Investigation of Microearthquakes
—On Seismicity—

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(Manuscript received August 30, 1969)

Abstract

On the hypocenters of microearthquakes observed by the seismological network attached to the Tottori Microearthquake Observatory of the Disaster Prevention Research Institute of Kyoto University. The outline of observed data, method of hypocenter determination and seismicity maps are presented.

About 1500 hypocenters of microearthquakes were determined using the data obtained at five seismological stations from Aug. 1964 to June 1968. About 98% of them were determined within the standard deviation 0.15 sec by the method of hypocenter determination discussed in this paper.

1. On the Observed Data

The observation of microearthquakes in the northern part of the Kinki and the western part of the Chugoku areas was started on July 1963 at Mikazuki station, Hyogo Pref. by the Tottori Microearthquake Observatory attached to the Disaster Prevention Research Institute of Kyoto University. After that, new

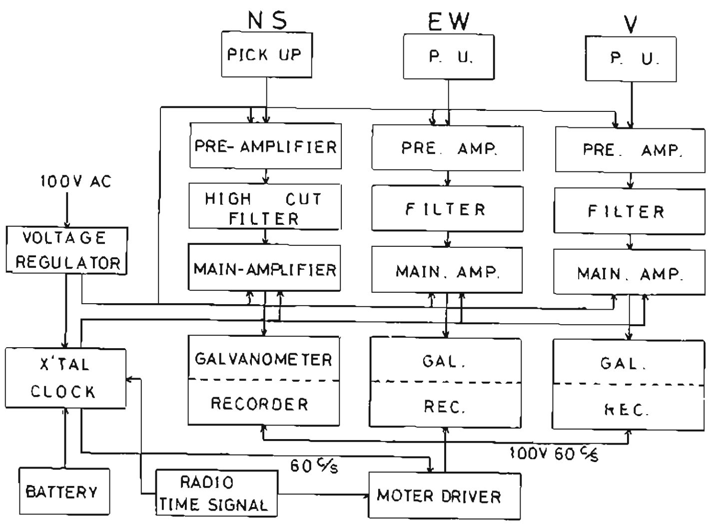

Fig. 1 Outline of observation system.

Table 1 Outline of observation stations attached to the Tottori Microeathquake Observatory.

	MIKAZUKI	FUNAOKA	OYA	HIKAMI	IZUMI
Abbreviation	MZ	FO	OY	HM	IZ
Latitude (E)	134°26′ 40″. 5	134°16′ 02″. 1	134°39′ 52″. 2	135°02′ 36″. 6	134°53′ 15″. 5
Longitude (N)	34°59′ 12″. 0	35°19′ 49″. 9	35°20′ 02″. 3	35°13′ 35″. 5	34°58′ 20″. 0
Height (m)	200	200	260	250	230
Lat. (km)	0. 33	38. 48	38. 86	26. 94	−1. 27
Long. (km)	−0. 10	−16. 26	19. 93	54. 50	40. 35
Components	1V 2H	1V	1V 2H	1V 2H	1V
Sensitivity (μ kine/cm)	400	400	400	400	400

Origin of coordinate : 134°26′ 44″. 6 (E)
34°59′ 01″. 3 (N)

1V means 1 set of vertical component.

2H means 2 sets of horizontal component, NS and EW.

(1968 July present)

Table 2 Outline of instruments.

Instrument				
Seismogram	Natural period (sec)	Voltage* sensitivity (v/kine)	Output** impedance (kΩ)	Internal resistance (kΩ)
Vertical	1. 0	2. 6-3. 1	3-6	1. 0
Horizontal	1. 0	3. 3-3. 6		1. 0
Amplifier & Galvanometer	Direct current amplifier Maximum amplitude : 40mm/p-p Natural frequency : 50c/s Overall sensitivity : about 1cm/mv			

* Voltage sensitivity is expressed as output voltage against velocity displacement of ground.

** Output impedance is adjusted as damping constant $h=0.707$.

(1968 July present)

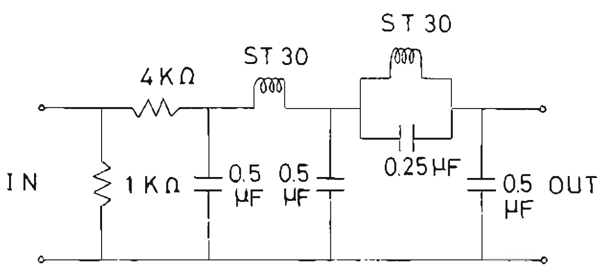

Fig. 2 Circuit of high cut filter.

stations were constructed by and by and the seismological network now consists of five stations. In this paper some descriptions are presented mainly on the data obtained by this network.

The outline of the present network, outline of

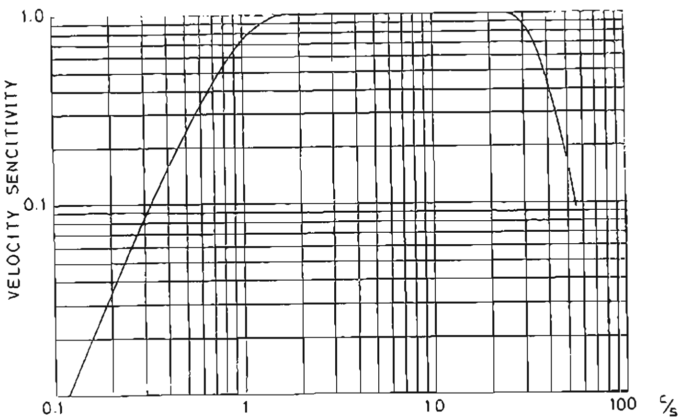

Fig. 3 Overall frequency response curve.

observation instruments and observation system diagram are shown in Table 1, Table 2 and Fig. 1 respectively. The natural frequency of seismometers is 1.0 c/s and it has strong higher mode near 30 c/s. This higher mode vibration stimulated by ground noise or sound disturbs the seismograms. And the high cut filter shown in Fig. 2 is used at the position shown in Fig. 1. The overall frequency response curve is shown in Fig. 3. The characteristics of the seismograms are kept constant during the whole period of observation.

A set of apparatus in an observation system and an example of a seismogram are shown in Photo 1 and Photo 2 respectively. The time mark is recorded every second and it is controlled by X'tal clock, and the X'tal clock is checked every hour by N. H. K (Japan Broadcasting Corporation) time signal.

The terms of seismograms reading are date, P and S time and their accuracy, P initial sense, P-F time, maximum amplitude and its period and some remarks. These terms are punched on a card, and the compilation and some corrections are done by electronic computer. There were some alternations of the observation system, and they are all shown in the volume of data on the same title[1].

There is the Takatsuki branch that belongs to the Tottori Microearthquake Observatory. The outline of the seismological network of the Takatsuki branch is shown in Table 3. The seismograms are all one component (V), and although they are soot writing the time accuracy is not so good because only minute marks are recorded on them. So, without any comment, no data are used in this paper or the succeeding papers.

All the data used were reexamined by the author himself, and the time accuracy is as for the data of P time ± 0.05, and as for S-P time ± 0.1 sec.

Data used in this paper are confined to maximum S-P times less than 18 sec, and they are shown in the volume of data[1]. Before June 1965 the observation system was not so complete, so only the data that were well observed at more than three stations are used.

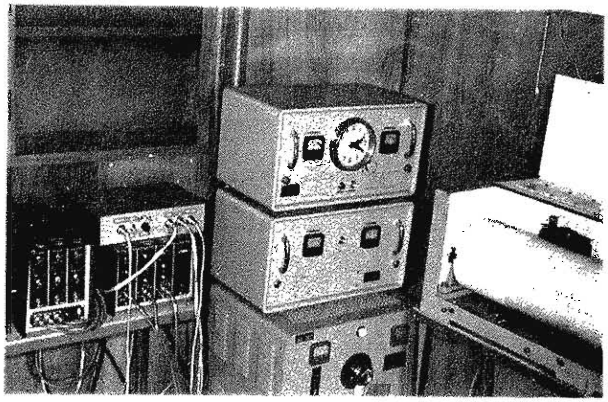

Photo 1 Apparatus of observation system.

