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Investigation of Microearthquakes
— On Earthquake Occurrence in the Crust —

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Abstract

1. The author has tried to consider that seismicity is expressed only by earthquakes occurring independently each other. This is accomplished fairly well by eliminating earthquakes that occur within the same hypocentral domains in a certain period. A seismicity map made in this way resembles more to that of large earthquakes than the original one.

2. The coefficient m of Ishimoto-Iida's relation varies minutely by space or by time.

3. The focal mechanisms of the earthquakes that occur in the northern part of Kinki District and the eastern part of Chugoku District are nearly the same, regardless of magnitude, focal depth and pattern of occurrence. The direction of the principal compressional axis is latitudinal and that of the intermediate principal axis is vertical, with slight variations. Then the majority of these earthquakes are considered to be the strike slip type, but a few earthquakes are the dip slip type. This is consistent with the neotectonic evidences.

4. In explaining the flactuation of focal mechanisms by conjugate sets of fractures, even when the ratio of the conjugate fractures is assumed as small as 1/4, the fracture angle exceeds 40°.

5. To explain the low stress drop actually observed in natural earthquakes and the uniformness of their focal mechanisms at the same time, these earthquakes must be considered to occur where the stress is not large but the strength is weak.

6. If this weak strength is attributable to interstitial fluid material, various geophysical or geological phenomena can be explained.

1. Introduction

Microearthquakes have two important properties. One is their abundance in number, and the other is hypocentral domains so narrow as to be considered as point sources. This latter property cannot be found in large earthquakes. Therefore we can obtain detailed informations on the crust without disturbances of the surface or upper mantle. This property of microearthquakes is very important for studying the nature of the crust.

The author has already published several papers under the same title, with the subtitles: —On Seismicity—, —On the Accuracy of Hypocenter Determination—, and —On the Nature of the Crust— (Hashizume, 1969, 1970b,c). In them he has discussed seismicity and the nature of the crust using these properties of microearthquakes.

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The seismicity map is presented in Fig. 7*); and most of the earthquakes are confined to the upper crust, the so-called granitic layer, as is shown in Fig. 11. The upper crust is found to be defined fairly sharply from the lower crust. He will continue this investigation of microearthquakes referring further the above results and properties.

2. Seismicity

From the seismicity map in Fig. 7, we can read that some regions are seismically active and other regions are not. But it is not certain whether the number of earthquakes in a region is directly connected to this tectonic activity or not. Various expressions for seismicity may be presented for various purposes. Large earthquakes are usually accompanied with many aftershocks, and sometimes foreshocks. In 1965, the Matsushiro earthquake swarm occurred and if we pay attention only to the

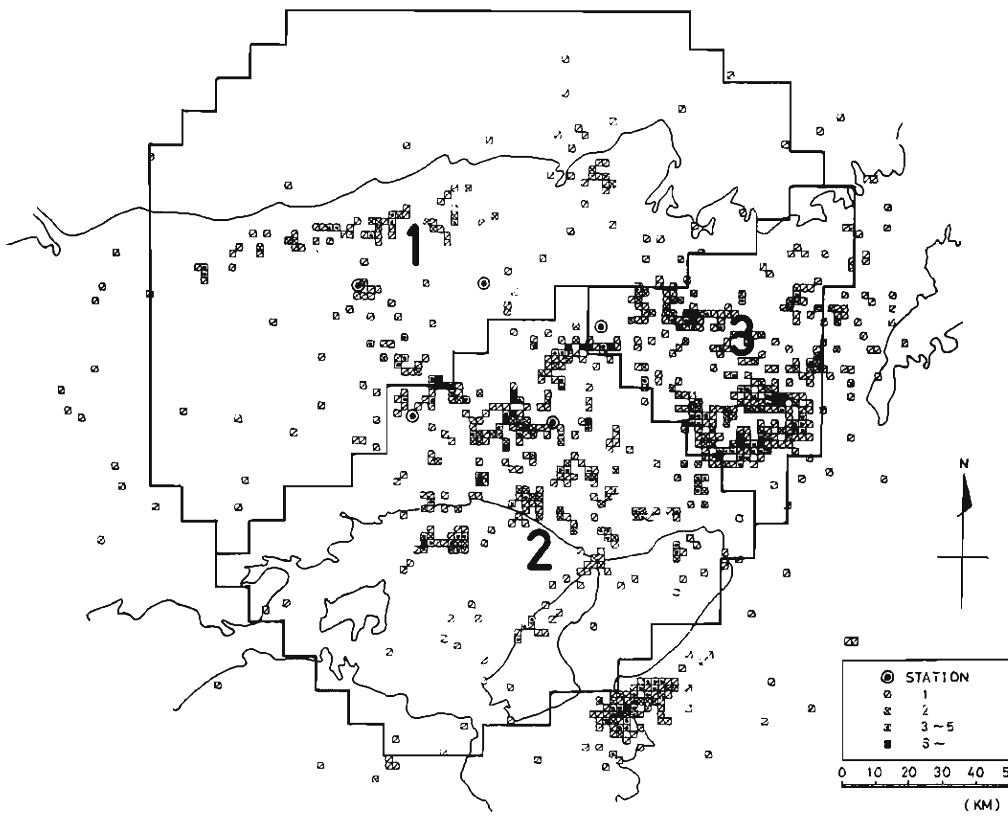


Fig. 27. Regional classification in studying seismic statistics.

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

number of earthquakes, the Matsushiro area becomes the most tectonically active area in Japan.

If we can pick up only those earthquakes that occur wholly uninfluenced by other earthquakes, that is, "mutually independent" earthquakes, these might be used as one of the indicators of the field of earthquakes. Much researches have been done on this problem. For example, Knopoff (1964) has said that if earthquakes that occurred more than a certain number in a certain time interval are excluded, the frequency distribution of the number of earthquakes in that time interval fits Poisson's distribution. The term "independent" is used as meaning to be statistically uniform and occurring at random; and this method is improved in the following way.

The data used are confined to the period April 1965 to June 1968, to magnitudes greater than 1.2, and their epicenters within regions 1, 2 and 3 in Fig. 27, for observational reasons. (The author has presented a question about the definition of magnitude. The magnitude scale can be less by about 1.0. However, here this scale is used only tentatively (Hashizume, 1969)). The number of earthquakes thus adopted was 403. Among these earthquakes, those which occurred in the same hypocentral domains in a given time interval t were subtracted. For example, taking $t=31.6$ days, if several earthquakes occur within the time interval of 30 days in the same hypocentral domain, the total number of these earthquakes is counted as one, and the onset time of the first earthquake of this series is defined as the beginning time of this earthquake series. Fortunately the hypocentral domain of a microearthquake is considered to be within the order of 100 meters, using Utsu-Seki's formula extrapolated for microearthquakes (Utsu *et al.*, 1955) ; thus each hypocentral domain can be confined to this dimension and

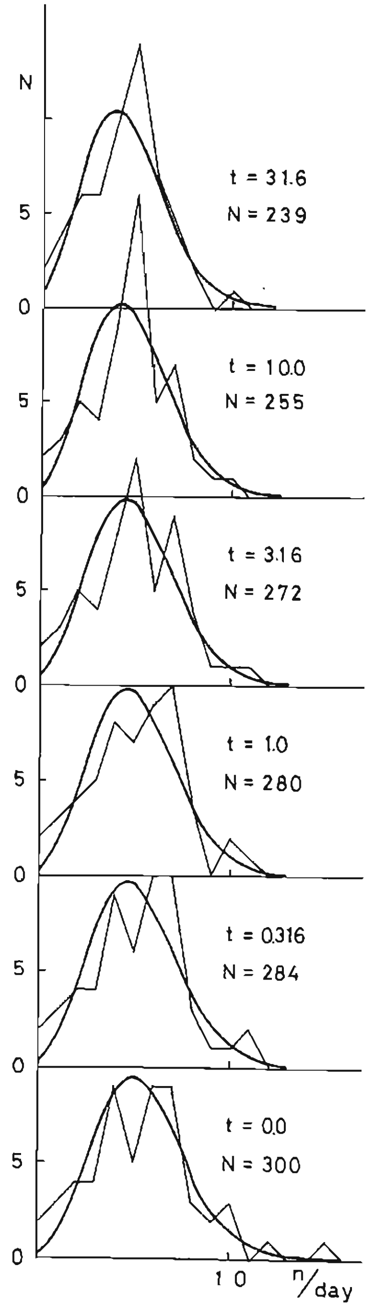


Fig. 28. Frequency histograms of the number of earthquakes per 20 days varying the time interval t. The sampling number is 55 and N is the number of earthquakes. Heavy lines show the expected Poisson's distribution.

