Investigation of Microearthquakes — On the Nature of the Crust —

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Abstract

1. The ambiguities in the interpretation of explosion seismic data can be compensated by studying the seismic data of microearthquakes. It was found that the upper crust and the lower crust were defined by a fairly sharp contrast in their compressional waves.

2. The Poisson's ratio of the upper crust is about 0.24, which is a rather lower value than that to be expected from rock experiments. There is also a sharp contrast in the Poisson's ratio between the upper and lower crust. This suggests that there must exist a change of chemical and/or mineral composition between these two layers.

3. By comparing the phase velocities of a Rayleigh wave in the central U.S. and one in the western part of Japan, areas which are infered to have nearly the same crustal structures according to the results of explosion seismology, it was found that the Poisson's ratio of the upper mantle and/or lower crust in the latter district was larger than that of the former district by 0.03-0.05 in average. The lower crust and the upper mantle of the island arc may be characterized as having a high Poisson's ratio. It is interesting relating to the formation of the crust and its development.

1. On the Crustal Structure

The author has already discussed in another paper the possibility of research into detailed crustal structure by using microearthquakes (Hashizume, 1970b). It has been believed, hitherto, that the most effective and accurate method for investigating crustal structure is by explosion seismology. With this method, nevertheless, there still remain some ambiguities in the interpretation. It has been pointed out that given the same travel times, two fairly different crustal models can be calculated with few errors even when the observation stations were densely enough spaced (Steinhart *et al.*, 1962). This is mainly due to the fact that the explosions were always conducted near the surface. For the study of detailed crustal structure, we now use microearthquakes whose hypocentral domains are as small as those of explosions, but which have much more variety in their focal depths than do artificial detonations.

Although in studying real crustal structure regional variations or inclinations of boundaries must be taken into consideration, horizontally homogeneous and only vertically variable structure is assumed, as our region under study is not so broad. The data used here are from earthquakes whose hypocenters were determined by

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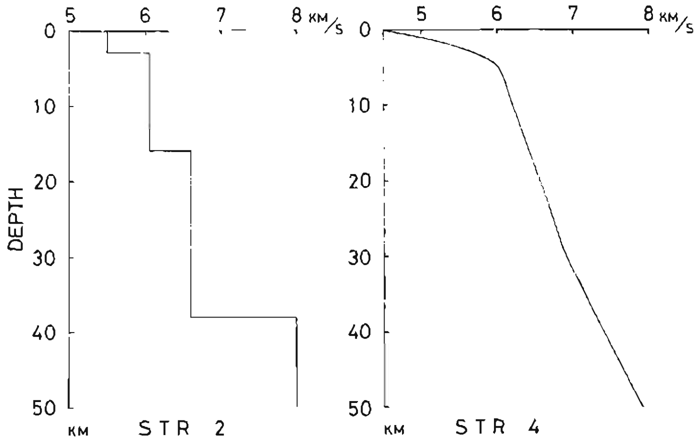

Fig. 21. Crustal structures used in this paper. Both of these two structures can be interpreted, within few errors, from the same seismic wave data of explosions.

more than four stations shown in Table 1*) during the period from June 1965 to June 1968 and whose initial motions of the *P* waves were clearly read by at least one station in Table 3. The total number of data is 313. The method of hypocenter determination has been described already (Hashizume, 1969). The time accuracies of the data from the stations in Table 3 were not generally as good as those in Table 1. It is supposed that the accuracy is sometimes much worse than the accuracy of origin time determination of ± 0.2 sec (Hashizume, 1970b). Two crustal structures are assumed as STR 2 and STR 4 in Fig. 21. The structure STR 4 satisfies the traveltimes of Kurayosi explosions on our district within few errors as well as STR 2 used for hypocenter determination (R.G.E.S., 1966). That is, using only the traveltime table of the initial *P* motions, these two structures cannot be distinguished.

