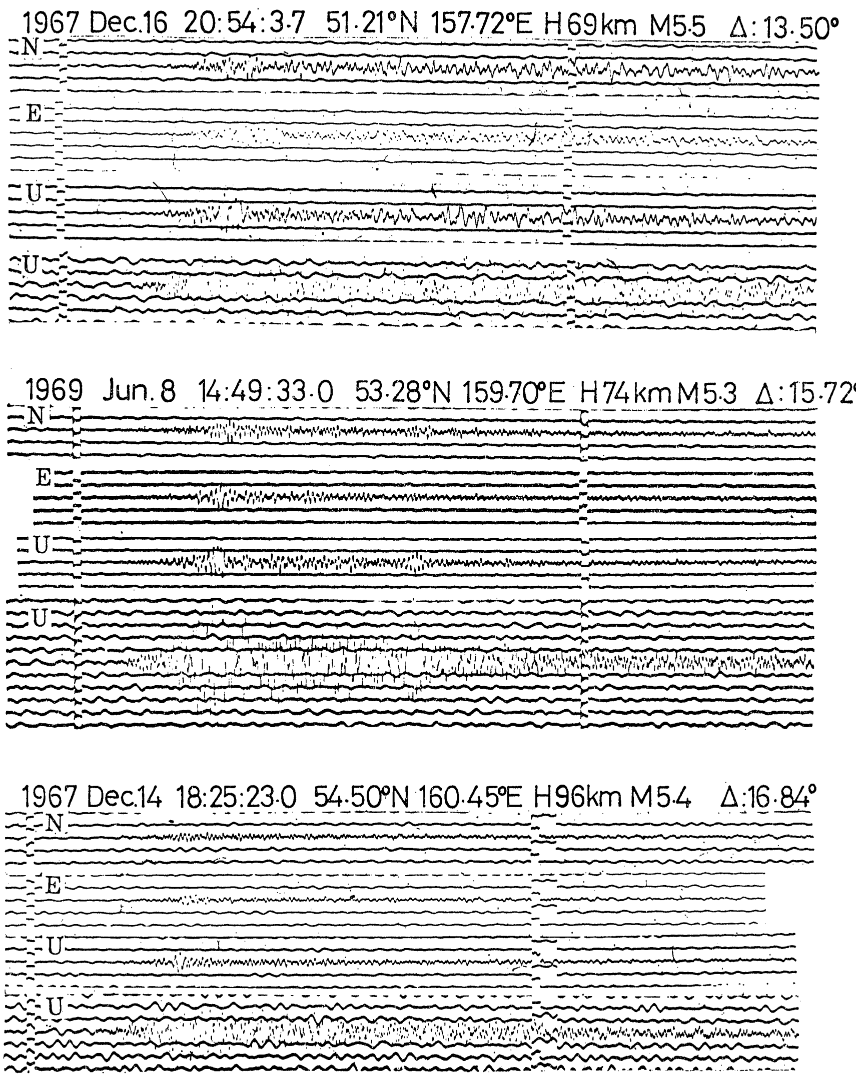

Fig. 16(b)

the northeast direction and negative for those in the southwest direction (Fig. 19). After KANAMORI (1963), the Mohorovicic discontinuity inclines to the northwest in the southwest Japan. This is elementarily consistent with the changes in the sign of $\Delta\Phi$. The strike of the inclined Mohorovicic discontinuity is about N45°E, which is also suggested from Fig. 19 because the deviations of the incident direction are maximum in the direction of the strike. When the seismic waves are incident from the direction of the strike, ΔV's vanish (NIAZI (1966)). Therefore, for this array, we take the effects due to the inclined Mohorovicic discontinuity out of consideration.

Large positive values of $\Delta\Phi$ in the observations of the Wakayama Microearthquake Observatory for the events in the northeast of the array seem to agree with the steep inclination of the Mohorovicic discontinuity which is also inferred from

