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Anomalies in Seismic Wave Velocity and Attenuation Associated with a Deep Earthquake Zone (I)

Tokuji UTSU

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Abstract

A travel time study of P and S waves from both near and distant deep earthquakes indicates the appreciable anomalies in wave velocities which seem to be associated with the inclined seismic layer extending from the earth's surface near the Japan-Kurile Trench to a depth of several hundred kilometers under the Seas of Japan and Okhotsk. This layer transmits seismic waves with higher velocity and smaller attenuation as compared with the adjacent portion of the upper mantle at comparable depths. A rough estimate of the velocity difference between the two portions of the upper mantle is 6% for both P and S waves. This fact may be important in the discussion of the composition, state, and processes occurring in the upper mantle.

1. Introduction

From an analysis of the distributions of seismic intensities due to large earthquakes in the vicinity of Japan, Utsu[1] concluded that the abnormally high seismic intensities observed on the Pacific coast side of northeastern Japan in the case of all large deep earthquakes and some large shallow ones are caused by a zone of relatively small absorption which extends in the upper mantle where absorption is generally high especially for S waves. It is a remarkable fact that this peculiar zone coincides with the seismic layer where intermidiate and deep earthquakes take place.

The fact that the seismic zone transmits seismic waves with relatively small attenuation was already pointed out by Okura[2] and Katsumata[3]. Utsu[1] made an estimate of Q-value contrast between this peculiar layer and the adjacent portion of the upper mantle at comparable depths.

Most deep and intermidiate earthquakes are associated with island arcs and other arcuate structures around the Pacific Ocean. They occur in dipping layers which make angles of 30 to 60 degrees with the earth's surface reaching depths as great as 700 km below the surface. On the other hand the almost world-wide existence of a low velocity layer in the upper mantle has been an established fact. Since the high attenuation of seismic waves in the upper mantle seems to be associated with the low velocity layer, the relatively high velocity is expected in the seismic layers.

It is supposed that these inclined seismic layers cut the low velocity layer as shown in Figure 1, A. However it is also conceivable that earthquakes occur in the marginal part of the less absorptive region as shown in Figure 1, B. It is impossible to tell which is the more realistic model from the seismic intensity observations in Japan.

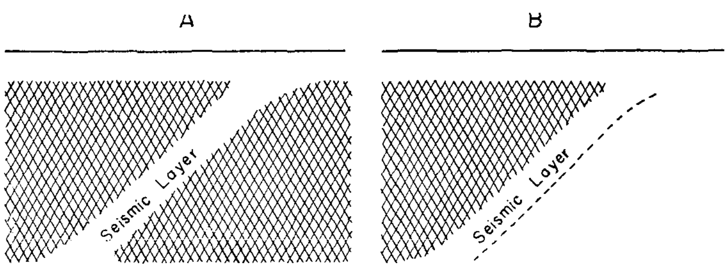

Fig. 1. Two types of the upper mantle model with a deep seismic zone. Hatchings indicate low-velocity and high-attenuation zones.

In this paper a preliminary study has been made on the regional differences in seismic wave velocity and attenuation connected with the deep seismic zone under northern Japan. The data used are mainly from the seismic observations by the Japan Meteorological Agency.

2. Some results from previous studies

Regional anomalies in travel time of P waves in Japan were investigated by many seismologists in the past forty years.[3)-17)] Most authors considered that the travel time residual (the actual arrival time substructed by the estimated arrival time from a standard (or smoothed) travel time curve) at each station is attributable to local velocity variations beneath the station. According to this idea the residual is approximately constant rather than proportional to the travel time for each station. Until recently no significant regional variations in seismic wave velocity in the upper mantle were recognized. The variations in the thickness of the crustal layers had been considered as the main cause of the travel time anomalies. Wadati[5)] who found a close relationship between the abnormal distributions of seismic intensities and travel time anomalies] in Japan tried to explain these phenomena based on this hypothesis.

Although the results obtained by most authors agree only very approximately, it is to be noted that the P waves arrive earlier by a few seconds at stations in the eastern part of Tôhoku and Kwantô districts. However most of these studies seems to involve the following difficulties.

(1) Owing to the low sensitivity of seismographs in use, the initial motion of P arrival which is usually very small in amplitude except in near-source region or in the case of large deep earthquakes often failed to be recorded. The timing errors due to irregular movements of chronometers and recording drums were not so large (less than 1 second) at most stations. They are random in nature and can be eliminated to some extent by taking the average of many events. The most authors assumed that the errors in travel time residuals are introduced by timing and not by misinterpretation of the initial motions.

According to independent studies by Haseba[15] and Iizuka[17] who used the JMA data after the establishment of the Matsushisro Seismological Observatory, the mean residual at Matsushiro is negative in contrast with the positive residuals at surrounding stations. If the high sensitivity of the instruments at Matsushiro accounts for this fact, it is suggested that the surrounding stations sometimes could not record the true initial motions. There is a possibility that the early arrivals generally observed in the anomalous zone of seismic intensity may be due to the relatively large amplitude of short period oscillations in P waves which is easily recorded by seismographs with mechanical magnification.

(2) The accurate epicenter location is needed to find travel time residuals at each station, but it is usually difficult except under favorable conditions. If the method of least squares or similar technique is used in determining epicenters assuming horizontally uniform crustal structure, epicenters are fixed at such points that the travel time residuals are in a sense minimized. If large scale horizontal variation in crust-mantle structure exists, the least-square epicenter may differ systematically from the true epicenter and the residuals based on the least-square epicenter may be smaller than the true residuals. A good example is the case of Gnome explosion in which the least square solution gives an epicenter about 16 kilometers apart from the actual explosion point.[18] This is due to the regional variations in P_n velocity rather than the observational errors.

Regional variations in P_n or S_n velocity within Japan have not been confirmed from explosion-seismic observations or surface wave studies. However, Morita[19] noticed that S waves from a deep earthquake in central Japan traveled with higher velocity and smaller attenuation under north-eastern Japan than under southwestern Japan. Similar result was obtained by Katsumata[3] who studied many deep earthquakes in central Japan. Hisamoto[20] found a high velocity region for S waves off the Pacific coast of northeastern Japan. These studies indicate the regional variations in S wave velocity in the upper mantle.

As explained in a previous paper[1], abnormal distributions of seismic intensities are typical manifestations of the regional differences in seismic wave attenuation in the upper mantle. In Figure 2 four examples of the intensity distributions of deep earthquakes are shown. All these earthquakes are used in travel time studies in a later chapter. Another source of evidence for regional variations in attenuation was provided in the course of earthquake magnitude studies.

Hayatu[21] described a systematic geographical distribution of the values of α and γ in the formula for calculating magnitude from the maximum amplitude A recorded at an epicentral distance Δ

$$M = \alpha \log \Delta + \log A + \gamma .$$

He determined the values of α and γ for 34 stations in Japan using data on shallow earthquakes. Since the variations of α and γ from station to station are much larger than those expected from statistical fluctuations[21], they are to some extent connected with regional vairations in attenuation. It is interesting that α-values at Tôhoku stations determined from data on earthquakes occurring east of the Morioka-Shirakawa line are substantially smaller than the values determined from the whole data (for details see also Tsuboi[23]).