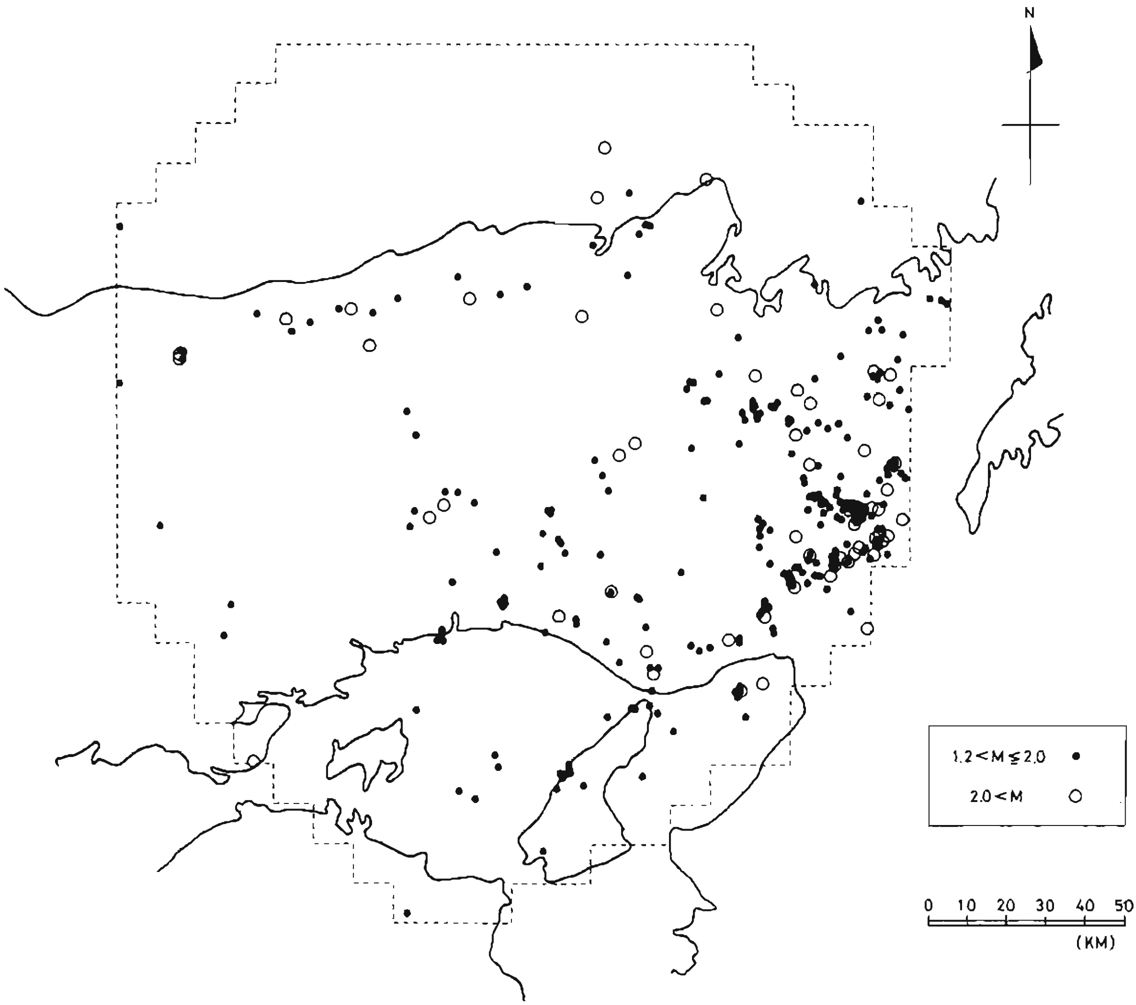


Fig. 29 a. Observationally uniform seismicity map of $M\geqq 1.2$. Dotted line is the limited boundary of uniform observation.

does not overlap with the hypocentral domains of other independent earthquakes so far as the seismicity level is not very high. This is the great advantage of microearthquakes compared to large earthquakes for simplifying the study. Here, the hypocentral domain is defined, regardless of the magnitude and focal depth, in 3×3 km square, taking into account any errors of hypocenter determination (Hashizume, 1970b).

The frequency histograms of the number of earthquakes thus defined per 20 days are shown in Fig. 28, varying the time interval t. The total number sampled was 55, and the total number of earthquakes is shown as N in the figures. If these earthquakes occur perfectly uniformly and at random, the frequency distribution

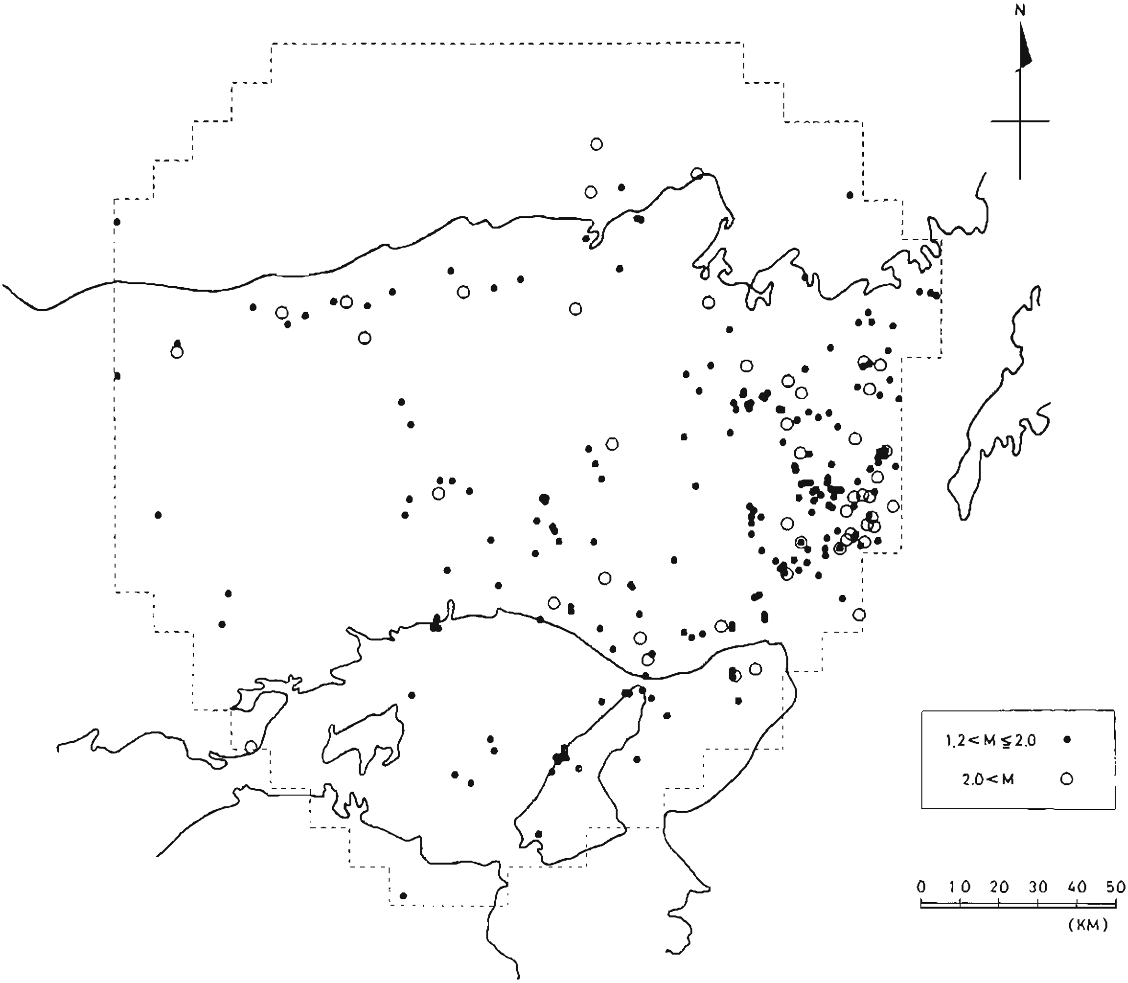


Fig. 29 b. Seismicity map of independently occurring earthquakes, in which earthquakes that occur in the same hypocentral domains within one month are omitted from those in Fig. 29 a.

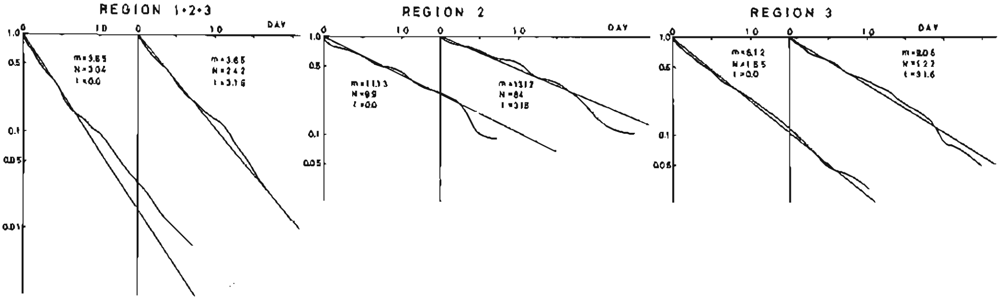


Fig. 30. Cumulative distribution of time intervals between adjacent earthquakes. N, m and l are the number of earthquakes, the number of earthquakes per unit time and the filtered time interval in the text respectively.

Table 13. χ^2 test for Poisson's distribution for the earthquake sequence treated in Fig. 27.

Hypothesis	$t=0.0$	$t=0.316$	$t=1.0$	$t=3.16$	$t=10.0$	Days $t=31.6$
Fitting to Poisson's distribution	× 0.0001	× 0.025	× 0.07	× 0.15	× 0.2	○ 0.25

For example ○ 0.25 : Hypothesis cannot be rejected in the significance level of 0.25.
× 0.25 : Hypothesis can be rejected in the significance level of 0.25.

must fit to the Poisson's distribution, although the opposite is not true. The expected Poisson's distribution is also shown with thick lines for reference. The Pearsen's X^2 test was applied to examine compatibility with Poisson's distribution and the results are shown in Table 13. For the time interval $t=0$, that is, the original earthquake sequence, we can see that earthquakes did not occur at random. For the time interval $t=31.6$ days most of the earthquakes in this sequence can be said to have the necessary conditions for independently occurring earthquakes, but for any time interval less than one month the sequence cannot be said to consist of independently occurring earthquakes. The resultant seismicity maps are shown in Figs. 29a and b. Fig. 29a is the original seismicity map only in uniform observation, that is, $t=0$. In Fig. 29b filteration of t as equal to 31.6 days is operated. The dotted lines in these figures are the limited boundary of the uniform observation. And it can be said that the earthquakes concerned here influence on the other earthquakes within a time interval of one month. As the observation period is limited it is difficult to estimate what happens in a time interval much longer than one month; however, in Fig. 29b many earthquakes are still plotted in the same hypocentral domain and it may be said that these earthquakes occur partly or completely independently after elapsing a period longer than one month, although this might not be said in general.

Another method to test uniformness and randomness is to examine the time interval between adjacent earthquakes. Setting m as the probability of occurrence of earthquakes in a unit time, then the probability function $f(\tau)$ of the occurrence of earthquakes at a certain time interval τ is,

$$f(\tau)=me^{-m\tau} \tag{1}$$

and the cumulative probability function $F(\tau)$ of a interval longer time than τ is,

$$F(\tau)=e^{-m\tau} \tag{2}$$

The cumulative distributions of time intervals are shown in Fig. 30 for region 2, region 3 and the whole regions 1, 2, 3 in Fig. 27. Region 1 was omitted because of the small number of earthquakes. Two filters of the time intervals $t=0$ and $t=31.6$, mentioned above, were applied. N and m are the number of earthquakes and the mean value of the time intervals respectively. Expected probability in eq.(2) is shown for reference. Pearsen's X^2 test was also applied and the results

Table 14. χ^2 test for eq (2) for the earthquake sequence treated in Fig. 30.