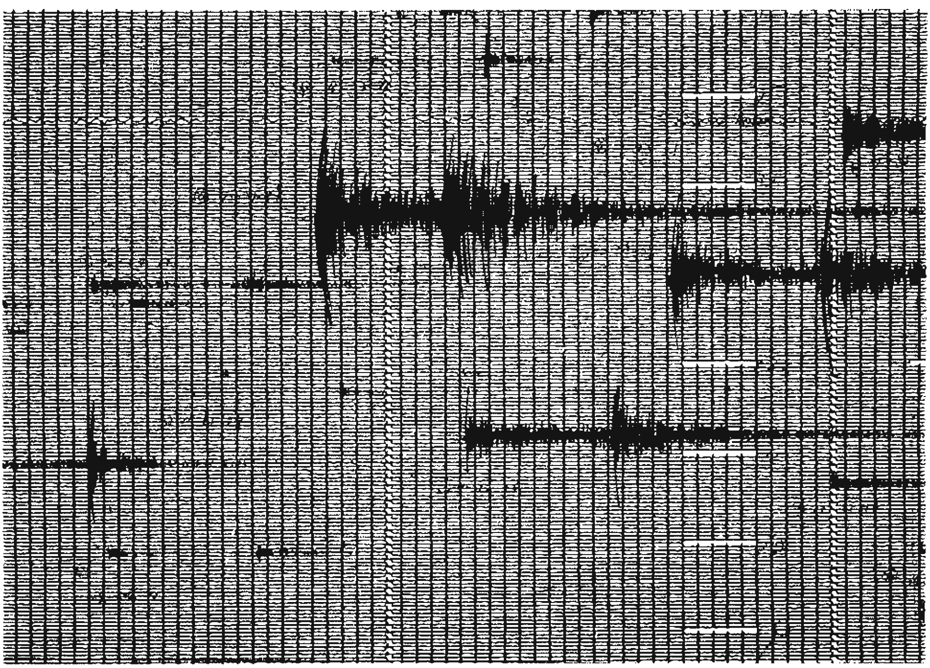

Photo 2 Example of a seismogram.

Table 3 Outline of observation stations attached to the Takatsuki branch of the Tottori Microearthquake Observatory.

	ABUYAMA	MYOKEN	YAGI	KAMI-GAMO	KYOHOKU	TANTO
Abbreviation	AB	MK	YG	KG	KH	TT
Latitude (E)	135°34′22″.4	135°28′18″.0	135°30′46″.0	135°45′56″.3	135°39′43″	134°58′38″.1
Longitude(N)	34°51′34″.0	34°55′32″.0	35°04′04″.0	35°03′32″.0	35°10′37″	35°30′23″.5
Lat. (km)	103.00	93.71	97.38	120.46	110.92	48.38
Long. (km)	−13.78	−5.76	9.33	8.34	21.44	58.00

	TANNAN	KOBE	TSUNA	IKEHARA	ROKKO	IWAO
Abbreviation	TN	KB	TU	IK	RK	IW
Latitude (E)	135°12′50″.0	135°10′34″.0	134°53′05″.6	135°59′05″.7	135°10′31″.2	136°07′56″.0
Longitude (N)	35°01′55″.8	34°44′05″.4	34°25′04″.8	34°02′00″.8	34°44′05″.4	34°52′57″.0
Lat. (km)	70.12	66.79	40.24	141.35	66.73	154.09
Long. (km)	5.38	−27.61	−62.76	−105.40	−27.61	−11.23

(1968 July present)

2. On the Method of Hypocenter Determination

The program for hypocenter determination is as follows :—

The principle is to search for the best fitted point in the plain of a varying depth as for the observed P times or P and S times and theoretical traveltime calculated from a known crustal structure.

In this paper the hypocenter domain is limited within ±250 km on the coordinate of Table 1. Depth is determined at the interval of 1 km up to 15 km and then 2 km up to 37 km.

As for the accuracy or some other discussions another paper is prepared[2].

Some notations are described for the following discussions :—

Suffix j : station identification
P^0 time : origin time
T time : traveltime
Z : depth
X^{st}, Y^{st} : coordinate of station
P time : onset time of initial P wave
S-P time : S-P duration time
T^c time : calculated travel time

(1) *In a case where more than three P times and more than one S time are available* :—

a) To determine origin time.

Assuming Poisson's ratio $\sigma=0.2402$ and using P_j and $(S\text{-}P)_j$ time we put,

$$P^0_j=P_j-1.408\times(S\text{-}P)_j. \tag{2.1}$$

We assume the mean value of P^0_j to be origin time P^0. Then the travel time is reduced as,

$$T_j=P_j-P^0 \tag{2.2}$$

If even only one datum of S-P time can be available the traveltime is determined by this method.

b) To determine initial value of X and Y.

Setting mesh on the surface of $Z=0$ of which the grid spacing is 12.5 km, the best fitted point to hypocenter is searched for on the cross point of the mesh by stage e) as the initial value of hypocenter determination. In this procedure the mesh can be rough in the area where the deviation is greater than a certain value.

c) To search for the epicenter.

We set nine new points X', Y' as,

$$\begin{cases} X'=X\pm d' & \text{or } X'=X \\ Y'=Y\pm d' & \text{or } Y'=Y \end{cases} \qquad (2.3)$$

$$d'=d/1.8$$

where X and Y is the initial value of the epicenter and d the grid spacing. The best fitted point to the epicenter is chosen by stage e) and it is newly set as the initial value X and Y. This procedure is repeated until the grid spacing constant d is 200 m.

d) To search for the depth.

Stage c) is repeated on the plane $Z=H$. H is varied by 1 km up to 15 km of depth, and more than 15 km of depth by 2 km, but if the deviation calculated by stage e) is greater than a certain value some steps can be jumped. For $Z=0$ the epicenter is determined by stage b) and stage c). For $Z>0$ stage b) is deleted and the calculated most probable point of the just upper plane is set as the initial value of X and Y. The point of the least deviation δ is searched as the most probable hypocenter.

e) To calculate the deviation.

We calculate the traveltime T_j^c for the distance D_j between stations X_j^{st}, Y_j^{st} and X, Y by stage f) at the depth H, and put as :—

$$\delta=(\sum(T_j-T_j^c)^2)^{1/2}. \qquad (2.4)$$

where δ is defined to be the deviation at X, Y and Z.

f) To calculate the traveltime.

Traveltime curves are calculated on the base of the crustal structure shown in Fig. 4. The structure is a slightly modified structure of Model II researched in this district by explosion seismology[3]. These traveltime curves are divided minutely and memorized. Traveltime is calculated by the interpolation of distance. The time accuracy is within 1/100 sec.

(2) *In a case where S-P time is not available and more than four P times are available.*

Calculating T_j^c by stage f) for

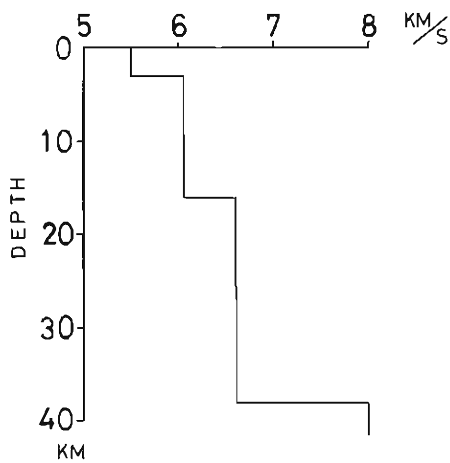

Fig. 4 Crustal structure used to calculate traveltime.

