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55. *Seismicity and Travel-Time Anomaly in and around Japan.*

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Abstract

This paper is concerned with the seismicity and related problems in and around the Japanese Islands. The area covered in this study is bounded by latitudes 25°N and 50°N, and longitudes 128°E and 150°E. About 410 hypocenters of earthquakes which occurred during the year 1967 have been relocated. The accuracy of the epicenter and depth determinations is improved by inclusion of accurate data from nearby stations. The absolute error of the epicenter and depth is estimated to be about 20 km. The thickness of the intermediate and deep seismic plane in the Kuril-Hokkaido region is determined to be about 20 km; this is significantly smaller than the value previously determined. The intermediate and deep earthquakes are confined to a thin zone about 20 km thick in the Honshu region and about 50 km in the Izu-Bonin Islands region. In the Izu-Bonin Islands region, the seismic activity is very low at depths from 50 to 250 km. This depth range coincides approximately with the depth of the low-velocity layer in this region. The distribution of intermediate and deep earthquakes is not uniform in space and time. They tend to swarm at a certain place on the seismic plane. The spatial distribution of intermediate and deep earthquakes varies from month to month. However, the seismicity over a one year period resembles closely that over a much longer period. The relation between the structural anomaly and the seismic activity is examined by using travel-time anomalies observed at Japanese stations. A plate-like structure exists beneath the Japanese Islands; it is more remarkable in the southern part than in the northern part. The low-velocity zone below the seismic plane seems to be delimited to a region parallel to the seismic plane.

1. Introduction

The distribution of earthquake hypocenters gives an important clue to the understanding of global tectonics [e.g., *Gutenberg and Richter*, 1954]. The recent development of the hypothesis of sea-floor spreading

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demands more and more detailed picture of spatial as well as temporal distribution of earthquakes. The seismicity at island arcs where the lithosphere, created at mid-oceanic ridges, is presumably sinking back into the mantle is of particular interest; various physical and chemical processes taking place beneath island arcs can be inferred from the detailed study of seismicity. The present work is concerned with the seismicity and its related problems in and around the Japanese Islands, a prominent arcuate structure of the Circum Pacific region. This region has been already studied by many workers [e.g., *Wadati*, 1935; *Katsumata*, 1956, 1966, 1967; *Miyamura*, 1962; *Katsumata and Sykes*, 1969].

However, several important questions do not seem to be answered completely. Among those are: (1) What is the absolute accuracy of the epicenter and depth determinations of the earthquakes in this region? (2) Does the thickness of the focal plane of intermediate and deep earthquakes vary from one place to another? (3) What is the relation between the depth distribution of earthquakes and the mantle structure beneath Japan? (4) How does the seismicity over a relatively short time interval, say 1 year, compare with that over a longer period of time? (5) What is the nature of the structural anomaly associated with the seismic activity?

The present paper attempts to answer, at least partially, these questions so that we can understand better the physics of island arcs.

The area covered in this study is bounded by latitudes 25°N and 50°N, and longitudes 128°E and 150°E (see Figs. 1 and 2). About 410 hypocenters of earthquakes which occurred during the year 1967 have been examined.

2. Accuracy of epicenter and depth determinations

About 410 earthquakes which occurred during the year 1967 were relocated (Figs. 1 and 2). The area covered is bounded by latitudes 25°N and 50°N, and longitudes 128°E and 150°E. For the relocation the standard method which employs the non-linear least-squares method and the Jeffreys-Bullen travel-time table was used. The effect of the earth's ellipticity and the station height correction were included [*Bullen*, 1963]. *P* wave travel-times reported in the Earthquake Data Report (EDR) of U.S. Coast and Geodetic Survey (USCGS) and the Seismological Bulletin of the Japan Meteorological Agency (JMA) were mainly used. In addition to these, the Seismological Bulletin of the Sakhalin Complex Science-Research Institute (USSR) and the preliminary reports of micro-earthquake observatories at Urakawa (KMU), Dodaira (DDR, TSK, KYS), Wakayama (WKU, OIS), Shiraki (SHK), Tottori (FO, HM, IZ, MZ, OY),

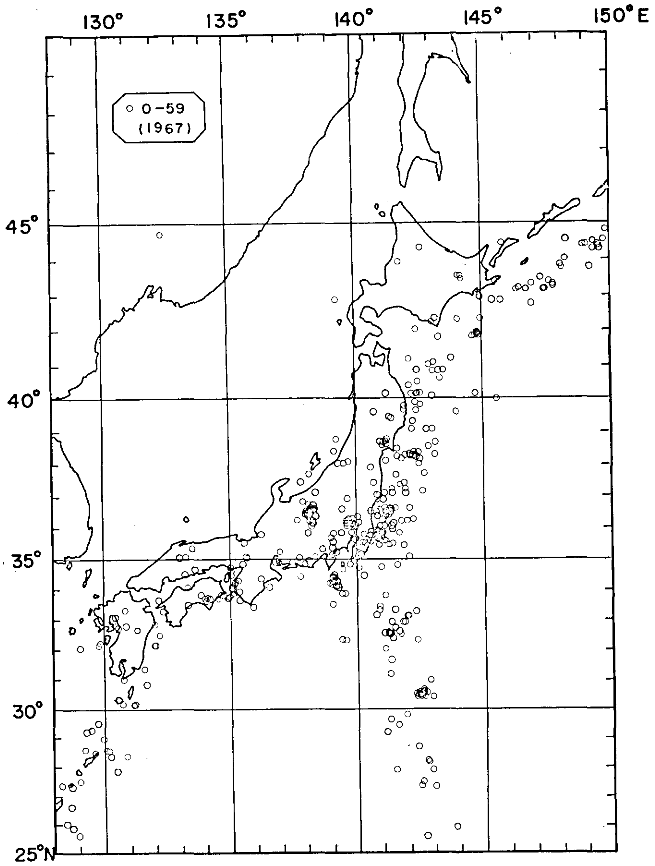

Fig. 1. Shallow earthquakes (depth 0 to 50 km) in and around Japan during the year 1967. The relocated hypocenters and the JMA hypocenters which are not relocated are plotted.

Kochi (URS, WMY, IHR), and Matsushiro (MAT) were used. The travel-times of the surface reflection *pP* were used to check the focal depths wherever available.

In general, the accuracy of the hypocenter determination depends on the travel-time table and the method of computation. However, the timing accuracy and the distribution of stations around the epicenter have most important effects upon the accuracy in earthquake locations. The standard error as a result of computation does not always indicate the real accuracy. Further, the large lateral heterogeneity at island arcs, suggested by many recent works, may introduce a significant azimuthal variation in the travel-time of seismic waves and may yield systematic errors in the location of hypocenters. It is difficult to estimate the real accuracy of the hypocenter location in the region where

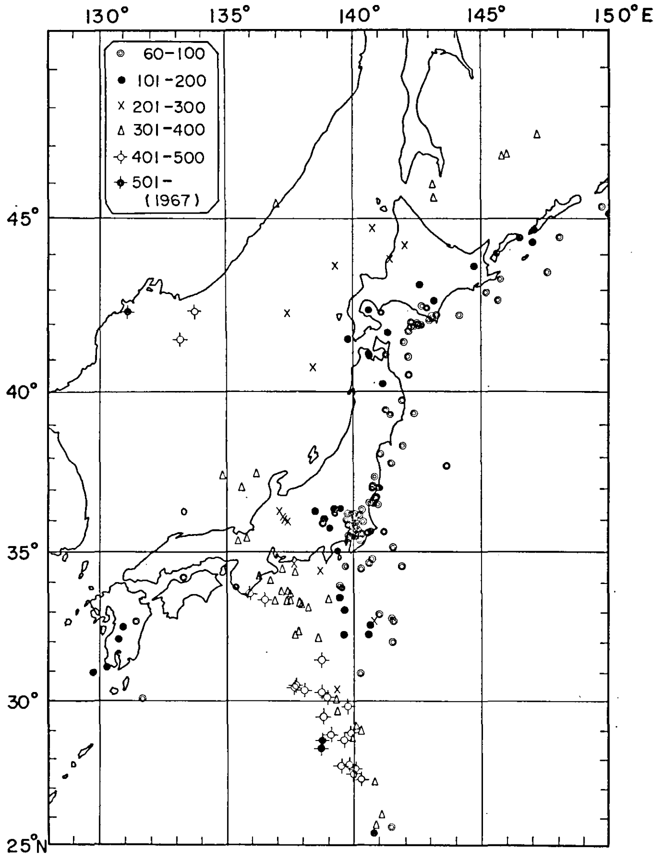

Fig. 2. Intermediate and deep earthquakes (depth>60 km) in and around Japan during the year 1967.

the mantle structure is not known in detail. However, if accurate readings at nearby stations distributed over a wide azimuthal range around the epicenter are available, the hypocenters determined with such data will not deviate very much from the true hypocenter even if the travel-time table is slightly in error. For this reason, it is suggested that the earthquakes in the Kuril-Hokkaido region are most accurately relocated; for this region the reports of micro-earthquake observatories and the USSR stations were included. The hypocenters in the Izu-Bonin Islands region could not be determined so precisely as those in the Kuril-Hokkaido region because of the lack of the data from nearby stations. The accuracy of the present determination was examined by using the earthquakes which occurred in the vicinity of Matsushiro and in the Kuril-Hokkaido region. During the Matsushiro earthquake swarm (1965-1969), a temporary observatory net of the Earthquake Research Institute

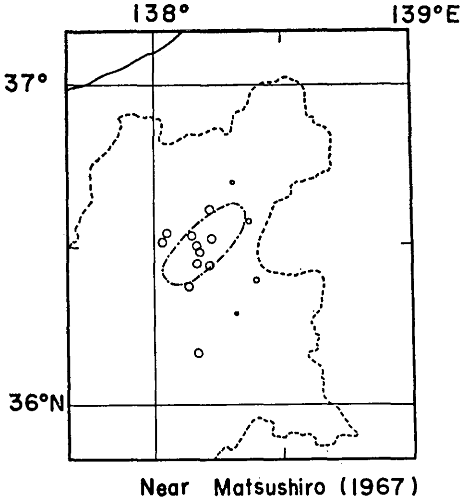

Fig. 3. The epicenters of all the earthquakes which occurred near Matsushiro during the year 1967; circles indicate the relocated epicenters. Larger symbols indicate more accurate determinations of hypocenters. The dot-dash curve indicates the aftershock area determined by the temporary seismographic network of the Earthquake Research Institute.