Hypothesis	Region	$t=0.0$	$t=31.6$ Days
Fitting to exponential distribution	1+2+3	$\times$ 0.0001	0.5
	3	$\times$ 0.0005	$\times$ 0.25

For example $\times$ 0.25 : Hypothesis can be rejected in the significance level of 0.25.

are shown in Table 14. In the case of $t=0$, the earthquakes did not occur uniformly and at random in the time domain; and in the case of $t=31.6$ also the earthquake sequence cannot be considered to have occurred perfectly uniform and at random. Utsu (1969) interprets that this phenomenon is caused by a few comparatively large earthquakes accompanied by aftershocks. Hamada (1968) said that the Matsushiro earthquake swarm occurred intermittently, although his treatment was different from the one in this study. Some reports show that, omitting the aftershocks, earthquake sequences fit eq. (1) or (2) (Gaiskii, 1961). In our case, the definition of hypocentral domain is clear and the filter to take out those aftershocks that occur within about one month is applied; nevertheless the earthquake sequence still cannot be considered uniform and at random. One reason may be that the filter was not complete, but even when the time interval t of the filter is lengthened this tendency seems to be unchanged. The second reason may be the problem of the data themselves. An alternative possibility is that a given earthquake brings about other earthquakes beyond its own hypocentral domain. It is reported that the Matsushiro earthquake swarm expanded its seismic active area from near Mt. Minakami after a lapse of time (J.M.A., 1968).

We must also examine the matter in spatial terms, in order to say about the uniformness and randomness of the occurrence of earthquakes. In Fig. 29b, howhowever, the spatial distribution of earthquakes is not uniform. Furthermore, if this earthquake activity were regulated by the fluctuations of the stress field that overwhelm this region, as will be mentioned later, some relations of occurrences of earthquakes must exist among the three regions shown in Fig. 27. The number of earthquakes per 100 days, with the above-mentioned filter of $t=31.6$ days, is shown in Table 15. Testing the hypothesis of mutual independence of the number of earthquakes of the three regions, the χ^2 test was also applied to the crosscorrelation. This hypothesis of mutual independence could not be rejected in the significance level of 0.2. Although we cannot say decisively, if enough spaces are taken, the influence of a given seismic activity upon other regions is considered to be negligible.

Thus we can, to a certain extent, pick up in the time field those earthquakes that occurred mutually independently, by omitting earthquakes that occurred at the same hypocentral domain within a certain time interval. On the other hand some earthquake occurrences may be affected by other earthquakes

Table 15. The number of earthquakes per 100 days filtered by $t=31.6$ days in the text. Hypothesis of mutual inde pendence could not be rejected in the significance level of 0.2.

Days \ Region	1	2	3	Total
0	2	7	11	20
100	5	5	9	19
200	1	5	10	16
300	2	13	15	30
400	5	7	12	24
500	3	6	13	22
600	4	11	7	22
700	1	6	12	19
800	5	12	11	28
900	3	8	12	23
1000	3	4	9	16
Total	34	84	121	239

beyond their hypocental domains. It can be considered, from one point of view, that seismicity consists only of these mutually independent earthquakes. Comparing the seismicity maps of relatively large earthquakes in Figs. 10a, b with those in Fig. 29b, we find they resemble each other better than when we compare Figs. 10a, b with Fig. 7 or Fig. 29a. It is interesting that the earthquake groups which seem to occur along or near the Quaternary active faults named Yamazaki and Mitoke in Fig. 43 fade out from the seismicity map of Fig. 29b. This may imply that primitive seismic activity must be viewed from another standpoint. In the area of this study, the western part and off the Japan Sea coast are low seismic active regions. And comparing Fig. 29a with Fig. 29b, we can find differences of regional earthquake occurrence types. These regional differences would result from differences in the nature of the rocks or the field of state that consits of seismic field.

3. Ishimoto-Iida's Relation

Relations between the frequencies and the maximum amplitudes observed at the five stations in Table 1 in the period from May 1965 to June 1968, are shown in Fig. 31. The sampling was done with a classification of two groups, that is, $S\text{-}P \leqq 4$ sec and $4 < S\text{-}P \leqq 16$ sec. The reading of $S\text{-}P$ times on seismograms was done on the principle that all earthquakes should be read as possible as we could, with remarks of their accuracies.

On those data, assuming that Ishimoto-Iida's relation,

$$n(a) = k\, a^{-m} \qquad (3)$$

a : maximum trace amplitude

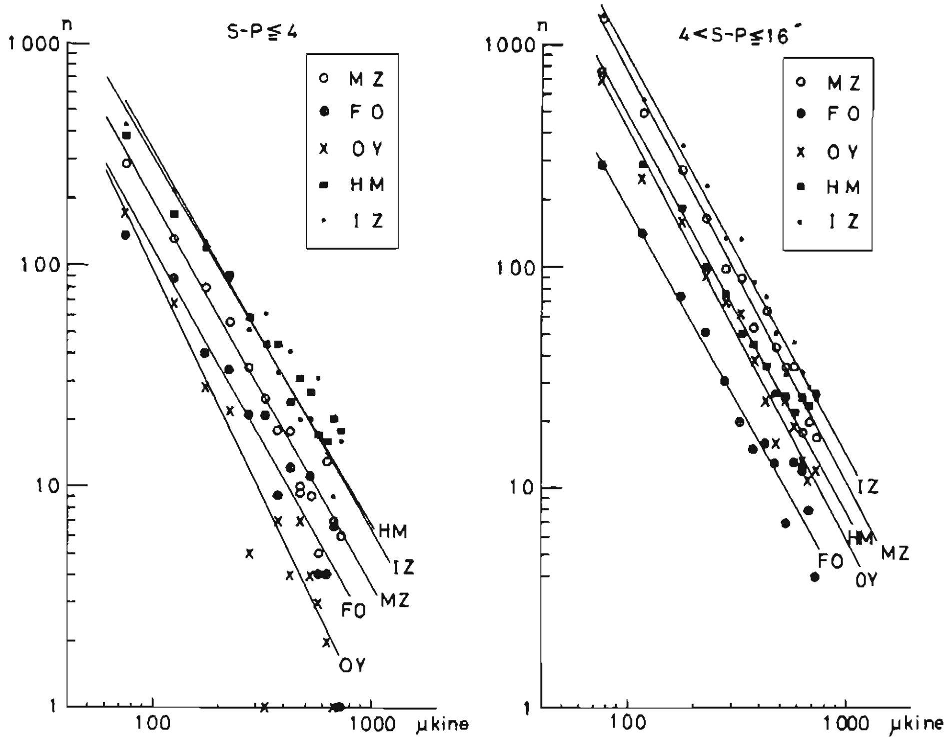


Fig. 31. Relations between the frequencies and the maximum amplitudes for each station. Data are divided into two groups of *S-P* time.

Table 16. Ishimoto-Iida's coefficient *m* for each station divided by *S-P* time and observation period. Numbers of earthquakes used in the calculation are shown in parenthesis.

Station	$S\text{-}P \leq 4^{sec}$	$4 < S\text{-}P \leq 16^{sec}$	Period
MZ	1.73 (786)	1.85 (2935)	Apr. '65 - July '68
FO	1.77 (423)	1.74 (773)	//
OY	2.04 (337)	1.84 (1598)	//
	1.67 (1210)	1.79 (1801)	//
HM	1.64 (367)		Apr. '65 - Dec. '66
	1.69 (841)		Jan. '67 - July '68
	1.70 (1297)	1.78 (3431)	Apr. '65 - July '68
IZ	1.74 (438)		Apr. '65 - Dec. '66
	1.63 (859)		Jan. '67 - July '68

holds, the coefficients *m* were calculated by the moment or most likelihood method (Utsu, 1965; Aki, 1965), and tablulated in Table 16. This method was confirmed to be the best by Utsu (1967a). The total numbers of earthquakes used in each calculation are shown in parentheses in the table. As far as the eq.(3) is supported, an earthquake number of 400 is enough to discriminate the differences of m by 0.15; and an earthquake number of 1000 by 0.10, in the confidence limit of 95% (Utsu, 1967b). As for $S\text{-}P \leqq 4$ sec, that is, when earthquakes occurred within about 30 km around each station, staions OY and HM are separated each other in the coefficient m. Station OY is distinguished from the other regions in high confidence limit. Station HM is also discriminated from the regional mean value of m in $4 < S\text{-}P \leqq 16$ sec in high confidence limit. Essentially eq. (3) becomes meaningless if the value of m differs from place to place or from time to time. In our data, m varies meaningfully at least in the range of 30 km. But when the data used contain systematic errors, the matter becomes complicated. In Fig. 31, we can find some examples that do not satisfy eq. (3) well. These deviations come mainly from the fact that comparatively large earthquakes, greater than 1000 μ kine, are counted less than the others. In the case $4 < S\text{-}P \leqq 16$ sec, which can be expected to cover our regions roughly uniformly, excluding the station FO which is relatively apart from the other stations, no meaningful differences can be found among the values of m.

Then the following points must be considered.

a) Either eq. (3) does not hold in these cases or else m varies with the amplitude.

b) m varies with a more minute scale.

c) Systematic errors in the observations and in the reading systems, such as: nonlinearity of the sensitivity of apparatus; processing of seismograms whose S-P time could not be read by any means; mixtures of noise with the data; processing of saturated seismograms or diminishing small earthquakes, etc......

The nonlinearity of sensitivity is not possible. A little fluctuation of sensitivity itself is not a problem. No earthquakes having amplitudes greater than 50 μ kine can be missed. But it is a problem if the ratio of the number of earthquakes whose S-P times are unreadable is related to the amplitude. Some saturated seismograms may be discarded in reading the S-P time; that is, the data of only large earthquakes fall off selectively. But in cases where $4 < S\text{-}P \leqq 16$ sec, no stations have such tendencies. Even if some parts of earthquakes had been missed and we could add those parts to our data, the existence of meaningful differences among the values m would be hard to eliminate.

There is many researches about Ishimoto-Iida's coefficient m or Gutenberg-Richter's coefficient b and their relations to regional geotectonics (Miyamura, 1962; Mogi, 1963b). Utsu (1967a) reexamined those investigations. According to them, the difference between the largest group of m or b and the least group is atmost 0.3. Some geological or geophysical differences are found between the Japan Sea side and the Pacific Ocean side of this district, but from a larger standpoint our field is considered as rather stable district in Japanese islands. Our results vary seemingly indifferent to the geotectonic activity up to the value of 0.3 even in the 30 km range. In the time field, if variations can be found as at station IZ in Table 16 or as in the case of Matsushiro earthquake swarm (Hamada *et al.*, 1966), we must be very careful to connect coefficient m directly to geotectonics.

It is said that there exist no relations between the coefficients k and m in eq. (3) (Aki, 1961). The coefficient k corresponds to seismicity and it suggests there exists no direct relations between m and main factors of earthquake occurrence.

