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40. *Travel Times to Japanese Stations from Longshot and their Geophysical Implications.*

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Abstract

Travel times to Japanese stations from the Longshot underground explosion of October 29, 1965, were studied. The arrival times of P waves are 1.5 to 3 seconds earlier than the Jeffreys-Bullen (J-B) time. A systematic variation of the J-B residual was found in relation to the seismic plane which dips downward from the Pacific Ocean side towards the continent; a station on the Pacific Ocean side registered a relatively early arrival (−3 sec), while stations on the Japan Sea side registered relatively late arrivals (−1.5 sec). This systematic variation can be interpreted in terms of a structural anomaly associated with the deep seismic plane. The simplest of all the models which explain this systematic variation is the one in which the P-wave velocity in the top 250 km of the mantle is reduced by 0.4 km/sec on the continental side as compared with that on the ocean side of the deep seismic plane. The amplitude ratio of PcP to P also provides a favorable evidence for this model. The average Q is estimated as 80 in the top 250 km of the mantle on the continental side of the deep seismic plane. The low velocity and Q can be explained in terms of a temperature excess of about 500°C coupled with a partial melting of about 2%. Such anomalous structure would have an important bearing on various island-arc tectonics such as the magma generation and convection. The arrival times at Japanese stations are compared with those at stations elsewhere in the world. The earliest arrival among the Japanese stations is comparable to that at stations on relatively old crust, and the latest to that at stations on relatively young crust. The residual at Japanese stations, on the average, is not anomalous as compared with the world average.

Introduction

The Longshot underground explosion of October 29, 1965, detonated on Amchitka Island, registered clear P-waves at several Japanese stations operating high sensitivity short-period seismographs. Since the location and the origin time are accurately known, this explosion afforded a good opportunity of establishing accurate travel times to Japanese stations from the Aleutian Islands. Accurate travel times are indispensable for determining the station correction, an important parameter in the problem of teleseismic location. One of the purposes of this paper is to present a complete data for P and, at some stations, PcP phases.

Another purpose of this paper is to study deep structures beneath the Japan arc on the basis of the Longshot data. It was shown by *Press and Biehler* (1964) that accurate P wave delays observed for explosions can be used for revealing subsurface structures. *Otsuka* (1966), and *Bolt and Nuttli* (1966) have shown that the azimuthal variation of P-wave residuals can effectively be used to study the lateral variation in mantle structures. Anomalous propagations of seismic waves in Japan have long been recognized by a number of workers. *Morita* (1937) found a distinct regional difference in travel times, amplitude, and period of S waves from a deep-focus earthquake. *Okura* (1959), in his study on the zone of abnormal seismic intensity, found that when seismic waves have a longer portion of the propagation path in a region having a previous seismic activity, they are felt more strongly. *Katsumata* (1960) arrived at a similar conclusion with more convincing data than had been previously presented. He found that the wave velocity is relatively high, and the attenuation relatively low in regions of high seismic activity. *Hisamoto* (1965*a*, *b*) found anomalously early arrivals of S waves when a part of the propagation path is just below a region of high seismic activity. Although his results are difficult to interpret quantitatively, they seem to reflect a peculiarity of the island arc structure. *Utsu* (1966) introduced a highly attenuating region in the upper mantle above the intermediate-deep seismic plane in order to explain the regional difference of seismic intensity. *Utsu* (1967) further studied travel times of several deep-focus earthquakes, and found a systematic variation of travel-time residuals among stations in Hokkaido. He attributed it to a lateral inhomogeneity of the mantle structure, and proposed a model in which P wave velocity is about 6% lower on the continental side than on the ocean side of the deep seismic plane. A number of other papers reporting abnormal propagations of seismic waves can be found in literature (e.g. *Honda*, 1932*a*, *b*; *Tamaki*, 1955), although they were not always considered to be related to the deep structure of the arc.

In the vicinity of the Tonga-Kermadec arc, *Oliver and Isacks* (1967) made an extensive study and found a high Q zone just below the deep seismic plane. They considered this anomalous structure as an evidence for the movement of the lithosphere.

It is now hoped to make more quantitative determinations of the structural features of the Japan arc in order to elicit possible structural difference either within the Japan arc or between the Japanese and other arcs. Since the Longshot data are one of the most reliable among those available for the Japan arc, this paper attempts to interpret them in terms of the structural peculiarity of the Japan arc.