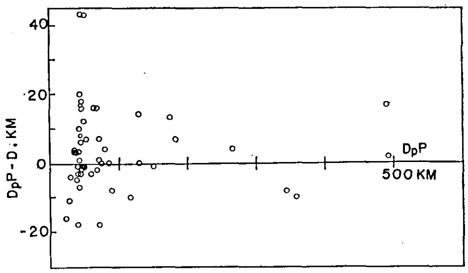

Fig. 4. Difference of depths determined by *P* and *pP* phases. *D* indicates the redetermined depth; *DpP* indicates the depth computed from *pP*.

of Tokyo University was installed [*Hagiwara and Iwata*, 1968; *Hamada*, 1969] in Matsushiro to locate the local shocks very precisely. The swarm area during the year 1697 was clearly defined as shown in Fig. 3. We relocated several events which occurred in the Matsushiro region by our method using the data from distant stations. The epicenters relocated are compared with the swarm area in Fig. 3. The maximum distance between the relocated epicenters and the swarm area is less than 20 km. This value may be regarded as the absolute accuracy of the present epicenter determination all over Honshu.

An Ocean-Bottom Seismograph field experiment (OBS) was conducted during the period October 22 to December 13, 1966, in the Kuril Islands region by Texas Instruments Incorporated [*Mcdermott et al.*, 1967]. The data obtained by this experiment provide a good means of testing the accuracy of the present relocation in the Kuril-Hokkaido region. About 13 earthquakes which occurred in the neighborhood of the experimental site during this period were relocated by adding the data obtained by OBS to those which are normally used in the present study. The hypocenters thus determined are probably very accurate and may be regarded

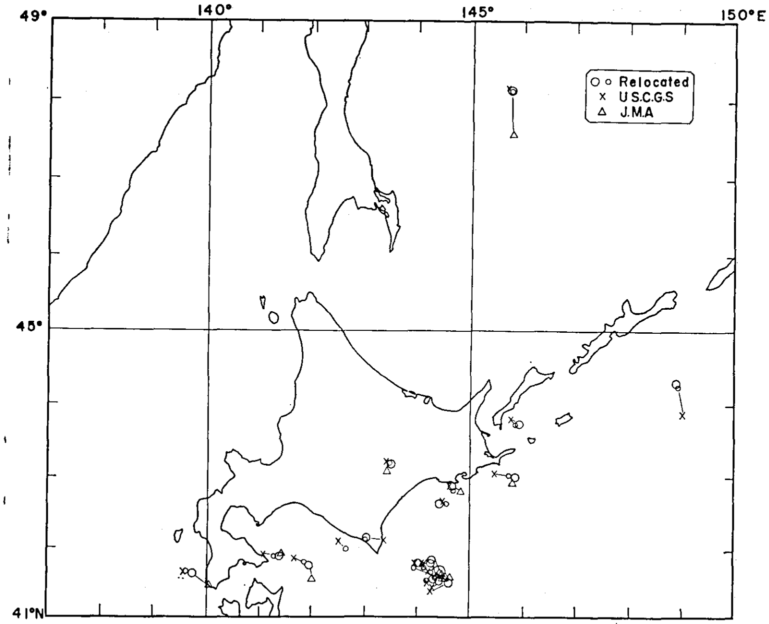

Fig. 5-*a*. Comparison of the relocated, the USCGS and the JMA epicenters in the Kuril-Hokkaido region (1966).

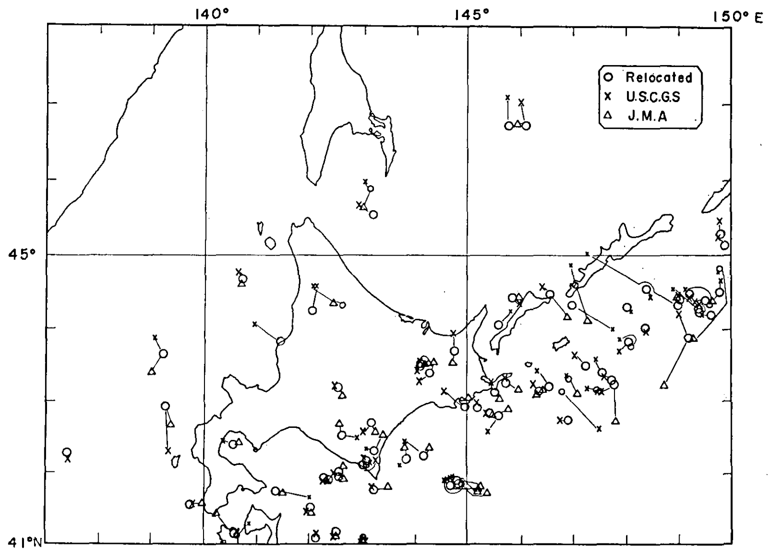

Fig. 5-*b*. Comparison of the relocated, the USCGS and the JMA epicenters in the Kuril-Hokkaido region (1967).

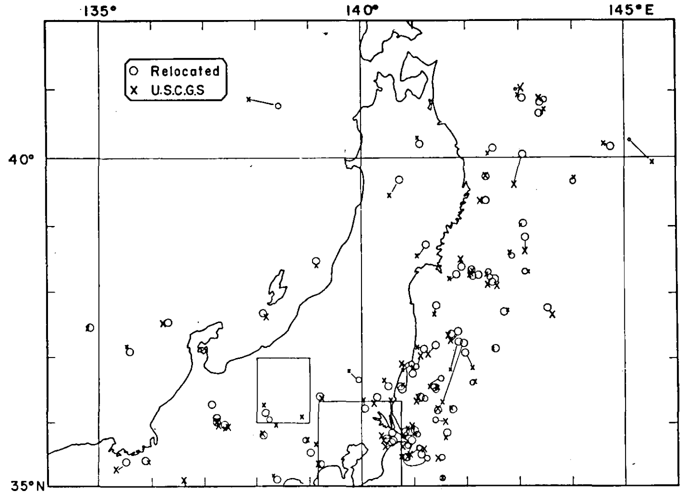

Fig. 6-*a*. Comparison of the relocated, and the USCGS epicenters in northeastern Japan. The areas enclosed by the rectangles are shown in Fig. 6-*b* and Fig. 6-*c*.

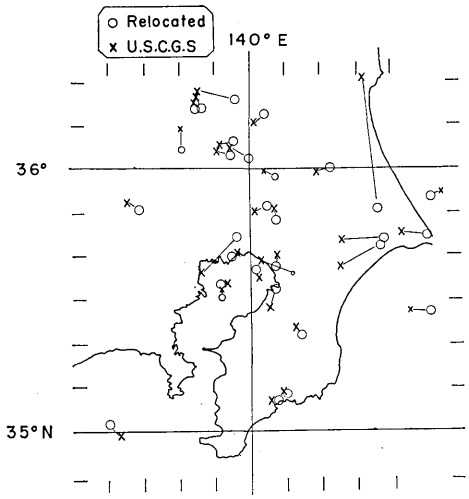

Fig. 6-*b*. The enlarged map of area enclosed by the rectangle in Fig. 6-*a* (Kanto region).

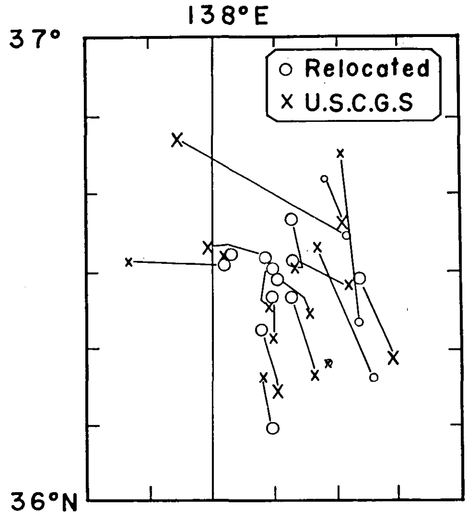

Fig. 6-*c*. The enlarged map of the area enclosed by the rectangle in Fig. 6-*a* (Matsushiro region).

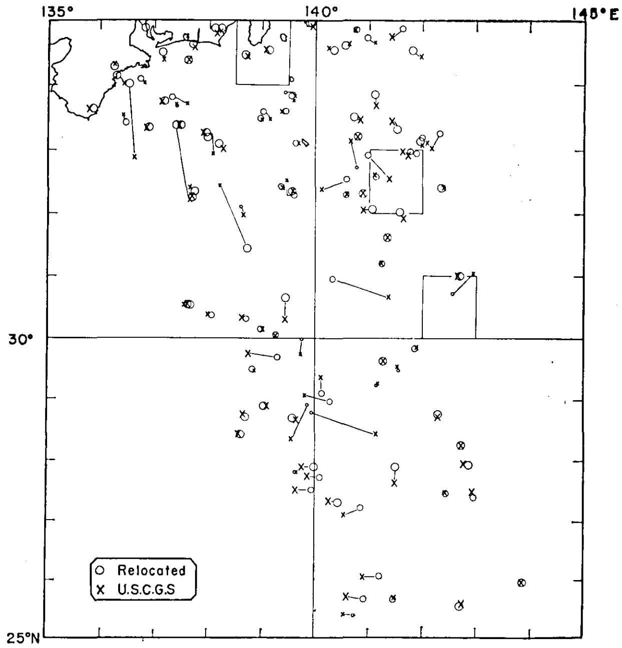

Fig. 7-*a*. Comparison of the relocated and the USCGS epicenters in the Izu-Bonin Islands region. The areas enclosed by the rectangles are shown in Fig. 7-*b*, Fig. 7-*c*, and Fig. 7-*d*.

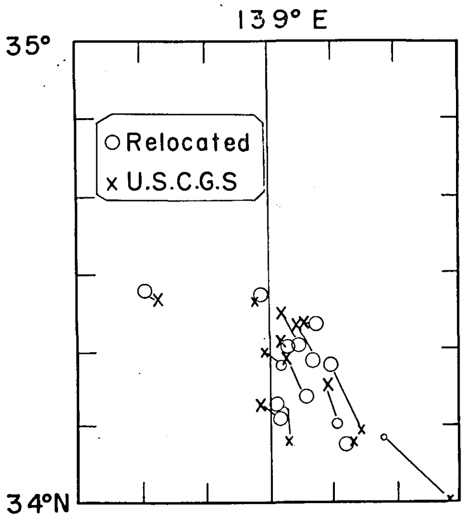

Fig. 7-*b*. The enlarged map of the area enclosed by the rectangle in Fig. 7-*a*.