Inouye[24], in his study of magnitude determination for deep earthquakes from near-by observations, obtained the station correction C for 74 Japanese stations. He made a map showing the distribution of $C-\Delta M$, where ΔM is the magnitude correction to be used for shallow earthquakes obtained by Ichikawa[25], According to this map stations in the eastern part of Honshû and in Kyûshû have generally negative values. This indicates that the seismic waves (perhaps S waves) reaching these stations suffer less attenuation than the average.

It has been noticed among seismogram interpreters in Japan that body wave phases recorded at certain stations contain less high-frequency oscil-

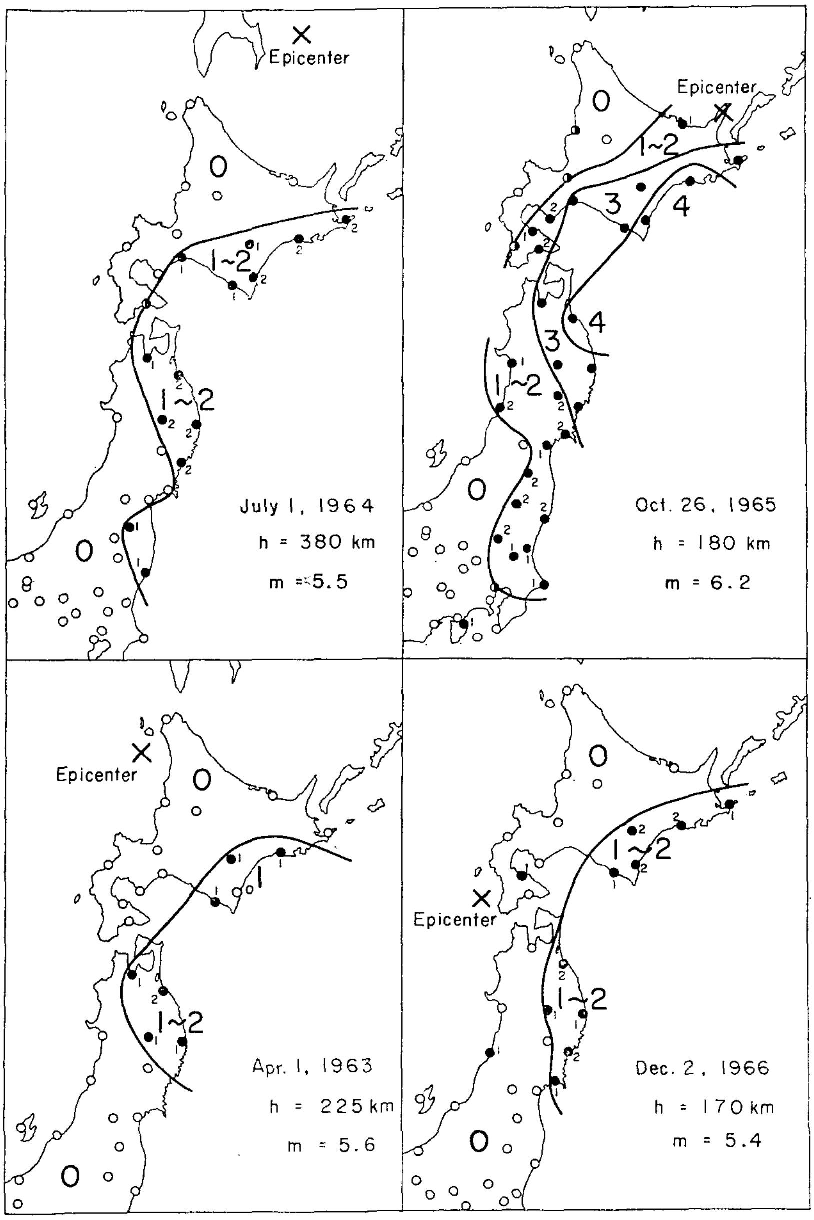

Fig. 2. Distributions of seismic intensities for some recent deep earthquakes.

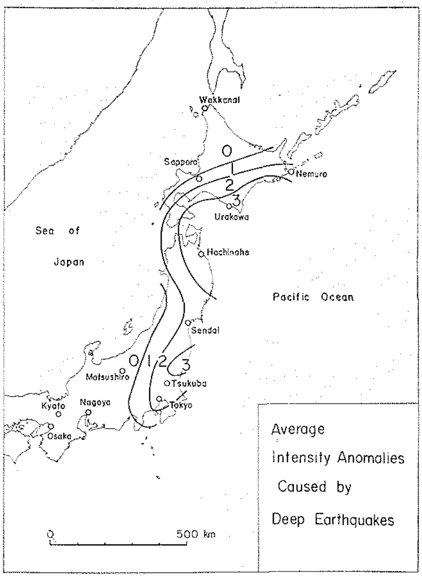

Fig. 3.

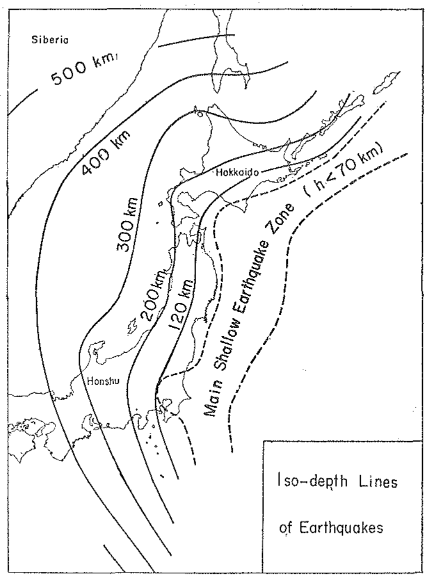

Fig. 4.

Fig. 3. Average seismic intensity anomalies for deep earthquakes.[1]
Fig. 4. Average focal-depth of earthquakes belonging to the deep seismic zone.[29)-30]

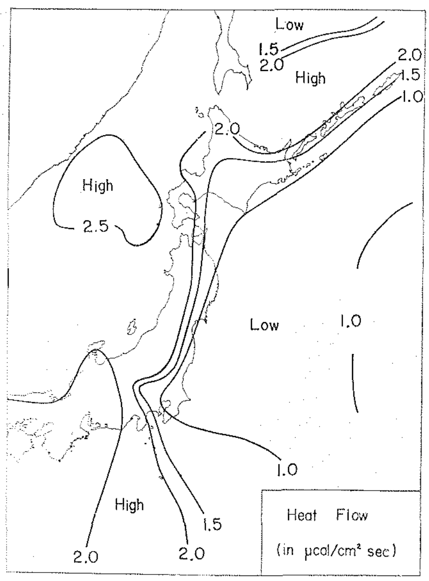

Fig. 5.