X, Y and Z, we put the origin time P^0 as :—

$$P^0=\sum(P_j-T_j^c)/\sum j \qquad (2.5)$$

Then we repeat the above procedure in the same manner as the case using S-P time. The least deviated point and origin time are searched as the most probable hypocenter and origin time.

In a case where S-P time is used and crustal structure is known, the equation to determine the epicenter at the depth H is :—

$$(T_jV_j)^2=(X_j^{st}-X)^2+(Y_j^{st}-Y)^2 \qquad (2.6)$$

where V_j is the mean apparent velocity from the assumed epicenter to the observation station. In the above method the determination of the most probable point at the depth H is to determine the minimun value of δ^2 that is :—

$$\delta^2=\sum(T_j-T_j^c)^2/\sum j=\sum\Big(T_j-\frac{1}{V_j}\sqrt{(X_j^{st}-X)^2+(Y_j^{st}-Y)^2}\Big)^2/\sum j \qquad (2.7)$$

$$\frac{\partial\delta^2}{\partial x}=\frac{\partial\delta^2}{\partial y}=0 \qquad (2.8)$$

In other words δ is the standard deviation of the observed data to the determined hypocenter.

Mathematically the minimum point of this equation is very complicated, and we have some possibility of mistaking another minimum point to be the most probable point. In this sense stage b) is important. If the mesh to give the initial value is rough, then in some cases a very large value of deviation δ and mistaken hypocenter is determined. The length of the grid spacing may depend on the seismic network size and the number of stations. In this case 12.5 km is given.

For the hypocenter that has a value of δ more than 0.10 sec, all data was checked by the graphical method or by varying the initial value slightly, and it was confirmed so as not to mistake epicenter more than 0.3 km or so.

In the case of using P time only it is more complicated, but in our case almost all the earthquake that can not be read with the S-P time are rather large earthquakes which are reported in the bulletin of the J. M. A. (Japan Meteorological Agency) and the results could be compared to them.

As was mentioned already, the time accuracy of data for P time is within ± 0.05 sec, but some unknown errors are left such as the origin time, crustal structure, regional travel time anomalies, so that more minute discussion than deviation less than 0.15 sec may be nonsense.

There are some data as to local traveltime anomalies by explosion seismology in Table 4. All the onset times of P wave are clearly read within ± 0.01 sec. From those results about ± 0.1 sec of traveltime correction is needed if confined only to those data, but since it does not affect the result so much, we have left it for further reseach work.

The inaccuracy of S-P time affects the origin time. In some cases residual ΔP_j^0 from mean origin time, that is:—

$$\Delta P_j^0=T_j-1.408\times(S-P)_j \qquad (2.9)$$

exceeds ± 0.3 sec, but it is elucidated that it does not affect the hypocenter determination so much[2].

The list of standard deviation δ and its number of earthquakes N, the number of earthquakes observed at each station and the number of stations used in

Table 4 Data for local traveltime anomalies by explosion seismology.

1st and 2nd Kurayosi Explosion[4)]

	Δ	P-O-Δ/6
FO	39.85	0.52
MZ	71.36	0.33
OY	75.28	0.53

1st Hanabusa Explosion[4)]

	Δ	P-O-Δ/6
HM	143.00	0.35

Tomieda quarry blast
1967 Nov. 30 06h 30m
Shot point : 134°22′03″.2 ($Y=41.687$)
35°21′43″.1 ($X=-7.122$)

	Δ	P-Δ/6
FO	9.69	32.38
OY	27.10	32.34
MZ	41.95	32.61
HM	63.36	32.58
IZ	64.02	32.55

Table 5

a) Number of earthquakes of which hypocenters are determined within a cirtain value of standard deviation δ.

$\delta<0.05$	$0.05\leq\delta<0.10$	$0.10\leq\delta<0.15$	$0.15\leq\delta$
941	438	132	19

Total 1530

(Among them the number of earthquakes determined by P time only is 8)

b) Number of earthquakes observed at each station.

MZ	FO	OY	HM	IZ
1346	722	1309	1260	1384

c) Number of stations used in determining hypocenter for each earthquake and its number of earthquakes.

Station	3	4	5
Earthquake	503	623	404

determining the hypocenter for each earthquake are in Table 5.

The most difficult point to level up in the accuracy of hypocenter determination is how to keep the accuracy of obsevation and reading seismograms of each station of each earthquake.

3. On the Determination of Magnitude

The determination of magnitude is after Muramatsu's formula for microearthquake[5)],

$$M=1.25\ \log(V)+2.50\ \log(R)-6.50 \tag{3.1}$$

where V is the maximum amplitude in μ kine, in our case the maximum amplitude of vertical component is used and R the epicentral distance.

There is some questions of applying this equation and the corresponding energy and magnitude relation,

$$E = 11.8 + 1.5M \tag{3.2}$$

to our network.

Then as the first step in applying equation (3.1), we calculate the residual ΔM_j, reduced magnitude determined by each station M_j from its mean value M, that is :—

$$\Delta M_j = M_j - M \tag{3.3}$$

In Table 6 the mean value of ΔM and mean epicentral distance R is shown for

Table 6. The mean value of residual magnitude ΔM and mean epicentral distance R for each station. N is the number of earthquakes used in this calculation.

Station	MZ	FO	OY	HM	IZ
ΔM	0.04	−0.04	−0.02	−0.10	0.09
R (km)	76.4	94.7	79.2	63.0	58.2
N	678	333	674	631	734

each station. N is the number of earthquakes used in this calculation. From this result it needs about 20% of amplitude correction in linear scale for our network.

As for FO the mean epicentral distance is not so great as that of the stations MZ or OY and the amplitude correction is the same as the others, but the number of data used is fairly small. The reason why is because there were often no observation periods due to some accidents. Then there is a possibility that the seismicity level is underestimated around station FO.

If ΔM_j is due to second coefficient of the right side of equation (3.1), it must be changed as follows by the least square curve fitting.

$$M = 1.25 \log(V) + 2.90 \log(R) - \text{const.} \tag{3.4}$$

But some consideration must be paid to the calibration of the instruments, deviation of seismicity from each station, focal mechanism, and so on.

As we have not so much data, then at the first step we do not use equation (3.4) but use (3.1) because the difference between the two equations does not seem to have a decisive effect on our discussion.

By the characteristics of the galvanometer, the amplitude is saturated more than 2.0 cm (about 800 μ kine), so some device is needed to estimate the real value of the scaleout amplitude. Tsumura found a certain relation between P-F time (from onset time of initial P time to Coda time) and the amplitude for microearthquakes[6]. But we find this relation does not hold in our case.

We find the relation shown in Fig. 5, of which the abscissa is the time interval in seconds from the onset time of S inferred from the hypocentral distance determined by the above method, to the time when the amplitude becomes less than 2.0cm, and the ordinate is the amplitude in m kine after equation (3.1) using the magnitude determined by the J. M. A. (open circle), or determined only by unsaturated seismograms (solid circle). In Table 6 the data thus determined is not used.

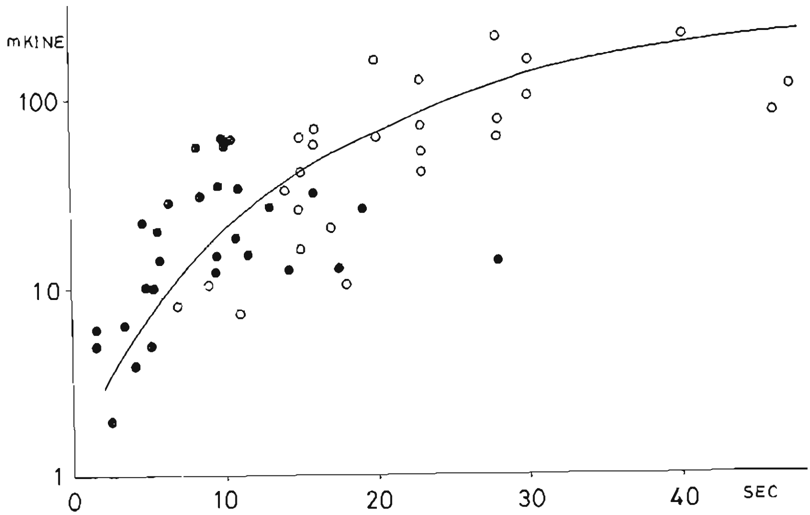

Fig. 5 Curve for estimating the amplitude of a saturated seismogram. Abscissa is the time interval in seconds from the onset time of S inferred from the hypocentral distance to the time when the amplitude becomes less than 2.0 cm, and the ordinate is the amplitude in m kine after equation (3-1) using the magnitude determined by the J. M. A. (open circle), or determined only by unsaturated seismograms (solid circle).