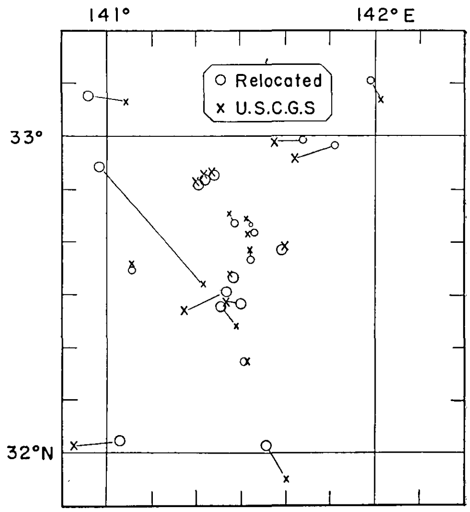

Fig. 7-*c*. The enlarged map of the area enclosed by the rectangle in Fig. 7-*a*.

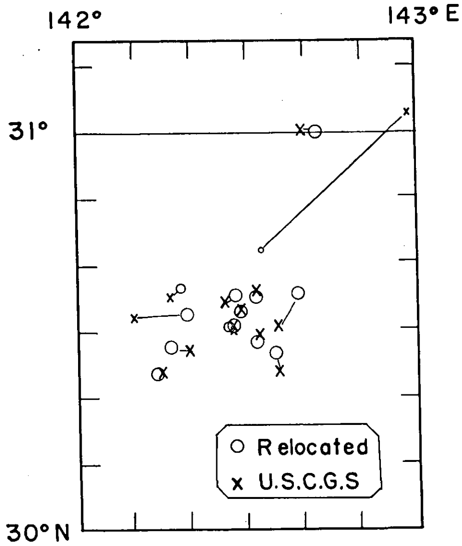

Fig. 7-*d*. The enlarged map of the area enclosed by the rectangle in Fig. 7-*a*.

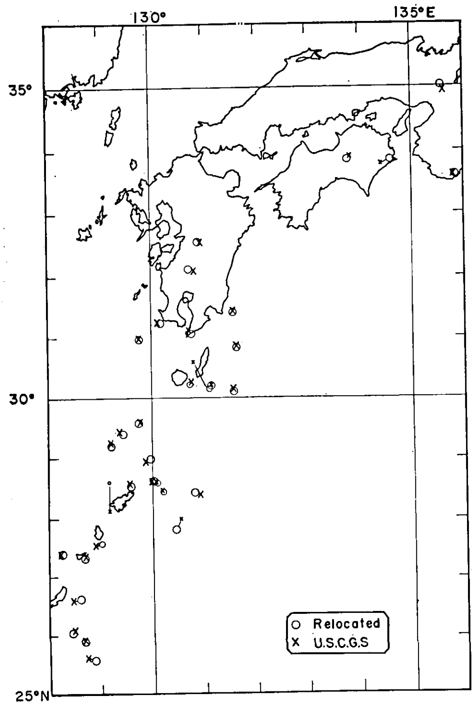

Fig. 8. Comparison of the relocated and the USCGS epicenters in the Kyushu region.

as the true hypocenters. These hypocenters are compared with those determined by our ordinary procedure without including the OBS data. The differences between the two kinds of hypocenters are less than 20 km (Fig. 5-*a*). In view of these results, the absolute accuracy of the epicenter determination in this region is probably greater than 20 km. Another important result is that the USCGS epicenter determinations seem more accurate, as a whole, than those of JMA.

The arrival times of the surface reflection *pP* were reported for many earthquakes. The focal depths determined by using *pP*, if identified correctly, are very accurate because the difference of arrival time between *P* and *pP* depends mainly on the depth but very little on the epicenter distance. The redetermined focal depths were compared with those determined by *pP* wherever they were available. The difference of the two depths was found to be less than 20 km (Fig. 4). This value, 20 km, can be regarded as a measure of the accuracy of the depth determination of earthquakes for which no *pP* data are available. The comparison of the relocated and the USCGS epicenters is shown in Figs. 5-*a* to 9. The maximum deviation of the USCGS epicenter and depth from the " true " (relocated) epicenter and depth amounts to 150 and

80 km respectively. When the focal depth became negative in the process of the computation, it was restrained at 33 km. These negative depths are obviously due to systematic errors in the assumed travel-time table or a lack of data from close stations. The number of earthquakes whose depth became negative decreased in the present determination as compared with the USCGS determination; this is a result of the inclusion of precise data from nearby stations. To give a rough estimate of the quality of the data the located hypocenters were grouped, following *Katsumata and Sykes* [1969], into four classifications, *A*, *B*, *C* and *D* according to the standard errors of the computations. Events for which the standard errors for all elements (latitude, longitude and depth) were less than 10 km were labeled *A*; events for which the standard errors for any element exceed 10 km but for which none exceed 20 km were labeled *B*; events for which the standard errors for any element exceed 20 km but for which none exceed 30 km labeled *C*; events for which the standard errors for any element exceed 30 km were labeled *D*, but events *D* were not plotted. The ratios of each class to the total number of events are as follows; *A*=71%, *B*=22%, *C*=6%, and *D*=1%.

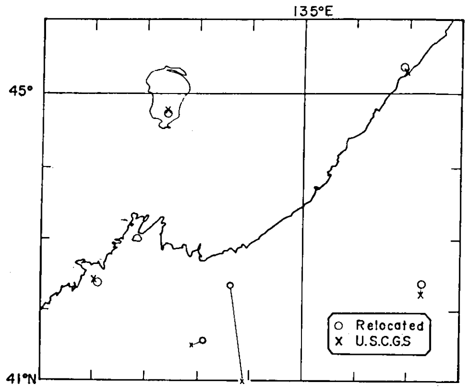

Fig. 9. Comparison of the relocated and the USCGS epicenters northwest of the Japan Sea.

3. Seismic plane

According to *Sykes et al.* [1969], it seems that the higher the accuracy of the hypocenter determination, the thinner is the deep seismic zone inferred. It was suggested that the thickness of the deep seismic zone appears to be less than 100 km and may, at least for some regions, be less than 20 km [*Isacks et al.*, 1968].

The locations of several vertical cross-sections perpendicular to the trench are shown in Fig. 10 as *A-A'*, *B-B'*, *C-C'*, *D-D'*, *E-E'* and *F-F'*. These profiles, shown in Figs. 11 to 16, contain the earthquakes which occurred within about 150 km on either side of the each section. The vertical cross-sections perpendicular to the Kuril trench are shown in Figs. 11 and 12. *Sykes* [1966] concluded that the hypocenters of nearly all well-located events are confined to a zone about 50 to 100 km thick

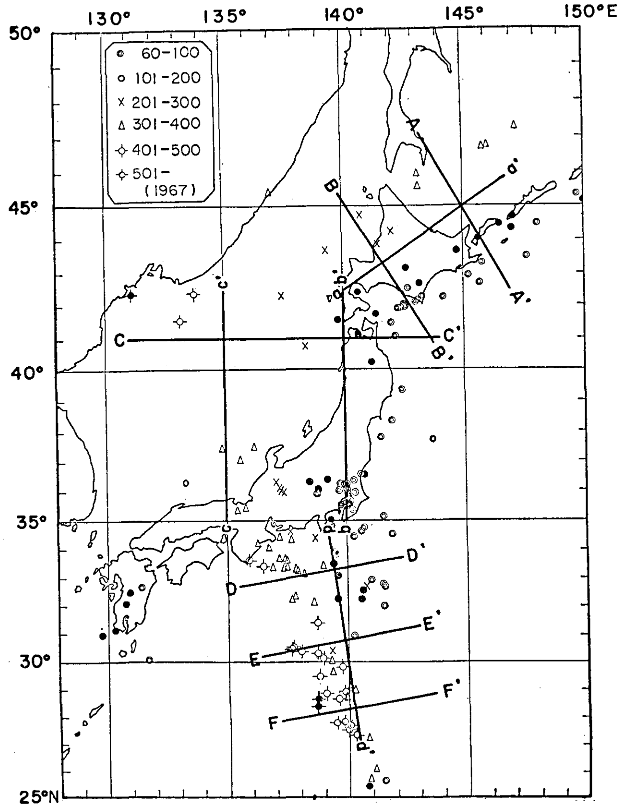

Fig. 10. Index map showing the location of cross-sections perpendicular to the trench shown in Fig. 11 to Fig. 16, and those parallel to the trench shown in Fig. 17 to Fig. 20.

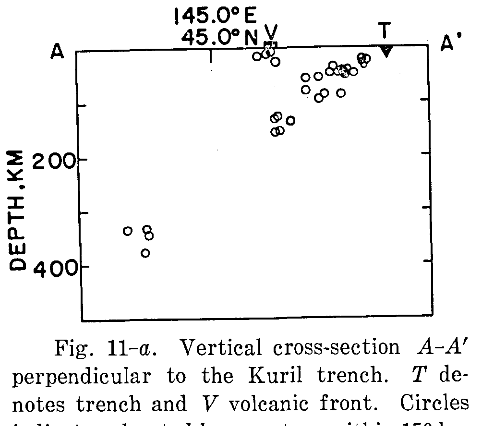

Fig. 11-*a*. Vertical cross-section *A–A'* perpendicular to the Kuril trench. *T* denotes trench and *V* volcanic front. Circles indicate relocated hypocenters within 150 km of the section. Larger circles denote more precise hypocenters. The horizontal scale is the same as the vertical scale.

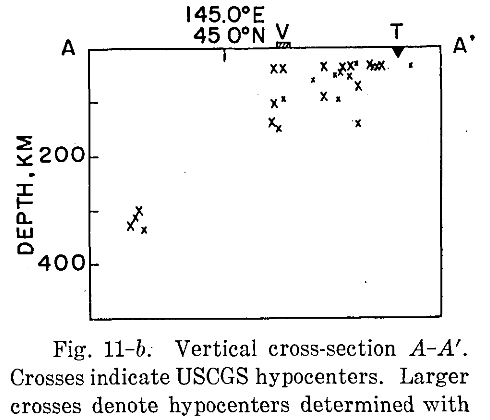

Fig. 11-*b*. Vertical cross-section *A–A'*. Crosses indicate USCGS hypocenters. Larger crosses denote hypocenters determined with smaller standard errors.

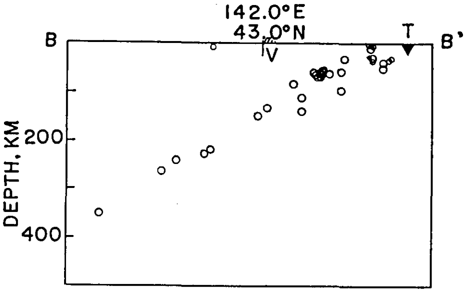

Fig. 12-*a*. Vertical section *B-B′* (relocated hypocenters).