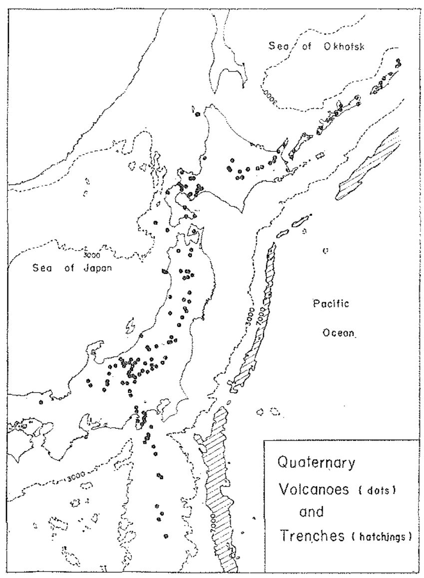

Fig. 6.

Fig. 5. Distribution of heat flow values.[31)-33]
Fig. 6. Distributions of Quaternary volcanoes and trenches.

lations than those recorded at some other stations. Asada and Takano[26] compared the frequency spectra of P waves from distant earthquakes recorded at Tsukuba and Matsushiro and found increased attenuation of high-frequency waves under the latter station. A similar result was obtained by Okada[27] who compared the spectra of P and S waves from two groups of earthquakes near Hokkaidô recorded at Urakawa and Wakkanai. Attenuation in high-frequency range is quite large at Wakkanai. These results indicate the lateral variations in attenuation and is in accord with intensity anomalies shown in Figure 3 considering the locations of these stations shown in the same figure.

3. Travel time anomalies for distant deep earthquakes

In 1960 the Japan Meteorological Agency began to replace old seismographs by new electromagnetic seismograph systems. The system includes three-component seismometers having natural period of 1.5 seconds connected directly to highly damped galvanometers having natural period of 3 cps. The motion of the galvanometers are recorded photographically on 35 mm films. The magnification of this instrument in the frequency range 1 to 10 cps is 3000 to 6000 when measured on the microfilm viewer. Both time signals and revolution of recording drums are crystal-controlled. The time accuracy of this instrument is ± 0.1 second and errors exceeding 0.3 second seledom occur for impulsive signals.[28].

The introduction of this system has improved the accuracy of data remarkably. However data on deep and intermidiate earthquakes are used in this paper, since initial motions of P and S waves from shallow origins are still apt to be missing. First, distant deep earthquakes observed at stations in Hokkaidô are studied for travel time anomalies. As seen in Figures 3–6[29)–33] geophysically interesting features which seem to be correlated one another are well developed in and near Hokkaidô, and 8 stations among the 15 stations in Hokkaido have been equipped with the new instruments since 1962–63. One station Hachinohe in northern Tôhoku which also has the new instruments is added.

A dozen of earthquakes during 1963–66 are selected as suitable for this study. They fall into three groups as listed in Table 1. Epicentral distances and azimuths to these earthquakes from the 16 stations are calculated, and the arrival times of P and S waves are plotted against the distance for each earthquake. The errors in epicenter locations do not affect the configuration of the travel time plots seriously since the small change in epicenter location

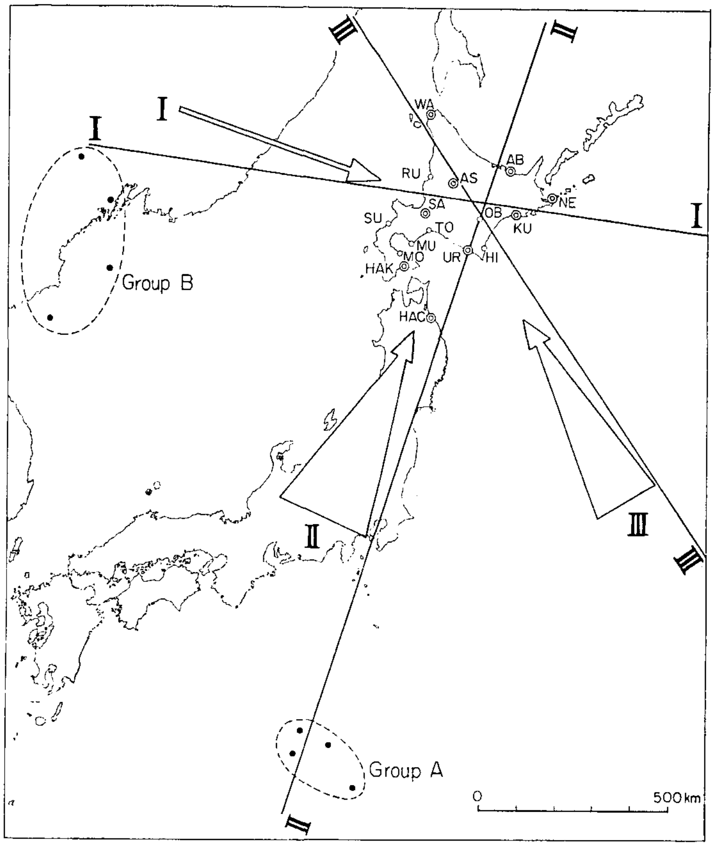

Fig. 7. Map showing the directions of approaching waves from earthquakes in groups I, II, and III, locations of the three profiles I, II, and III in Fig. 8, locations of earthquakes in groups A and B, and stations used in this study. Double circles indicate the improved stations.

causes an almost constant change in epicentral distance at each station.

In Figure 7 directions of the propagation of waves from each group of earthquakes are shown together with the locations of the stations used in this investigation. The double circles denote the stations equipped with the new seismographs. The first two or three letters in station names are used as station codes. In Figure 8 three vertical cross-sections along the lines indicated in Figure 7 are shown. The lines I, II, and III are alomost parallel to the directions of wave propagation from earthquakes in groups I, II, and III respectively. Earthquake foci occurring within 100 km from

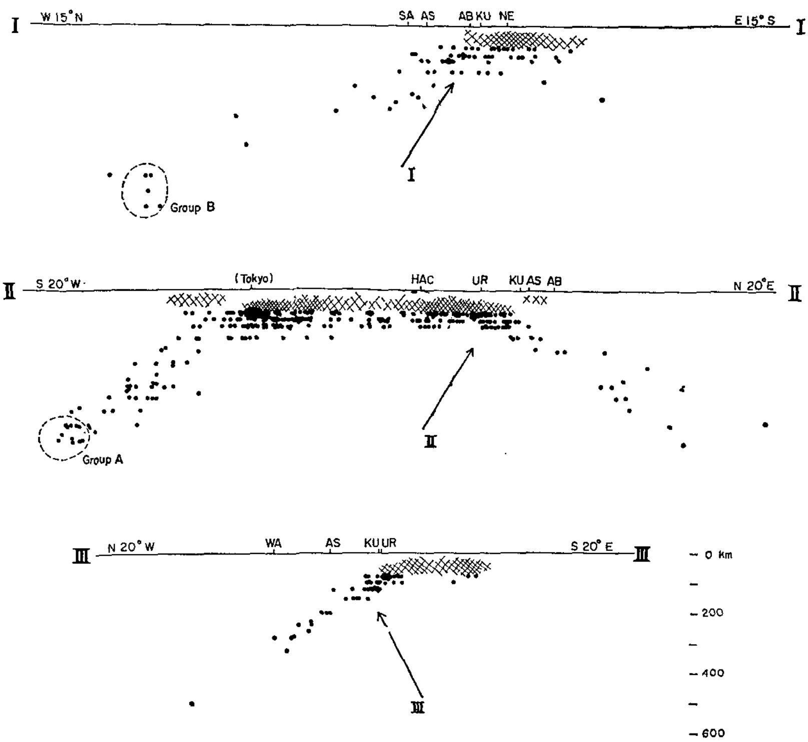

Fig. 8. Vertical cross-section along the lines shown in Fig. 7. Dots represent earthquake foci deeper than 70 km. Hatchings are shallow earthquake domains.