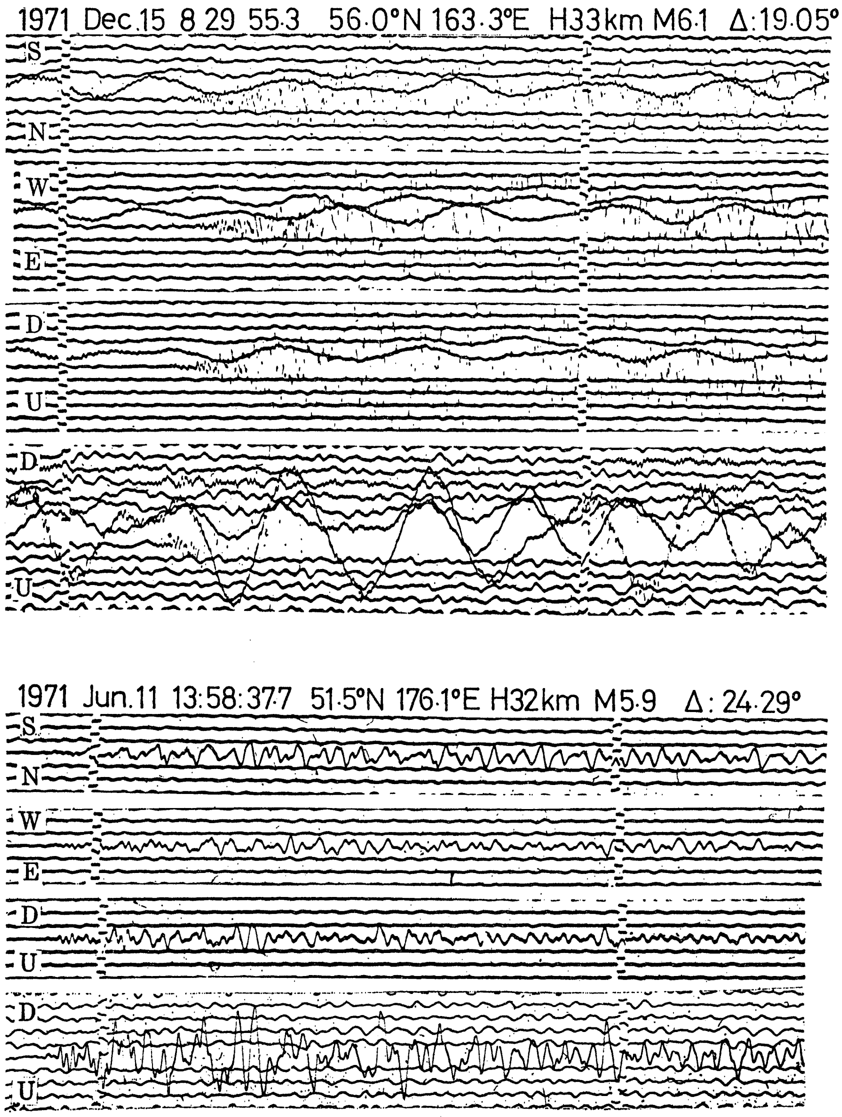

Fig. 17. Examples of seismograms at KMU for the shallow events in the Kamchatka-Kurile and Aleutian regions. Focal parameters are those by NOAA.

KANAMORI (1963). It is, however, contrary to our expectations that not only the signs but absolute values of $\Delta\Phi$ for the events in the southwest direction do not differ from the former case.

If we proceed to analyze the smoothing curves with neglect of high values of $dT/d\Delta$ at small distances, the residuals of travel times are serious. Fig. 8 also suggests that these high values of $dT/d\Delta$ are significant. Taking account of these data, we have the model of K-4 type, which coincides with a slightly inner section of UTSU's model (UTSU and OKADA (1968)) in its nature and does not contradict with the model named as Model II by HASHIZUME *et al.* (1966) and SASAKI *et al.* (1970).

For the analyses of the lower part than the upper surface of the low velocity layer,

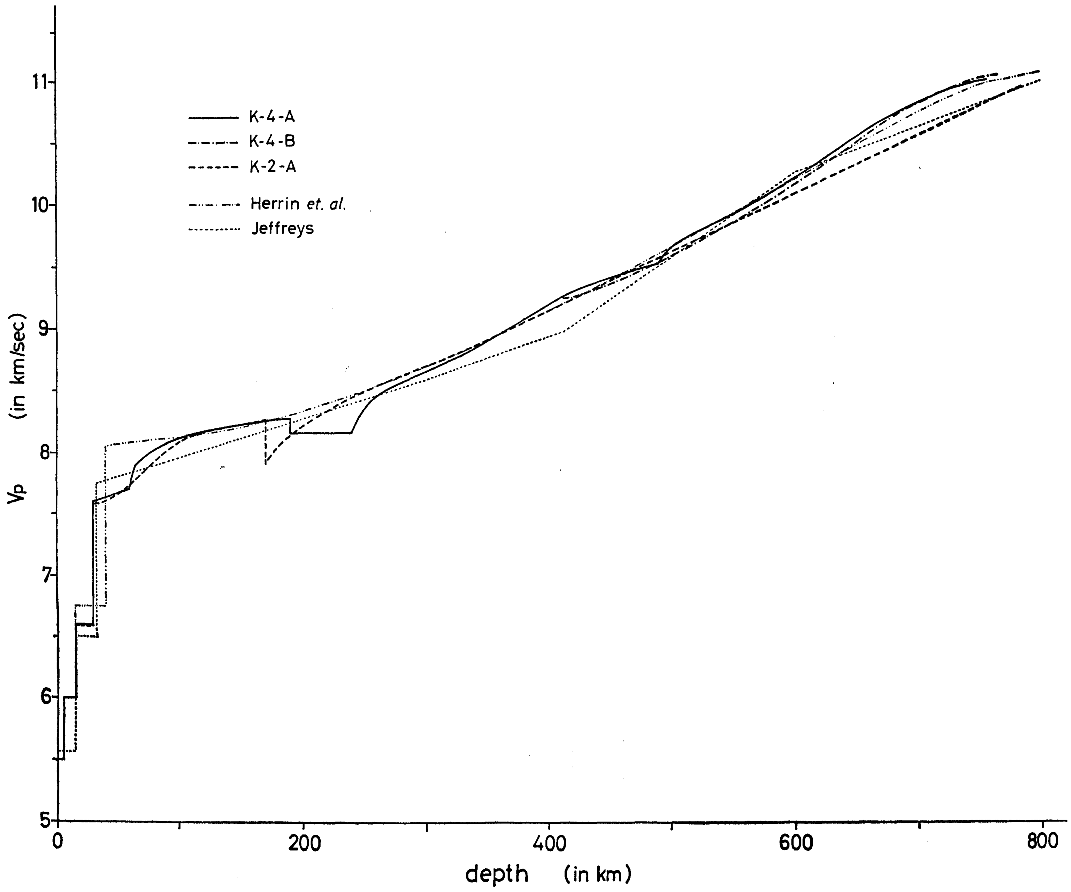

Fig. 18 The velocity structure in the section along the axis of the Kurile arc. For reference, those of HERRIN *et al.* and JEFFREYS are entered.