From the seismicity maps Fig. 7, Fig. 11 or Fig. 29b, it is evident that near station FO, there are relatively few earthquakes except for a large earthquake in 1943. Within 30 km from station OY, there is almost a seismic vacant area; and about half of the data used in this study come from the earthquake swarm which occurred in 1965 near Hamasaka in Hyogo Prefecture (Kishimoto *et al.*, 1966). The hypocentral areas of this earthquake swarm were confined to very narrow area. Near station HM many little earthquake swarms occurred frequently. And near stations MZ and IZ also remarkable seismicity maps are developed. Thus each separate Ishimoto-Iida's coefficient m seems to correspond to the seismicity characteristics of their particular regions.

On the other hand, calculation of coefficient m is possible only when the sampling number attains a certain level, or else it will be hidden by other large numbers of data with different coefficient m. That is, m is not defined or cannot be calculated for an area with few earthquakes. When a few large earthquakes occur associated with fore- or aftershocks or earthquake swarms, which are known empirically to satisfy Ishimoto-Iida's relation, the system of this relation can be constructed approximately by these few earthquake groups. By extending the space or time field, and accumulating data of earthquakes caused by complex factors, it might be possible to approach a cirtain value of Ishimoto-Iida's relation statistically. One of the reasons why the results of an essentially meaningless superposition of a different value of m in $4<P\text{-}S\leqq16$ sec well support the eq.(3), may come from the above. It would be useful to know the coefficient m or b in limited narrow area or short time interval for simplifying the factors of earthquake occurrence.

Studying the physical aspects of Ishimoto-Iida's coefficient m, Mogi(1962) said that the m of heterogeneous rocks is greater than that of homogeneous rocks. And he also found by experiments on some heterogeneous rocks that the difference between m of fore- and aftershocks amounts to 0.3 (Mogi, 1963a). Nagumo (1969) studied the relation between the spectrum of plastic deformation and the coefficient m. Suzuki *et al.* (1966) showed by experiments on tempered glass that at the place where stress is focused m is disturbed. In our case m might have varied from place to place or from time to time according to some heterogeneous unknown factors.

4. Focal Mechanism

We have already found, by the study of microearthquakes, that the minimum principal axis of compression in the northern part of Kinki District directs nearly latitudinally (Hashizume *et al.*, 1966).

In this paper, the author has added further data and investigated this problem from various aspects. The data used here are the initial motions of the *P* waves of microearthquakes which occurred from August 1964 to June 1968 and whose hypocenters were determined. The observation stations were those in Table 1 and Table 3. Indistinct initial motions were not included in the data. It is impossible to determine focal mechanisms for each earthquake as the number of stations is limitted, so the data of each station were superposed putting the epicenters as origin. The epicenters are classified into eight regions, as shown in Fig. 32, in which the observation stations are marked as solid circles. The results are plotted in Fig. 33 using open circles for the push sense and solid circles for pull sense. The numerals shown outside the figures correspond to the regions in Fig. 32 and those shown inside the figures signify as follows:

Period	Aug. 1964 - Dec. 1966	1, 3, 5, 7
	Jan. 1967 - June 1968	2, 4, 6, 8
Magnitude	$M \geqq 1.2$	1, 2, 5, 6
	$M < 1.2$	3, 4, 7, 8
Epicentral	$D \geqq 20$	1, 2, 3, 4
distance	$D < 20$ (km)	5, 6, 7, 8

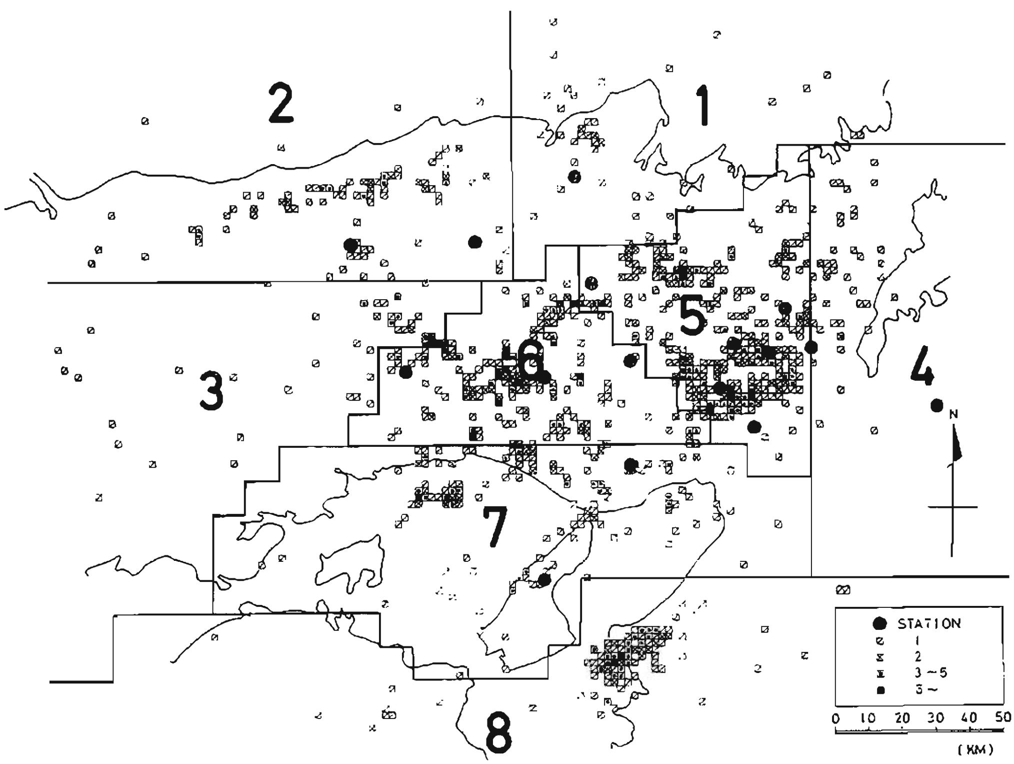
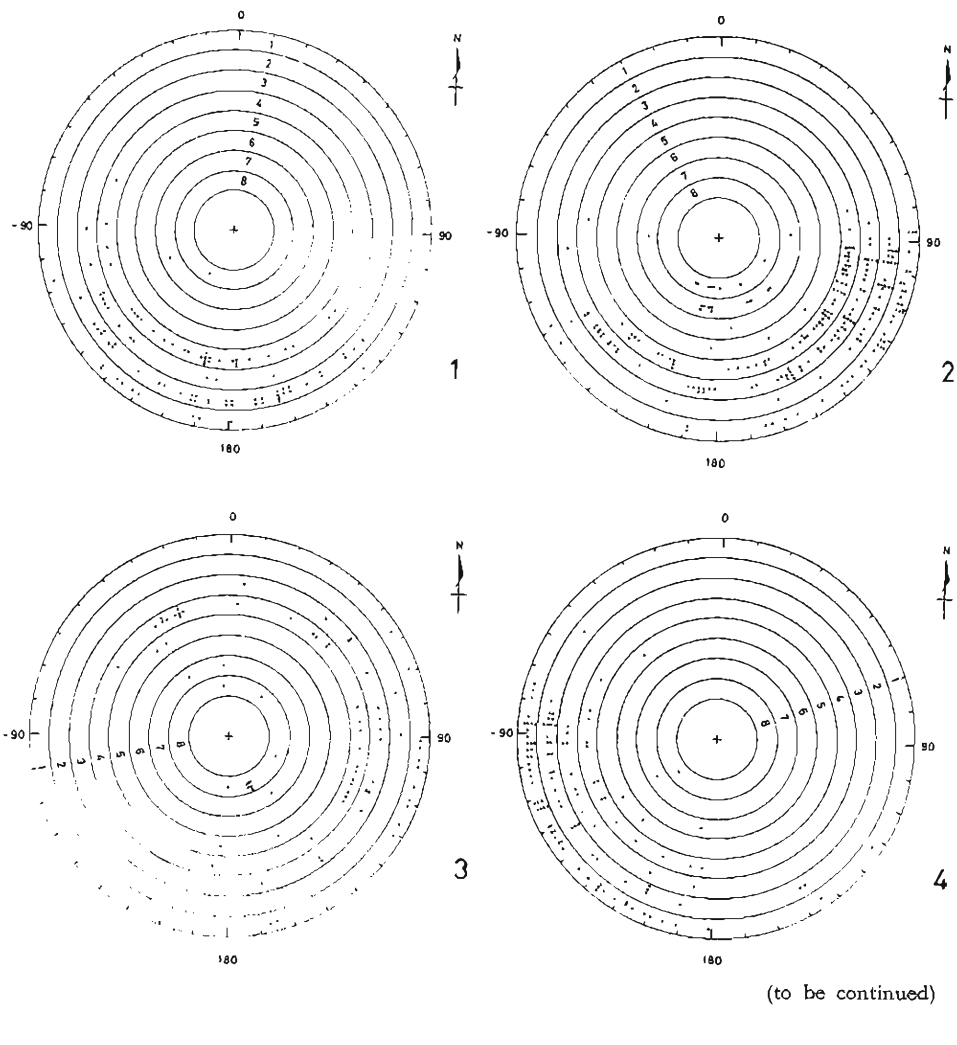


Fig. 32. Regional classification in studying focal mechanisms.

From these figures we can find each figure is roughly divided into four quadrants of push and pull regions whose axes are perpendicular each other. In analysing the patterns of the stress fields, the crosscorrelation is calculated between Fig. 33 and the step function in Fig. 34 varying the Φ and θ in the figure. Push is set as $+1$ and pull as -1. Data whose epicentral distances are less than 20 km are excluded.