In the paper cited above, two methods are presented for correcting the calculation of the crustal structure, compensating the data by explosion seismology (Hashizume, 1970b). One is the method to search optimum structure to put minimum the standard deviation in hypocenter determination. For this method many accurate data are required. As the hypocenters should not be determined from inaccurate data, they are determined only by the stations in Table 1. The results are shown in Table 12, with the number of earthquakes used in the calculation. From this table only, because of insufficient data, it cannot be said which structure model represents well the real crustal structure. However, the structure STR 2 seems more probable than the structure STR 4 for very shallow parts of the crust.

* The numbers of the Tables and Figures are in series in this series of articles with the same title (Hashizume, 1969, 1970a, b).

Table 12. Mean standard deviations in determining hypocenters by assuming the two crustal models STR 2 and STR 4 in Fig. 21. This is classified into three parts according to focal depth.

Focal depth (km)	Number of data	STR 2	STR 4
$Z \leqq 5$	83	0.068	0.078
$5 < Z \leqq 10$	106	0.060	0.064
$10 < Z \leqq 15$	124	0.057	0.059

The second point is as follows. Setting,

$$T_j = P_j - P^\circ \tag{1}$$

$$(O-C)_j{}^2 = T_j - T_j{}^2 \tag{2}$$

$$(O-C)_j{}^4 = T_j - T_j{}^4 \tag{3}$$

$$\Delta T_j{}^{2\cdot 4} = T_j{}^2 - T_j{}^4 \tag{4}$$

j : observation station index
P°: origin time
$T_j{}^n$: calculated traveltime from hypocenter determined assuming STR n, to station j passing through this structure.

If there are any differences between the real and the assumed crustal structures, $(O\text{-}C)_j$, that is, the differences between the calculated traveltimes $T_j{}^n$ using the assumed crustal structure and the observed traveltime T_j, should be as shown in Figs. 20 a and c. The $(O\text{-}C)_j$ calculated from the stations which are used for the hypocenter determination is not available because there is no definite tendency to indicate the difference of the two structures (Hashizume, 1970b). The hypocenters were determined by the stations in Table 1, then $(O\text{-}C)_j{}^n$ and $\Delta T_j{}^{2\cdot 4}$ were calculated by the data observed at the stations in Table 3 which were not used for hypocenter determination.

If there are no errors by factors such as observation, hypocenter determination, local traveltime anomaly and so on, $(O\text{-}C)_j$ should vary complicatedly according to the location of the epicenter, the observation station or the focal depth; and theoretically it is possible to know a minute structure. Now, the data were arranged and summed up roughly in the following intervals referring to the characteristics from Fig. 20, although this may not be the best.

$Z \leqq 5$, $5 < Z \leqq 10$, $10 < Z \leqq 15$,
where Z is the focal depth in kilometers and

$D \leqq 50$, $50 < D \leqq 100$, $100 < D$,
where D is the epicentral distance in kilometers.