In Fig. 6 the magnitude (M)—accumlated frequency (N) relation is shown. At a glance on this figure we find Gutenberg—Richter's relation,

$$\log n(M) = a - bM \qquad (3.6)$$

which does not hold in this district. If we manage to fit straight lines, their coeficientsf b are culculated by the method of Utsu[7],

$1.2 < M < 2.2$; $b = 0.89$

$2.2 \leq M$; $b = 0.52$

and for M less than 1.2 hypocenter could not be determined uniformly.

Seismograms begin to saturate more than $M = 2.0$, for M more than 4.0 magnitude is determined by the J. M. A. And the magnitude between 2.2 and 4.0 the amplitude is estimated by Fig. 5. The value $b = 0.89$ is the same as the value in this district[8].

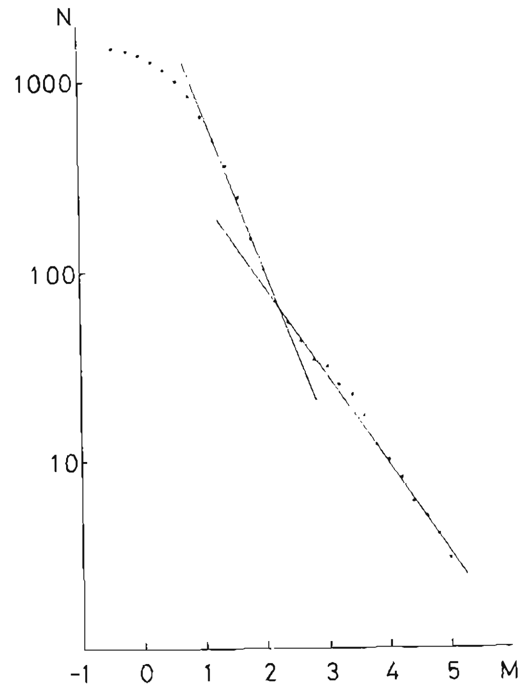

Fig. 6 Magnitude (M) and accumulated frequancy(N) relation.

If in equation (3.1) the constant term is less than 6.5 by 1.0, then Fig. 6 is little changed and our curve may be one straight line connecting the microearthquake to the large earthquake. We have left this problem for further research.

In the volume of data[1] all the earthquake from Aug. 1964 to June 1968 is listed for date, origin time, epicenter X, and Y focal depth Z and its probable domain, standard deviation, magnitude and station observed.

4. On Seismicity

The seismicity map of the northwestern part of the Kinki and the eastern part of the Chugoku districts is shown in Fig. 7. In this figure the end of shore line is nearly the limited boundary on determining the hypocenter in this paper. Fig. 8 a, b and c is the seismicity map, from Aug. 1964 to June 1966, from July 1966 to June 1967 and from July 1967 to June 1968 respectively. As was mentioned already before the June 1965 observation system was not complete so the seismicity is much underestimated.

Fig. 9 a, b and c is the seismicity map of magnitude M as $1.0 \leq M$, $2.0 \leq M$ and $3.0 \leq M$ respectively.

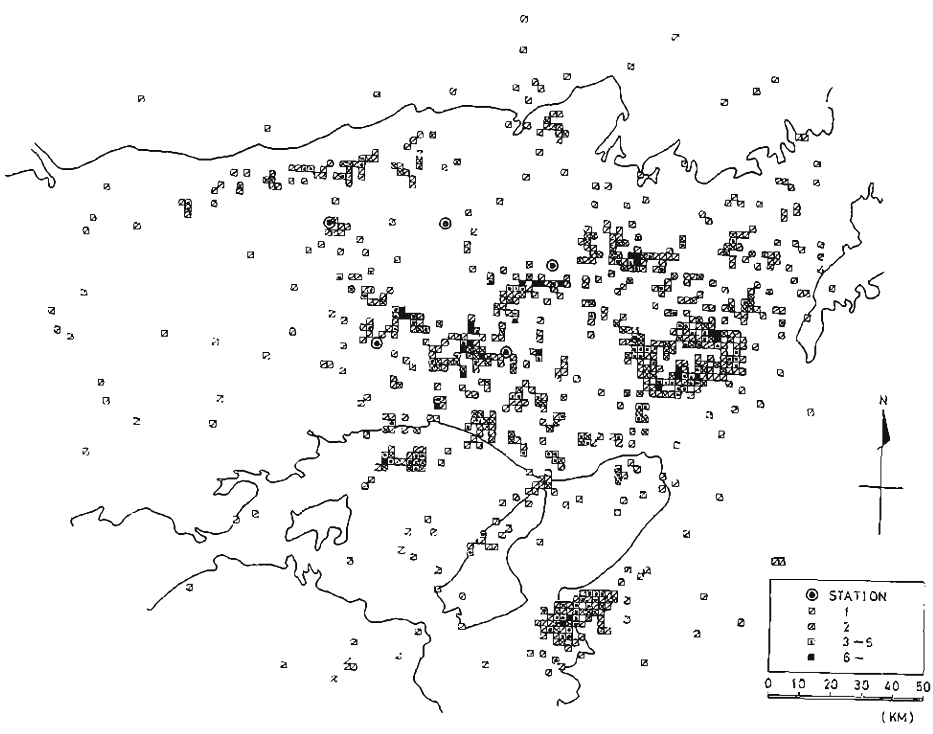

Fig. 7 Seismicity map of the northwestern part of the Kinki and the eastern part of the Chugoku districts. The end of shore line is nearly the limited boundary on determining the hypocenter.

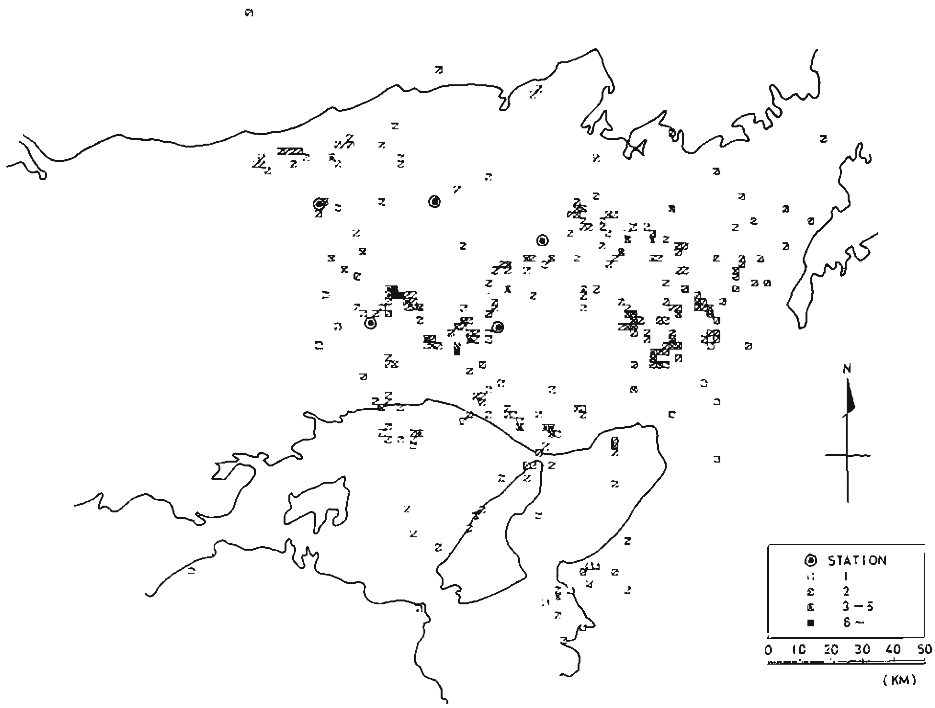

Fig. 8 (a) Seismicity map from Aup 1964 to June 1966. In this period seismicity is under estimated.