Regional dependence. As the first approximation, the nodal lines are set to intersect at right angles, that is, the case $\theta=45°$. The results are shown in Fig. 35. In regions 4, 5, 6 the patterns are simple and the coefficients of crosscorrelation are as high as nearly 0.7–0.8. The direction of the principal axis of compression is nearly N 85°E. In region 7 the matters are similar to regions 4, 5, 6 but the direction of the principal axis of compression is nearly N 85°W. Among regions 1, 2, 3, regions 1, 3 are not far from the patterns of the above regions but the direction of the



(to be continued)

principal axis of compression of region 2 much deviates from the others and is nearly N 60°W. Region 1 shows an intermediate pattern between region 2 and regions 4, 5, 6, 7 but the arrangement of the observation stations is not suitable for this region and the number of data is deficient so that nothing can be stated in detail. Although the above regions have the maximum coefficient of 0.7–0.8, this

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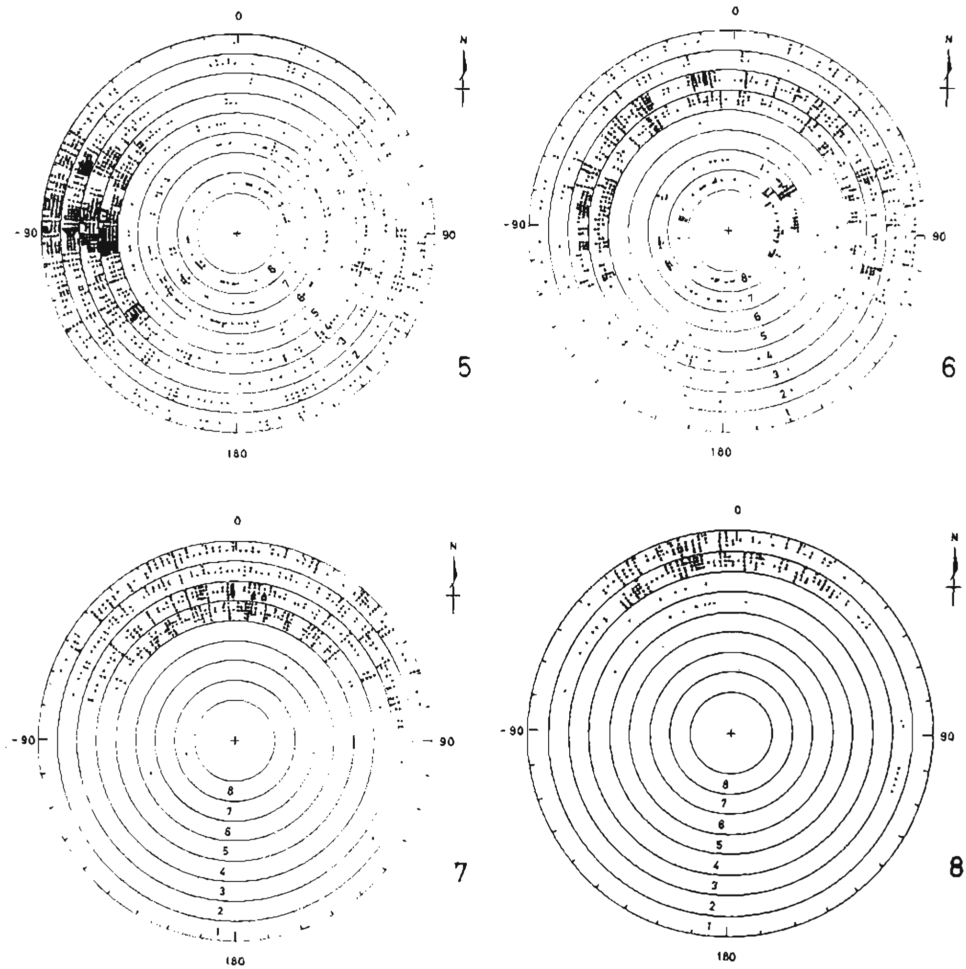

Fig. 33. Push-Pull distribution of the initial motions of P waves, setting the epicenters as origins (cross). Open circle : Push, Solid circle : Pull. Numerals inside the figures are as follows:

Period	Aug., 1964 Dec., 1966	1, 3, 5, 7
	Jan., 1967 June 1968	2, 4, 6, 8
Magnitude	$M \geqq 1.2$	1, 2, 5, 6
	$M < 1.2$	3, 4, 7, 8
Epicentral distance	$D \geqq 20$	1, 2, 3, 4
	$D < 20$ (km)	5, 6, 7, 8

Numerals outside the figure correspond to regions in Fig. 32.

value reduces to about 0.3 in region 8 and the direction of the principal axis of compression is nearly N 60°W. The focal mechanism in this region must be different from those in other regions.

Magnitude dependence. According to the above data the focal mechanisms of earthquakes are similar each other in regions 4, 5, 6 and 7, thus these regions will be discussed collectively. The earthquakes are divided into two groups on a base of magnitude M=1.2. The resultant crosscorrelation calculated, setting $\theta=45°$, is shown in Fig. 36. The numerals in parentheses show the number of data used in each calculation. This is the same for the followings. The maximum values of the coefficients of crosscorrelation are about 0.7 and the patterns look quite similar. It is found that earthquakes occur in surprisingly the same focal mechanisms for rather small as well as for large earthquakes. It is natural that the coefficient is a little higher for greater magnitudes considering the degree of easiness in reading the initial motions. Furthermore, pressure direction are consistent to the general tendency of the focal mechanisms of very shallow large earthquakes determined by Ichikawa (1966), as shown in Fig. 37; but it seems that the fluctuations of focal mechanisms of large earthquakes are slightly greater than those of microearthquakes.

Depth dependence. It is very important to study focal mechanisms in relation to depth in order to know the spatial variation of stress patterns. The crosscorrelation is shown in Fig. 38 in the same manner as the above. The calculations were done on regions 5, 6, 7 excluding 4 because of the difficulty of hypocenter determination (Hashizume, 1970 b). The focal mechanisms are nearly the same within a depth of 15km. For depth greater than 15 km, the sampling data were not sufficient and the result can easily be affected by special earthquake series.

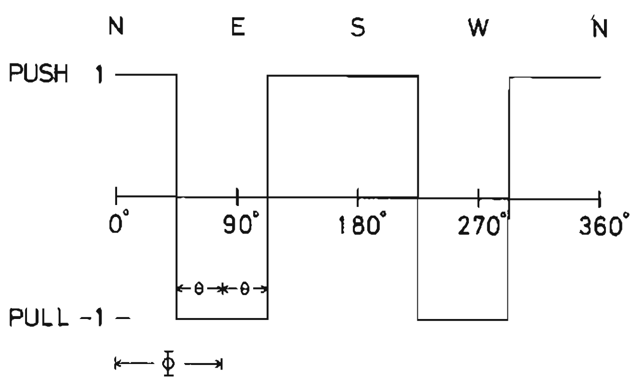

Fig. 34. Function used in calculating crosscorrelation for the study of focal mechanisms.

Effect of preceding earthquakes on focal mechanisms. Next, we shall study the effect of preceding earthquakes on focal mechanisms. The focal mechanisms of earthquakes in Fig. 29 b which are thought to consist mainly of mutually independently occurring earthquakes omitting those earthquakes which occurred within the same hypocentral domain within a certain time interval as mentioned in the foregoing section, are shown for regions 4, 5, 6, 7 in Fig. 39 and Fig. 40 in the same manner as above. For reference, crosscorrelations for all earthquakes in these regions are shown by dotted lines. Although the data are not abundant, the coincidence of the two patterns is very good. Thus it is possible to say that the focal mechanism does not vary according to the sequence of earthquakes. This was recognized also in the Matsushiro earthquake swarm (J.M.A., 1968).

Intermediate principal axis. Hitherto, it has been assumed that two nodal lines intersect at a right angle,

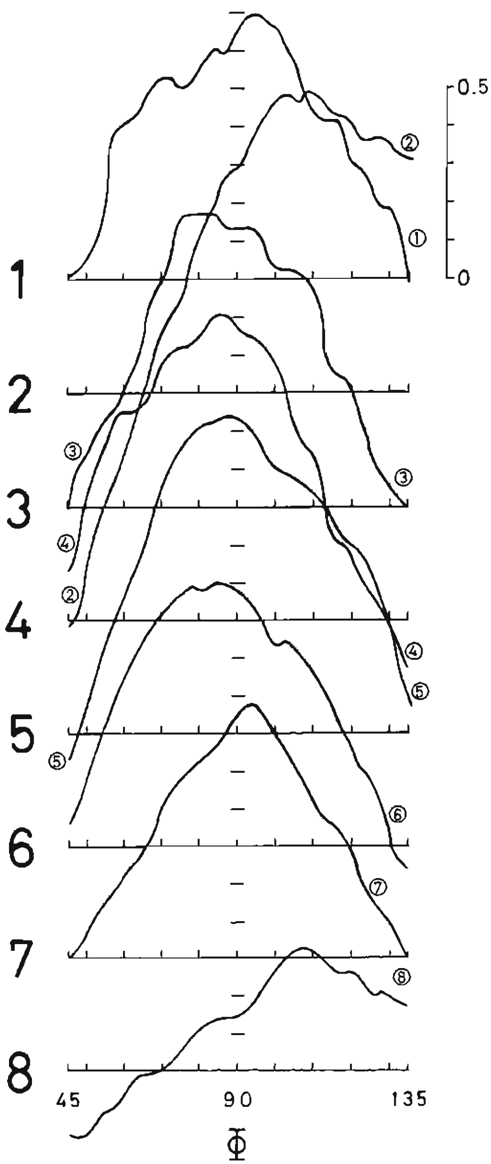

Fig. 35. Crosscorrelation between the push-pull distribution in Fig. 33 and the step function in Fig. 34 setting $\theta=45°$. Numerals on the left side of the figures correspond to regions in Fig. 32.