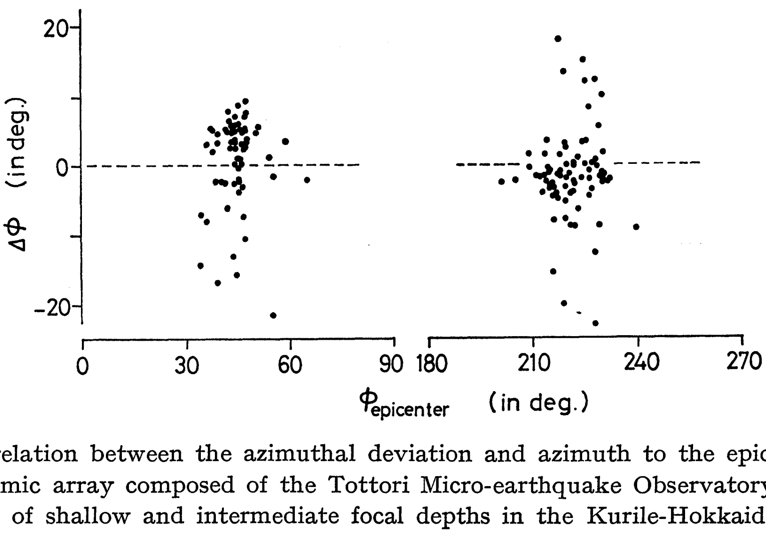

Fig. 19 The relation between the azimuthal deviation and azimuth to the epicenter observed at the seismic array composed of the Tottori Micro-earthquake Observatory and ABU for the events of shallow and intermediate focal depths in the Kurile-Hokkaido and Kyushu-Taiwan-Philippine regions.

it is the same as another cases to take off the known parts lying over the surface. There is, however, no possibility to fill up the blank in the corrected $dT/d\Delta$ curves even if later phases are investigated in detail. Then, in this case, we must assume either the velocity distribution or the shape of $dT/d\Delta$ curve corresponding to the blank. It is tested by the observed travel times and its physical possibility.

K-2-A is the model solved under the assumptions of the shape of $dT/d\Delta=a+b\Delta$ in the blank and no discontinuity in the velocity gradient. In this model, the gradient of velocity is seriously high in the uppermost part of the low velocity layer. The hasty increase of velocity may be interpreted as a result of an intensive concentration of heat produced by friction between the high-V and low-V zones. Such friction may be caused by a relative slippage of high-V zone which is carried by the convection of the low-V zone with suffering reaction against its going down, though there remains a doubt whether it is so sufficient as to cause the intensive concentration of heat.

If the low velocity layer is sustaining cooling by the high-V zone overlying it, the velocity in it has its minimum in a certain depth. When the high-V zone is moved by the mantle convection, it increases gradually from a limited value at the boundary

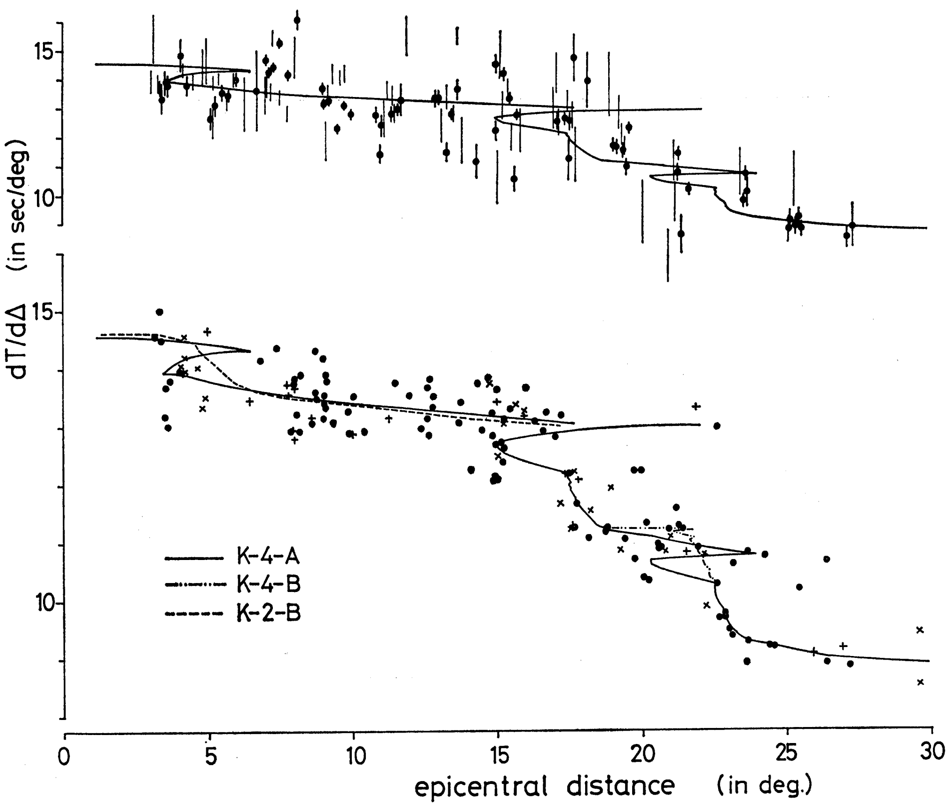

Fig. 20 $dT/d\Delta$ curves for the models of the velocity structure in the section along the axis of the Kurile arc. In the lower figure, in addition to the data in Fig. 13, those for which absolute values of $\Delta\Phi$ are less than 20° and those from the arrays of the Wakayama Micro-earthquake Observatory and JMA stations are also plotted. The data from the latter array are indicated by x and others are by cross.

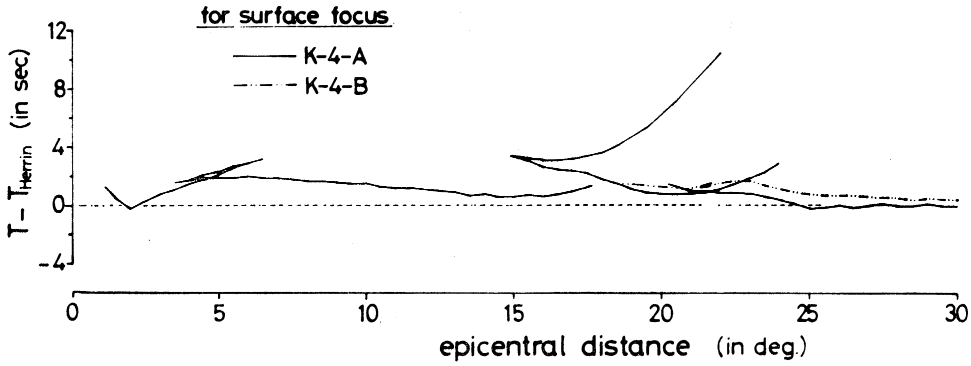

Fig. 21 P travel-time curves from which travel times of HERRIN *et al.* are subtracted based on the models of K-4 type for surface focus.

between both zones with increasing depth. In any case, gradual variation of velocity is most probable in the uppermost part of the low velocity layer.