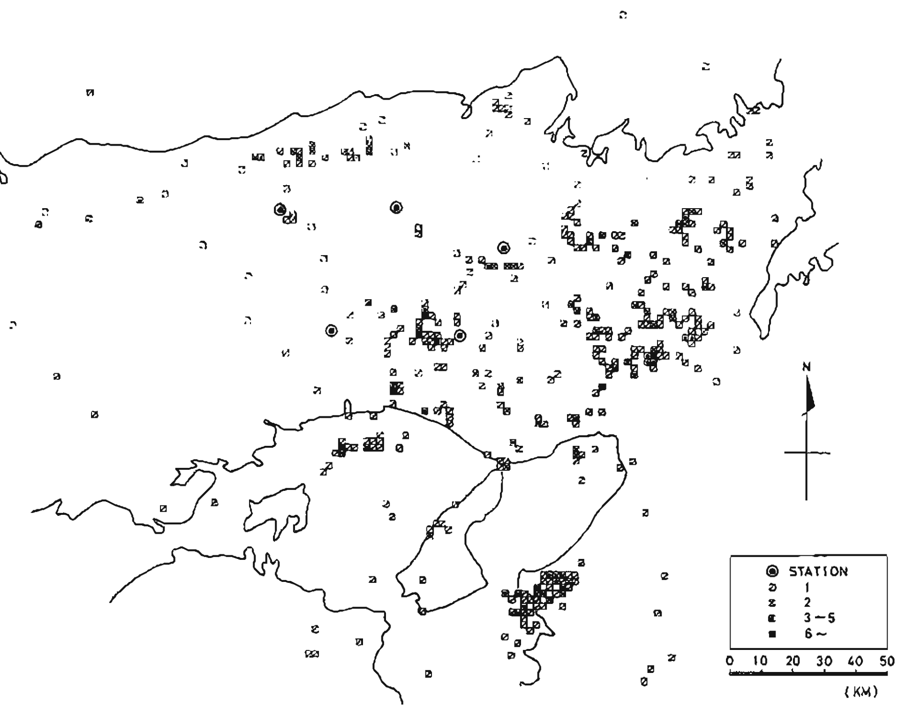

Fig. 8 (b) Seismicity map from July 1966 to June 1967.

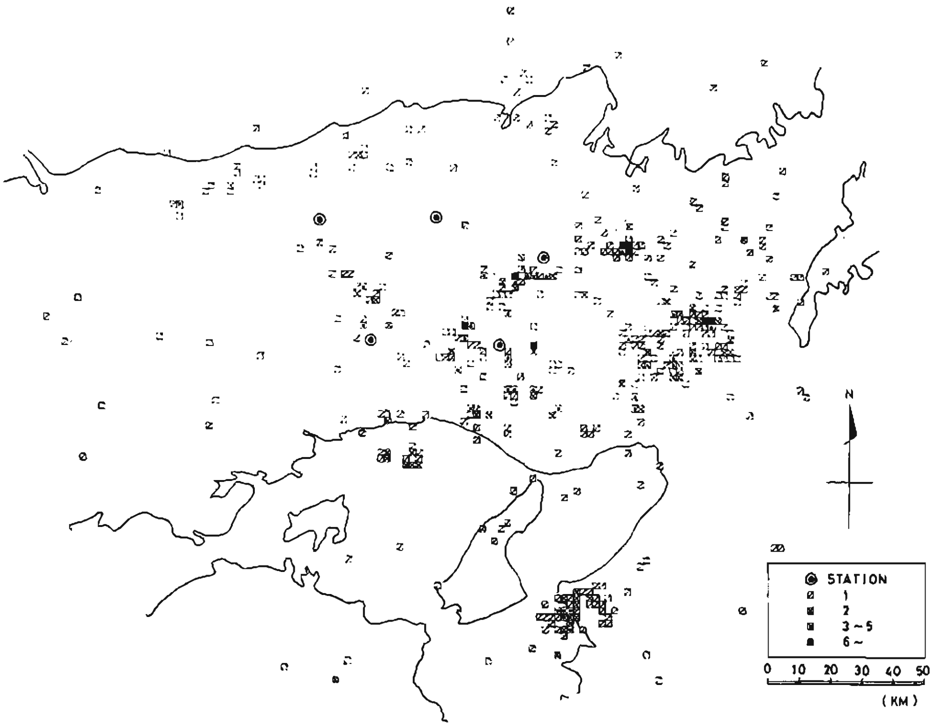

Fig. 8 (c) Seismicity map from July 1967 to June 1968.

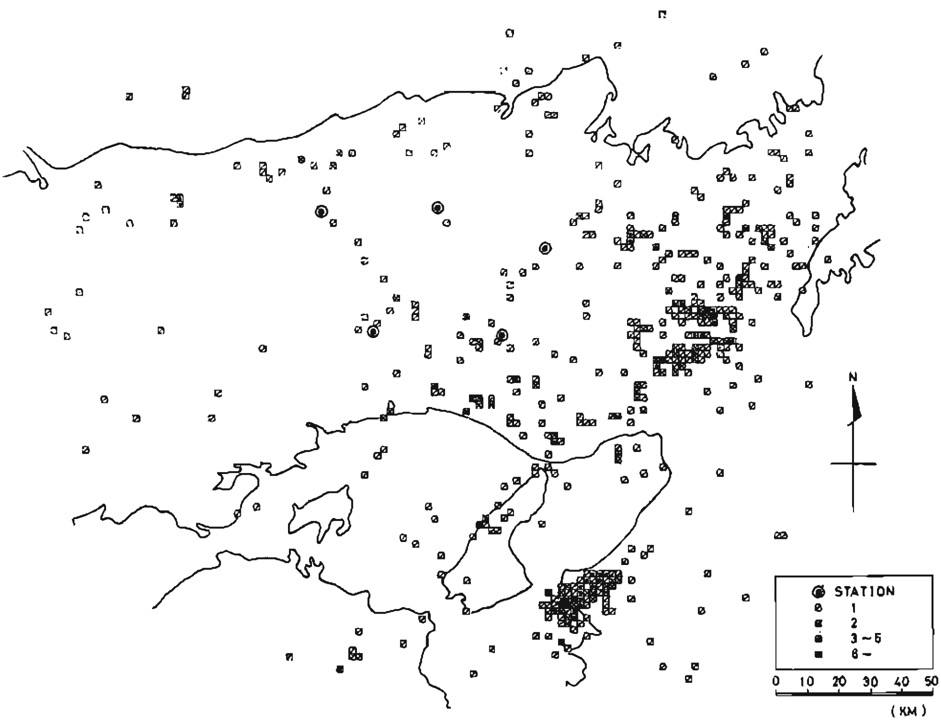

Fig. 9 (a) Seismicity map of magnitude M as $1.0 \leq M$.