Table 1.

Group	Date and Time (J.S.T.)	Location	CGS Epicenter		Depth	m
I	1962 July 7 08:05	Hindu Kush	70°.4E	36°.6N	203km	
	1964 Jan. 28 23:09	Hindu Kush	70°.9E	36°.5N	207	6.1
	1965 Mar. 15 00:53	Hindu Kush	70°.7E	36°.3N	219	6.6
	1966 June 6 06:55	Hindu Kush	71°.2E	36°.3N	225	6.3
II	1963 Dec. 16 04:34	Java Sea	108°.0E	4°.8S	650	6.4
	1964 July 8 20:55	Banda Sea	129°.8E	5°.5S	165	6.5
	1965 Aug. 20 14:54	Banda Sea	128°.6E	5°.7S	326	6.2
	1966 June 23 05:37	Banda Sea	124°.6E	7°.2S	507	6.1
III	1963 July 4 19:58	Tonga Is.	177°.7W	26°.3S	158	6.5
	1964 July 10 01:39	New Hebrides Is.	167°.6E	15°.5S	121	6.6
	1964 Aug. 13 09:31	Solomon Is.	154°.8E	5°.4S	383	6.0
	1964 Dec. 29 01:16	Fiji Is.	179°.6W	22°.1S	611	6.2

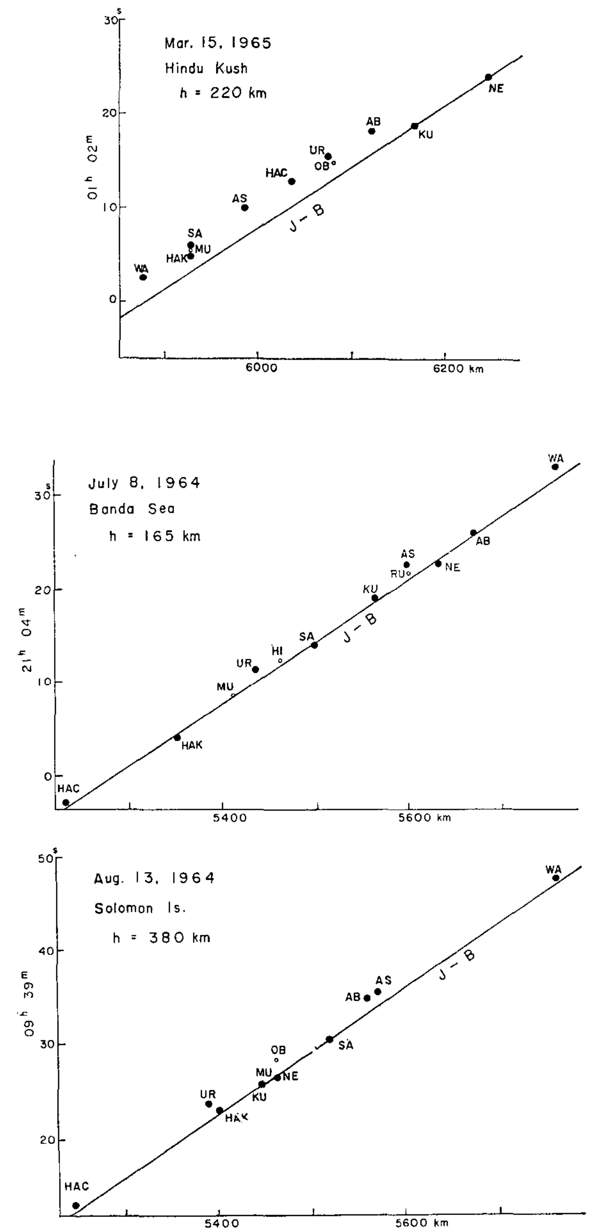

Fig. 9. Examples of travel time diagrams for P waves from distant deep earthquakes observed in Hokkaido.

each vertical sectional plane are projected on the plane. The foci thus plotted in Figure 8 are those located by JMA from 1962 through 1965 at focal depths 80 km and larger. Earthquakes with focal depth larger than 150 km and magnitude 6 or above occurring in the years 1935 to 1957 inclusive are supplemented from Katsumata's table[34]. Distributions of shallow earthquakes which outnumber the deeper ones are indicated by hatchings.

Figure 9 shows the travel times of P waves plotted against epicentral distance for a sample earthquake in each group. S wave travel times are not shown, since they are scattered rather randomly and not usable in the analysis. This may be due to small and flattened onset of shrot period S phase at large distances. The large circles in Figure 9 represent the improved stations. The solid lines marked with J-B are the Jeffreys-Bullen travel time curves best fitting to Nemuro and Kushiro arrival times which are usually earlier than those expected from the average curve. The travel time residuals at the improved stations are obtained by referring to such J-B curves. The mean residuals for each group are listed in Table 2.

The distributions of the mean residuals among the stations are more clearly represented in Figure 10 in which the stations are arranged in order of average focal depth of earthquakes occurring just beneath the respective station

Table 2. Mean residuals (in seconds) for P waves with standard deviations.

Class	Distant			Moderate distanc e		Near-by		
Group	I	II	III	A	B			
Epicenter	CGS			CGS		JMA	CGS	Combined
Reference Curve	J-B			J-B		W-M		
Abashiri	2.22±0.73	1.00±0.81	0.57±0.78	0.97±0.55	2.32±1.03	1.50±1.18	2.82±0.97	2.26±0.85
Asahikawa	2.40±0.55	1.27±0.54	1.22±0.85	0.27±0.51	0.75±0.15	0.90±1.10	2.95±1.06	2.16±1.00
Hachinohe	1.13±1.00	0.50±1.21	−0.08±0.44	−0.12±0.62	−0.80±1.02	0.58±1.44	0.75±0.91	0.68±1.18
Hakodate	1.12±0.23	0.05±0.57	0.22±0.30	0.30±0.63	0.47±1.02	0.22±1.06	1.48±1.13	0.72±0.78
Kushlro	0.20±0.07	0.30±0.10	0.15±0.11	0.15±0.18	−0.30±0.23	0.24±0.28	0.32±0.24	0.28±0.24
Nemuro	−0.20±0.07	−0.30±0.10	−0.17±0.16	−0.23±0.13	0.30±0.23	−0.24±0.28	−0.32±0.24	−0.28±0.24
Sapporo	1.65±0.65	0.20±0.12	0.12±0.04	−0.80±0.50	0.30±0.55	0.49±0.86	2.53±0.96	1.66±0.84
Urakawa	2.57+0.99	1.80±0.50	1.57±0.50	0.90±0.92	0.05±0.77	1.84±1.04	1.74±0.52	1.82±0.95
Wakkanai	2.90±0.35	1.18±0.44	0.92±0.56	2.97±1.00	3.37±0.65	−0.16±1.77	4.17±1.54	2.46±1.47