K-4 type is obtained under the assumption that the velocity is constant in the range of depth from 190 km to 240 km. The velocity drop at the upper surface of the low velocity layer is only 0.12 km/sec, which may not increase so much because the thickness of the layer is rather thin in this model. The data in the range of epicentral distance from 20° to 25° are interpreted in two ways: that is, an averaged curve is fitted in K-4-A for the data of $\mathrm{d}T/\mathrm{d}\Delta$ between 10.5 and 11.5 sec/deg, whereas, in K-4-B, special weight is given to the data of about 11.2 sec/deg at the distance between 21° and 21.5° in disregard of the data of $\mathrm{d}T/\mathrm{d}\Delta$ in the range stated above at the distance larger than 23°.

In Fig. 20, the $\mathrm{d}T/\mathrm{d}\Delta$ curves recomputed for the models of K-4 type and K-2-B are shown with the data in Figs. 9 and 13. The crosses in the figure are the data of $10° \leq |\Delta\Phi| < 20°$ from the above-mentioned seismic array and those from the Wakayama Micro-earthquake Observatory. In this figure, the data from JMA network are also plotted. They are indicated by x, which is, however, not taken into consideration in smoothing of $\mathrm{d}T/\mathrm{d}\Delta$. The rugged $\mathrm{d}T/\mathrm{d}\Delta$ curves are resulted from that the structure is divided into layers of 5 km in thickness. In each layer, the velocity varies such that $V(r) = V(r_i)\,(r/r_i)^{\zeta_i}$, where r_i is the radius to the upper surface of the i-th layer from the earth's center and ζ_i is an constant peculiar to the layer. The travel-time curves for surface focus from which those of HERRIN *et al.* (1968) are subtracted are shown in Fig. 21 for the models of K-4 type.

IV. Travel-time residuals for the events in the Izu-Mariana and Kyushu-Taiwan-Philippine regions

KEBEASY (1969) examined travel-time residuals for the events not only in the Izu-Mariana arc but in the Ryukyu arc. He reported that such pronounced anomalies in travel-time residuals as observed for the events in the Kurile arc could not be noticed in these arcs. His examinations are, however, based on CGS parameters which are inferior to ISC ones in their accuracies. As residuals are subject to the influence of accuracies of focal parameters, they must be reexamined for ISC parameters.

4.1 Events in the Izu-Mariana arc

The Izu-Mariana arc belongs to the same arc system as the Kurile arc and has many typical features of the island arc, such as a distinguished seismic zone, a deep oceanic trench, a chain of active volcanoes and etc. (UYEDA and SUGIMURA (1970)). Anomalous distributions of seismic intensities which closely correlate with travel-time anomalies are also observed for the deep events in this arc (UTSU (1966)). Then, the travel-time anomalies similar to those shown in Fig. 7 may well be expected for this case. Indeed, YAMAMIZU (1971) related travel-time anomalies at Chichi-jima, the Bonin Islands, with the high-V zone in this arc and KEBEASY (1969) noted that the negative residuals are observed at TSK and DDR for $\Delta \leq 23°$.

P travel-time residuals based on ISC parameters at TSK, KMU and YSS are shown in Fig. 22 as a function of T_{cal} (J-B). These are very similar to the relations in Fig. 7 though the data are fewer than the latter.

It is noteworthy that such a relation as shown in Fig. 22 is observed at YSS. Such a relation may be explained if we assume that the high-V zone in the Izu-Mariana arc extends to Sakhalin without turning at Hokkaido to the Kurile Islands. Thus, this relation suggests that some modification of a general notion concerning the figure of the high-V zone may possibly be needed, although high attenuations of seismic waves at SAP, ASA and WAK are not consistent with such an assumption without any complicated conditions.

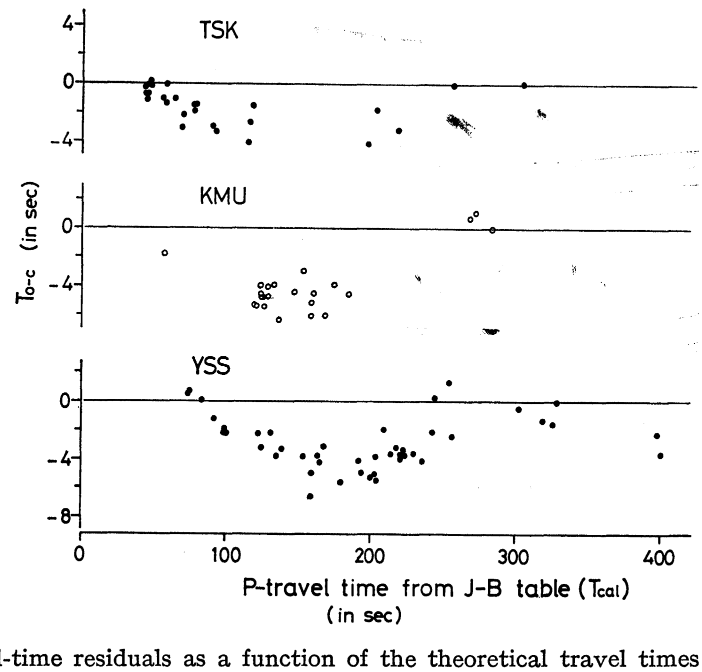

Fig. 22 P travel-time residuals as a function of the theoretical travel times for the events of shallow and intermediate depths in the Izu-Mariana region. YSS is the station in USSR.