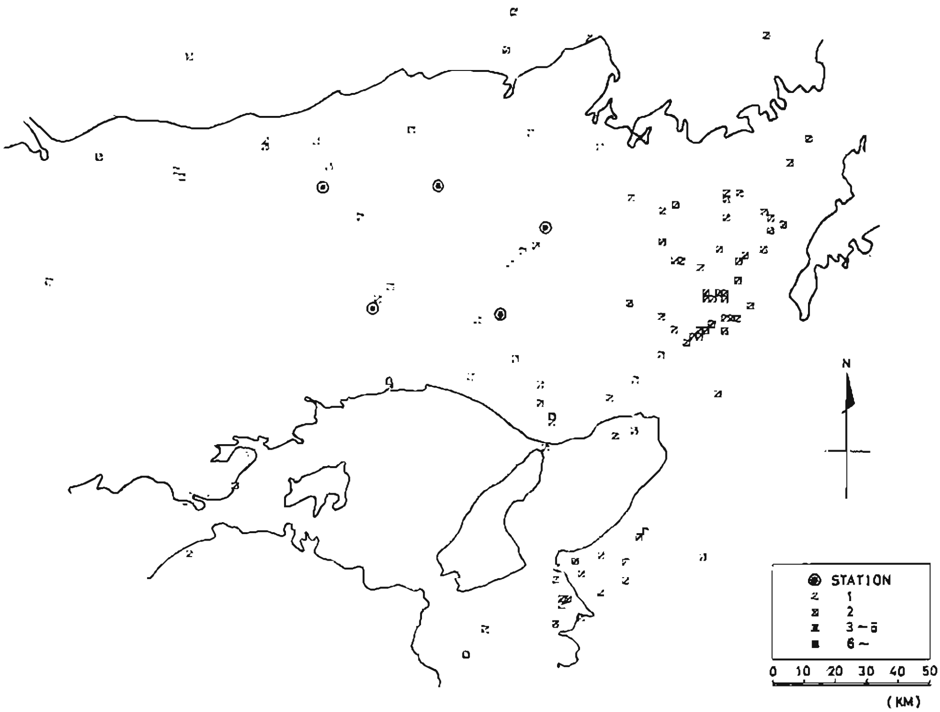

Fig. 9 (b) Seismicity map of magnitude M as $2.0 \leq M$.

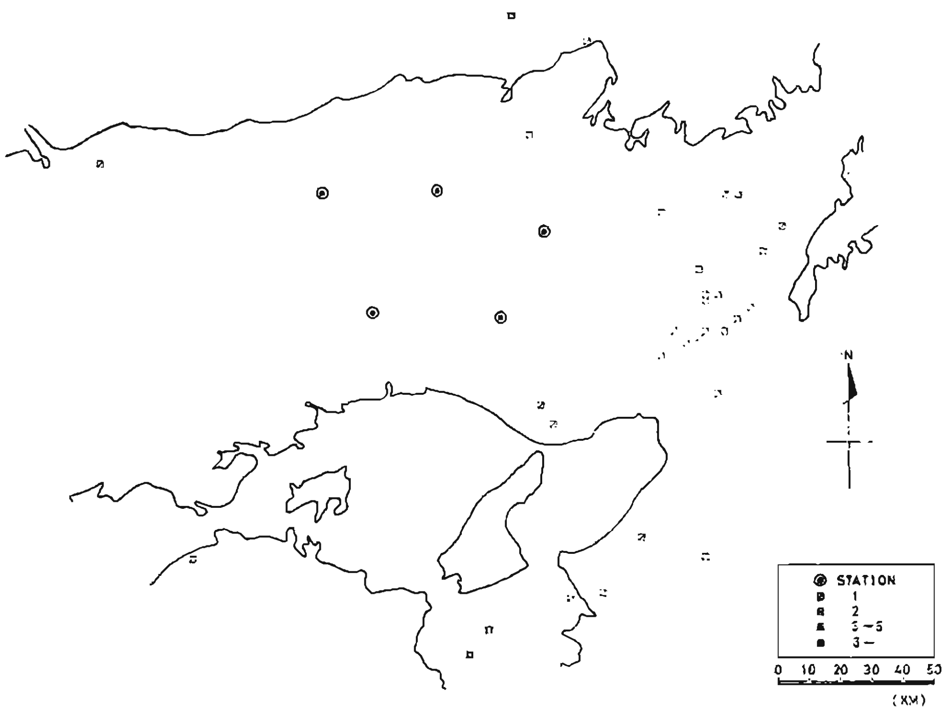

Fig. 9 (c) Seismicity map of magnitude M as $3.0 \leq M$.

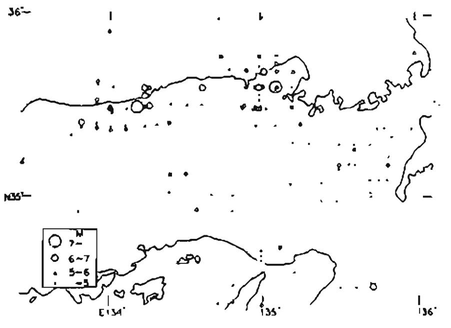

Fig. 10 (a)
Distribution of epicenters determined by J. M. A. from 1926 to 1954.

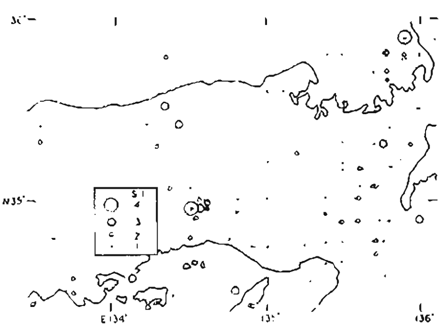

Fig. 10 (b)
Distribution of epicenters determined by J. M. A. from 1955 to 1964.

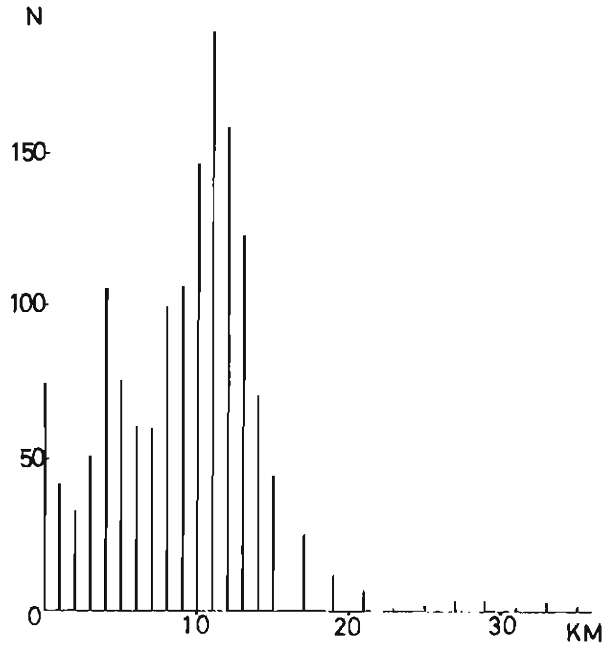

Fig. 11 Histogram of frequency (N)-depth.

For reference the seismicity map of rather large earthquakes by the J. M. A. is shown in Fig. 10 a and b. The coincidence between the microearthquake seismicity map and the rather large earthquake seismicity map is not necessarily good.

The histogram of frequency-depth is shown in Fig. 11.

Fig. 12 a, b, c, d, e and f is the seismicity map of the focal depth d(km) as $0<d\leqq 5$, $5<d\leqq 10$, $10<d\leqq 15$, $15<d\leqq 20$, $20<d\leqq 30$, and $30<d\leqq 40$ respectively.

The details of seismicity and quarternary geotectonics and related problems will be discussed in succeeding papers[9].