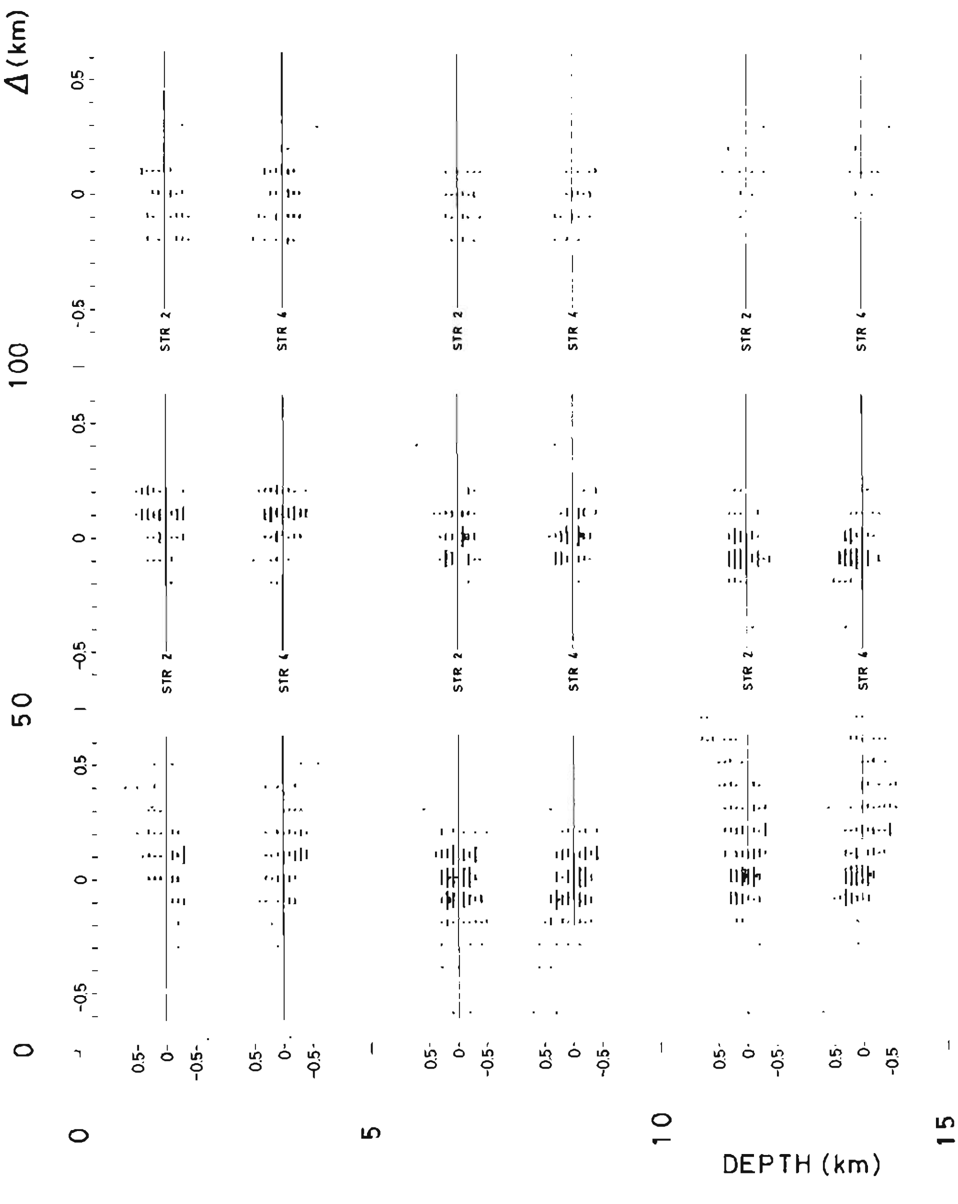

Fig. 22. $(O\text{-}C)^{n}{}_{j}$ is plotted against $\Delta T^{2\text{-}4}{}_{j}$ as abscissa. The figure is divided into nine blocks according to epicentral distance and focal depth at the interval as shown. Each figure shows the summed up results in the blocks for STR 2 and STR 4. Scales are in seconds for $(O\text{-}C)^{n}{}_{j}$ and $\Delta T_{j}{}^{2\text{-}4}$.

Results are shown in Fig. 22. Scales are in seconds for $(O\text{-}C)_j^n$ and $\Delta T_j^{2\cdot4}$. If the unknown real structure coincides with STR 2, it becomes $(O\text{-}C)_j^2=0$ and $(O\text{-}C)_j^4=-\Delta T_j^{2\cdot4}$, while if the real structure coincides with STR 4, it becomes $(O\text{-}C)_j^2=0$ and $(O\text{-}C)_j^4=\Delta T_j^{2\cdot4}$. Further, if the traveltimes expected from the real structure are systematically larger or smaller than those from the assumed structure, $(O\text{-}C)_j^n$ should be systematically larger or smaller than zero near $\Delta T_j^{2\cdot4}=0$ as evident from Fig. 20. That value of $(O\text{-}C)_j^n$ near $\Delta T_j^{2\cdot4}=0$ indicates mainly the difference of accumulated traveltime along the path between the real and the assumed structures. On the other hand, the distribution pattern of $(O\text{-}C)_j^n$ indicates mainly the characteristics of the structure at the depth concerned.