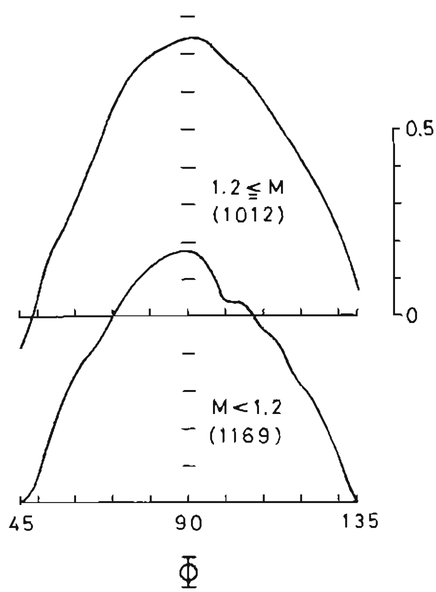

Fig. 36. Crosscorrelation relating to magnitude. Number of data is shown in parentheses. (Regions 4, 5, 6, 7)

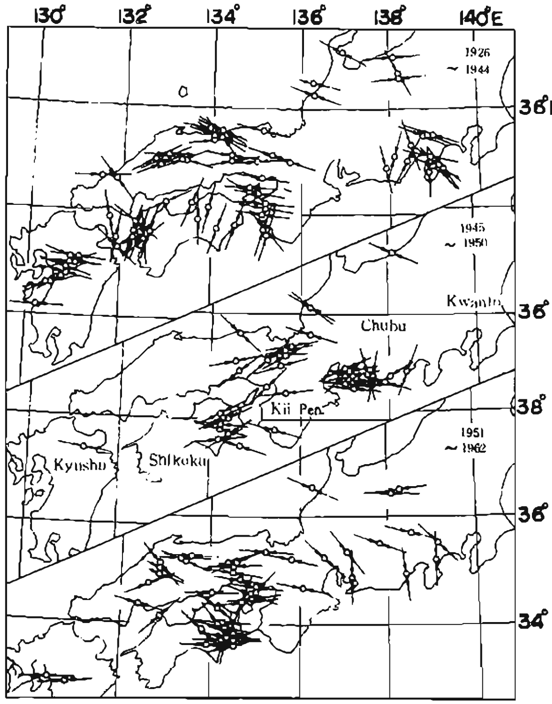

Fig. 37. Pressure directions of very shallow earthquakes occurring in the period 1926–1962. (after Ichikawa, 1966)

that is, the orientation of the null vector is vertical. So, we shall calculate the crosscorrelations by varying the intersect angle 2θ of the two nodal lines, for regions 1, 2, 3 and regions 4, 5, 6, 7. This analysis is valid when focal depths are confined within 15 km and the epicentral distances within 20 km. The results are shown in Figs. 41 a and b. In these figures, numerals in circles are the acute intersect angles 2θ of the two nodal lines; and this angle θ corresponds to θ in Fig. 34. It can be easily seen that the two patterns resemble each other. Thus, on a fairly large scale, the pattern of a stress field differs little relating to the regions. The direction of the principal axis of compression is nearly latitudinal in regions 4, 5, 6, 7 while it rotates clockwise about 10° from that in regions 1, 2, 3. The maximum coefficient of crosscorrelation is about 0.7 at $2\theta=90°$. The preceding analyses, putting $2\theta=90°$, were found to be suitable. In this northern part of Kinki and Chugoku districts, the major focal mechanisms of earthquakes take a vertical null vector or strike slip fault whose trend is nearly NE or NW. As the acute intersect angle 2θ of the two nodal lines decreases, the value of the coefficient of crosscorrelation near $\Phi=90°$ decreases rapidly and shows contrast to that of the coefficient near $\Phi=180°$. This tendency does not come from the deviation of the location of stations, epicenters, misreading, or any other effects except that the

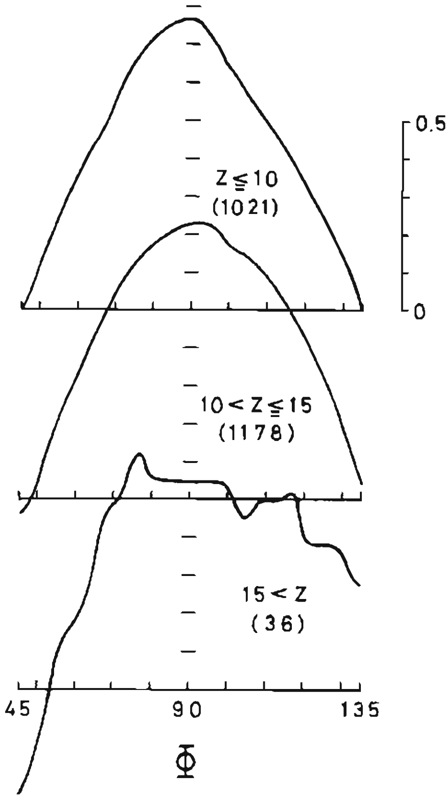

Fig. 38. Crosscorrelation relating to focal depth. (Regions 5, 6, 7)

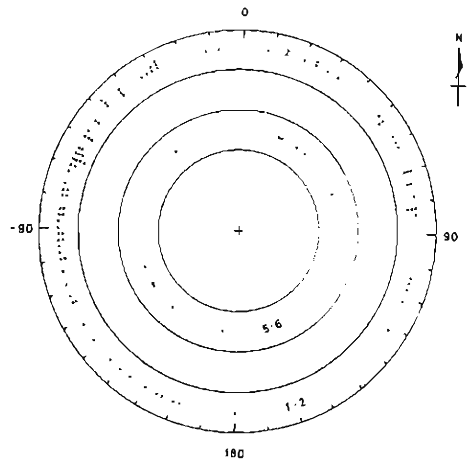

Fig. 39. Push-Pull distribution of the earthquakes in Fig. 29b, from which aftershocks and swarm earthquakes are omitted. (Regions 4, 5, 6, 7) Notations are as same as Fig. 33.

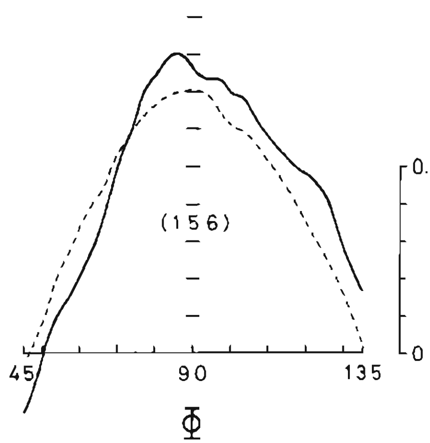

Fig. 40. Crosscorrelation for the earthquakes in Fig. 39. (Regions 4. 5, 6, 7)

orientation of the minimum principal axis is not so stable. Thus some earthquakes may occur by thrust faults with a longitudinal trend. On further inspection, we found that there were some earthquakes caused by stresses whose direction rotates a little clockwise from that of the major stress field, with some dip components. This type of earthquake may be dominant in some areas.

Summing up the aboves, the following conclusions are presented;

1. The stress field in the northern part of Kinki and the eastern part of Chugoku districts have a nearly latitudinal principal axis of compression. The great majority of earthquakes are caused by this major stress field.

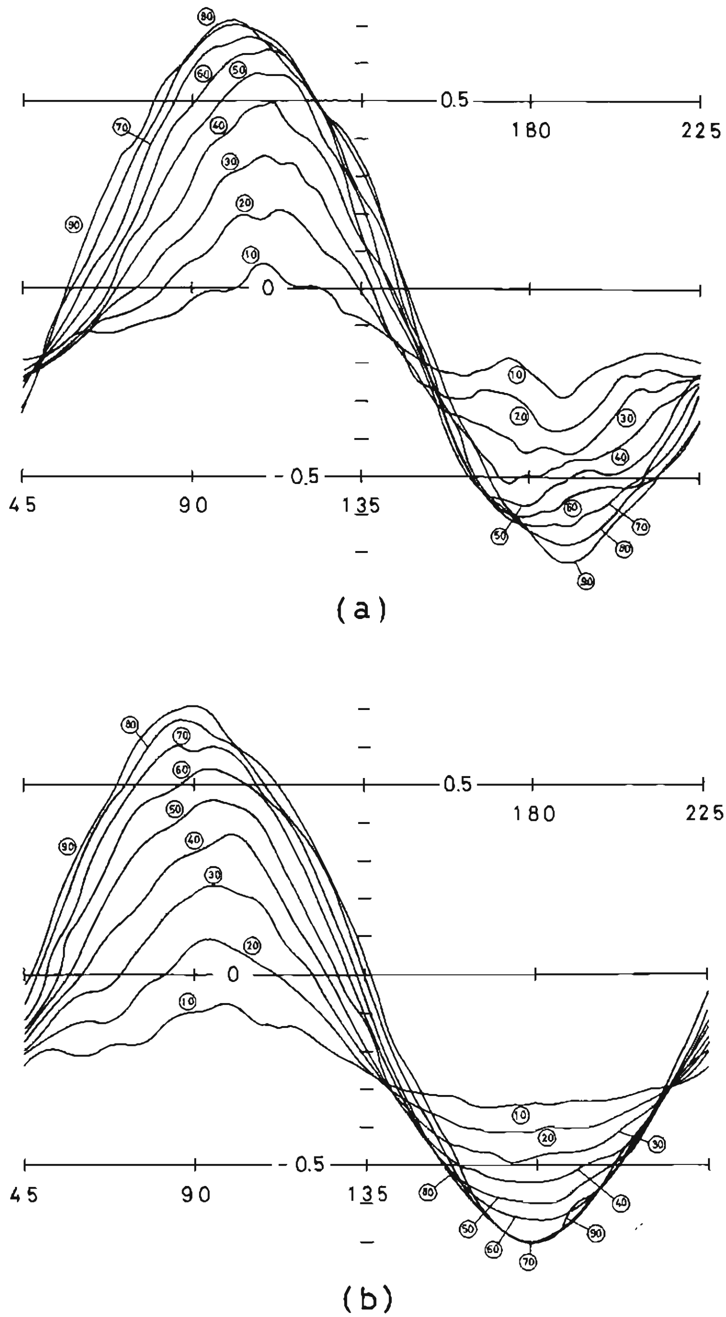

Fig. 41. Cross correlation varying the intersect angle θ of the two nodal lines in Fig. 34. (a) : Regions 1, 2, 3 (b) : Regions 4, 5, 6, 7

2. Most of these earthquakes are considered to be of the strike slip type, but there exist some dip slip type earthquakes. Putting maximum differential stress (dilatational) σ_3, intermediate differential stress σ_2, minimum differential stress (compressional) σ_1, the major stress field is,

$\sigma_3 > \sigma_2 > \sigma_1$
σ_3 : longitudinal
σ_2 : vertical
σ_1 : latitudinal

The difference between σ_2 and σ_3 is estimated to be not so large, and sometimes σ_2 becomes equal to σ_3 and the maximum and intermediate principal axis becomes unstable.

3. The focal mechanism does not vary remarkably according to either the magnitude or the earthquake type.

4. The above stress field is nearly uniform in the upper crust, 15 km thick; but it remains unknown for the lower crust.

Now some geological sketches of our seismic field will be described. The northern parts of Kinki and Chugoku district belong to the inner zone of Southwest Honshu arc. They are bounded by a Median line at the southern end and by a Fossa Magna at the eastern end adjoining Northeast Honshu arc. This southwestern arc is considered to be more stable than the eastern arc and has medium characteristics between those of the continent and of the island arc (Matsuda *et al.*, 1967). Mizouye (1967) said that the type of crustal deformation found in this southwestern arc is also midway between that of the continent and of the island arc. He concluded this by spectrum analysis of the crustal deformation. On the other hand, the activity characteristic of the Cenozoic era called Green Tuff activity was seen in these districts but less vigorous than in the eastern arc. Other geophysical data such as gravity, heat flow, volcanism, and crustal structure show that these are not so complex but not simple at all.