Data

Table 1 lists the data. The data for Seoul, Korea, are added for comparison. All the seismograms were collected and reread. Some of the seismograms are reproduced in Fig. 1. At stations, TSK, MAT, MYK, SHK and SEO, clear *PcP* phases were recorded. At other stations, they were not clear enough to be used for analysis. The recording instruments are all short-period vertical seismographs, although they differ in minor respects. At TSK and SHK, the HES 1–0.2 seismograph (see e.g. *Shiraki Micro-Earthquake Observatory*, 1965) was used. Readings at MAT and SEO were made on the standardized Benioff short-period seismograms supplied by the United States Coast and Geodetic Survey (USCGS). The recordings at ISE, HBR, HMI, OYA, IZM, SRT, FOK, MZK, TDR, WKU, KKW, ARD and HDK were made by electromagnetic seismographs which are described in the Seismological Bulletin of Wakayama Micro-Earthquake Observatory and Its Substations, January-June, 1965 (*Wakayama Micro-Earthquake Observatory*, 1966). Ellipticity corrections for P and PcP phases were calculated by the simplified formula given by *Bullen* (1937; 1963, p. 181). The ellipticity correction is about -0.20 sec for the P phase and -0.35 sec for the PcP phase. The corrections are to be subtracted from the observed times. Corrections were also made for the depth of the source and the heights of the stations. Except for MAT and SEO the readings are believed to be accurate to ± 0.2 sec for P and ± 0.3 sec for PcP phases. The residuals from the Jeffreys-Bullen times (*Jeffreys* and *Bullen*, 1958) (observed minus calculated time) were calculated for both P and PcP phases; they are given in Table 1, and plotted in Fig. 2. The earliest arrival was registered at TSK, a station on the Pacific coast, for both P and PcP phases. The western stations recorded relatively late arrivals (see also Fig. 3). It is to be noted that the general trend of the variation of the J-B residual is about the same for P and

Table 1. Travel-time data for Longshot.

LONGSHOT: Date 29 Oct. 1965, Origin time 21:00:00.08, Depth 0.7 km
Lat. 51.4381°N, Long. 179.1826°W, m(USCGS)=6.1

Station	Abbr.	Latitude* (deg N)	Longitude (deg E)	Height (km)	Δ (deg)	Azimuth from source (deg)	Azimuth from station (deg)	Arrival time (GMT) h m s	Residual from J-B time† (sec)
Tsukuba	TSK	36.2108	140.1100	0.286	31.606	256.6	48.8	*P*, 21 06 23.40	−3.01
								PcP, 21 09 14.70	−4.08
Matsushiro	MAT	36.5383	138.2083	0.440	32.556	258.9	49.7	*P*, 21 06 32.60	−2.15
								PcP, 21 09 18.10	−3.25
Ise	ISE	34.4585	136.7740	0.440	34.805	257.5	47.7	*P*, 21 06 52.30	−1.95
Haibara	HBR	34.5028	135.9934	0.390	35.253	258.3	47.9	*P*, 21 06 56.47	−1.59
Myoken	MYK	34.9272	135.4996	0.600	35.274	259.4	48.5	*P*, 21 06 56.70	−1.57
								PcP, 21 09 26.00	−3.04
Abuyama	ABU	34.8568	135.5729	0.220	35.276	295.2	48.4	*P*, 21 06 56.70	−1.51
Hikami	HMI	35.2265	135.0435	0.250	35.358	260.2	48.9	*P*, 21 06 56.83	−2.09
Oya	OYA	35.3339	134.6645	0.260	35.522	260.7	49.1	*P*, 21 06 58.17	−2.13
Izumi	IZM	34.9722	134.8858	0.230	35.623	260.0	48.6	*P*, 21 06 59.39	−1.77
Sarutani	SRT	34.1762	135.7460	0.470	35.624	258.2	47.6	*P*, 21 06 59.50	−1.73
Funaoka	FOK	35.3331	134.2718	0.160	35.765	261.0	49.1	*P*, 21 07 00.35	−2.01
Mikazuki	MZK	34.9837	134.4457	0.200	35.886	260.4	48.7	*P*, 21 07 01.13	−2.27
Todoroki	TDR	34.1597	135.3069	0.050	35.904	258.6	47.7	*P*, 21 07 02.33	−1.20
Wakaura	WKU	34.1879	135.1730	0.010	35.967	258.7	47.8	*P*, 21 07 02.70	−1.36
Kainokawa	KKW	33.8988	135.4418	0.260	35.997	258.1	47.4	*P*, 21 07 02.73	−1.63
Arida	ARD	34.0859	135.1617	0.040	36.043	258.6	47.7	*P*, 21 07 03.22	−1.48
Hidaka	HDK	33.9259	135.1390	0.030	36.164	258.4	47.5	*P*, 21 07 03.90	−1.82
Shiraki	SHK	34.5322	132.6775	0.285	37.277	261.5	48.6	*P*, 21 07 13.50	−1.60
								PcP, 21 09 32.30	−2.71
Seoul (Korea)	SEO	37.5667	126.9667	0.086	38.907	270.0	52.0	*P*, 21 07 26.80	−1.91
								PcP, 21 09 36.60	−3.40

* Geographical latitude.
† Corrected for ellipticity, station height, and focal depth.
Note: Stations ISE, HBR, SRT, TDR, WKU, KKW, ARD, and HDK belong to the Wakayama Micro-Earthquake Observatory. Stations HMI, OYA, IZM, FOK, and MZK belong to the Tottori Micro-Earthquake Observatory.

PcP phases. Consequently the difference, the residual of *PcP* minus *P* time, is almost constant, -1.0 sec, as shown in Fig. 2.