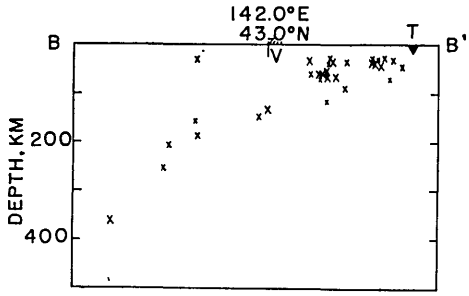

Fig. 12-*b*. Vertical section *B-B′* (USCGS hypocenters).

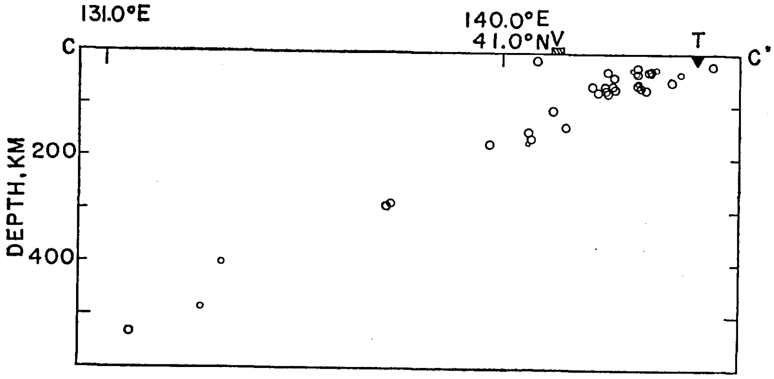

Fig. 13-*a*. Vertical section *C-C′* (relocated hypocenters).

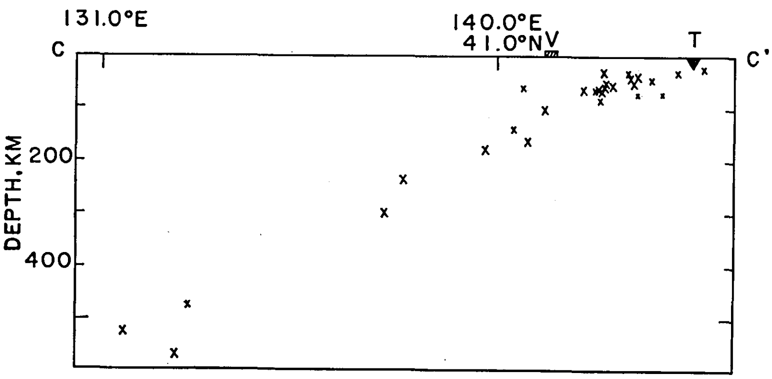

Fig. 13-*b*. Vertical section *C-C′* (USCGS hypocenters).

in the Kuril-Kamchatka region. In this study, however, the seismic zone in this region appears to be thinner than 50 to 100 km. The thickness as inferred from the USCGS hypocenters is about 100 km (see Fig. 11-*b*), but that by the present study is about 20 km (see Fig. 11-*a*). *Sykes* [1966] suggested that the hypocenters near 44°N, 149°E seem to delineate at least two very narrow zones of activity. The present data also show two similar narrow zones of seismic activity (Fig. 11-*a*). The scatter of the hypocenters at the depth of about 350 km in Fig. 11-*a* is due to the difference in the locations of these hypocenters in the direction perpendicular to the cross-section. These are about 300 km apart from one another; the actual seismic zone is probably thinner than Fig. 11-*a* indicates.

Figs. 12-*a* and 12-*b* show the distribution of the hypocenters along the profile *B-B′* as determined by the present study and USCGS respectively. The hypocenters determined by the present study are confined to a remarkably narrow zone. This result is probably due to the improved accuracy in the present study. Figs. 13-*a* and 13-*b* show the profile *C-C′*. A remarkable linearity should be noted. The thickness of the seismic zone along these profiles is probably less than 20 km.

For Izu-Bonin Islands, *Katsu-*

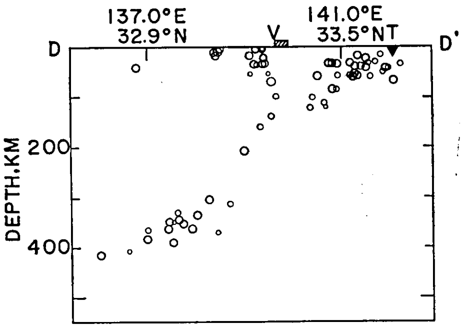

Fig. 14. Vertical section *D-D′* perpendicular to the Izu-Bonin trench. *T* denotes trench and *V* volcanic front. Circles indicate relocated hypocenters within 130 km of the section. Larger circles denote more precise hypocenter locations. The horizontal scale is the same as the vertical scale.

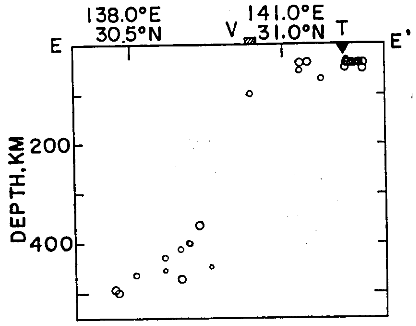

Fig. 15. Vertical section *E-E′*.

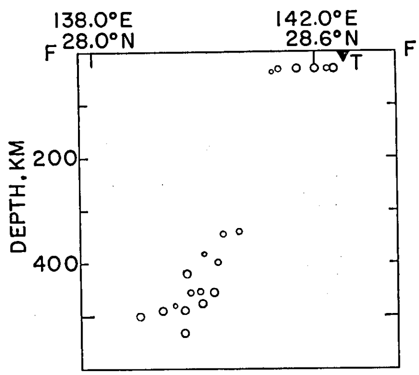

Fig. 16. Vertical section *F-F′*.

mata and Sykes [1969] concluded that the hypocenters are confined to a zone thinner than 50 km in the northern portion of the Izu Islands. In the Bonin Islands region, they found that the zone extends over a thickness of more than 100 km. They suggested that this apparently large thickness of the seismic zones might reflect the error in the depth determination. Since the accuracy of the hypocenter determination in the Izu Islands is higher than that in the Bonin Islands, we might conclude that the earthquakes are confined to a zone probably thinner than 50 km in the Izu-Bonin Islands.

The dip of the seismic plane varies from place to place. In the Kuril-Hokkaido region the dip of the seismic plane shown in Figs. 11 and 12 is approximately 30° to 40° and is nearly constant over the depth from 0 to 400 km. The dip of the seismic plane under the Japan Sea shown in Fig. 13 is slightly less than 30°. In the Izu-Bonin Islands there is an indication that the dip angle decreases from about 45° to 30° at a depth of approximately 300 km.

4. Depth distribution

Although the earthquakes appear to be distributed uniformly in space

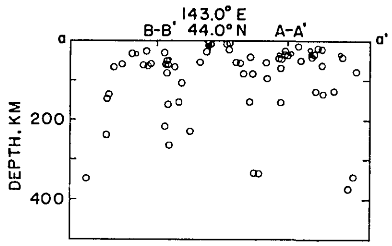

Fig. 17. Vertical section *a–a'* parallel to the trench. Larger circles denote more precise hypocenter locations. The horizontal scale is the same as the vertical scale.

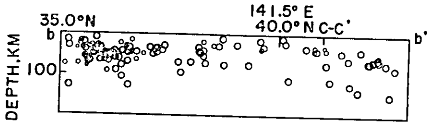

Fig. 18. Vertical section *b–b'*.

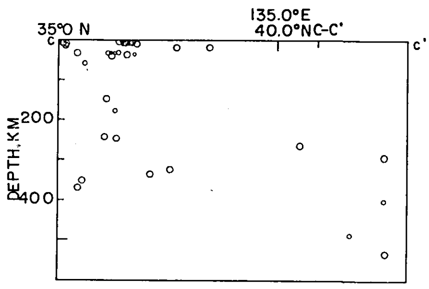

Fig. 19. Vertical section *c–c'*.

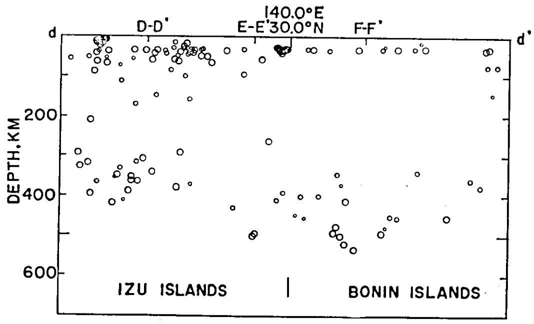

Fig. 20. Vertical section *d–d'*.

in Figs. 1 and 2, they actually tend to swarm at certain places on the seismic plane. The locations of several vertical sections parallel to the trench are shown in Fig. 10; each profile, *a–a′*, *b–b′*, *c–c′* and *d–d′*, is shown in Figs. 17 to 20.

Fig. 17 shows the distribution of the hypocenters along the profile *a–a′*. The activity of intermediate and deep earthquakes is extremely low in the region beneath the region where the activity of shallow earthquakes is high, and *vice versa*. Fig. 19 shows the distribution of the hypocenters under the Japan Sea; deep earthquakes are rare but appear to have the same trend as that in the Kuril-Hokkaido region (profile *a–a′*). In the Izu-Bonin Islands region shown in Fig. 20, deep earthquakes swarm at the depth between 300 to 400 km off the east coast of Kii peninsula (near 33.4°N, 137°E). The lower bound of the seismic activity gradually deepens from about 400 km in the northern portion of the Izu Islands to about 500 km near the Bonin Islands. Seismic activity is very low at the depth between 50 and 250 km in the Bonin Islands region. This zone coincides with the low-velocity layer in this region which was suggested by *Kanamori and Abe* [1968].

A zone of shallow earthquakes can be traced along the axis of the trench from the Kuril Islands to Bonin Islands. This activity is not uniform all over these regions. Shallow earthquakes swarm at several localities. Fig. 18 shows the distribution of the hypocenters in the Honshu region. The shallow earthquakes from 60 to 80 km cluster in a region near latitude 36°N, in the Kanto region. The earthquakes from 0 to 10 km depth cluster in the Matsushiro area and its neighborhood, near latitude 36.5°N, as shown in Fig. 19. In the Izu-Bonin Islands the shallow earthquakes swarm at several places at latitudes about 34°N, 33°N and 31°N as shown in Fig. 20. The regions where shallow earthquakes swarm are enclosed by the rectangles in Figs. 6–*a* and 7–*a*.