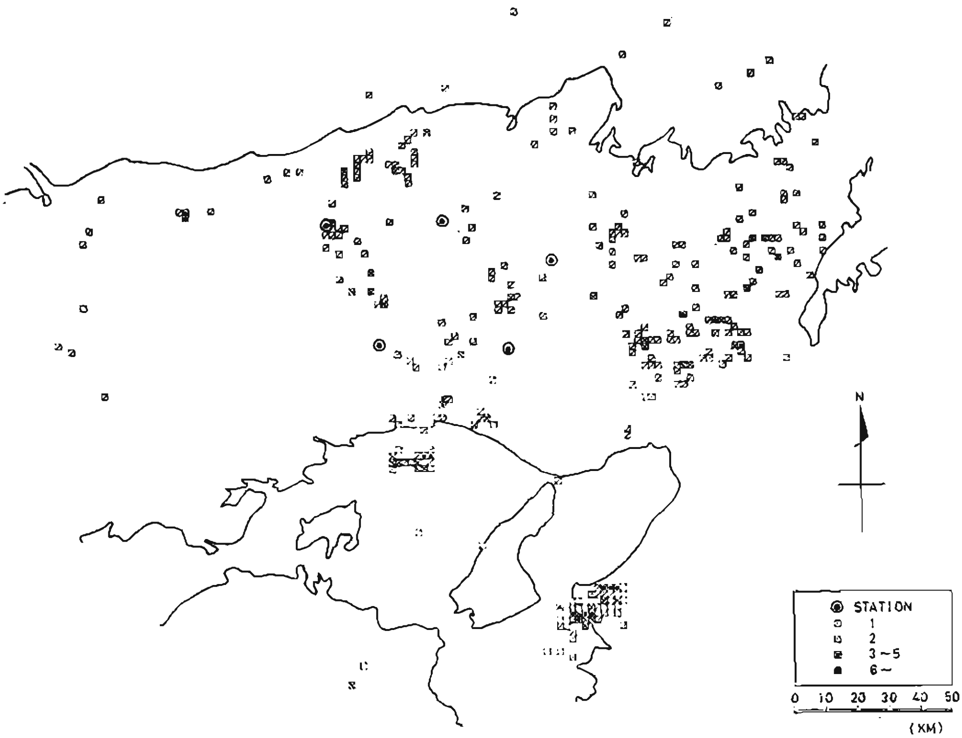

Fig. 12 (a) Seismicity map of the focal depth d as $0<d\leq5$.

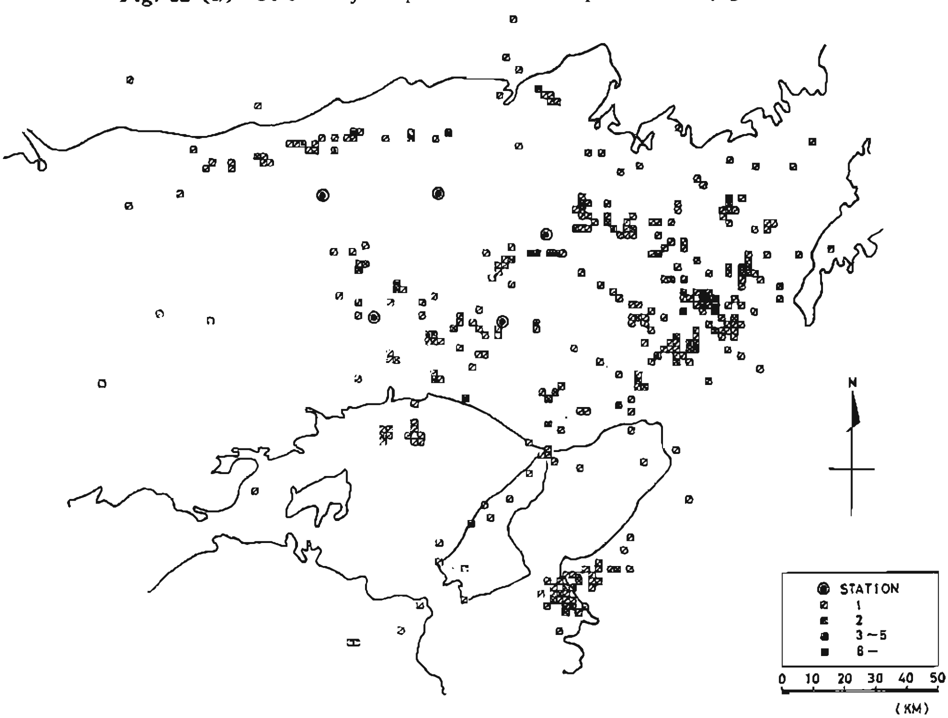

Fig. 12 (b) Seismicity map of the focal depth d as $5<d\leq10$.

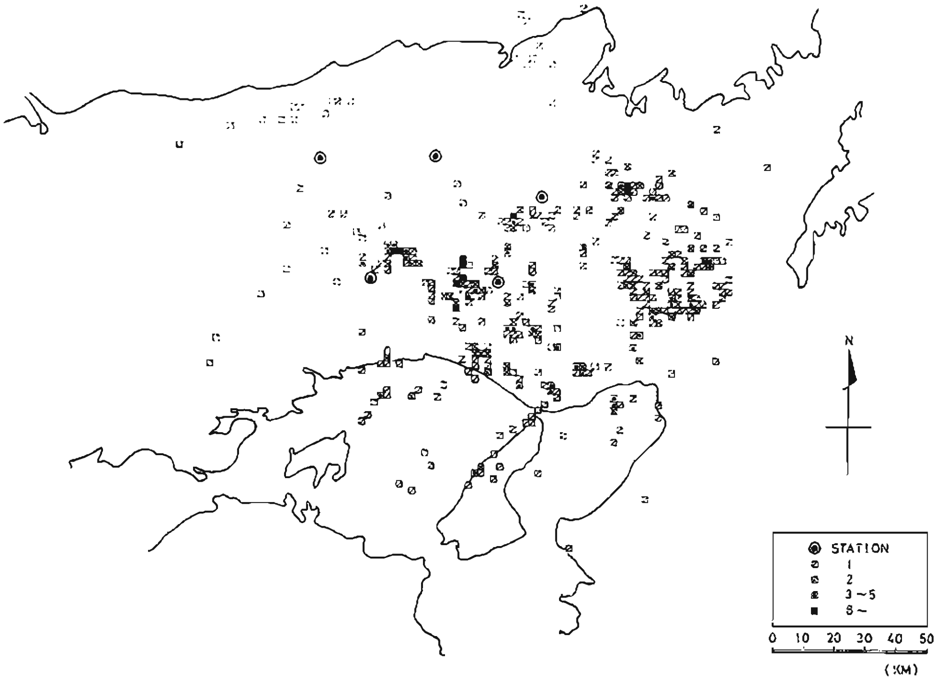

Fig. 12 (c) Seismicity map of the focal depth d as $10<d\leqq15$.

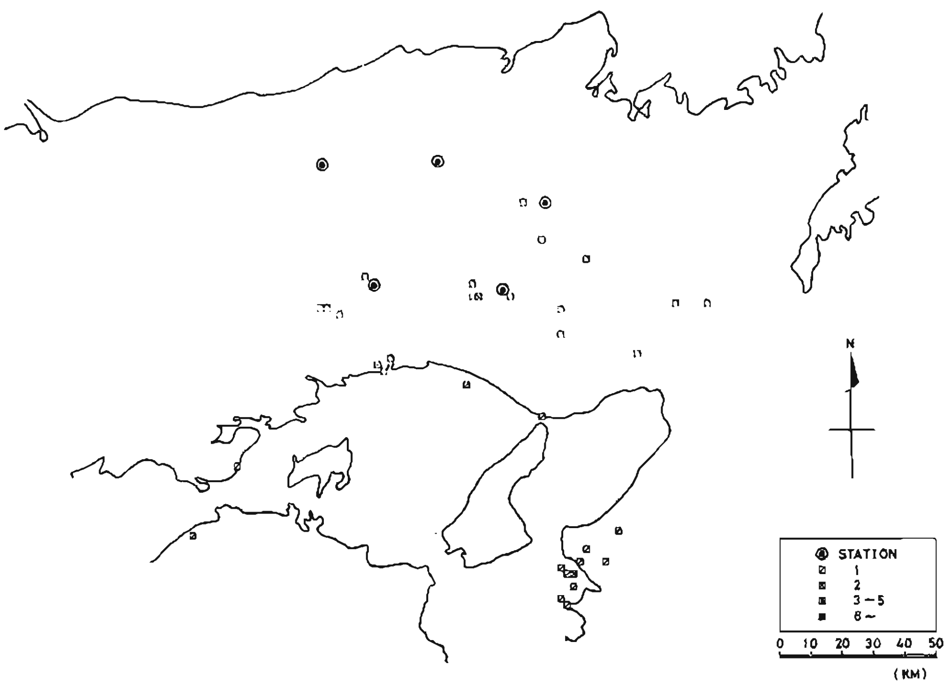

Fig. 12 (d) Seismicity map of the focal depth d as $15<d\leqq20$.