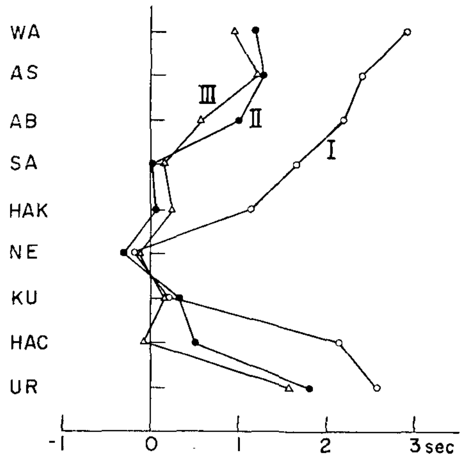

Fig. 10. Mean residuals at each station for earthquakes in groups I, II, and III.

(see Figure 4). It is readily seen in this figure that the residual distributions for groups II and III are approximately the same, whereas they are significantly different from that of group I. The residuals for group I are considerably larger than those for groups II and III. This fact may be explained by the difference in azimuth of the source. As seen in Figure 8 waves from group I earthquakes approach the stations in Hokkaidô from the continental side (the deeper side of the inclined seismic layer) making small angles with the seismic layer. Waves from groups II and III earthquakes approach from the oceanic side making nearly right angles with the seismic layer. When the model B in Figure 1 is adopted, it is easily understandable from Figure 11 (lower right) that the residuals are positive at Wakkanai, Asahikawa, and Abashiri and particularly large in group I. The large positive residuals at Urakawa and Hachinohe (group I only) are difficult to explain on the basis of this simple model. They are probably associated with more local conditions peculiar to the stations. If model A in Figure 1 is assumed, it is difficult to explain the variations in residuals with station in groups II and III. However small modification of this model like model A′ in Figure 11 makes an explanation possible.

In order to explain the travel time residuals quantitatively, the residuals at each station for each earthquake group are calculated on the basis of the models shown in Figure 11, A′ and B. (Both models yield the same result.) In this calculation the actual location of seismic layer relative to the seismic

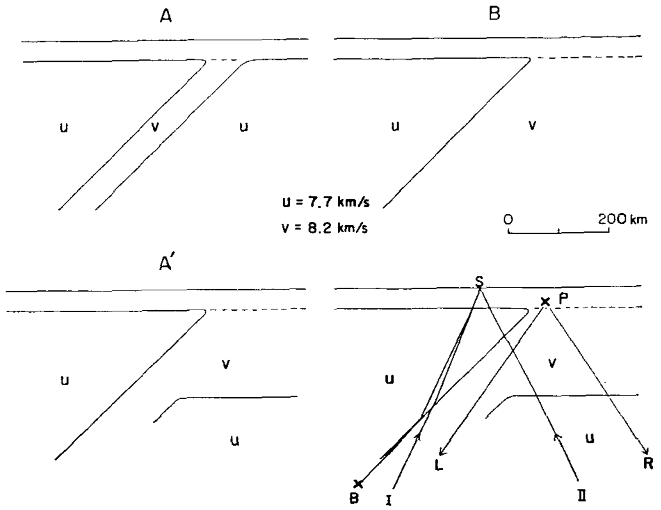

Fig. 11. Upper mantle models used in calculation of the travel time residuals. The lower-right diagram show ray paths to a station S from distant earthquakes (I and II) and near-by deep earthquakes (B), and those from a focus P.

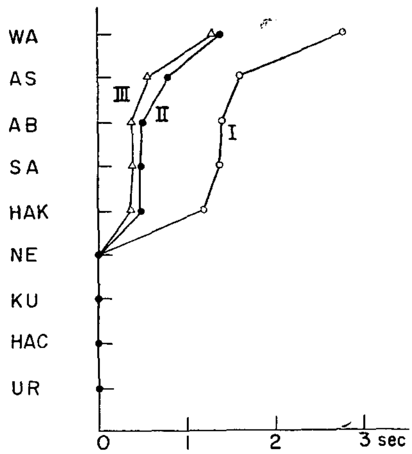

Fig. 12. Calculated residuals at each station based on model A′ or B for groups I, II, and III.

rays reaching at each station is considered. The models in Figure 11 are only indexes of the types of structure. The calculated residuals are plotted in Figure 12 in the same manner as in Figure 10. A comparison of these two figures indicates that the model A′ or B is an acceptable one, though this is only a rough approximation. P wave velocities of 7.7 km/sec and 8.2 km/sec in the models are only assumptions, but a velocity difference of about 6% between the high and low velocity portions is a conclusion not affected by allowable changes in the assumed velocities.

4. Systematic differences in epicenter location between JMA and CGS

Earthquake hypocenters in the vicinity of Japan are determined by the Japan Meteorological Agency and the U. S. Coast and Geodetic Survey independently. CGS locates hypocenters from the world-wide observations, while JMA uses the data from Japanese stations only. The two hypocenters for the same earthquake often differ appreciably as illustrate in Figure 13. The

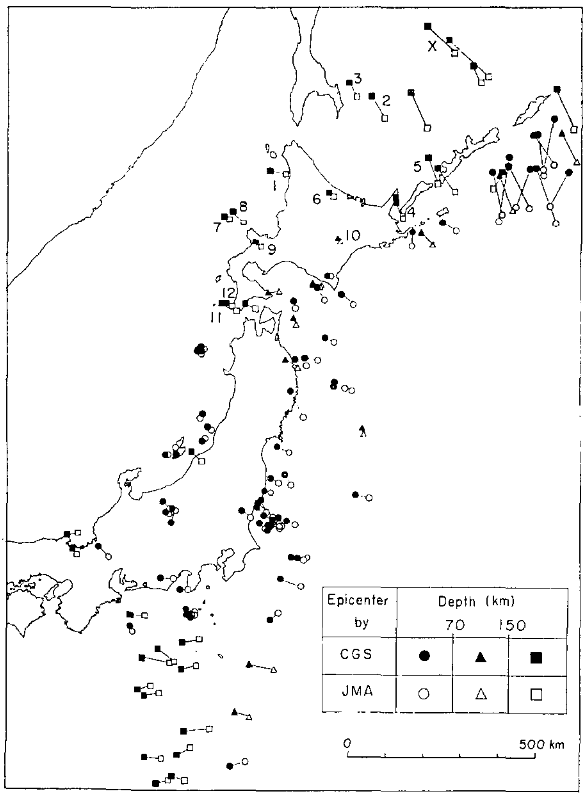

Fig. 13. Comparison of epicenters determined by JMA and CGS. The numerals 1,2,3, etc. refer to those in Table 4.

earthquakes plotted in this figure are taken from "Seismological notes" for 1963–65 in Journal of the Seismiological Society of Japan (*Zisin*). Most earthquakes which occurred near Matsushiro are neglected to avoid the confusion. Only four largest Matsushiro shocks are plotted.