5. Temporal variation

The seismicity of Japan has heretofore been discussed on the basis of the data obtained over a relatively long period of time. Because of the low magnification seismographs of the Japan Meteorological Agency stations, a period of at least several years was necessary to accumulate the earthquake data adequate for discussions of seismicity. In this case, the quality of the data is not uniform, and the variation of the seismicity over a short period of time cannot be detected. In the present work, owing to the use of the data from high sensitivity stations, it is possible to compare the seismicity over a relatively short period of time.

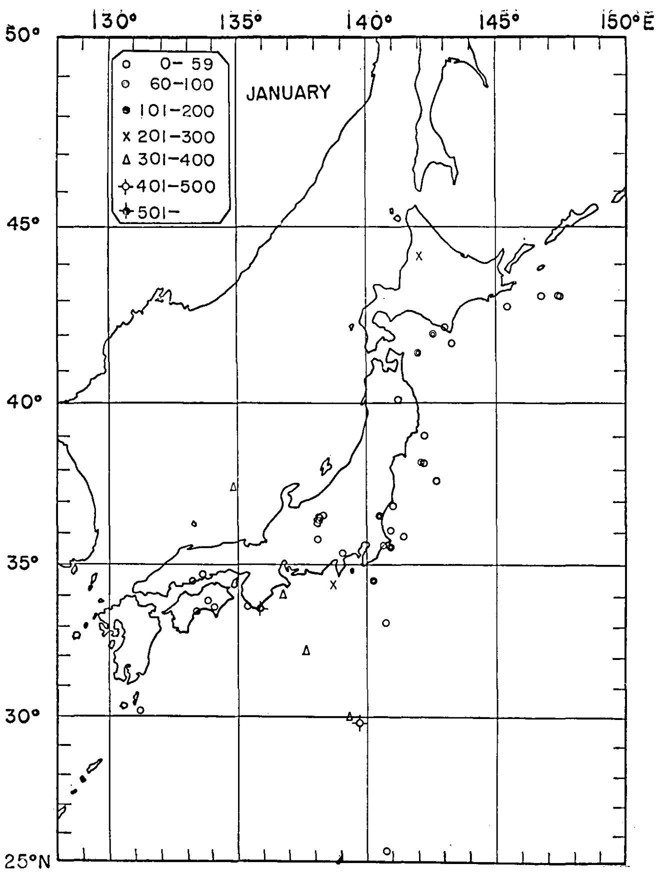

Fig. 21. Distribution of hypocenters in January. Figs. 21 to 32 showing the relocated hypocenters and the JMA hypocenters which are not relocated are plotted for each month.

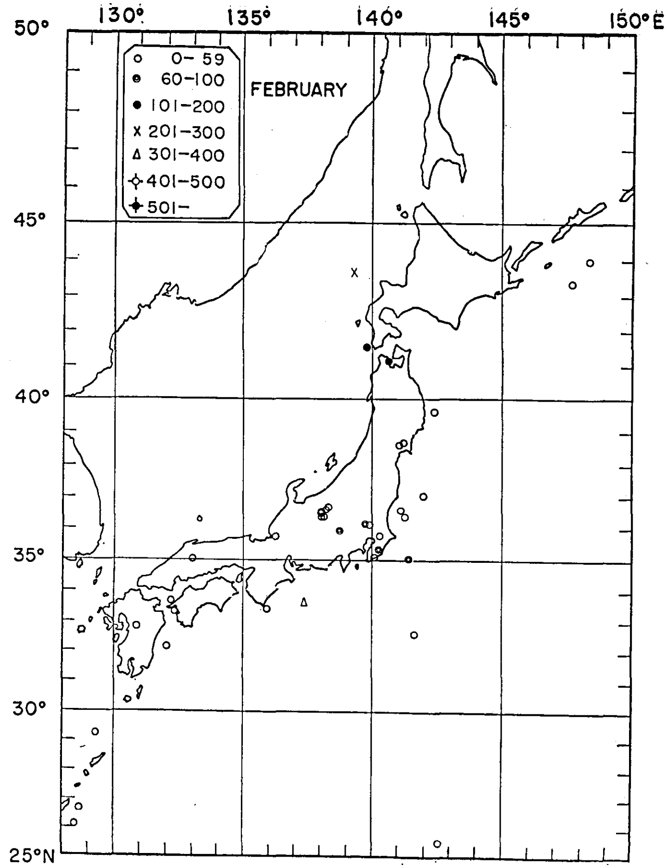

Fig. 22. Distribution of hypocenters in February.

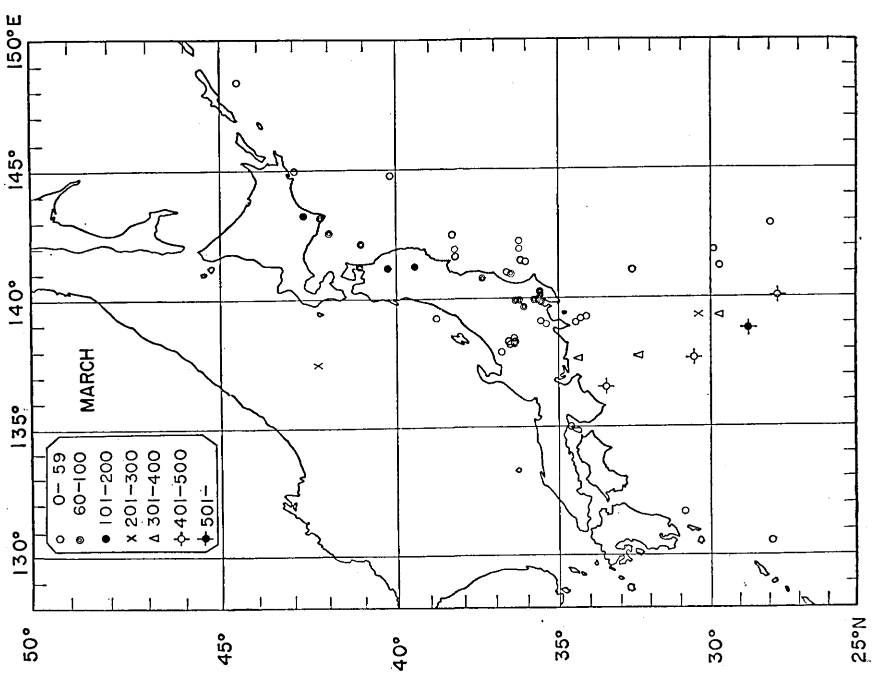

Fig. 23. Distribution of hypocenters in March.

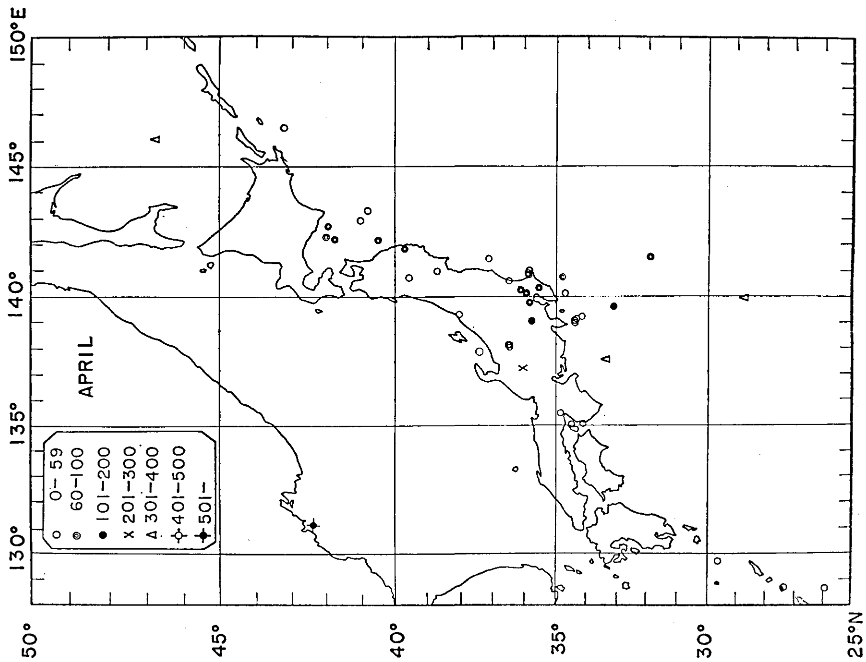

Fig. 24. Distribution of hypocenters in April.

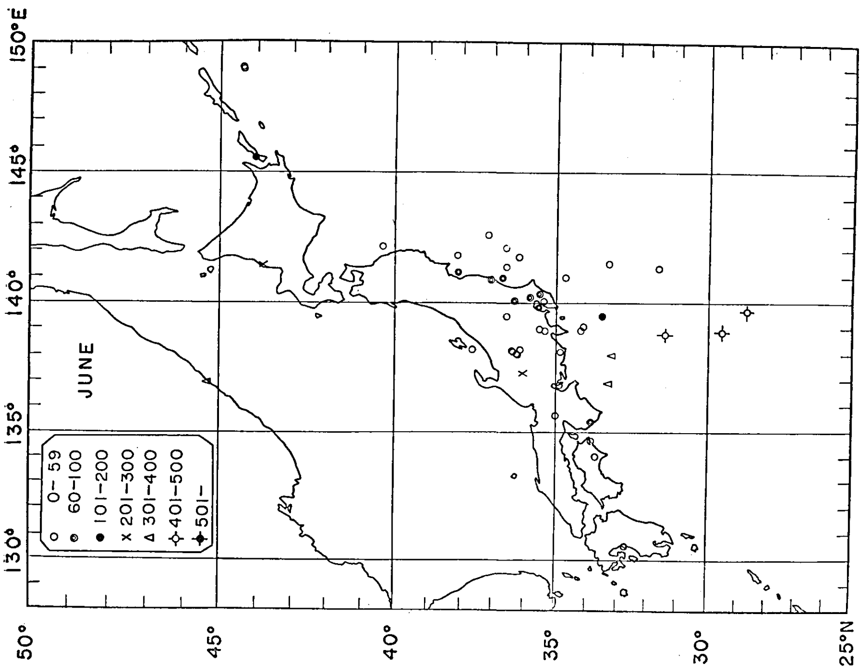

Fig. 26. Distribution of hypocenters in June.