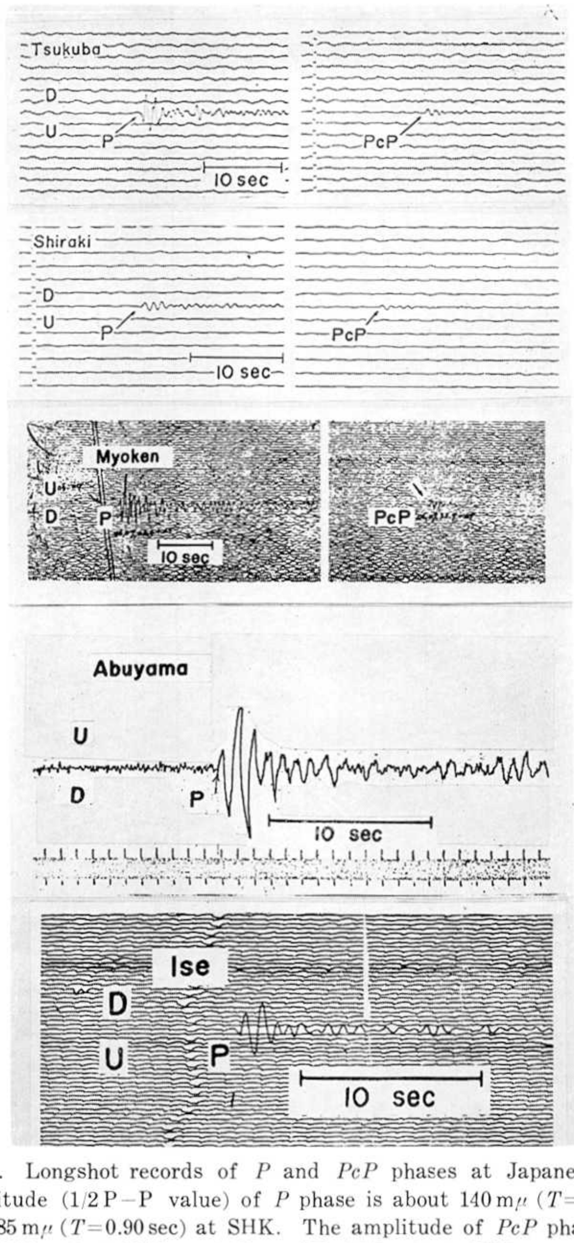


Fig. 1. Longshot records of *P* and *PcP* phases at Japanese stations. The amplitude (1/2 P−P value) of *P* phase is about 140 mμ (T=0.86 sec) at TSK, and 85 mμ (T=0.90 sec) at SHK. The amplitude of *PcP* phase is about 27mμ (T=0.80 sec) at TSK, and 50 mμ (T=0.86 sec) at SHK.

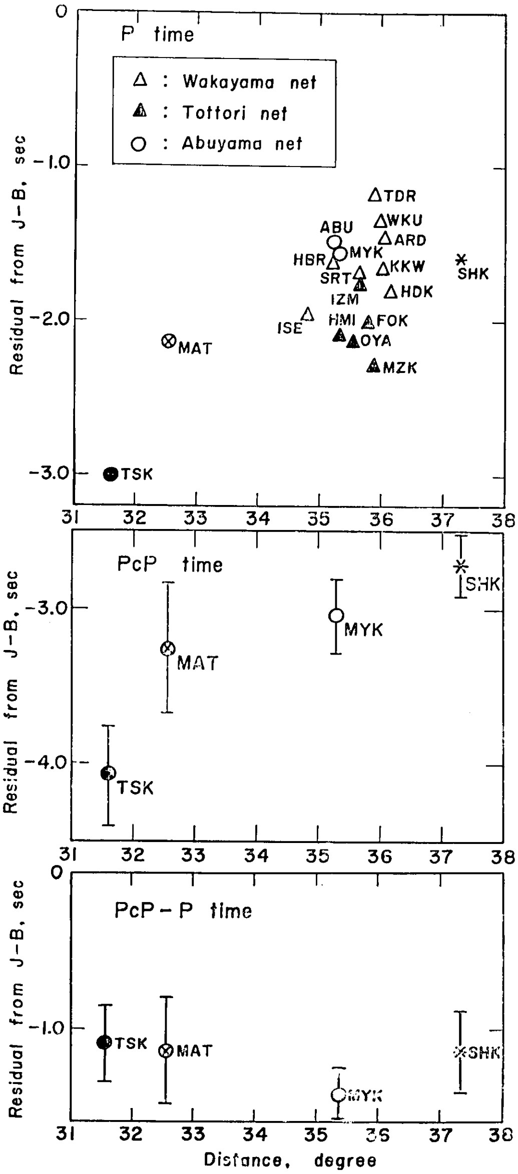


Fig. 2. Travel-time residuals of P and PcP phases from the Jeffreys-Bullen time. Negative values indicate early arrivals.