Thus the conclusions that can be drawn from Fig. 22 are as follows:

1. We can say nothing from the data near $\Delta T_j^{2\cdot4}=0$, because the time accuracy of the available data was not good. However, the assumed structure itself is adjusted at the surface, to the data from explosion seismology, and the accumulated traveltimes may not deviate so much.

2. Each figure shows generally a plus gradient for STR 2 and a minus gradient for STR 4. This tendency is especially remarkable for the figures whose epicentral distances are less than 50 km. This suggests that the real crustal structure is intermediate of the two assumed structures for the crust shallower than 15 km. It is notable that at a depth of 10–15 km this tendency is particularly clearly shown. The real structure near the depth of 15km is estimated to have a fairly steep velocity gradient, although it is not sharply defined.

From our data no definite conclusions can be derived for the feature of deeper structure. More accurate and numerous observation data are needed for the investigation of more detailed crustal structures.

At this time we can conclude from the two methods described above that the crust is divided into two parts: upper and middle or lower crust, that is, the so-called granitic layer and the gabbroic one; and although these two layers might not be sharply defined, their boundary could not be as indistinct as in STR 4. In explosion seismograms, the boundary can scarcely be recognized from the reflected waves. This may be partly due to its indistinctness and partly due to technical difficulties.

2. On Poisson's Ratio Variation in the Crust

Hitherto, in determining hypocenters, a uniform Poisson's ratio structure has been assumed. Of course this is the first approximation, and naturally it is possible that there exist regional or depth-dependent variations. Here, it is assumed that there exist no regional variations and the Poisson's ratio is a function only of depth. Set the equations as,

$$P_j^\circ=P_j-A\times(S-P)_j \tag{5}$$

$$A=\sqrt{1-2\sigma}/(\sqrt{2-2\sigma}-\sqrt{1-2\sigma}) \tag{6}$$

σ : Poisson's ratio

j : station index

The best fitted coefficient A is a function of the weighted mean of the Poisson's ratio along the ray path. Principally, by determination of the coefficient A's for earthquakes of various epicentral distances or S-P times and focal depths, the Poisson's ratio can be calculated as a function of depth, if the ray path is known. However, S-P times contain considerable reading errors, so they need to be rounded by a fairly great number of data. And this is the same for focal depth (Hashizume, 1970b). To make the matter more complicated, the ray path varies according to the assumed crustal structure. Therefore we must abandon detailed discussions.