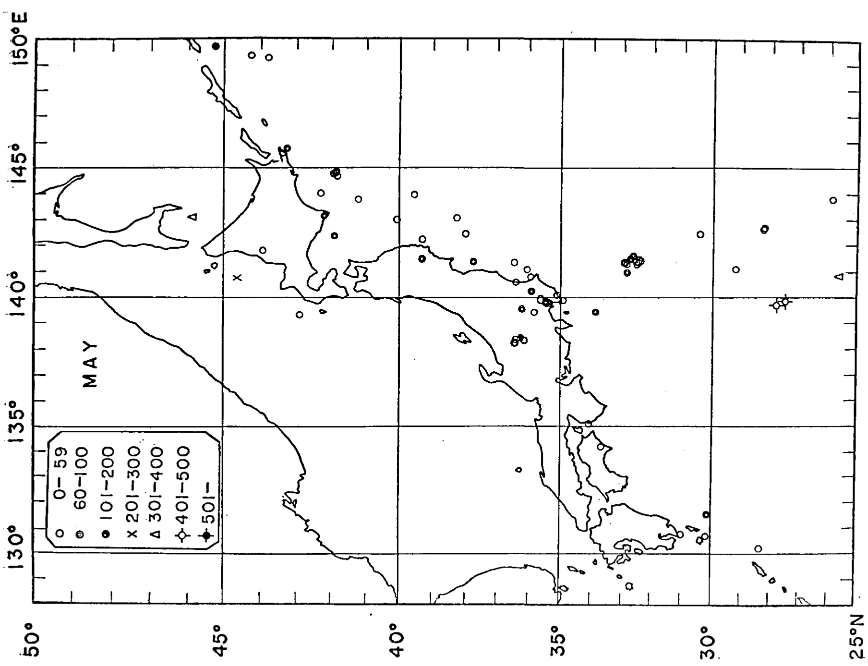

Fig. 25. Distribution of hypocenters in May.

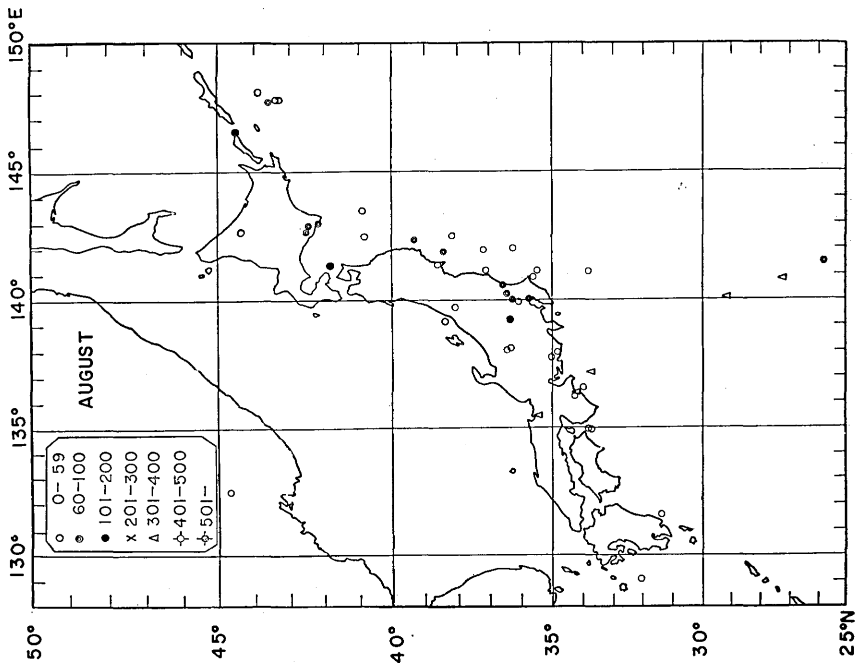

Fig. 28. Distribution of hypocenters in August.

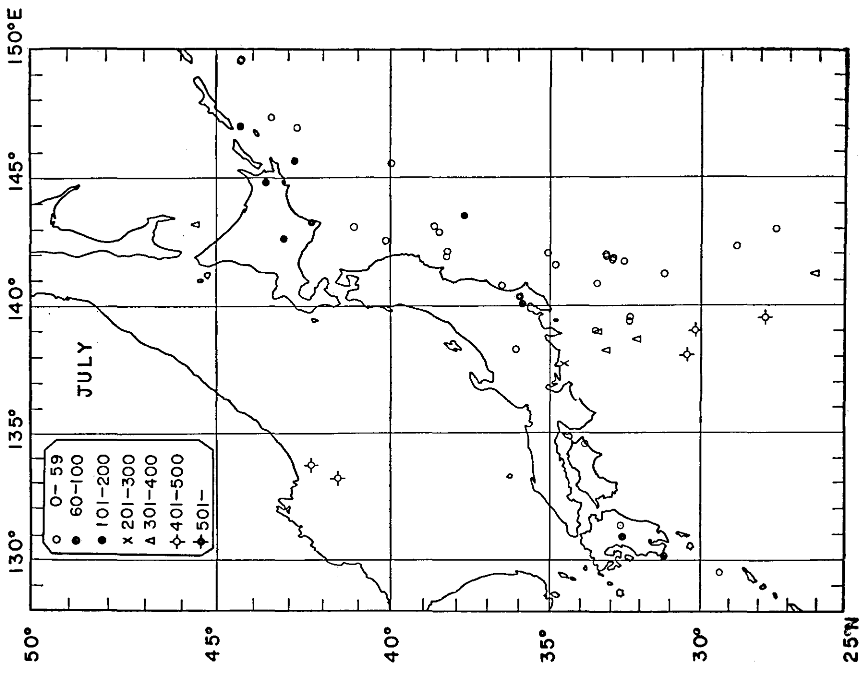

Fig. 27. Distribution of hypocenters in July.

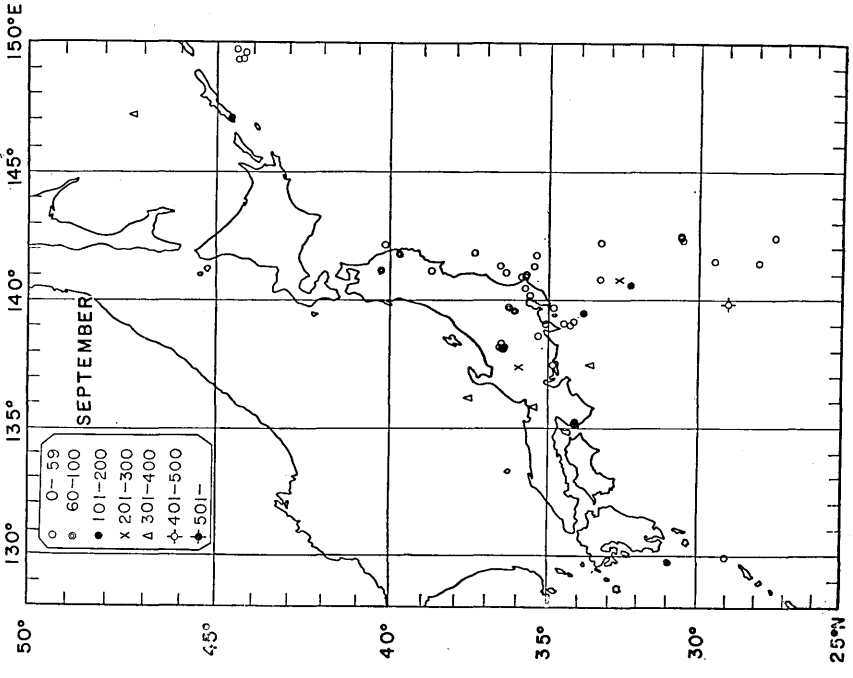

Fig. 29. Distribution of hypocenters in September.

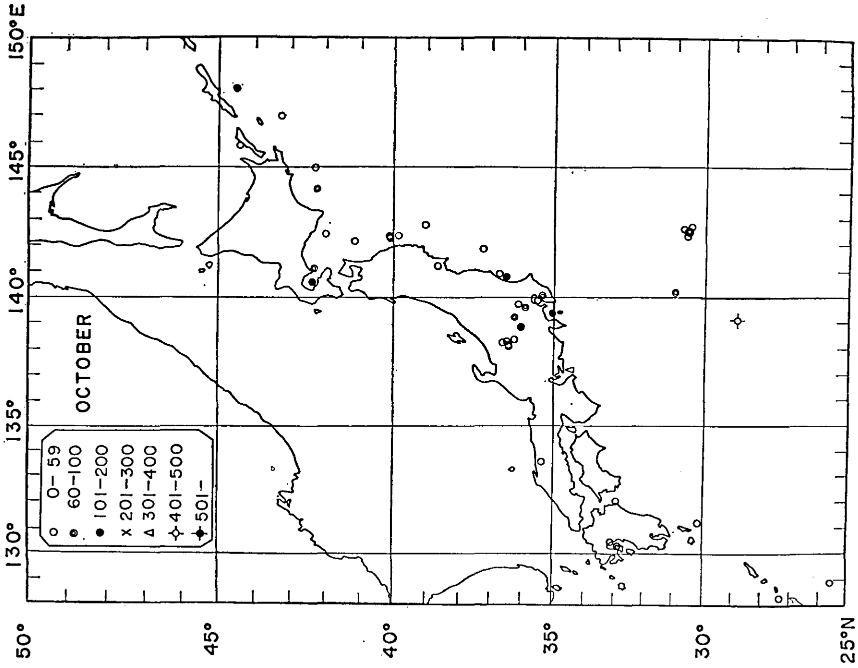

Fig. 30. Distribution of hypocenters in October.

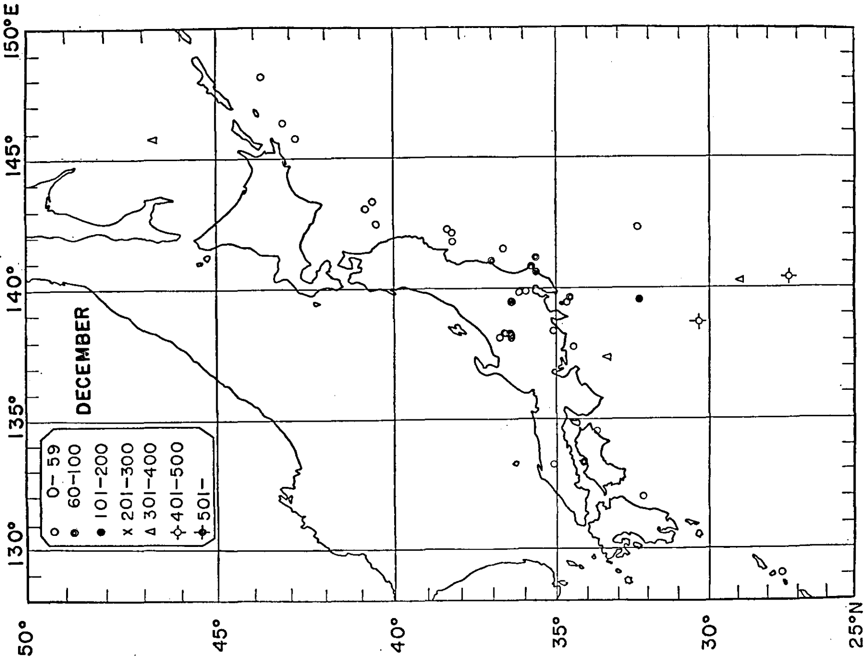

Fig. 32. Distribution of hypocenters in December.