4.2 Events in the Kyushu-Taiwan-Philippine region

In the west Japan arc system which the Ryukyu arc belongs to, the features of the island arc are not so remarkable as those in the east Japan arc system (UYEDA and SUGIMURA (1970)). Nevertheless, KATSUMATA and SYKES (1969) found a thin planar seismic zone which dips 35° to 45° northwest to a depth of about 280 km in the Ryukyu arc and UTSU (1969) showed the existence of anomalous distributions of seismic intensities for the deep events in the west Japan. Thus, for the events in

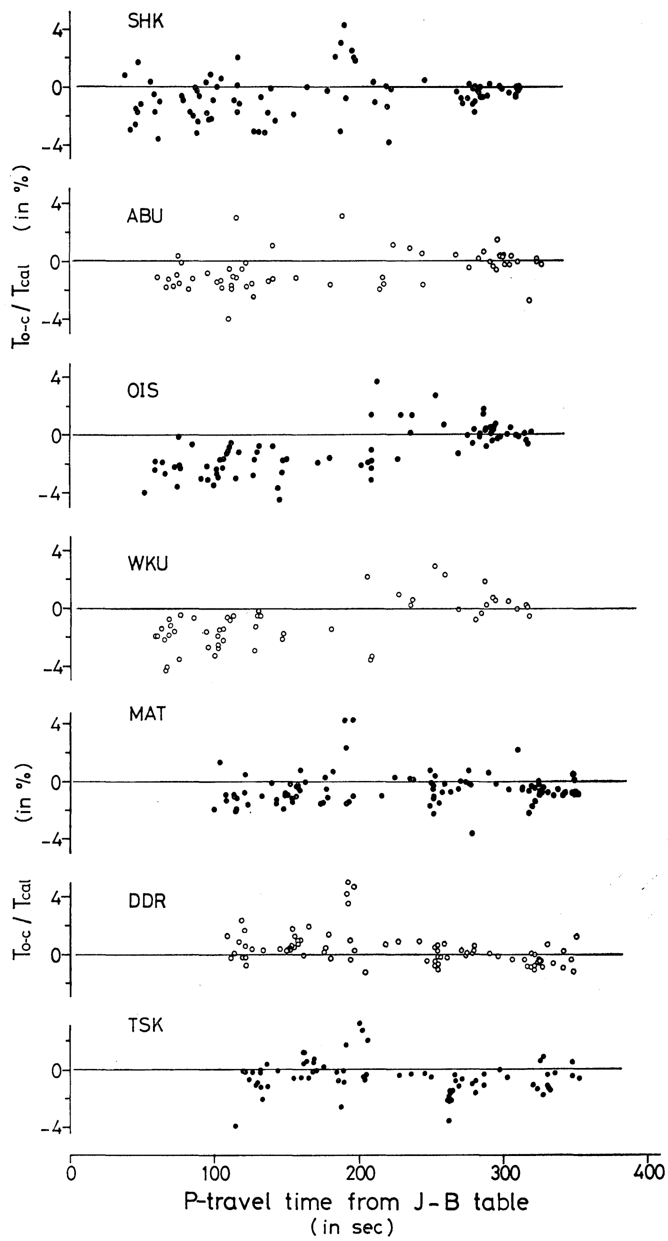

Fig. 23 The relation of P travel-time residuals to the theoretical times for the events of shallow and intermediate focal depths in the Kyushu-Taiwan-Philippine region.

this region, travel-time anomalies which are closely connected with the existence of the high-V zone may be observed.

Fig. 23 in which the ratio of T_{o-c}/T_{cal} is plotted as a function of T_{cal} shows that similar relations to those in Fig. 7 are distinctly visible at ABU, OIS and WKU. At SHK, most residuals are negative for $T_{cal} \leq 220$ sec and there is an evident difference between the mean values of T_{o-c}/T_{cal} for the range of T_{cal} below and over 220 sec. Such variation of the mean value with T_{cal} is almost the same as those at ABU, OIS and WKU. The difference between the mean values for the range of T_{cal} below and over 220 sec is small at MAT, but the tendency of variation of the mean value is not so different from those at the former stations. These all may be concerned with the existence of the high-V zone in the Ryukyu arc.

On the other hand, at TSK and DDR, each curve of T_{o-c}/T_{cal} shows a slight decrease with the increasing of T_{cal}. This is quite different from the former cases.

The seismic rays which come to TSK and DDR are emitted from the focus in the Ryukyu arc in a fairly different direction from the strike of the arc. Thus, if there is no thick high-V zone in the southwest Japan arc, travel-time residuals at TSK and DDR may be positive or, at the utmost, slightly negative. In the fact, such a thick one may be deniable, because attenuations of seismic waves are high at TSK and DDR for the events in the Ryukyu region (KEBEASY (1969)) and no deep events related with the southwest Japan arc have ever been observed.

The residuals for large T_{cal} are negative at TSK, DDR and MAT. These are probably concerned with the deep thrusting high-V zone in the Izu-Mariana arc. Then, the decreases of T_{o-c}/T_{cal} with the increase of T_{cal} at TSK and DDR indicate that the effects of the zone on seismic rays increase with increasing epicentral distance.

For all stations, the averaged curves of T_{o-c}/T_{cal} are nearly continuous to T_{cal} of about 220 sec. This means that the discontinuities of these curves do not result from that the arc bends in the neibourhood of Taiwan.

V. Comparison of various models

Some models of the velocity structure in a certain section of the Kurile arc are obtained by Herglotz-Wiechert method in Chapter III. They are relatively suitable for the $dT/d\Delta$ data but not always fit for the observed travel times. The $dT/d\Delta$ data are determined from the observations at some appointed seismic arrays or the superposed travel times for the events of various depths. Then the epicentral distances where they are observed must be corrected to the values for the surface focus under an appropriate assumption of the structure. These are the reasons why the unfitness of such models for the observed travel times should occur.

In this chapter, to know which model is better for the structure in each arc, we examine to what extent various models are fitted for the observed travel times. At first, the residuals of the superposed travel times are examined. This is an ordinary way of examination. Next, a new method is introduced. In this method, the existence of the station correction is taken into consideration and, therefore, it is not always a necessary condition that the residuals are small.