The process in analysis is as follows. At first, set the equations as,

$$(S-P)^c=\sum(S-P)_j/n \tag{7}$$

$$P^\circ=\sum P_j^\circ/n \tag{8}$$

n : the number of observation data $n \geqq 2$

j : station index

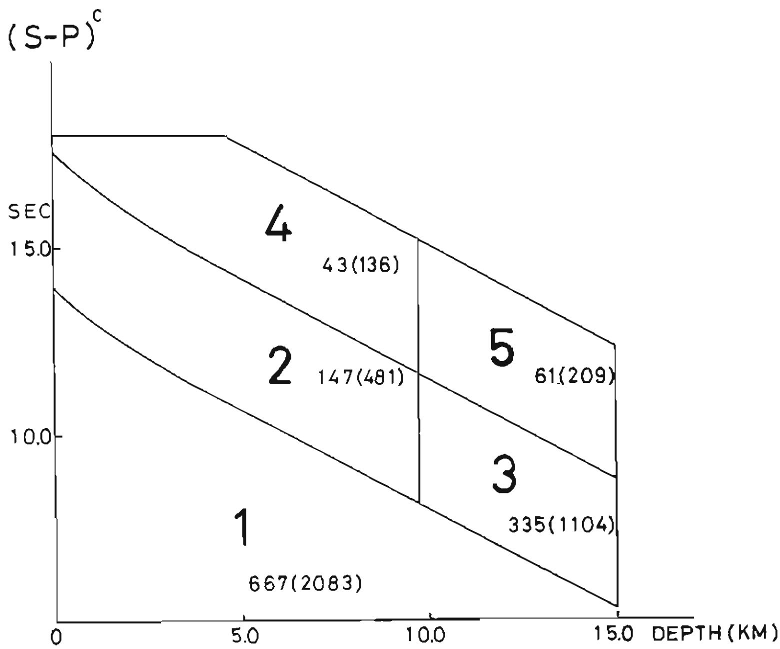

Fig. 23. Classification for studying the Poisson's ratios. The large numerals in each block are the classification numbers. The small numerals in and out of parentheses are the numbers of the S-P data and of the earthquakes respectively used in the relevant blocks.