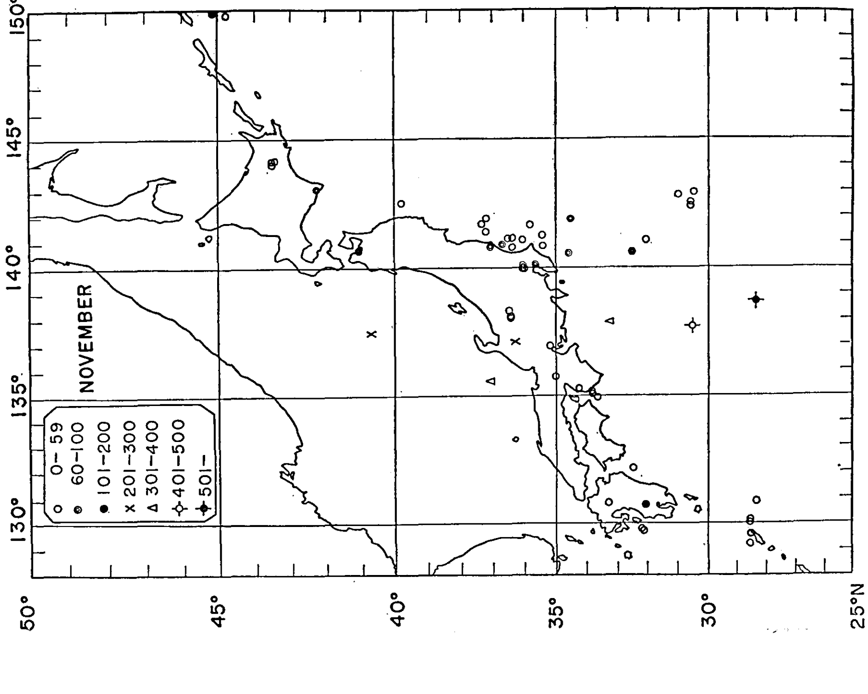

Fig. 31. Distribution of hypocenters in November.

Figs. 21 to 32 show the distribution of earthquakes for each month from January, 1967, to December, 1967. A comparison of these figures indicates that the time variation of the deep earthquake activity appears to be larger than that of shallow earthquakes. For example, in the Izu Islands region, 25°N to 35°N and 137°E to 141°E, the deep earthquake activity which is very high in March (Fig. 23) becomes low through April (Fig. 24) and May (Fig. 25), and again becomes high in June (Fig. 26). The shallow earthquake activity is more uniform than that of deep earthquakes; an example is the shallow earthquake activity along the Sanriku coast (38°N to 41°N and 141°E to 142°E). In the Izu-Bonin Islands region the shallow activity is not as uniform as that of the Sanriku coast, for example the activity is high in July, low in August, and again high in September (Figs. 27, 28 and 29). In general the seismic activity varies considerably from month to month.

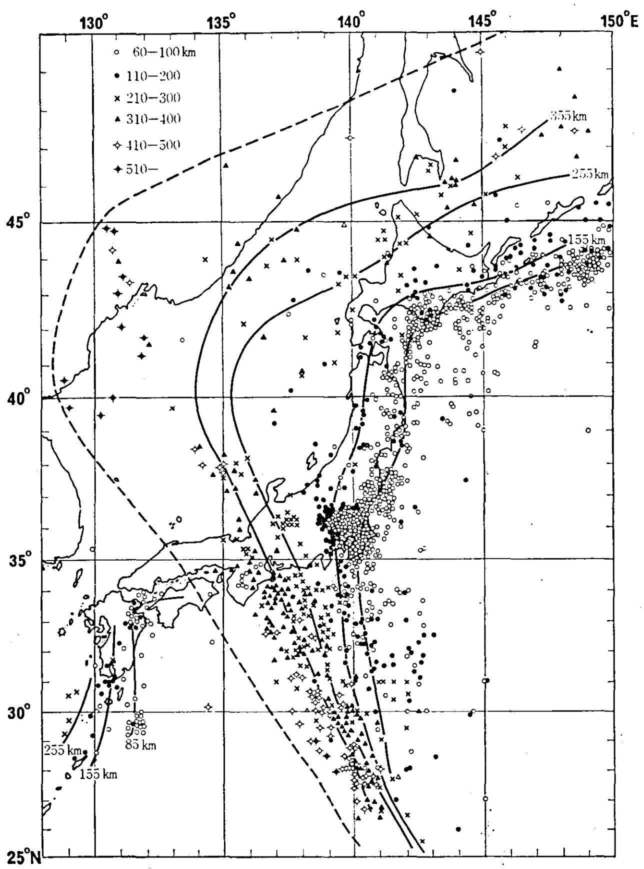

Fig. 33. Intermediate and deep earthquakes in and around Japan for the period 1926 through 1962 as reported by Japan Meteorological Agency. The deep seismic plane is contoured by *Sugimura* (1960). The dashed line defines the western limit of the seismic zone.

In spite of this variation, the distribution of the earthquakes during the year 1967 as a whole resembles closely that during 37 years from 1926 to 1962 [Fig. 33, *Sugimura and Uyeda*, 1968]. The localities where shallow earthquakes cluster during the year 1967 are the same as those during the 37 years. A similar trend is observed for intermediate and deep earthquakes. The earthquakes at depths between 300 and 400 km are numerous off the coast of Noto peninsula, near 35.5°N and 135.5°E, and in the region bounded by 32°N to 35°N and 136°E to 140°E; this trend is common to both periods, the year 1967, and from 1926 to 1962. In both periods the earthquakes at 400 to 500 km depth are numerous in the region bounded by 27°N to 31°N and 137°E to 141°E, and the earthquakes at 200 to 300 km depth are numerous in the region near 36.2°N and 137.3°E. The seismic activity is low off the west coast of northeastern Japan near 38°N to 41°N during both periods. In view of these results, if the data from recent sensitive stations are used, data over a relatively short period of time suffice to describe the general feature of seismicity of Japan.

6. Travel-time anomaly

The nature of the structural anomaly associated with the seismic activity has already been examined in several regions [*Katsumata*, 1960, 1966; *Utsu*, 1966, 1967; *Oliver and Isacks*, 1967; *Kanamori*, 1968; *Utsu and Okada*, 1968; *Davies and McKenzie*, 1969; *Mitronovas et al.*, 1969; *Wadati et al.*, 1969]. *Oliver and Isacks* [1967] concluded that the seismic body waves, particularly S waves, propagating in the anomalous zone roughly defined by the foci of deep earthquakes, are much less subjected to attenuation than are waves of the same type propagating in parts of the mantle at similar depths elsewhere; the average velocities for both phases propagating along the anomalous zone are higher by an average of 1 to 2% than the values for both phases propagating through the aseismic zone in the Tonga-Fiji region. *Utsu* [1967] proposed a model, a model A' (see Fig. 35), in which the velocity of both P and S waves is higher in the deep seismic zone by about 6% than in other portions of the upper mantle beneath the Hokkaido region.

In this section comparisons of travel-times for various paths from focus to station will be made. The hypocenters of earthquakes used were located by using more than twenty stations. The standard errors of the hypocenters of these earthquakes are less than 20 km for all elements; we used such earthquakes as labeled A or B. The observed minus computed travel-time at a station can be regarded as an anomaly of travel-time of a seismic wave from the hypocenter to the station.

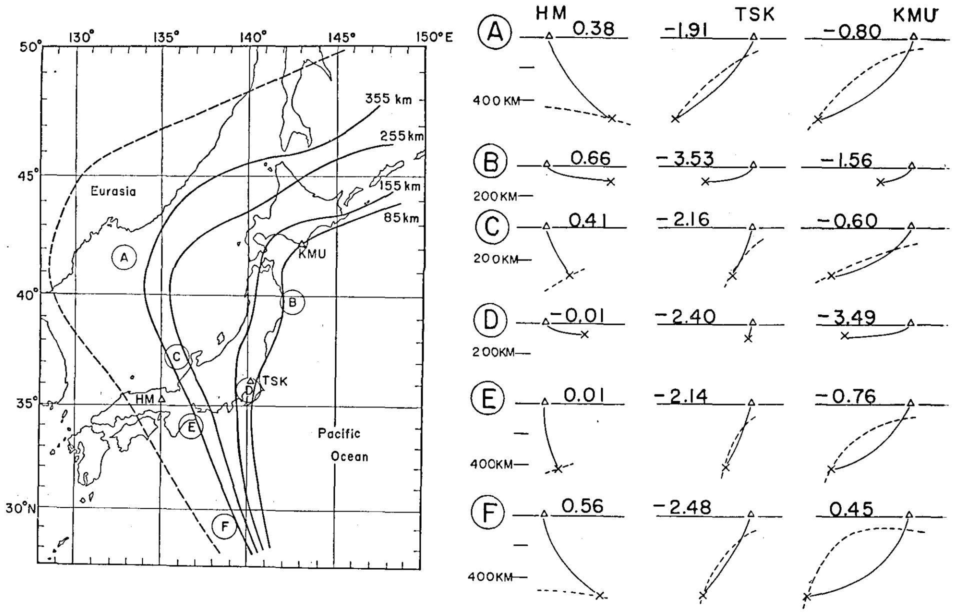

Fig. 34. Relocation between P wave travel-time residuals and the ray paths. Epicenters of earthquakes and stations used for the analysis are shown on the map on the left. The cross-sections of the structure and the ray paths to the stations are shown on the right. The vertical exaggeration is about 2:1. The solid lines indicate the ray paths and the dotted lines indicate the seismic planes.

If a large number of stations distributed over a wide azimuthal range around the epicenter are used to locate the hypocenter, the hypocenter location will not be very much affected by inclusion of one or two anomalous stations. For this reason, the travel-time anomaly as a result of the present computation probably reveals the structural anomaly fairly well.