Interpretation

Fig. 3 shows the locations of stations, TSK, MAT, ABU, WKU, FOK and SHK. The station WKU represents those belonging to the Wakayama net; FOK those of the Tottori net. Superposed upon the location map are the contours of the deep seismic plane derived from *Sugimura and Uyeda* (1968) who considerably updated *Wadati's* (1935) work. The arrows towards each station indicate the direction of the wave path from the Aleutian Islands. Comparing the path lengths over the deep seismic zone with the J-B residuals given in Table 1, we readily find a strong correlation between the path lengths and the J-B residuals; the longer the path across the seismic zone, the later the arrival. This situation can be more clearly demonstrated if we take vertical cross sections including the paths as shown in Fig. 4. We let L be the length of the path above the deep seismic plane (Fig. 4). Fig. 5 shows a plot of the J-B residuals (δT_{JB}) against L for the six stations shown in Fig. 3. An approximately linear trend is seen. It may be argued, however, that the variation of δT_{JB} is caused by a systematic deviation of the J-B travel-time curve from the true travel time rather than by the difference of L. A systematic deviation of the travel time from the J-B time has been reported by several workers (e.g. *Carder et al.*, 1966; *Cleary and Hales*, 1966; *Chinnery and Toksöz*, 1967; *Johnson*, 1967). The residuals given in Fig. 2 might be interpreted as due to this systematic error in the J-B time. However, for the following two reasons, I do not believe that this is the case. First, the range of the residuals given in Fig. 2, about 1.5 sec, is much larger than that predicted, for example, by the curve by *Carder et al.* (1966); it gives only 0.3 sec over the range $\Delta=31.5°$ to $37°$. Secondly, as mentioned earlier, the *PcP* residuals have about the same trend as the *P* residuals. This strongly

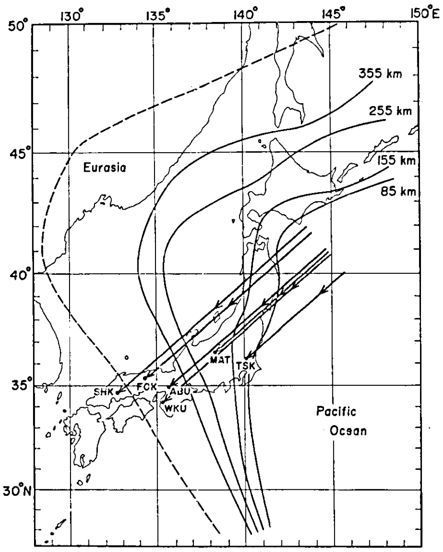


Fig. 3. Location of stations. The deep seismic plane is contoured according to *Sugimura and Uyeda* (1968). The dashed line defines the western limit of the seismic zone. The arrows to each station indicate the wave paths from the Aleutian Islands.

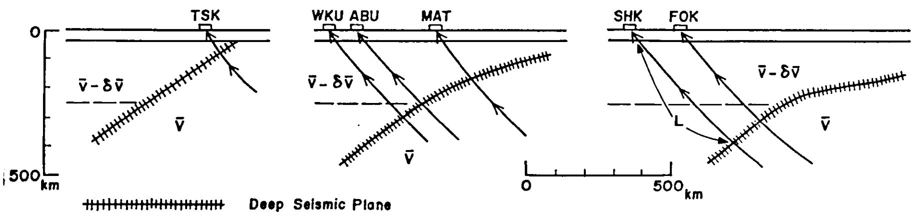


Fig. 4. Cross sections of the structure along the seismic ray.

suggests that the variation of the residuals shown in Fig. 2 is caused by structures just under the station which affect both *P* and *PcP* phases rather than by the error in the standard travel-time curve, the J-B curve. The possible azimuthal source term (*Cleary*, 1967*b*) is not responsible for the variation of δT_{JB} either; all the stations used here are in a narrow azimuthal and distance range from the source so that the range of the take-off angle at the source is too small (less than 1°) for the azimuthal source term to manifest any effect.

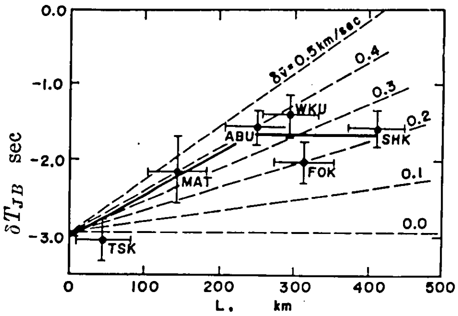


Fig. 5. Jeffreys-Bullen residuals of P phase plotted against the path lengths above the deep seismic plane. The dashed lines give the slopes for various values of the velocity contrast $\delta\bar{v}$ in Fig. 4. The solid line gives the relation when the "$\bar{v}-\delta\bar{v}$" region is bounded at the depth of 250 km as shown in Fig. 4.