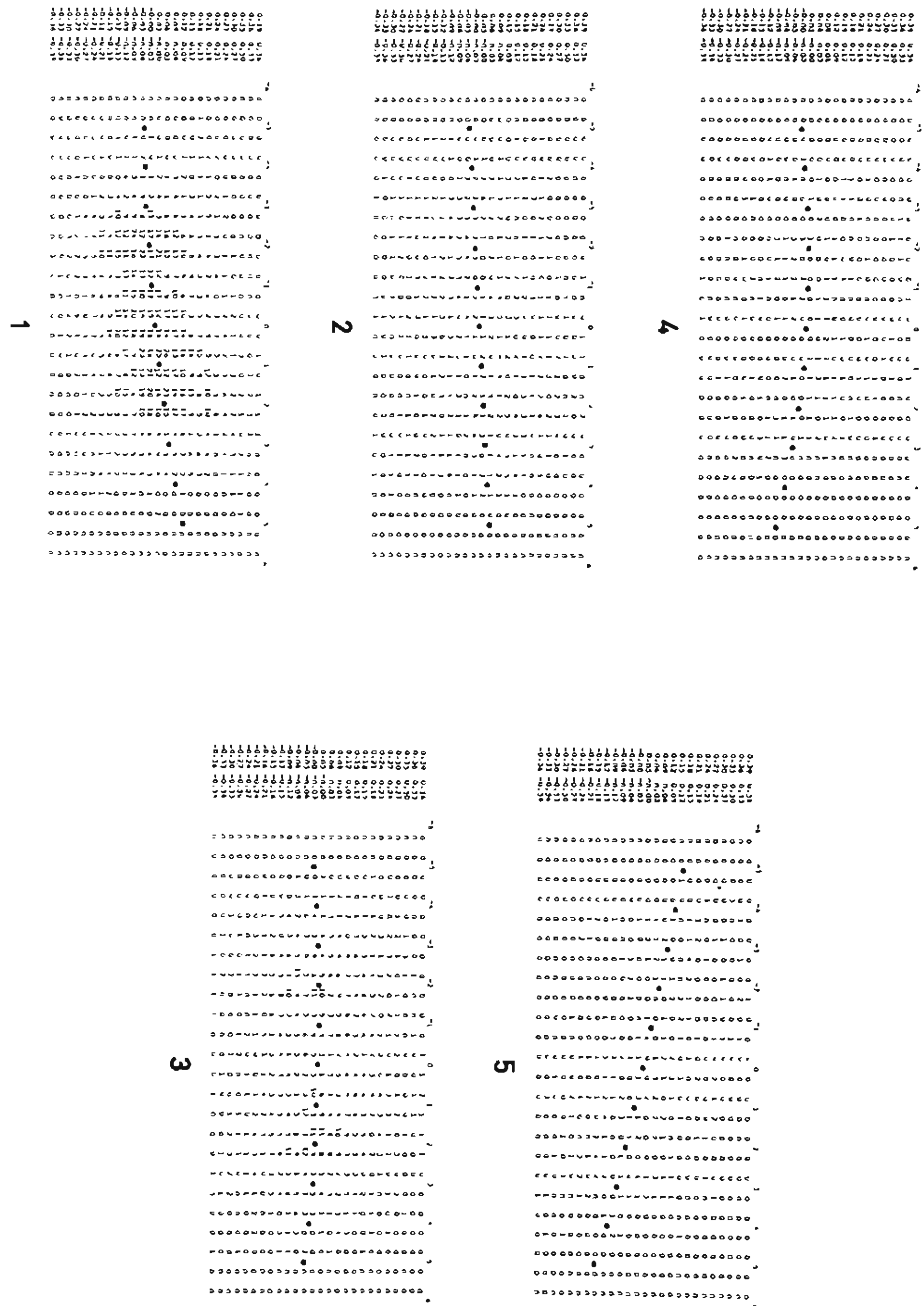

Fig. 24. Setting $A=1.4083$, the summed up numbers of ΔP°_{j} for the respective intervals are shown against $(S\text{-}P)^{r}$ as abscissa. Solid points indicate the discrete values of the approximated quadratic functions.

$$(S-P)^r=(S-P)_j-(S-P)^c \tag{9}$$

$$\Delta P_j^\circ=P_j^\circ-P^\circ \tag{10}$$

The data are classified into five blocks according to focal depth and $(S\text{-}P)^c$, as shown in Fig. 23. At the boundary between the blocks 2, 3 and 4, 5 the seismic waves are considered to pass through about half a way in the so-called gabbroic layer. The ray paths of the earthquakes in block 1 do not enter into the gabbroic layer at all. The numbers of S-P data and earthquakes used in the calculation in relevant blocks are shown in the figures in and out of parentheses respectively. For the first approximation A is set as 1.4083 (Hashizume, 1970b). The data are summed up, taking $(S\text{-}P)^c$ as origin, $(S\text{-}P)_j^r$ as abscissa and ΔP_j° as ordinate for each block in Fig. 23, as shown in Fig. 24. The earthquakes used in this analysis are shown in Table 7; they were observed in the period from August 1964 to June 1968 at five stations shown in Table 1. ΔP_j° greater than 0.4 sec is omitted from the data because it is considered to come from reading error of S-P time. The earthquakes with focal depths greater than 16 km are very small in number and are omitted from this analysis.

Now, the distribution of ΔP_j° is approximated as a quadratic equation,

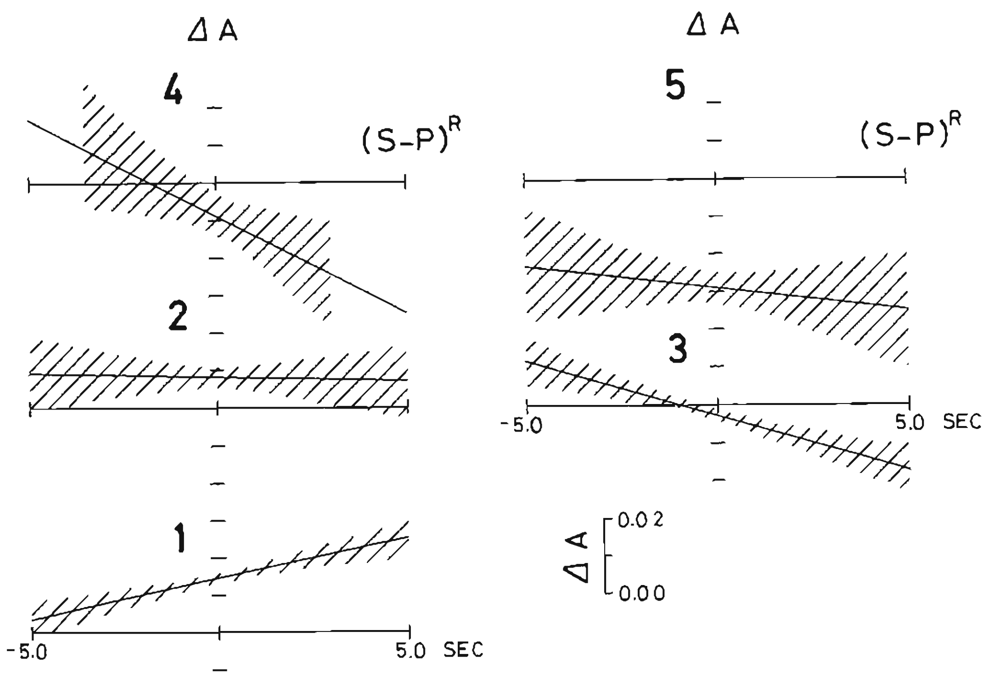

Fig. 25. ΔA is plotted against $(S\text{-}P)^r$ as abscissa. Ruled areas indicate the error bands.