It is apparent from this figure that the two epicenters for the same earthquake belonging to the inclined seismic layer are systematically separated, that is, in most cases the direction of a CGS epicenter from the corresponding JMA epicenter is nearly normal to the strike of the seismic layer and the other features shown in Figures 3–6. CGS epicenters are always on the inner (continental) side of JMA epicenters. Another important fact is a good agreement of the two epicenters for most of the shallow earthquakes not belonging to the inclined seismic layer. This fact suggests that both JMA and CGS hypocenter programs are reliable. The systematic differences between the two epicenter locations cannot be caused only by the regional velocity anomalies beneath distant stations reporting their readings to CGS. If some azimuthally non-uniform distributions of the anomalies associated with the distant receiving stations is the cause of the deviation of CGS epicenters, the deviation must be in almost the same direction throghout the area covered by Figure 13 which is rather small considering the distances to most receiving stations.

According to this idea CGS epicenters for very deep earthquakes (depth 500–600 km) appear to be fixed close to the true epicenters, since large lateral variation in seismic velocities seems unlikely beneath these levels. JMA epicenter locations for these earthquakes are affected by the upper mantle irregularities described before. All models in Figure 11 can be used to explain the fact that all JMA epicenters of deep earthquakes are shifted toward the outer (oceanic) side of the seismic zone. A rough estimate of the deviation of a JMA epicenter from the true epicenter is 20 to 60 km for these models. The deviation depends on the location of the epicenter relative to the observation networks.

JMA epicenters for shallow earthquakes occurring in and near Japan (within about 100 km from the coast) seem to be very close to the true epicenters, but most CGS epicenters which are more than about 20 km from these JMA epicenters seem to be incorrect because they are inconcistent with the data from nearby stations. In this case CGS epicenters are affected by the upper mantle irregularities under Japan. As seen in Figure 11 (lower-right) only the A′-type model provides an explanation of the deviations of

CGS epicenters from the true (JMA) epicenters. The seismic waves from a focus P to a direction PL arrive earlier at a distant station than the waves to a direction PR which arrive at another station in opposite direction at the equal distance. Therefore the CGS hypocenter-program will fix the epicenter on the left (continental) side of the true epicenter with an estimated separation of about 20 km.

For earthquakes at intermidiate depths both JMA and CGS epicenters are affected by the upper mantle irregularities. The two epicenters are shifted in opposite directions. Therefore the true epicenters for these earthquakes may lie somewhere between the two epicenters.

Travel time studies by Husebye[35] show a geographical distribution of the travel time correction for Uppsala station for earthquakes in the Japanese area, but there seems to be no simple relation between this distribution and the upper mantle models discussed in this paper.

Table 3.

Group	Date and Time (J.S.T.)				JMA Epicenter		Depth	CGS Epicenter		Depth	M
A	1956	Feb.	18	16:34	138°.5 E	29°.9 N	480km	$137^{\circ 1}/_2$E	30° N	450km	$7^1/_4$-$7^1/_2$
	1962	Dec.	7	23:04	139°35′	29°12′	400	139°.2	29°.2	411	$6^3/_4$-7
	1964	May	7	20:11	138°08′	30°37′	450	137°.7	30°.6	469	m: 5.1
	1965	Apr.	13	05:41	138°54′	30°08′	450	138°.5	30°.2	421	m: 5.8
B	1957	Jan.	3	21:48	$131^{\circ 1}/_2$	$43^{\circ 1}/_2$	600	130°	44°	600	7
	1959	Oct.	29	23:30				131°	43°	550	$6^1/_4$
	1960	Oct.	8	14:53	130°	40°	650	129°.7	40°.0	608	$6^1/_2$-$6^3/_4$
	1965	Aug.	7	03:15				131°.2	41°.4	560	m: 5.3

5. Travel time anomalies for deep earthquakes at moderate distances

The effect of errors in epicenter locations on the results of travel time analyses is more serious for near earthquakes than for distant ones. Before dealing with near-by deep earthquakes, deep earthquakes at epicentral distances of 800 to 1800 km are studied here. The earthquakes are those occurring in two regions as indicated in Figure 7 (groups A and B). The CGS epicenters are probably more accurate for these shocks for the reason mentioned before.

Unfortunately only a few such shocks occurred since 1962 which were large enough to be recorded clearly at Hokkaidô stations. Accordingly several large shocks before 1962 are added in this investigation. These shocks

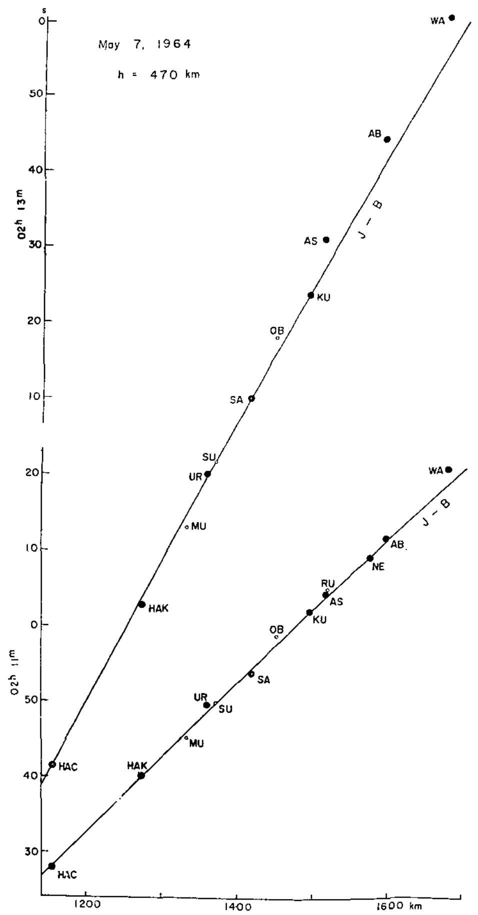

Fig. 14. Example of travel time diagrams of P and S waves from earthquakes at moderate distances.

are listed in Table 3 and plotted in Figure 7. An example of the travel time diagrams of P and S waves referring to the CGS epicenters are shown in Figure 14. The J-B curves are drawn in the same manner as in Figure 9, and the travel time residuals are obtained from these curves. The mean residuals for P waves at each station for each group are listed in Table 2. They are plotted in Figure 15. It can be seen in Figure 8 along what paths seismic waves from these earthquakes reach the Hokkaidô stations.