The simplest model that explains the systematic variation of δT_{JB} with L (Fig. 5) is the one which introduces a velocity contrast between the two regions of the mantle separated by the deep seismic plane. We let $\bar{v}$ and $\bar{v}-\delta\bar{v}$ be average P velocities in the two regions of the mantle separated by the deep seismic plane; the lower velocity is assigned to the continental side and the higher velocity to the ocean side (see Fig. 4). With this model, the travel-time anomaly δT_{JB} is approximately related to L by

$$\frac{\partial(\delta T_{JB})}{\partial L}=\frac{\delta\bar{v}}{\bar{v}^2}$$

This slope is shown by dashed lines in Fig. 5 for several values of $\delta\bar{v}$, $\bar{v}^2$ being assumed 70 (km/sec)^2. A velocity contrast, $\delta\bar{v}$, of 0.2 to 0.4 km/sec evidently best explains the observations. A closer inspection of the data gives the impression that the observed travel-time anomaly levels off at $L\sim250$ km as shown by the solid line in Fig. 5. A model which explains the solid line can be made by bounding the "$\bar{v}-\delta\bar{v}$" region at a depth of about 250 km as shown by the dashed lines in Fig. 4. The difference of the velocity is taken at 0.4 km/sec. Although this detailed model is not unique, I want to point out that the existence of a gross regional difference of this magnitude in P velocity beneath Japan is definite.

Another evidence for the above model can be derived from the amplitude ratio of PcP to P phase observed at TSK and SHK. At these two stations, seismographs with identical characteristics were operated. As seen in Fig. 1, the amplitude ratio PcP/P at SHK ($\Delta=37.3°$) is 0.58 and is remarkably larger than that at TSK ($\Delta=31.6°$), 0.19. This difference can be partially explained in terms of the variation of PcP/P with distance. However, the major factor which causes this difference is probably the regional difference of Q associated with the geometry of the deep seismic plane.

The ratio PcP/P is affected by the source parameter, the reflection at the core boundary, the ray spreading, and the differential attenuation for P and PcP phases (e.g. *Kanamori*, 1967*a*, *b*). If the mantle beneath a station is attenuating, P suffers a greater attenuation than PcP does because of its oblique trajectory (*Anderson*, 1967*a*, p. 400). Since the difference of Δ between TSK and SHK is small, we can assume that the source effect and the effect of the differential attenuation over the path from the source to the bottom of the deep seismic plane are the same for those two stations. We let r and r' be the ratios of the vertical component of PcP to that of P at TSK and SHK respectively. Then,

$$\frac{r}{r'}=\frac{c}{c'}\exp\left[-\frac{\pi f(\Delta L-\Delta L')}{\bar{Q}(\bar{v}-\delta\bar{v})}\right]$$

where f is the frequency, and c and c' are the "theoretical" ratios for TSK and SHK as given by *Dana* (1945): from Dana's Table 3, we have $c=0.116$ and $c'=0.163$. In the above, ΔL and $\Delta L'$ are the path lengths of PcP minus those of P in the "$\bar{v}-\delta\bar{v}$" region beneath TSK and SHK respectively. The symbol $\bar{Q}$ denotes the average Q in the "$\bar{v}-\delta\bar{v}$" region. We now try to estimate $\bar{Q}$. The frequency read on the seismograms is about 1.3 c/sec. The difference $(\Delta L-\Delta L')$ can be estimated as 120 km from the geometry of the ray paths of P and PcP in the "$\bar{v}-\delta\bar{v}$" region. Substituting $r=0.19$, $r'=0.58$, $c=0.116$, $c'=0.163$ and $(\bar{v}-\delta\bar{v})\sim 8$ km/sec into the above expression, and solving it for $\bar{Q}$, we have $\bar{Q}=80$. This is the average for the mantle above 250 km on the continental side of the deep seismic plane. Although the value of Q for P waves for the normal mantle over this period and depth ranges has not been established, a number of works have suggested that the average Q is larger than 80, the value for $\bar{Q}$. For example, the MM8 Q_α (Q for P waves) model proposed by *Anderson et al.* (1965) gives about 200 for this depth range. The value of Q_α for short-period P waves was determined at Tonto Forest, Arizona (*Kanamori*, 1967*b*); the average value ranged from 130 to 180. Since Tonto Forest is in the basin and range province, a tectonically active area, this value may be slightly biased towards low Q side. Yet it is larger than Q estimated here. From the evidence stated above, it can be concluded that the velocity and Q for P waves are significantly lower in the mantle above, and on the continental side of, the deep seismic plane than in the average mantle at a corresponding depth.