The division of the basement rocks in southwest Japan is shown in Fig. 42 (Huzita, 1962). Here, our regions 1, 2, 3 and regions 4, 5, 6, 7 are bounded by the Maizuru zone, and the basement rocks of the western part consist mainly of Cretaceous granites of the San'in type, while those of the eastern part are Palaeozoic formations and Ryoke gneisses. Huzita (1969) said that the stress pattern in these districts had changed perpendicularly from nearly longitudinal compression to nearly latitudinal compression in the late Pliocene or early Pleistocene. Thus the former period is called the First Setouchi Series and the latter the Second Setouchi Series according to morphological and structual investigations. At this time of stress field change, the tectonic features called Kinki triangle were formed, creating a complex topography by the mutual influences of the above basement rocks. The apex of this triangle is Turuga Bay and its base is the Median line connecting the Hira and Awaji mountains on the western side and

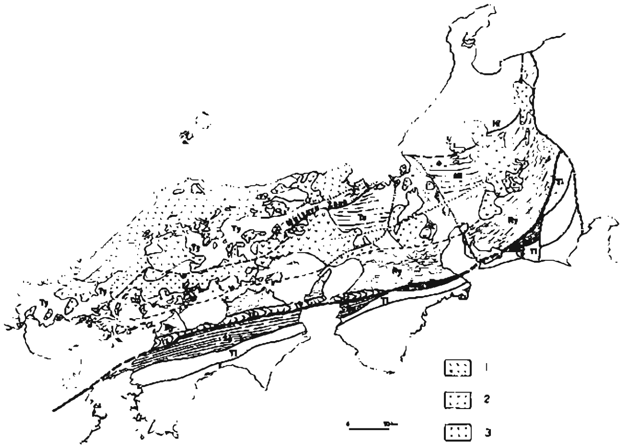

Fig. 42, Division of the basement rocks in Southwest Japan. 1 : Cretaceous acid volcanic rocks and their pyroclasts. 2 : Cretaceous granites of the San'in-type. 3 : Cretaceous granites of the Hirosima-type. Hi : Hida Complex. Ta : Tamba Paleozoic Terrain. Mi : Mino Paleozoic Terrain. Ry : Ryoke Gneiss zone. Ty : Tyugoku zone of Paleozoic. Ti : Titibu zone of Paleozoic. Sa : Sambagawa metamorphic zone. Iz : Cretaceous Izumi group. (after Huzita, 1962)

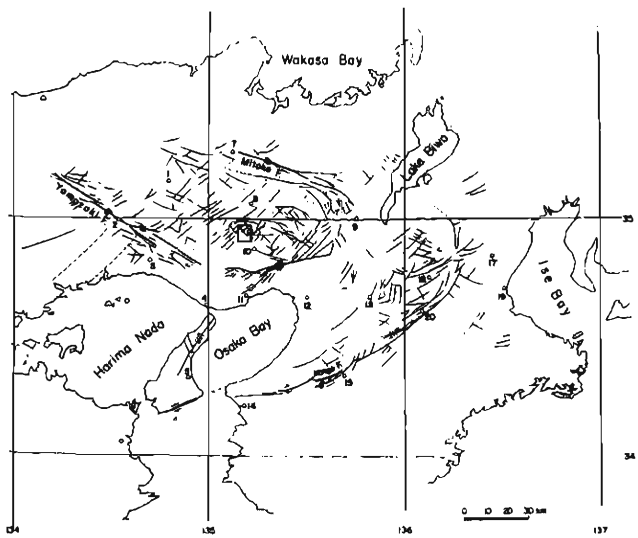

Fig. 43. Map showing the youngest fractures along which no distinct vertical displacement can be recognized. (after Huzita, 1969)

the Yoro mountains on the eastern side. We have no data to explain the former pattern, but the more recent pattern is well explained with our seismic data.

Region 7 faces the Seto Inland Sea and the province is suffered depression through the Second Setouchi Series, whose degree of depression is 5 mm/y, according to recent triangulation (Dambara, 1968). On the other hand, the Rokko mountains in this region show recent upheaval activity measuring 500 m/my, according to geological data (Ikebe *et al.*, 1966). Any movement, however, must proceed maintaining the relation of $\sigma_3 > \sigma_2 > \sigma_1$ where σ_1 is latitudinal. The joint system analysis made by Hirano (1969) at the Rokko mountains well supports this relation. Due to this latitudinal stress field, the rounded topography of this district shows saw-toothlike undulations with a longitudinal trend with a wave length of about 80–100 km. Their western side forms a gentle slope, and their eastern side forms a steep slope accompanied by longitudinal thrust faults (Huzita, personal communication). This also follows from our findings that some earthquakes are accompanied by thrust faults. The trend of the Median line is roughly latitudinal and it is known to be undergoing right lateral slippage recently (Kaneko, 1965). This movement might correspond to the above latitudinal stress field. We can easily suppose that a compressional stress source exists at the western side of the inner zone of Southwest Honshu arc. Huzita (1969) also shows the joint system map which is considered to be created in the Second Setouchi Series (Fig. 43). The focal mechanisms calculated by earthquakes and joint systems correspond very well. Kasahara *et al.* (1964) calculated the minimum principal axis from triangulation data but it does not necessarily coincide with our results. Maybe the characteristics of his data are somewhat different from ours.

5. Earthquakes and Fractures of Rocks

We do not know the real mechanisms of earthquake occurrence; however, it seems no doubt that earthquakes in the crust are caused by brittle fractures of rocks in the crust. It is known from structual geology that conjugate sets of faults are usually observed in fracture systems and that each of them makes an angle of about 30° with the minimum compressional principal axis. Then, if earthquakes occur by this fracture system, the expected focal mechanisms must be as shown in Fig. 44 in which the ruled area indicate the mixed regions of push and pull sense when many earthquakes are summed up, as above. We shall investigate this problem using the preceding data.

According to the principal axis determined in Fig. 35 for each region, we set the two nodal lines which intersect perpendicularly each other. The angle α is taken clockwise or counterclockwise from each principal axis of stress ($\alpha=0°$) to the two adjacent nodal lines ($\alpha=45°$). On thus assumed focal mechanism the consistent and the inconsistent data with the expected sense of initial motions are summed up separately as for the angle α. The number of consistent data is set as n_1 and of inconsistent data as n_2; then the ratio $n_2/(n_1+n_2)$ is plotted in Fig. 45, grouping every two degrees. The numbers of data n_1+n_2 are classified into three ranks in this figure. If the push-pull distribution is restricted only by the two conjugate faults and each conjugate fault occurs at the same ratio and suppose the fracture angle α is 30°, then the angle made by the ruled area should be 30°, that is, $n_2/(n_1+n_2)=0.5$ for angle greater than $\alpha=30°$, and $n_2/(n_1+n_2)=0$ for less than $\alpha=30°$. The data are confined to regions 5, 6, 7 in Fig. 32, epicentral distances greater than 20 km and focal depths less than 15 km. Set the ratio to be 1 : 2, then $n_2/(n_1+n_2)$ should be 0.33 for the angle greater than fracture angle. Therefore, in this case, the fracture angle should be nearly 45° from Fig.4 5 and even in the case that the ratio is set as 1 : 3, that is, $n_2/(n_1+n_2)=0.25$, the fracture angle does not decrease to less than 40°.

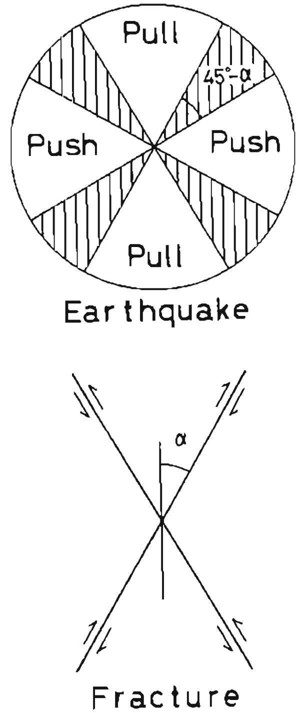

Fig. 44. Assuming that earthquakes are caused by conjugate sets of fractures, the expected focal mechanism is shown. Ruled area are mixed areas of push and pull of initial motions of *P* waves when many earthquakes are summed up as in the previous discussions. The ratio is controlled by the ratio of two conjugate faults.

Now, we must consider the following items for the above estimation of the fracture angle.

a) Reading errors.

b) Mixing of data of earthquakes which have completely different patterns of focal mechanism.

c) Fluctuation of stress vector.

d) Declination of null vector.

e) Mutual interaction among fore-, main, and aftershocks or swarm earthquakes.

f) Preferred orientation in the crust.

For *a*, *b*, *c* and *d* the above estimation of fracture angle indicates minimum value. There are some data that contradict each other and there is a high possibility of reading errors near the nodal lines. Some earthquakes have focal mechanisms that clearly deviate widely from the majority. Naturally these earthquakes should be omitted from the data. The reason why $n_2/(n_1+n_2)$ does not decrease to zero for small angles of α, may come from the above. As stated before, some earthquakes are caused by dip slip faults, although the majority are of the strike slip type. It can be concluded, explaining the data in Fig. 45

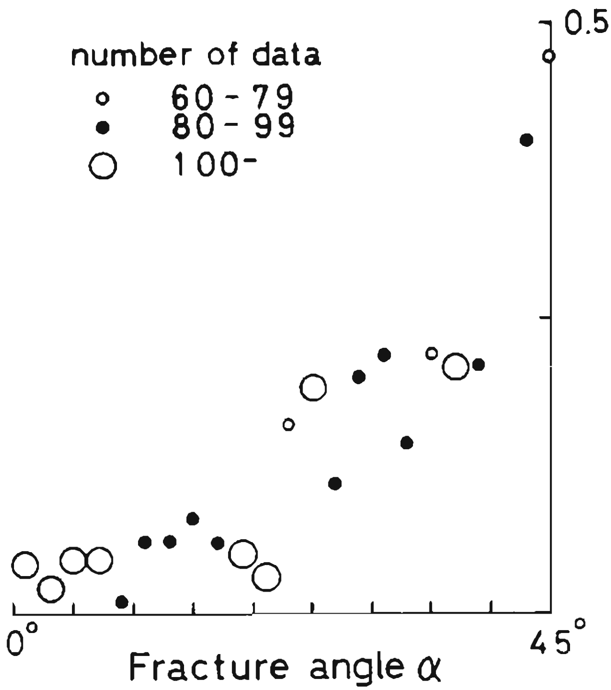

Fig. 45. The ratio of inconsistent data to the sampling number for the assumed pattern of Push-Pull distribution. Principal axes or nodal lines are determined from Fig. 35 and set as $\alpha=0°$ and $\alpha=45°$ respectively. Assuming the ratio of two conjugate faults, the fracture angle α can be read.