The calculated residuals for groups A and B are critically dependent on the actual location and shape of the seismic layer, the depth to the upper

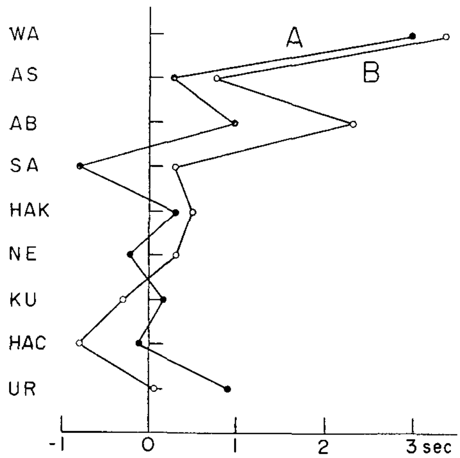

Fig. 15.

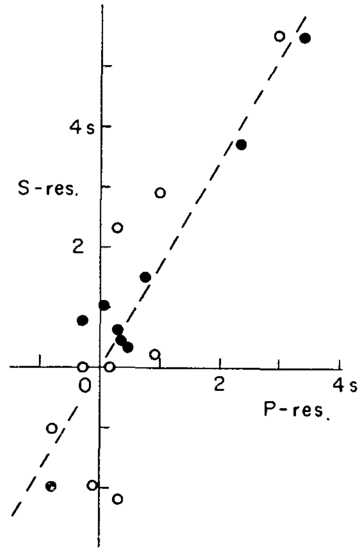

Fig. 16.

Fig. 15. Mean residuals at each station for earthquakes in groups A and B.
Fig. 16. Relation between travel time residuals for P and S waves at nine stations. open circle: group A, solid circle group B. The broken line indicates the theoretical relation when $\sigma = 1/4$.

boundary of the low-velocity portion, etc., owing to the small angles between the seismic rays and the earth's surface. A graphical calculation of the residuals for A group using model A′ or B gives about 1.8 sec for Wakkanai and less than 0.5 sec for the other stations. For group B the calculated residuals are larger than the observed ones. The calculated value is about 5.5 sec for Wakkanai, about 3.5 sec for Asahikawa, about 2.5 sec for Abashiri, about 2.0 sec for Sapporo and Hakodate, and zero for the other stations. Since the agreement between the calculated and the observed residuals is not very satisfactory, the modification of the model comes into question, but it is not made here considering the accuracy of the data.

In Figure 16 the S wave travel time residuals at each station for each group are plotted against the P wave residuals. However no conclusion is obtained from this figure to indicate the regional variations in the velocity ratio of P to S waves, which will be discussed in the next chapter.

6. Velocity ratio of P waves to S waves in the upper mantle

Vertical or lateral variations in velocity ratio between P and S waves v_P/v_S or Poisson's ratio σ $\left(\sigma=\{(v_P/v_S)^2-2\}/2\{(v_P/v_S)^2-1\}\right)$ in Japan have been investigated by Yoshiyama[36], Kamitsuki and others[37-38]. Since there are observations suggesting that the low velocity layer in the upper mantle is more prominent in S wave velocity (for Japanese region, see Inouye[39] and Kakuta[40],) it is supposed that v_P/v_S or σ is larger in the low velocity layer than in the inclined seismic layer.

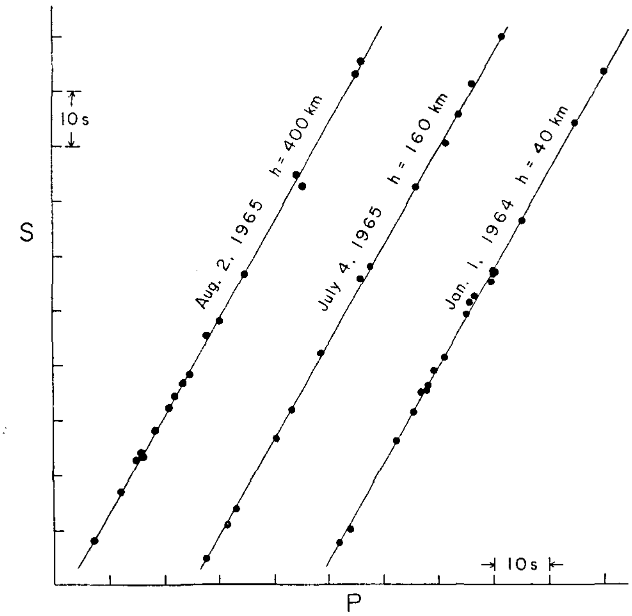

Fig. 17. S wave arrival times plotted against P wave arrival times.

It is not necessary to know the exact locations of the foci in a study of v_P/v_S or σ. When S wave arrival times are plotted against P wave arrival times, the plotted points must lie on a straight line if there is no variation of v_P/v_S or σ in the concerned portion of the earth. Such diagrams are prepared for fifty shallow (depth 40 km or greater) and deep earthquakes which occurred in the northern Japanese region from 1963 through 1965 using data from all Hokkaido stations and four northern Tôhoku stations. Only three examples are illsutrated in Figure 17. No large systematic deviations from a straight line having a slope of $v_P/v_S=1.77\pm0.03$ fitted to the plotted points are found in this figure and the other figures not reproduced here. This indicates that v_P/v_S or σ are almost constant in the upper mantle as far as the present study concerned. Hisamoto's discovery[20] of the anomalously high velocity region for S waves off the Pacific coast of northeastern Japan does not contradict the present conclusion since no wave path studied here traverses this region.

Table 4.

No.	Date and Time (J.S.T.)				JMA Epicenter		Depth	CGS Epicenter		Depth	m
1*	1963	Apr.	1	13:29	141°38′E	44°42′N	280km	141°.1E	44°.8N	255km	5.6
2*	1964	July	1	05:09	145°02′	46°02′	360	144°.6	46°.6	383	5.5
3	1965	Aug.	2	00:03	144°06′	46°37′	400	143°.8	46°.9	400	5.7
4*		Oct.	26	07:34	145°31′	43°44′	160	145°.3	44°.2	180	6.2
5		Nov.	29	18:00	146°42′	44°26′	200	146°.5	45°.1	153	5.3
6	1966	Feb.	19	04:02	143°18′	44°11′	240	143° 1	44°.3	225	5.2
7		Feb.	28	11:02	139°48′	43°38′	240	139°.6	43°.7	225	5.5
8		June	23	14:02	140°15′	43°33′	240	139°.9	43°.8	218	5.5
9		Aug.	20	18:32	140°50′	42°59′	160	140°.6	43°.1	161	5.8
10		Nov.	1	16:01	143°26′	43°06′	120	143° 4	43°.2	127	4.8
×		Nov.	22	15:31	147°47′	47°35′	500	146°.7	48°.2	453	5.6
11*		Dec.	2	03:56	140°05′	41°27′	160	139°.6	41°.6	173	5.4
12	1967	Feb.	3	01:24	139°59′	41°31,	180	139°.7	41°.6	176	5.4

* Seismic intensity distributions for these earthquakes are shown in Fig. 2.