Comparison with world-wide data

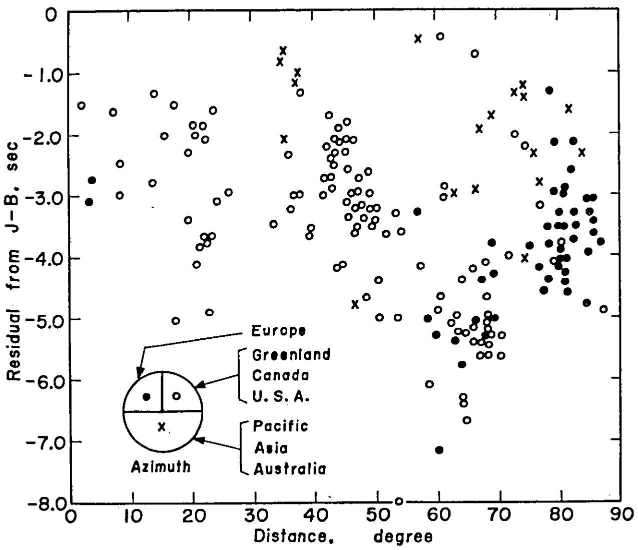
The travel times from the Longshot have been compiled by USCGS in Earthquake Data Report (*USCGS*, 1965). Special travel time studies and analyses have been made, for example, by *Dubois* (1966), *Carder et al.* (1967) and *Cleary* (1967*a*, *b*).

Fig. 6 shows plots of the J-B residuals as a function of distance. In Fig. 6a, the data are classified according to the azimuth from the source. The data are taken from the EDR of USCGS and *Carder et al.* (1967). The readings at UKI, DUG and RAM were rejected because of too large a deviation from the average. In Fig. 6b, the data for $\Delta > 30°$ are shown; the data are classified according to the chronological difference of the crust beneath individual stations. Open circles represent Precambrian and Paleozoic folded belts, dots are for folded belts of Mesozoic and later times, and crosses, mid-oceanic islands. The classification used here is based upon the Plate 5 of *Umbgrove* (1947). Although, as *Cleary* (1967*b*) pointed out, some azimuthal variation seems to exist, a relatively large regional difference seems to have obscured the azimuthal variation. *Dubois* (1966) also interpreted the Longshot data in terms of a regional difference with a somewhat different classification.

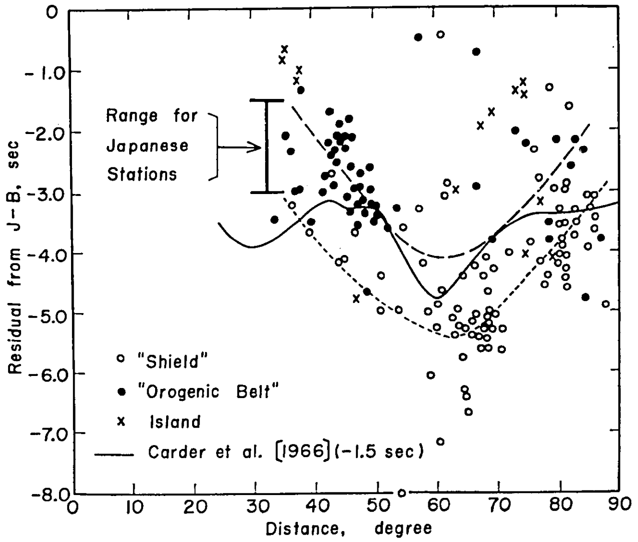
Despite the large scatter of the data, Fig. 6b suggests that the arrivals at stations on the "shield" are about 1.2 sec earlier than those at stations on the "orogenic belt", which, in turn, are about 1.2 sec earlier than those at stations on midoceanic islands. It is to be noted that the range for Japanese stations extends over the difference between the curves for the "shield" and "orogenic belt". This suggests that the deep structure beneath the Japanese Islands which lie on the deep seismic zone is not very different, in its gross velocity structure, from that of the world orogenic belts.

The deep structure of the ocean side of the deep seismic plane is characterized by such extremely high velocity as that under shields. This may be compared with the finding by *Oliver and Isacks* (1967) who introduced a high Q and high velocity layer outside the deep seismic plane of the Tonga-Kermadec arc.

The regional difference in deep mantle structures has been demonstrated by surface wave studies (e.g. *Dorman et al.*, 1960; *Brune and Dorman*, 1963; *Anderson and Toksöz*, 1963; and *Toksöz et al.*, 1967, see also *Press*, 1966; and *Anderson*, 1967*a*). The difference of the travel times for a vertical trajectory above 300 km calculated



(a)



(b)

Fig. 6a, b. The Jeffreys-Bullen residuals for the Longshot. In Fig. 6a, different symbols are used for different azimuths from the source. In Fig. 6b, stations are classified according to the chronological difference of the crust under the station. "Shield" includes Precambrian shields and Paleozoic folded belts. "Orogenic Belt" includes folded belts of Mesozoic and later times. The classification is based upon *Umbgrove* (1947). The crosses are for mid-oceanic islands. The dashed and dotted curves are approximate averages for "Orogenic Belt" and "Shield". The solid curve is the average travel-time residual given by *Carder et al.* (1966) displaced by -1.5 sec which takes account of the source term of the Longshot as given by *Gordon* (1966) and *Cleary* (1967*b*).

for the "shield" and "tectonic" models given by *Toksöz et al.* (1967) is about 2.3 sec for S waves; this suggests a difference of 1.3 sec for P waves. This value is consistent with that indicated by the Longshot data.