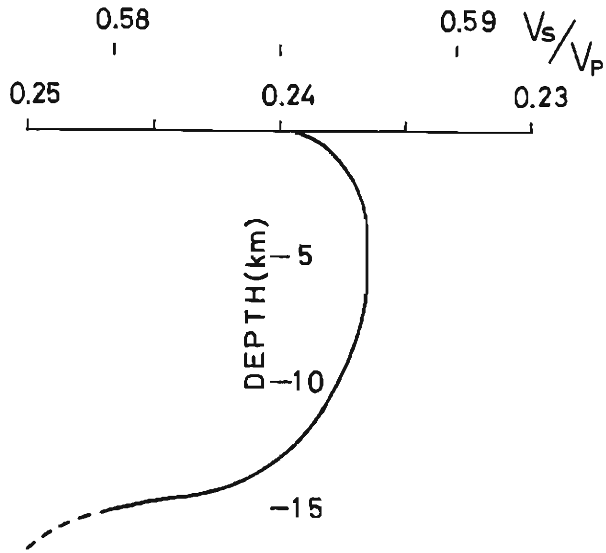

Fig. 26. Poisson's ratio variation in the crust. The abscissa is Poisson's ratio and V_s/V_p

$$Y = aX + bX^2 \tag{11}$$

This curve is shown in Fig. 24 as discrete points. From the first derivative of this function, the correction for the first approximated A_0 is calculated as follows,

$$Y' = a + 2bX$$
$$A = Y' \tag{12}$$
$$A_0 = 1.4083$$
$$\Delta A = A - A_0 = a + 2bX - 1.4083 \tag{13}$$

This correction ΔA is shown in Fig. 25 with the error band shown as a ruled area. The number of data is not yet sufficient but so far as these data are concerned, the Poisson's ratio variation is estimated consistently by these correction values of ΔA, and this is shown in Fig. 26.

The Poisson's ratio in the upper crust shallower than 15 km is 0.235–0.24, while it increases abruptly to 0.25 at the depth nears 15 km. V_s/V_p is shown together in the figure.

3. Discussions

Now some subjects must be considered from the above results. First, hypocenters have been determined assuming $A = 1.4083$ uniformly for all the earthquakes. As described in the preceding paper, the error in origin time directly affects the calculation of depths for earthquakes inside the seismic networks, while it does not affect the calculation of focal depths but only of epicenters for earthquakes outside the networks (Hashizume, 1970b). In this case, origin time calculations may show a delay of at most 0.2 sec for earthquakes having *S-P* time of about 10 sec. The effect of this error on a hypocenter is trifling.

Secondly, Kaminuma (1966) estimated the crustal structure in Japan by using surface waves with periods from 20 to 40 sec. According to his discussions, the phase velocity of Rayleigh waves in the central part of Japan is less than that of Canadian Shield with standard crustal structure by 6%; and he concluded that the Poisson's ratios of the upper crust, lower crust and upper mantle are 0.25, 0.26 and 0.27 respectively. Kanamori (1963) also estimated the Poisson's ratio of the upper crust to be 0.27 and that of lower crust 0.3–0.35 from both the phase velocity of Rayleigh waves and gravity data. Both of these discussions were based on the assumptions that the *P* wave velocity of the upper mantle or the intermediate layer was much less than 8.0 km/sec, and the crust in the Japanese islands was thin compared with the continental one. These assumptions come mainly from the poor data of explosion seismology at that time. Since then, fairly detailed investigations of crustal structure by explosion seismology have been done on the western part of Japan, and one of the models presented is our model of STR 2 with slight modification (Hashizume *et al.*, 1966). This crustal structure is very similar to that in the central U.S. except for a little lower *P* wave velocity in the upper mantle.