5.1 Residuals of superposed travel times

For each model, the residual curves of the smoothed travel times which are obtained by the method of summary values are exhibited in Fig. 24 for the events of various depths in the Kurile region. In any case, it is not seen in the figure that the curves of different focal depths separate with each other. Thus it is inferred that each model does not deviate highly from the actual velocity distribution. These travel times are derived from the observations at micro-earthquake observatories and JMA stations. If we take them as the averaged travel times for the events in the Kurile region, the travel-time curves of HERRIN *et al.* must be increased by about two seconds within the epicentral distance of 17°. For the model of JEFFREYS (see KANAMORI (1967)), the residual curves decrease gradually with the increasing distance within 15°. Then we must decrease the gradient in travel-time curve for this model.

It seems from Fig. 24 that K-2-B is the better fitted model. If we separate the data of micro-earthquake observatories from those of JMA stations, we find that K-4-A is fit for the data of micro-earthquake observatories (Fig. 25). For this model, the

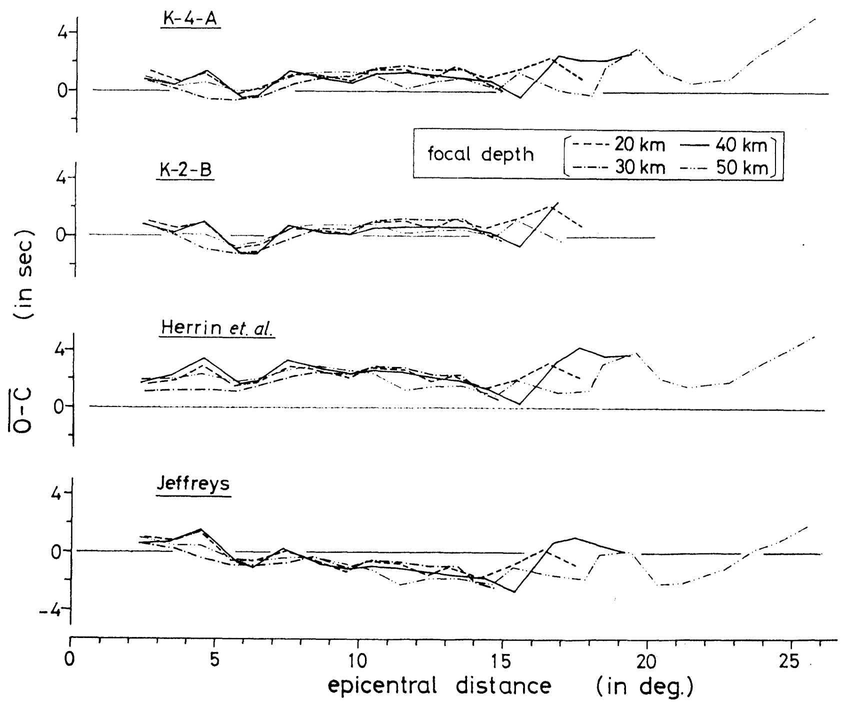

Fig. 24 Smoothed travel-time curves obtained by the method of summary values from the data of JMA stations and micro-earthquake observatories for the events of nearly equal focal depths in the Kurile-Hokkaido region. They are shown as residual curves for each model.

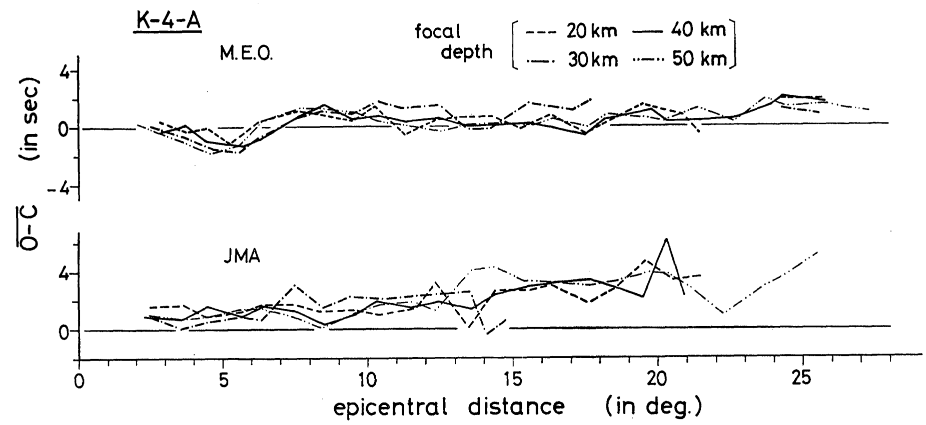

Fig. 25 The residual curves of smoothed travel times for K-4 models. Upper: for the data of micro-earthquake observatories. Lower: for the data of JMA stations.

travel-time residuals at JMA stations are positive and increase as the epicentral distance increases. These probably come of low sensitive seismographs of JMA stations. Then it appears to us that the delay of travel time increases with the increasing epicentral distance for JMA stations, because the onset of the initial phase is more missed owing to its fading out as the distance increases.