For comparison, we will use the fractional travel-time anomaly $\Delta t/t$, where t is the travel-time of P wave and Δt is the travel-time residual of P wave. Evidently $\Delta t/t=-\Delta v/v$ where v is the velocity averaged over the ray path from a focus to a station and Δv is the average velocity anomaly along this ray path. The residual Δt is negative (positive) where the average velocity of the P wave along the path is higher (lower) than that for the Jeffreys model.

Fig. 34 shows the relation between $\Delta t/t$ given in percent and the ray paths. The earthquakes used for the analysis are classified into six groups according to the regions, A, B, C, D, E and F, as shown on the map on the left. The cross-sections of the structure and the ray paths to Kamikineusu (KMU), Tsukuba (TSK) and Hikami (HM) and the values of $\Delta t/t$ (in percent) are shown on the right. The location of each station is shown on the map on the left. For each station, the values

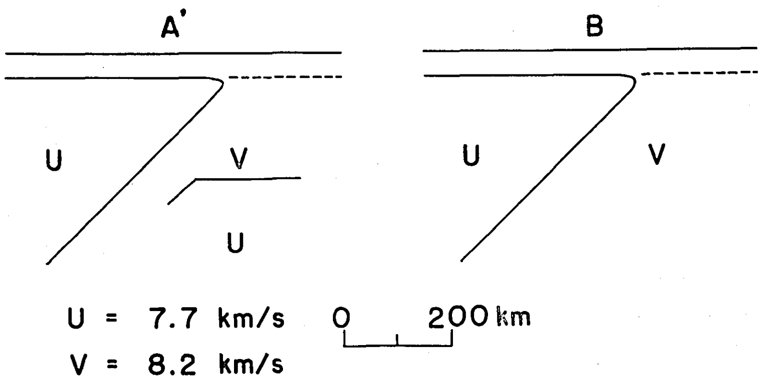

Fig. 35. Upper mantle models used in the calculation of the travel-time residuals [*Utsu*, 1967].

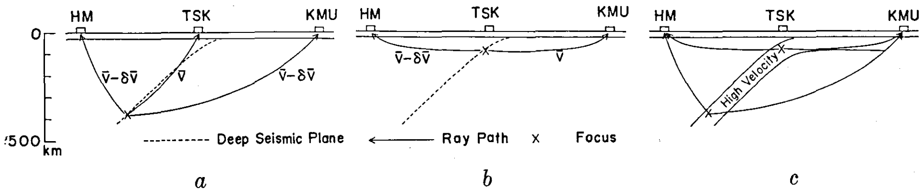

Fig. 36. Cross-sections of the structure along the seismic ray.

of $\Delta t/t$ for all the earthquakes belonging to the same region are averaged. Three earthquakes are averaged for the region *A*, seven for the region *B*, four for the region *C*, nine for the region *D*, ten for the region *E* and ten for the region F.

For the deep earthquakes in the region *F*, the ray path to TSK is on the whole in and near the seismic zone; the value of $\Delta t/t$ is -2.48%. The ray path to HM, is in the mantle above the seismic zone; for this path $\Delta t/t$ is 0.56%. Comparison of the residuals at TSK and HM for other regions clearly suggests that the average *P* wave velocity in the mantle above the deep seismic zone is lower by 2 to 3% than that in the mantle along the deep seismic plane.

This result may be compared with those obtained by *Utsu* [1967] who used the travel-times from teleseismic events, and by *Kanamori* [1968] who used the LONGSHOT data. Both Utsu and Kanamori reached the same conclusion that the average *P* wave velocity in the mantle on the inward (continental) side of the deep seismic zone is about 6% lower than that in the mantle on the outward (oceanic) side. The direction of the incident ray path with respect to the seismic zone is different among the present study and those by Utsu and Kanamori; yet a similar conclusion is obtained. This agreement strongly suggests that the overall velocity contrast between the mantles bounded by the deep seismic zone is real.

The next problem is whether the plate-like structure as proposed by *Oliver and Isacks* [1967] exists beneath Japan or not. *Utsu* [1967] on the basis of the systematic difference between epicenters of earthquakes determined by the JMA net and the USCGS net, preferred a plate-like model as shown by model A' in Fig. 35 rather than model B. It is to be noted, however, that the essential feature of Utsu's model A' is the relatively high velocity over a substantial thickness (probably thicker than 200 km) of the mantle on the outward side of the seismic zone. Thus, although Utsu's model A' represents a plate-like structure, it is not in the context of the ordinary plate model which requires a relatively " thin ", 70 km or so, high-velocity lithosphere beneath the ocean basin. In this connection, we examined the travel-time residuals at KMU. For deep earthquakes in region F, $\Delta t/t$ at KMU is 0.45% and is about the same as that at HM (see Fig. 34). The ray path to KMU is considerably removed from the seismic zone (see Fig. 36-*a*), although it is more or less parallel to the seismic zone. For region E, the ray path to KMU is closer to the seismic zone and $\Delta t/t$ becomes slightly negative. For shallow earthquakes (see Fig. 36-*b*), $\Delta t/t$ shows a very large negative value. These observations therefore suggest that the high velocity region is more or less confined to a narrow zone as shown in Fig. 36-*c*. This model is very close to the ordinary plate model. For regions A and C, the ray path to KMU is almost parallel to the seismic zone, and $\Delta t/t$ at KMU shows a negative value, though not as large as that at TSK. Thus we can conclude, to the first approximation, that a plate-like structure exists beneath the Japanese Islands; it is more remarkable in the southern part than in the northern part.

Two problems, however, must be considered here: (1) the ray path from region F to KUM is not so simple as the cross-section in Fig. 34 implies. It is more or less parallel to the seismic plane and may be affected by the large structural heterogeneity associated with the seismic plane; (2) the simple plate-like model shown in Fig. 36-*c* cannot explain the LONGSHOT data studied by *Kanamori* [1968]. The ray paths from the LONGSHOT to the stations HM and TSK are as shown in Fig. 37. The average P velocity along the ray path to HM was found to be lower by 4 to 6% than that along the ray path to TSK. The simple

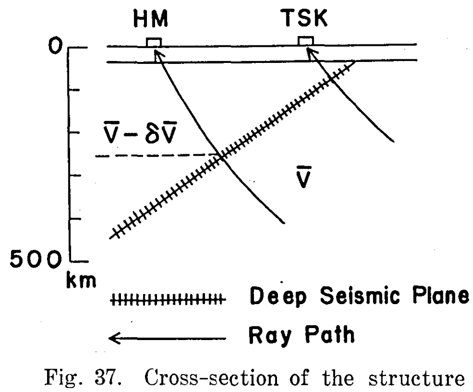

Fig. 37. Cross-section of the structure along the seismic ray [*Kanamori*, 1968].

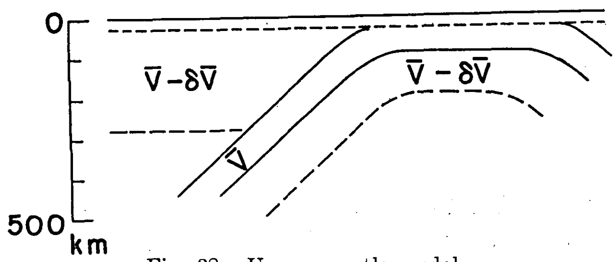

Fig. 38. Upper mantle model.

plate-like model as shown in Fig. 36-*c*, however, cannot explain this observation because the high-velocity plate equally affects the travel-times along both ray paths; the averaged *P* velocity should be the same for these two ray paths. The first problem is rather difficult to solve precisely. In order to resolve this problem, installation of ocean bottom seismographs at appropriate locations is absolutely necessary. It is to be noted, however, that for deep earthquakes in Kuril-Hokkaido region (not shown in Fig. 34) a similar result was obtained for the ray path to TSK. The nature of the ray path is similar to that for the ray path from region *F* to KMU; the ray path is considerably removed from the seismic plane but is almost parallel to it. The travel-time anomaly of 0.12% is obtained for this ray path. The second problem may be solved by delimiting the low-velocity zone below the seismic plane to a region parallel to the seismic plane as shown in Fig. 38. It is evident that this model can explain, at least qualitatively, both the LONGSHOT and the present data. Implicit in this model is the frictional heating of the asthenosphere caused by the sinking lithosphere. However, the above model is not unique and further discussions must await more detailed analysis of travel-times.

7. Conclusion

The absolute accuracy of the epicenter and depth determinations is estimated to be about 20 km. The intermediate and deep earthquakes are confined to a thin zone about 20 km thick in the Kuril-Hokkaido and the Honshu regions and about 50 km in the Izu-Bonin Islands region. In the Izu-Bonin Islands region, the seismic activity is very low at depths from 50 to 250 km. This depth range coincides approximately with the depth of the low-velocity layer in this region. The distribution of intermediate and deep earthquakes is not uniform in space and time. However, the seismicity over one year period resembles closely that over a much longer period. A plate-like structure exists beneath the Japanese

Islands; it is more remarkable in the southern part than in the northern part. The low-velocity zone below the seismic plane seems to be delimited to a region parallel to the seismic plane.

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Hypocenter data of 410 earthquakes relocated in the present study are available on request in the form of IBM Cards.

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55. 日本付近の地震活動と走時異常

地震研究所 石田瑞穂

日本列島付近の地震活動とそれに付随する問題を調べる為，北緯 25° から 50°，東経 128° から 150° の範囲に 1967 年 1 年間に起きた約 410 の地震について震源を再決定した．世界中の観測所のデータとともに，近距離にある高精度の観測所のデータをも用いた為，この範囲での震源決定の絶対精度は，非常に良くなり，約 20 km と推測できる．その結果，中深発地震面の厚さは，Kuril-Hokkaido, Honshu では約 20 km, Izu-Bonin Islands では約 50 km 以内と考えられる．Izu-Bonin Islands では，50 km から 250 km の深さで地震活動は非常に減少する．この深さは，この地域での low-velocity layer の深さと一致する，中深発地震の震源分布は，空間的時間的に一様でなく，震源面のある部分に集中する傾向があり，地震の起こり方も月別に非常に変化する．しかし，1 年間の *P* 地震活動と，長期間にわたる地震活動とは，全体として見れば，かなり良く似ている．観測された波の走時異常から，日本の下のマントルの構造異常と地震活動の関係を調べた．深発地震面より上 (陸) 側のマントルと，深発地震面より下 (海) 側のマントルでは，深発地震面を含む深発地震面に平行なマントル内でより，P 波の平均的な速さが 2〜3% 遅いと考えられる．