only by the term c, that more than about 80% of earthquakes are of the strike slip type whose intermediate principal axis inclines less than 25° to the vertical direction. Moreover it is hard to consider that all earthquake faults occur under closely uniform stress conditions in space and time. When the fracture angle was estimated, the above factors are not considered, but the real focal mechanism must be influenced by those various factors. Then it can be safely said fracture angle should be more than 40° and more than 80% of earthquakes are of the strike slip type, so far as the influence by the term e and f is neglected. The effect by term e of the above list may not be so great, because according to Fig. 40 the focal mechanisms are the same both for mutually independent earthquakes and for a whole earthquake series. Even if one of the conjugate faults is preferred by the influence of preceding earthquakes, those earthquakes must be evenly distributed over the region. That is, the effect from term e can be neglected. As for term f, there must a priori exist a preferred orientation, seismologically or geologically over wide range. If this is a seismological preferred orientation, as it is clear from the seismicity map in Fig. 7 that the earthquakes do not occur in the same manner as those along the San Andreas fault, the problem reduces to the earlier occurrences of many faults with a uniform trend over a wide area. On the joint system map of Fig, 43, when we pay attention to rather small fractures, the ratio of two conjugate faults seems nearly unity as a whole although areal tendency to take one side can be seen. There are some examples of restoration of stress field using nearly uniformly developed conjugate sets of faults with 10–100cm order of length (Hirayama *et al.*, 1965). The fault length of microearthquakes is considered to be of the 100 meter order, and this is the intermediate order of fault length of the two cases. Eevn if there exist some preferred orientations, the ratio would not be less than 1/4, so far as not to set an unnatural supposition. So it can be concluded that the fracture angle of the faults caused by the earthquakes concerned is larger than 40°.

It seems, on the other hand, that the focal mechanisms determined for the large earthquakes in Fig. 37 and for the joint system in Fig. 43 fluctuate more than those expected from the study of microearthquakes. This fluctuation may come from the fact that the faults of large earthquakes tend to take a fracture angle near 30° at the surface and this increases with depth to 40° or more on the same fault. Moreover they are affected by surface geology or topography, in addition to errors or ambiguity of determination of focal mechanism or location of fault, and by the differences of the mechanisms of occurrence of large earthquakes.

The fracture angle of a dry sample does not exceed 35° under a pressure of 3 kb, as shown by experiments of pressure effect on probable rocks (Mogi, 1966). Byerlee (1967, 1968) studied the effect of the nature of contact surface and of interstitial water on the frictional coefficient by already fractuated rock; and he also studied this coefficient on the assumption that brittle-ductile transition is related to internal friction. All of these studies showed that the fracture angle did not exceed much more than 30° and that remarkable effects could be observed neither from the presence of water nor from the nature of the surface. Taking into account the data from natural earthquakes which occurred at an average hydrostatic pressure of about 3 kb, the coefficient of Coulomb-Mohr internal friction in the real crust may be less than that expected from experiments on rocks. This problem needs further investigations.

6. Earthquake Occurrence and the Nature of the Crust

The author has found that the most of the earthquakes in the district under discussion occur in the upper crust and the main factor controlling their occurrence is the uniform stress field over the seismic field.

Now, according to the experiments on the fracture of probable dry rocks, the differential stress necessary to cause fracturing attains 10 kb under a confining pressure of 5 kb. And the pattern of fracture of some rocks shifts from brittle fracture to ductile flow even in the upper part of the crust (Griggs *et al.*, 1960). On the other hand, Chinnery (1964) estimated the stress drop of large earthquakes to be at the most 100 bars, applying dislocation theory to the faults. And Aki(1966), using surface waves, estimated the stress drop of the Niigata earthquake kinetically to be 100 bars. For most earthquakes the stress drop is now estimated to be within $10-10^2$ bar. Stress drop of one bar order or less is also observed (Brune *et al.*, 1967). To explain the difference of stress drop between experimental results and that estimated from natural earthquakes, such effects as size effect, temperature effect, strain rate effect and so on are considered; but it seems difficult to explain these gaps of 10^2 times or more only by those effects on brittle fracture. The idea that by such stress drop only a part of the stress is released as a stick slip fault (Brace *et al.*, 1966) also faces the difficulty of considering the order of magnitude. The so-called "Griffith crack" is proposed to explain the strength of the material (Griffith, 1921). The orientation of this growth must be consistent with the data in terms of focal mechanisms. According to the derivation of this theory, it is very difficult to show the growth of such a crack takes the same orientation and there must exist other processes to account for many Griffith cracks to concentrate along a fault in a definite trend expected from the stress field. The same thing is true when numerous minor fissures grow into one large fault.

There is a hypothesis that earthquakes are caused by fractures due to stress concentration on the heterogeneous parts of the medium. As we have already shown most earthquake mechanisms are determined by the stress field over a wide area. If earthquakes are due to concentration of stress on the heterogeneous parts of the material in the crust, there must exist great concentration of stress to cause fracture of $10-10^2$ bar order in the stress field. On the natural reduction the concentrated stress field is in disorder to a great extent, and the focal mechanism must fluctuate regardless of the major stress field. It is very difficult to find a model in which mutually independently occurring earthquakes satisfy the two conditions, that is, to have the same focal mechanisms in a unform stress field and very low stress drop at the same time. Another explanation for this difficulty is that the crust has already had a preferred orientation by which most of the earthquakes are controlled. But this explanation also raises many difficulties. The first is that such a uniform preferred orientation must exist over a wide area. The second is that almost all the focal mechanisms are controlled by very strongly preferred orientations which are indifferent to the stress field. We can observe everywhere in the field many epigenetic faults or fissures whose trends are not related to the schistosity or geological trend. In extreme cases we can observe faults that cut across pebbles included in unconsolidated sedimentary rocks, splitting them into two to the direction of the fault. Furthermore, it is interesting to note that the neotectonic stress field agrees with the data from the study of focal mechanisms.

Then for a model of earthquake occurrence ,we may consider that earthquakes occur where stress is not large but strength is weak.

The low value of internal friction estimated from natural earthquakes is concordant to this model. But in this case some lubricators must exist. Griggs *et al.* (1965) found that the strength of the material dropped sharply when water was introduced into a quartz crystal. This phenomenon was explained by the H_2O molecule weakening the silicon oxygen bond. Raleigh *et al.* (1965) also found a remarkable strength drop of material by the dehydration of serpentinite. He explained this phenomenon by an increase of pore pressure. It is known that the surface energy of a Griffith crack in a material decreases by the presence of H_2O (Orowan, 1944). Of course we need much more research to apply these findings to earthquake phenomena in a real crust, so far as the state and content of H_2O or fluid material such as CO_2 is not sufficiently known. However, this may be very pertinent to the generation of earthquakes. In general, from the modification of the Coulomb-Navier equation,

$$\tau=\tau_0+(\sigma-P)\ \tan 2\alpha \tag{4}$$

τ_0 : cohesive strength

$\tan 2\alpha$: coefficient of internal friction,

when the fluid pressure P approaches to the solid pressure σ, the strength τ decreases to the cohesive strength τ_0 (Hubert *et al.*, 1959). The value of the cohesive strength is estimated to be less than 1 kb for dry rocks and the order of stress drop of natural earthquakes for water saturated materials (Byerlee, 1967). When the system is closed, the precipitation of solid material in the fluid phase proceeds, then pore liquid pressure may increase up to solid pressure.

Thus, one of the important factors in weakening material strength may be the existence of water in the crust, beyond the various other effects, such as, strain rate, size effect, temperature effect and so on. The earthquake swarm that occurred at Denver in 1965 because of the presence of water supports this hypothesis. Ishimoto-Iida's relation comes from heterogeneity in the earthquake field; and the coefficient m in eq.(3) indicates heterogeneity. Mogi (1962) said occurrence patterns of earthquakes including fore- or aftershocks or swarm earthquakes are related to the heterogeneity of crustal material, where the heterogeneity should be read as heterogeneity of strength. Further, if this is replaced by heterogeneity of P_{H_2O}, it may be possible to explain brittle fracture with low stress drop without disturbing the stress field. And if the heterogeneity of P_{H_2O} varies in a short range, the coefficient m may vary with it. And also the time variation of m is easily induced by the effect of fluid material. After the fractures are formed H_2O moves along them. Thus an earthquake may induce other earthquakes beyond its hypocentral domain. In most cases the focal mechanisms of aftershocks are similar to those of the main shocks. And this is the same for earthquake swarms. If the earthquakes are caused by a concentration of stress, the stress field must change greatly after the release of that stress. Considering the distribution of earthquakes, it seems improbable that they all slip on the same main fault. Our hypothesis is that after the main shock H_2O is released and fault surface is fixed with increasing resistance to slippage and the stress field can recover elastically to a certain degree of its initial state, whereupon the unstable parts caused by the main shock fractuate as aftershocks.

Much information has been found regarding the changes in the nature of rocks, changes such as in the melting point or in electrical conductivity, that are produced by adding small quantities of water. These studies have been done in relation to real earth models by many investigators. The solubility of silicate in H_2O increases drastically as temperature or pressure increases (Sidney *et al.*, 1966). This is important in connection with the transportation of material and earthquake phenomena. It has been shown that it is impossible to constitute regional metamorphism only by solid diffusion without material transportation by H_2O. It is interesting in connection with the release of water that most of metamorphic rocks are considered to be open to H_2O (Miyashiro, 1965). Recently, it has been found that H_2O changes its nature greatly when it permeates into capillary tubes (Derjaguin, 1967). We do not yet know about the effects on rocks in the crust as the field of earthquakes. A study of the nature of water-containing rocks in the crust should yield very important data.

Then where does the water come from? The larger part may have been stored in the crust since the time when the crust was formed, stored either as free water or as constituents of minerals. According to Fig. 11, most of the microearthquakes occur within a depth of 15 km. It has already been noted that the upper crust and the lower crust have different natures and they are fairly sharply defined from each other at about this depth (Hashizume, 1970c). Whether this is a general case or not, it will require further researches but it seems that there may exist a very important relationship between earthquake phenomena and the development of the crust with regard to its formation, metamorphism and the release of water. We have discussed earthquakes that occur in the crust. The problems of tectonic activities in the deeper parts of the earth are remained for further study.

Acknowledgments

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