Comparison with the crust-mantle structure beneath Japan

Jeffreys (1966) found a relatively low P_n velocity (7.870 ± 0.024 km/sec) over a distance range 2° to 10° for Japan. A direct measurement of the apparent velocity in the vicinity of Japan (*Kanamori*, 1967*c*) gave a value of about 7.9 km/sec to $\Delta = 12°$ with a few exceptions. From the location of the stations relative to epicenters in these studies, it is considered that the major part of the propagation path is above the deep seismic plane. These low velocities are therefore consistent with the present results. *Fedotov and Kuzin* (1963) found a relatively low mantle velocity, 7.7 km/sec, beneath the South Kuril Islands to a depth of 80 km. They considered that island arcs in general are characterized by this low velocity mantle just below the Moho discontinuity.

A recent reinterpretation by *Aki* (1968) of the phase velocities of Rayleigh and Love waves determined for various areas of Japan (e.g. *Aki*, 1961; *Kaminuma and Aki*, 1963; *Kaminuma*, 1964; and *Aki and Kaminuma*, 1963) suggests a crust-mantle model having 2% soft material with shear velocity as low as 1.1 km/sec. Although his result is difficult to compare with the present result because of the difference of the depth range involved, it is qualitatively consistent with the present result in that both introduce an extremely soft mantle beneath Japan. The analyses by *Saito and Takeuchi* (1966) have also shown that mantle models in which the low-velocity layer extends up to the Moho discontinuity best explain the surface wave data given by Aki, Kaminuma and *Santo* (1961*a*, *b*, 1963). The Research Group for Explosion Seismology (e.g. *Research Group for Explosion Seismology*, 1966) has often reported considerably low P_n velocities beneath Japan. Although these studies are related to the mantle above 50 km, they probably have bearings on deeper structures.

The present model (Fig. 4) is, broadly speaking, consistent with the structure beneath Hokkaido derived by *Utsu* (1967). However, the present study cannot reveal such detailed structures along the deep seismic plane as *Utsu* (1967) has speculated. For example, *Oliver and Isacks* (1967) and *Utsu* (1967) introduced a wedge-like high Q or high velocity layer along the deep seismic plane. The present study can neither support nor reject the existence of such a layer, though a structure with a gross high velocity below the seismic plane is favored.

The recent measurements of the group velocity of surface waves with periods to 100 sec around Japan (*Kanamori and Abe*, 1968) have also suggested a low velocity mantle on the continental side, and a high velocity mantle on the ocean side, of the deep seismic plane. It was concluded that the "lid" of the low velocity layer is either very thin or even lacking in the mantle above the deep seismic plane.

Comparison with heat flow

Comparison of P-wave delays (δT) with heat flow (Q) has been made by *Toksöz and Arkani-Hamed* (1967) and *Horai and Simmons* (1968). Horai and Simmons found a relation $Q=0.45\delta T+2.95$ for world-wide data. The measurements of heat flow in and around Japan have been extensively made during the last decade. The latest summary of the results was published in *Vacquier et al.* (1966). Comparing the heat flow map (Vacquier *et al.* Fig. 4) with the residuals at stations given in Fig. 3, we find a distinct correlation between δT and Q; Fig. 7 shows the correlation. Since we considered that the travel-time anomaly is caused in the top 250 km of the mantle, beneath the station we took the range of the heat flow over a horizontal extent of 200 km taken backwards along the path from the station. The dashed line in Fig. 7 gives $(\partial\delta T)/(\partial Q)=1.2\times 10^6\ \mathrm{cm^2\,sec^2/cal}$ which may be compared with the value by *Horai and Simmons* (1968), $2.2\times10^6\ \mathrm{cm^2\,sec^2/cal}$. Despite the distinct correlation, the interpretation of the above relation is not straightforward. The reason is that the heat-flow anomaly is caused by the anomalies of heat source, temperature, and conductivity, while the travel-time anomaly is presumably caused by a temperature anomaly. *Horai* (1964), *Uyeda and Horai* (1964) and *Watanabe* (1968) calculated the temperature in

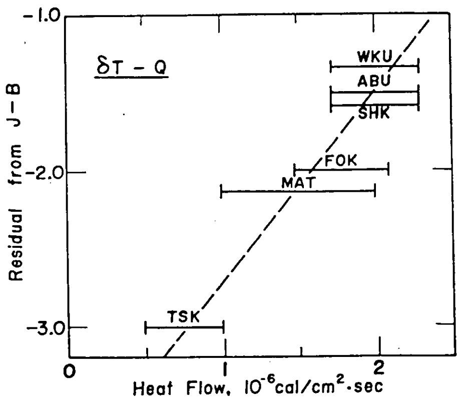


Fig. 7. The relation between the J-B residual and heat flow. The values of heat flow are the ranges over 200 km extent taken backwards along the path from the station.