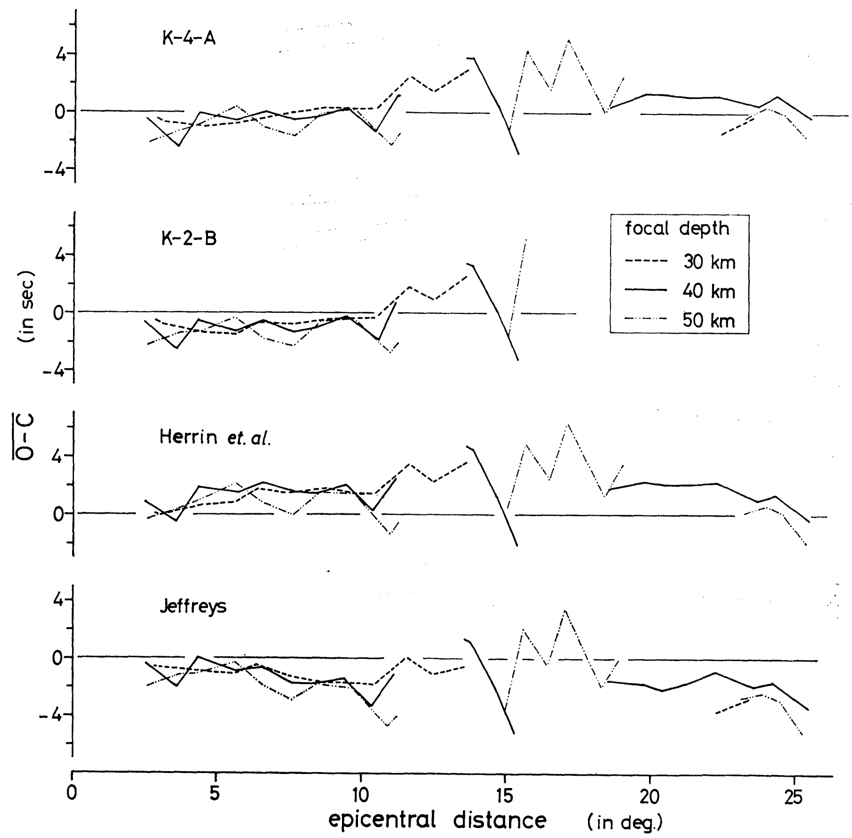

Fig. 26 The residual curves of smoothed P travel times for various models from the observations of the events of nearly equal focal depths in the Kyushu-Taiwan-Philippine region.

For the Kyushu-Taiwan-Philippine and Izu-Mariana regions, any models of the velocity structure are not derived because of insufficient data. Then the construction of some model is postponed to future. Nevertheless, in order to form some idea of the structure, it may be useful to examine the residuals for each model mentioned above.

The residual curves for the events in the Kyushu-Taiwan-Philippine region are shown in Fig. 26. These curves are so rugged in the distance from about 13° to 19° that examinations must be restricted to the residuals within the distance of 13°.

Only from the mean value of residuals, it is inferred that K-4-A is better. But residuals increase with the increase of distance. On the other hand, for the model of JEFFREYS, they decrease slightly as the distance increases. In the distance between about 5° and 10°, they are nearly constant for the models of K-2-B and HERRIN *et al.* Thus it may be a short way for construction of the model corresponding to the range of distance for this region to improve on the $dT/d\Delta$ curve for K-2-B or HERRIN *et al.*

For Izu-Mariana region, not only events but observatories distributed along the axis of the arc are too insufficient to derive superposed travel times for various focal depths.

5.2 Scrutiny of various models under the existence of station anomalies

In a usual manner to know whether a model is fit for the structure, they are inspected whether the distribution of the residuals for the model is represented by the normal distribution and whether the residuals vary systematically with epicentral distance. If there are station anomalies, such a method is not applied without any knowledge of station corrections, because of no criterion for the suitability of the model. In this case, it is required to know the values of station corrections at the same time when the model is examined.

When there is no error in focal parameters, the residual for the i-th event at the j-th station is expressed in the form of

$$(T_{\mathrm{o-c}})_{ij} = -(\delta V/V_c)(T_{\mathrm{cal}})_{ij} + \gamma_j ,$$

where $(\delta V/V_c)$ is the mean deviation ratio from the assumed velocity distribution in the greater parts of ray paths. If the assumed model is perfectly fitted for the structure, $(\delta V/V_c)$ vanishes. Then, solving simultaneous equations of this type, which are already given in (2.4.3) in a matrix expression, for some sets of events and observatories, we are possible to know whether the model is appropriate for the region. That is, it is a criterion for the suitability of the model whether the value of $(\delta V/V_c)$ is small or not.

We must pay attention in this method to that the solutions are not arbitrary and that each of γ_j is constant. The former will be satisfied if a reasonable extent of seismic zone is set up, and the latter stands if the values of $dT/d\Delta$ at each station are kept constant for all events in the extent of seismic zone.

For three limited seismic zones in each of the Kurile and Ryukyu regions and a narrow zone in the Izu-Mariana region, this method is applied (Fig. 27). The data of

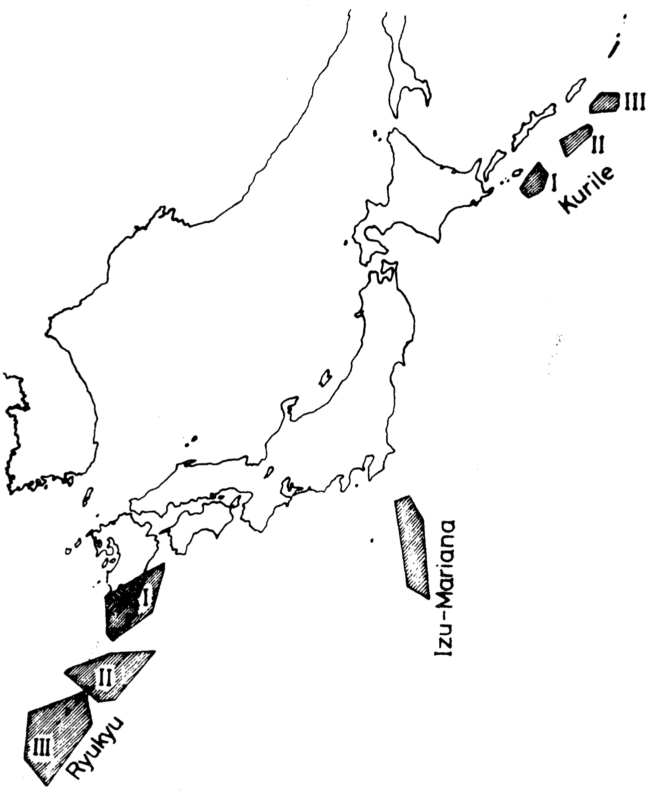

Fig. 27 The seismic zones for which inspections on the suitability of various models are made by the new method.