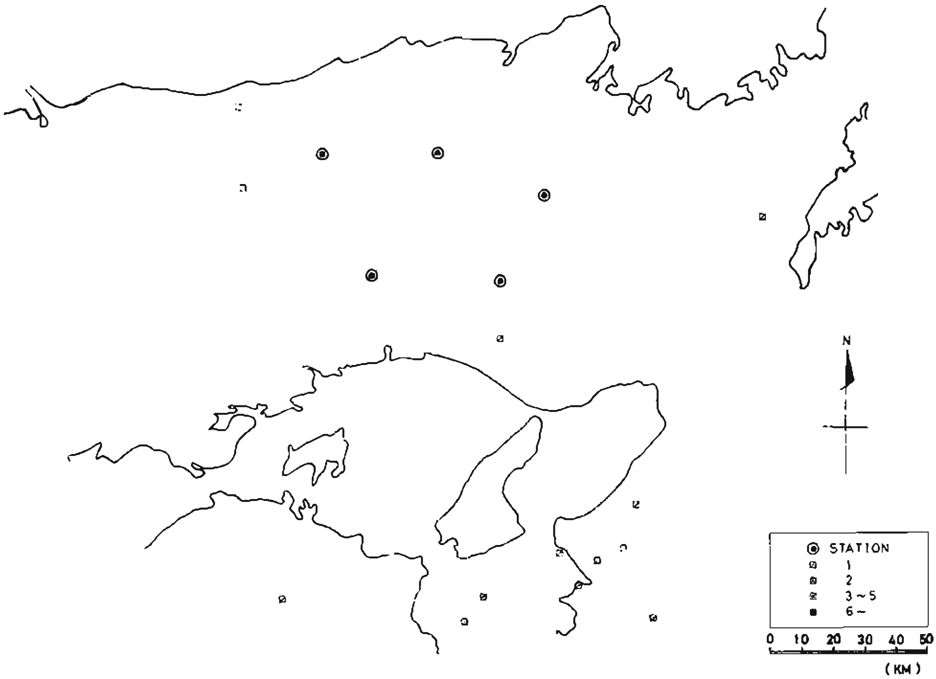

Fig. 12 (e) Seismicity map of the focal depth d as $20<d\leq 30$.

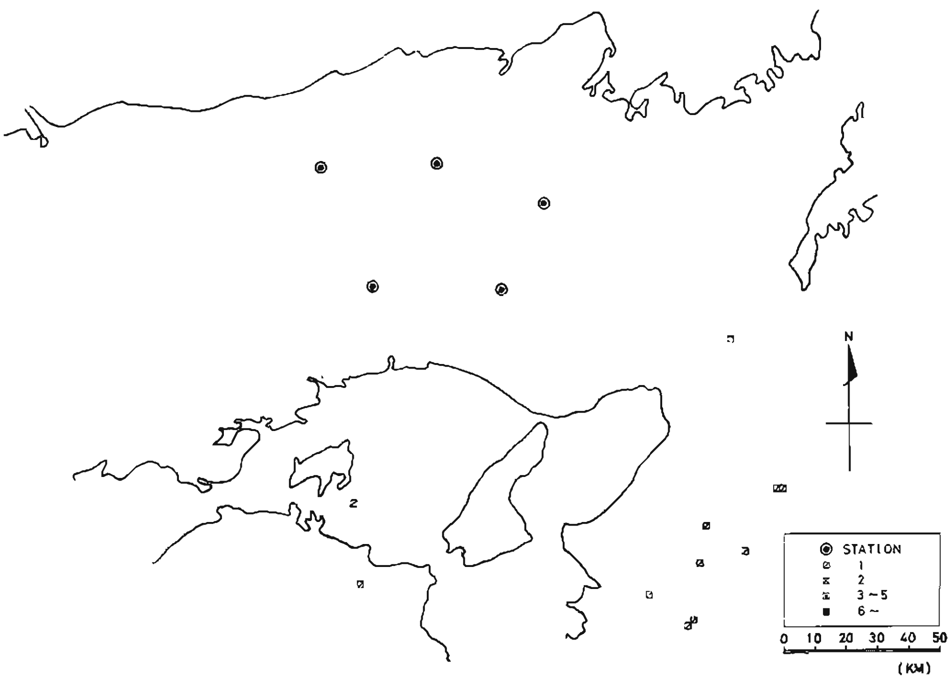

Fig. 12 (f) Seismicity map of the focal depth d as $30<d\leq 40$.

Acknowledgements

Most of the data in this study was provided by the Tottori Microearthquake Observatory attached to the Disaster Prevention Research Institute of Kyoto University.

The author wishes to express his sincere thanks to the following members :—

Director	Dr. Y. KISHIMOTO
Seismologists	Mr. K. MINO
	Mr. K. OIKE
	Mr. R. NISHIDA
Technicians	Mr. S. NAKAO
	Mr. Y. TSUSHIMA
	Mr. S. MATSUO
Seismogram	Mrs. M. KOGA
interpreters	Mr. S. YABE
Station	Mr. M. YABE
keepers	Mr. R. OGURA
	Mr. T. SHIKATA
	Mr. G. HOSOMI
	Mr. Y. UENAKA

Financial aid was granted by the fund for special work of this Institute.

References

1) Hashizume, M. : Investigation of Microearthquakes - Data - (in press).
2) Hashizume, M. : Investigation of Microearthquakes -On the Accuracy of Hypocenter Determination -(in press).
3) Hashizume, M. et al, : Crustal Structure in the Western Part of Japan Derived from the Observation of the First and Second Kurayosi and the Hanabusa Explosions, Part 2, Crustal Structure in the Western Part of Japan, Bull. Earthq. Res. Inst., Vol. 44. 1966. pp. 109-120.
4) The Research Group for Explosion Seismology : Crustal Structure in the Western Part of Japan Derived from the Observation of the First and Second Kurayosi and the Hanabusa Explosions, Part 1, Observation of Seismic Waves Generated by the First and Second Kurayosi and the Hanabusa Explosions, Bull. Earthq. Res. Inst., Vol. 44. 1966. pp. 89-107.
5) Muramatsu. I. et al : Observation of Microearthquakes in Mino District in Gifu Prefecture, Central Japan, Jour. Phys. Earth, Vol. 11, No. 2, 1963. pp. 35-48.
6) Tsumura, K. : Determination of Earthquake Magnitude from Total Duration of Oscillation, Bull. Earthq. Res. Inst, Vol. 45. 1967. pp. 7-18.
7) Utsu, T. : A Method for Determining the Value of b in a Formula $\log n=a-bM$ Showing the Magnitude-Frequency Relation for Earthquakes, Geophys. Bull. Hokkaido Univ., Vol. 13, 1965, pp. 99-103 (in Japanese).
8) Hamamatsu, O. : Seismicity of Shallow Earthquakes in and near Japan During 1926 to 1956, Quatr. Jour. Seisms., 25, 1960, pp. 97-108 (in Japanese).
9) Hashizume, M. : In preparation with the same title.