the mantle beneath Japan based on the heat-flow data. They assumed a distribution of radioactivity and thermal conductivity in the crust and the mantle. It was found that a temperature difference of about 500°C is necessary to account for the difference (1.0μ cal/cm^2 sec) between high and low heat flows in Japan (*Watanabe*, 1968). This temperature difference will cause a difference in velocity of approximately 0.3 km/sec, if we use a value 6×10^{-4} km/sec deg for the temperature coefficient of P velocity at high temperatures, a value based on Fig. 5 of *Soga et al.* (1966). It is also possible that this temperature excess may cause a partial melting which would also decrease the velocity. According to *Hashin* (1962), the P velocity in a material having a liquid inclusion of concentration c can be given by $V=V_0(1-0.58c)$ where V_0 is the velocity for $c=0$. In the above, it is assumed that $\lambda=\mu$ in the solid and that λ is the same for the solid and liquid. From the above expression, we see that a 2% partial melting would cause about 1% velocity decrease.

It has been experimentally shown that the Q of a material suddenly drops at the onset of melting (e.g. *Mizutani and Kanamori*, 1964; *Spetzler and Anderson*, 1968; see also *Anderson*, 1967*b*). In order to explain the relatively low velocity and Q beneath Japan, it seems reasonable to consider that the mantle beneath Japan is partially molten by a small fraction. From the above discussion, we see that a 500°C temperature contrast coupled with a 2% partial melting can explain the proposed velocity contrast, 0.4 km/sec. However, this conclusion greatly depends on the interpretation of heat flow and the temperature coefficient of P velocity. *Toksöz et al.* (1967) also suggested a temperature difference of 100° to 500°C between the mantles beneath oceans and shields.

It was found that a relatively rapid increase in P velocity exists beneath Japan at a depth slightly less than 200 km (*Kanamori*, 1967*c*). This rapid velocity increase was attributed to a "rehardening" of the material that is partially molten in the low-velocity layer. Recent studies of mantle structures beneath continents (e.g. *Niazi and Anderson*, 1965; *Archambeau et al.*, 1966; *Lehmann*, 1967; *Johnson*, 1967) all suggest that this rapid increase in P velocity is a common feature of the mantle structure. *Archambeau et al.* (1968) discussed the importance of the low-velocity zone and the high-velocity "lid" in understanding the tectonic processes beneath continents. Beneath the Japan arc too, these features appear to control deep activities.

Discussions

In the preceding discussions, the effect of the crust on the travel time was ignored. The difference of the crustal thickness beneath the stations used here ranges over about 6 km. The travel-time anomaly caused by this difference is probably about 0.16 sec and does not affect the present conclusions.

The deflection of the ray paths caused by the velocity anomaly associated with the deep seismic plane was neglected. A simple calculation shows that, in the present case, the neglect of the deflection causes an error of only 0.1 sec in the travel time.

The arrival times of P and PcP at Seoul, Korea, are -1.9 and -3.4 sec earlier than the J-B times respectively. These values are about the same as those at the western stations in Japan. It may be suggested that the "$\bar{v}-\delta\bar{v}$" region extends at least to the continental margin.

Raleigh (1967) discussed that the dehydration of serpentinite can reduce the mantle velocity at island arcs. This mechanism may be responsible for the low velocity at a relatively shallow part of, and in the neighborhood of, the deep seismic plane. However, this alone may not be a cause of the low velocity over the entire region above the deep seismic plane.

Conclusion

From the P wave delays at Japanese stations, it is strongly suggested that a P-velocity contrast of about 0.4 km/sec exists beneath Japan between the top 250 km of the mantle divided by the deep seismic plane; the velocity is low on the continental side, and high on the ocean side. From the amplitude ratio of PcP to P, the average value of Q in the mantle on the continental side is estimated as 80. These features can be explained in terms of a temperature contrast of about 500°C coupled with a partial melting of about 2%.

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40. Longshot の日本への走時と島弧の構造

地震研究所　金　森　博　雄

1965 年 10 月 29 日に行なわれた地下爆発 Longshot の日本への走時をしらべた．*P* 波の走時は Jeffreys-Bullen の走時にくらべて筑波で約 3 sec 早く，近畿，中国地方の観測点では約 1.5 sec 早い．*PcP* についても同様の傾向がみられた．このような系統的な走時異常は深発地震面に関係する構造異常によつて説明できる．もつとも簡単なモデルは，深発地震面を境にして深さ約 250 km より浅いマントル内での *P* 波の平均的な速さが陸側では海側においてより約 0.4 km/sec 遅いとするものである．筑波と白木で観測された *PcP* と *P* の振幅比は著しく異なる．この事実は陸側のマントルでの *Q* が海側での値にくらべて小さく，約 80 であるとすれば説明できる．このような低速度と小さい *Q* を説明するためには陸側のマントルでの温度が海側にくらべて平均的に 500°C 高く，同時に約 2% 程度の部分溶融があるとすればよい．日本の走時を世界の他の観測所での値と比較すると，日本におけるもつとも早い走時はシールド地域のものに対応し，もつとも遅い走時は中生代以後の比較的若い造山帯での値に対応する．