Table 4. Deviation ratio for each model derived from some sets of events and observatories.

Seismic Zone	HERRIN *et al.*	JEFFREYS	K-2-B	K-4-A
Kurile I	2.7±1.4%	4.0±1.3%	0.9±1.4%	1.2±1.4%
II	−2.4±2.3	−1.0±2.3	−2.3±2.3	−2.2+2.3
III	6.4±1.9	6.2±1.9	5.1±1.9	5.1±2.0
Ryukyu I	−4.6±0.7	−2.3±0.6	−3.8±0.6	−4.9±0.6
II	−2.8±1.3	−0.9±1.3	−2.8±1.2	−3.0±1.2
III	3.7±1.0	3.4±1.1	2.2±1.1	2.4±1.0
Izu-Mariana	−0.3±0.7	1.5±0.7	0.8±0.8	0.2±0.8

T_{cal} from 60 sec to 240 sec are used in these examinations. These correspond to the rays of which greater parts are in the high-V zone. The value of $(\delta V/V_c)$ for each model is tabulated in Table 4 with its standard deviation for every seismic zone. When there are no errors in focal parameters and readings of initial phases, it is expected that the values of $(\delta V/V_c)$ do not vary in a region if the model is fitted for the region. Contrary to such expectation, they vary from zone to zone even if the zones belong to the same region. Such variations may probably be concerned with errors in focal parameters or reading errors. These errors have much influence on the results when the extent of seismiz zone is improper or only a small number of events or observatories are used in computations. Nevertheless, the values of $(\delta V/V_c)$ in Table 4 are not so anomalous and the pattern of their variations from zone to zone is not so different if the model is exchanged. Thus it holds still now as a criterion for the suitability of the model that the absolute value of $(\delta V/V_c)$ is tiny.

Then K-2-B and K-4-A are more suitable for the Kurile region than others though their suitabilities are not necessarily satisfactory. For the Izu-Mariana region, K-4-A and HERRIN *et al.* are most adequate. As not only the shallow events in this region but observatories distributed along the axis of the arc are not so abundant, we cannot always be satisfied with these results. For all that, these tiny values suggest that the velocity distribution in the high-V zone in this arc is very similar to K-4-A or HERRIN *et al.* For the Ryukyu region, JEFFREYS' model is superior to any other model.

VI. Summary

It is thought to be resulted from the anomalous structure in the upper mantle by many authors that JMA epicenter always locates on the oceanic side of ISC epicenter for the event near and in Japan. The difference between JMA and ISC epicenters seems, however, to be to a considerable extent concerned with another type of travel-time table being employed. If we relocate JMA epicenter by using a straight line which is accepted as a simplified travel-time curve, the more the inverse of its gradient increases, the more the relocated epicenter is on the continental side. Moreover, the relocated epicenters for an event are distributed in a direction nearly orthogonal to the axis of the island arc. The same explanation applies to why an epicenter moves systematically in a certain direction as the focal depth is varied.

For the shallow events in the Kurile region, it is often noticed that the travel-time curves of "*S-P*" are separated one another. The arithmetic mean of *S-P* residuals for the part of *S-P* which is represented by a nearly linear function of *P* arrival is usually negative and increases with the increasing epicentral distance to the gravity center of the observational network except less than 700 km. Such a tendency is not seen for ISC epicenter. This is peculiar to JMA epicenter and is resulted from improper weights for *P* and *S* waves in location.

At a station which is in the direction along the axis of the island arc, the travel-time residual for the shallow event in the arc is represented as a function of the theoretical *P* travel time in a certain range of epicentral distance. For Δ less than about 17°, the relation is expressed by $(T_{o-c})_i = -(\delta V/V_c)\ (T_{cal})_i + \gamma$, where i is the suffix for the event. This is comprehended in the qualitative sense if we assume the upper mantle structure similar to UTSU's model which is composed of the low-V and high-V zones. The computed value at a station by the method of least squares is not usually concordant with that at another station. This probably comes from errors in focal parameters or from unsuitability of the standard velocity structure. Then, errors in focal parameters are investigated by a similar method to the joint epicenter method by DOUGLAS (1967). Our method is, however, different from the latter in having the term of $(\delta V/V_c)$. These simultaneous linear equations are solved by the method of conjugate gradients, which is proper to find the solutions of these types. Computations are made for the Kurile, Izu-Mariana and Kyushu-Taiwan regions by setting up some sets of events and observatories. These results show that anomalously high errors in focal parameters are frequently contained and that the value of $(\delta V/V_c)$ for one set differs from that for another even if the events of both sets belong to the same region. Thus it is inferred that the standard velocity structure is under the necessity of being modified.

For the Kurile region where many shallow events occur, the analyses are performed by Herglotz-Wiechert method. The used data are travel-time differences between two stations of which azimuths are equal on the bulletins of ISC, differences between the first and second summary points (ARNOLD (1968)) of the superposed travel times, and the seismic array measurements. The structure model obtained by the analyses is only a section of the anomalous mantle because of restrictions imposed by the method, but this may be useful for the comprehension of the entire image of the upper mantle by combining this with another analyses.

In the last part, various models are inspected by two types of methods. One of them is applicable even if there exist station anomalies. K-4 type is better fitted model for the high-V zone in the Kurile and Izu-Mariana regions, though it needs further revisions.

Acknowledgments. This work was performed at the Department of Geophysics, Faculty of Science, Hokkaido University by the support of Japan Society of Promotion of Science. The author expresses his heartfelt thanks to Drs. T. UTSU and I. YOKOYAMA who provide much convenience to him in his work. He also thanks Messrs. M. SEINO and Y. MOTOYA for their kind helps to his collection of data. The numerical computations were carried out by FACOM 230-60 at the Hokkaido University Computing Center (Problem No. 1001CJ0355).

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