7. Travel time anomalies for near-by deep earthquakes

Thirteen deep earthquakes which occurred in the vicinity of Hokkaidô since 1963 have been chosen in this investigation. They are listed in Table 4 and their locations are indicated in Figure 13. Travel time diagrams of P and S waves using both JMA and CGS epicenters have been prepared. The P wave travel time curves by Wadati, Sagisaka, and Masuda and the S wave travel time curves by Sagisaka and Takehana best fitting to the Nemuro and Kushiro data have been used as reference curves instead of the Jeffreys-Bullen curves in the previous diagrams. Mean travel time residuals for P waves thus obtained are shown in Table 2.

The residuals based on the JMA epicenters are usually smaller than those based on the CGS epicenters. However both residuals at each station vary randomly from earthquake to earthquake and no correlation between the residual and the hypocenter location can be found.

In fact, both JMA and CGS epicenters deviate from the true epicenters by the influence of the lateral velocity variations in the upper mantle. Consequently a new epcienter is chosen for each earthquake by using an equation

$$r = \frac{h}{500}(r_C - r_J) + r_J$$

where r, r_J, and r_C denote position vectors of the new, JMA, and CGS hypocenters respectively, and h is the focal-depth in kilometers. Some examples of the travel time diagrams using these new epicenters are illustrated in Figure 18. The mean residual at each station is also listed in the last column of Table 2.

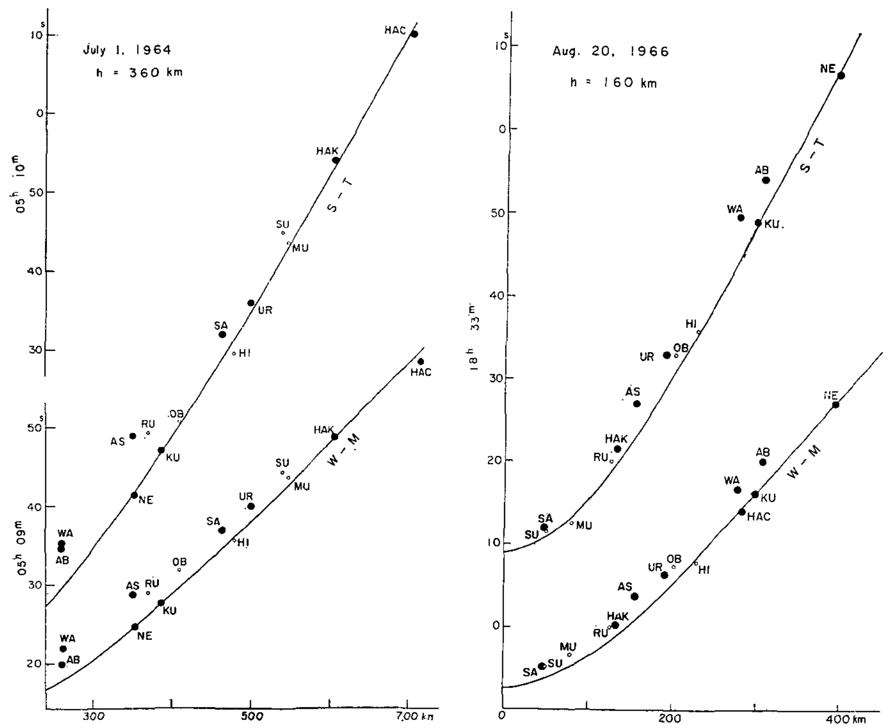

Fig. 18. Examples of travel time diagrams of P and S waves from near-by deep earthquakes.

In Figure 19 the P wave residuals at Wakkanai, Asahikawa, Abashiri, and Sapporo are plotted against the hypocentral distance. The points exhibit a considerable scatter, but a tendency for residuals to increase with distance is seen. The calculation of the residual at a station for an earthquake using models in Figure 11 is somewhat complicated, because the propagation path from the focus to the station is affected by the location of the focus

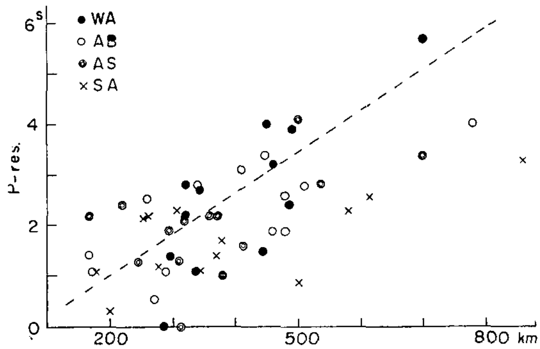

Fig. 19. P wave residuals plotted against hypocentral distance.

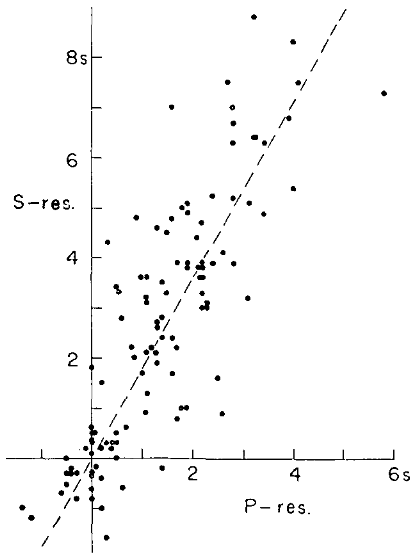

Fig. 20. Relation between travel time residuals of P and S waves for near-by deep earthquakes. The broken line indicate the theoretical relation when $v_P/v_S = 1.77$.

and the three-dimentional shape of the seismic layer. The residuals at Wakkanai are the simplest to calculate for the earthquakes used here. Roughly speaking, the residual at Wakkanai is proportinal to the hypocentral distance substracted by a constant, if the earthquake is assumed to occur at the top boundary of the seismic layer. The borken line in Figure 19 indicate the calculated residuals at Wakkanai based on the models in Figure 11.

The result of the previous chapter indicating the constant velocity ratio of $v_P/v_S=1.77\pm$ predicts that the residual for S wave is about 1.77 times as large as that for P waves. Figure 20 shows this relation, though the scatter of the data is fairly large.

The conclusion of the present paper will appear in the last chapter of the whole article, the second half of which will be published in the next number. The main result of the first half is the proposal of the model A′ in Figure 11, in which the velocities of both P and S waves are higher in the deep seismic layer by about 6% than in the other portion of the upper mantle. The regional variation of the thickness or velocity of the crust may cause the travel time anomalies, but the main cause of the anomalies at most stations studied in this paper can be attributed to the peculiar structure of the upper mantle.

Note added in proof: In a recent issue of J. G. R. (Aug. 15, 1967) Oliver and Isacks published a result from intensive seismic studies of the Tonga-Kermadec arc. The anomalous upper mantle structure associated with the deep seismic zone under the island arc has been clearly demonstrated.

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