Nowadays, the *P* wave velocity of the upper mantle is considered to be nearly 8.0 km/sec and the crustal thickness is as thick as the continental one, at least for the western part of Japan. This velocity of 8.0 km/sec for the upper mantle is confirmed also by the apparent velocity method with high accuracy observations of natural earthquakes by our microearthquake networks (Hashizume *et al.*, 1970). Nevertheless, the phase velocity of Rayleigh waves observed in the central U.S. is a little higher than that for the western part of Japan (McEvilly, 1964). This discrepancy is conspicuous for periods larger than 30 sec and it amounts to 0.1–0.2 km/sec. Of course this is out of the error band. From the partial derivatives calculated by McEvilly, it is impossible to explain the data of Rayleigh waves by the slight difference of *P* wave velocity of the upper mantle. So far as setting the Poisson's ratio to be the same for both regions, by changing the interpretation of the explosion seismic data of one of the two regions concerned, as in model II presented in the above cited paper, or in STR 4 in this paper, the matter is the same as above. A slight change of the assumption on density is also useless. McEvilly, in his study, set the Poisson's ratio as 0.27 for the upper crust and as 0.25 for the lower crust and upper mantle. Although it is fruitless to discuss this problem concerning only one of the multivariate parameters, it can be safely said that the Poisson's ratio of the upper mantle and/or lower crust in the western part of Japan is larger than that of Central U.S. by probably 0.03-0.05 in average by the partial derivative estimation, fixing the Poisson's ratio of the upper crust as in Fig. 26. This difference can, of course, be attributed to the difference of the material between the two regions. But it may not be indifferent to the geotherm difference between the continent and the island arc. It is reasonable that the ratio of V_s/V_p decreases with increasing temperature on the same material (Soga *et al*, 1966). And this must correspond to the low Q zone beneath the island arc (Ttsu, 1968).

Kanamori *et al.* (1965) determined the Poisson's ratios of various rocks by laboratory experiments. They investigated the effect of confining pressure, and showed that the Poisson's ratio increases with the increase of confining pressure up to 2 kb and then becomes constant; and generally the Poisson's ratio is about 0.26 in granitic rocks and about 0.3 in gabbroic rocks at 10 kb. Comparing these results with ours, it is noted that the Poisson's ratio decreases in the very shallow part of the crust, in contravention of the results from rock experiments. But this can be attributed to errors of seismic data. Poisson's ratio of less than 0.24 for granitic rocks is considerably less than that expected from rock experiments. It is supposed there exists, in the real crust, different circumstances from those of the experiments. At a depth of about 15 km, there exists a Poisson's ratio gap. Of course this gap cannot be caused by temperature or pressure effects.

It has also been confirmed that there exists a steep compressional velocity gradient at this depth. Therefore it is inferred that there is a fairly clear change of chemical and/or mineral composition between the upper crust and the lower crust. It is very interesting that this boundary also corresponds to a seismicity gap, as shown in Fig. 11. This problem is to be studied in another paper (Hashizume, 1970c). From explosion seismology on land, it is confirmed with very few exceptions that the apparent P wave velocity of near 6.0 km/sec is observed up to an epicentral distance of 100—150 km everywhere regardless of the geological structure. The configuration of our crustal model that the crust is composed of two parts, upper part that consists of acidic rocks and middle and lower parts that consist probably of basic or intermediate rocks, and that their volume ratio is about 1 : 2 and their boundary defined fairly clearly, may be adopted as a universal crustal model. Then how was this crustal structure formed?

The difference between the continental crust and of the island arc may be characterised by the difference of the Poisson's ratios of the lower crust and upper mantle. If this is due to temperature effects, then the nature of the crust of the island arc can shift to the continental one when the heat source is removed.

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