Investigation of Microearthquakes
—On Seismicity—

By Michio HASHIZUME

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

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>MIKAZUKI</th>
<th>FUNAOKA</th>
<th>OYA</th>
<th>HIKAMI</th>
<th>IZUMI</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude (E)</td>
<td>134°26'40&quot;.5</td>
<td>134°16'02&quot;.1</td>
<td>134°39'52&quot;.2</td>
<td>135°02'36&quot;.6</td>
<td>134°53'15&quot;.5</td>
</tr>
<tr>
<td>Longitude (N)</td>
<td>34°59'12&quot;.0</td>
<td>35°19'49&quot;.9</td>
<td>35°20'02&quot;.3</td>
<td>35°13'35&quot;.5</td>
<td>34°58'20&quot;.0</td>
</tr>
<tr>
<td>Height (m)</td>
<td>200</td>
<td>200</td>
<td>260</td>
<td>250</td>
<td>230</td>
</tr>
<tr>
<td>Lat. (km)</td>
<td>0.33</td>
<td>38.48</td>
<td>38.86</td>
<td>26.94</td>
<td>-1.27</td>
</tr>
<tr>
<td>Long. (km)</td>
<td>-0.10</td>
<td>-16.26</td>
<td>19.93</td>
<td>54.50</td>
<td>40.35</td>
</tr>
<tr>
<td>Components</td>
<td>1V 2H</td>
<td>1V 1V 2H</td>
<td>1V 2H</td>
<td>1V</td>
<td></td>
</tr>
<tr>
<td>Sensitivity</td>
<td>400</td>
<td>400</td>
<td>400</td>
<td>400</td>
<td>400</td>
</tr>
</tbody>
</table>

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.

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Seismogram</th>
<th>Natural period (sec)</th>
<th>Voltage sensitivity (v/kine)</th>
<th>Output impedance (kΩ)</th>
<th>Internal resistance (kΩ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vertical</td>
<td>1.0</td>
<td>2.6-3.1</td>
<td></td>
<td>3-6</td>
<td>1.0</td>
</tr>
<tr>
<td>Horizontal</td>
<td>1.0</td>
<td>3.3-3.6</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Amplifier & Galvanometer
- 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)

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
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.  

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. 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.

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>ABUYAMA</th>
<th>MYOKEN</th>
<th>YAGI</th>
<th>KAMI-GAMO</th>
<th>KYOHOOKU</th>
<th>TANTO</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude (E)</td>
<td>135°34'22&quot;</td>
<td>135°28'18&quot;</td>
<td>135°30'46&quot;</td>
<td>135°45'56&quot;</td>
<td>135°39'43&quot;</td>
<td>134°58'38&quot;</td>
</tr>
<tr>
<td>Longitude (N)</td>
<td>34°55'32&quot;</td>
<td>35°04'04&quot;</td>
<td>35°03'32&quot;</td>
<td>35°10'37&quot;</td>
<td>35°30'23&quot;</td>
<td></td>
</tr>
<tr>
<td>Lat. (km)</td>
<td>103.00</td>
<td>93.71</td>
<td>97.38</td>
<td>120.46</td>
<td>110.92</td>
<td>48.38</td>
</tr>
<tr>
<td>Long. (km)</td>
<td>-13.78</td>
<td>-5.76</td>
<td>9.33</td>
<td>8.34</td>
<td>21.44</td>
<td>58.00</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>TANNAN</th>
<th>KOBE</th>
<th>TSUNA</th>
<th>IKEHARA</th>
<th>ROKKO</th>
<th>IWAO</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude (E)</td>
<td>135°12'50&quot;</td>
<td>135°10'34&quot;</td>
<td>134°53'05&quot;</td>
<td>135°59'05&quot;</td>
<td>136°07'56&quot;</td>
<td></td>
</tr>
<tr>
<td>Longitude (N)</td>
<td>35°01'55&quot;</td>
<td>34°44'05&quot;</td>
<td>34°25'04&quot;</td>
<td>34°02'00&quot;</td>
<td>34°44'05&quot;</td>
<td></td>
</tr>
<tr>
<td>Lat. (km)</td>
<td>70.12</td>
<td>66.79</td>
<td>40.24</td>
<td>141.35</td>
<td>66.73</td>
<td>154.09</td>
</tr>
<tr>
<td>Long. (km)</td>
<td>5.38</td>
<td>-27.61</td>
<td>-62.76</td>
<td>-105.40</td>
<td>-27.61</td>
<td>-11.23</td>
</tr>
</tbody>
</table>

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.

Some notations are described for the following discussions:—

Suffix $j$ : station identification
$P^o$ time : origin time
$T$ time : traveltime
$Z$ : depth
$X^u$, $Y^u$ : coordinate of station
$P$ time : onset time of initial $P$ wave
$S-P$ time : $S-P$ duration time
$T^o$ 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 $(S-P)_j$ time we put,

$$P^o_j = P_j - 1.408 \times (S-P)_j$$  \hspace{1cm} (2.1)

We assume the mean value of $P^o_j$ to be origin time $P^o$. Then the travel time is reduced as,

$$T_j = P_j - P^o$$  \hspace{1cm} (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,

\[
\begin{align*}
X' &= X \pm d' \\
Y' &= Y \pm d'
\end{align*}
\]

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 re-
petitioned 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 \(\delta\) is searched as the most probable hypocenter.
e) To calculate the deviation.

We calculate the traveltime \(T_{f+}\) for the distance \(D_i\) between stations \(X'_i\),
\(Y'_i\) and \(X, Y\) by stage f) at the depth \(H\), and put as:

\[\delta = (\sum (T_i - T_{f+}))^2 14.\]  

(2.4)

where \(\delta\) is defined to be the deviation at X, Y and Z.
f) To calculate the traveltime.

The traveltime curves are calculated
on the base of the crustal struc-
ture shown in Fig. 4. The struc-
ture is a slightly modified struc-
ture of Model II researched in
this district by explosion seismo-
logy^3). These traveltime curves
are divided minutely and memo-
rized. 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 \(T_{f+}\) by stage f) for

\[
\begin{array}{c}
5 \\
6 \\
7 \\
8 \text{ KM/S}
\end{array}
\]

\[
\begin{array}{c}
0 \\
10 \\
20 \\
30 \\
40 \text{ KM}
\end{array}
\]

Fig. 4 Crustal structure used to
calculate traveltime.
We put the origin time \( P \) as:

\[
P = \frac{\sum (P_j - T_j)}{\sum j}
\]

(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 the crustal structure is known, the equation to determine the epicenter at the depth \( H \) is:

\[
(T_jV_j)^2 = (X_j^H - X)^2 + (Y_j^H - Y)^2
\]

(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 minimum value of \( \delta^2 \) that is:

\[
\delta^2 = \sum (T_j - T_j^*)^2 / \sum j = \sum (T_j - \frac{1}{V_j} \sqrt{(X_j^H - X)^2 + (Y_j^H - Y)^2})^2 / \sum j
\]

(2.7)

\[
\frac{\partial \delta^2}{\partial x} = \frac{\partial \delta^2}{\partial y} = 0
\]

(2.8)

In other words, \( \delta \) 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 \( \delta \) 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 \( \delta \) more than 0.1 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 earthquakes 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 research work.

The inaccuracy of \( S-P \) time affects the origin time. In some cases residual \( \Delta P_j^o \) from mean origin time, that is:

\[
\Delta P_j^o = T_j - 1.408 \times (S-P)_j
\]

(2.9)

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

The list of standard deviation \( \delta \) 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.

<table>
<thead>
<tr>
<th></th>
<th>( \Delta )</th>
<th>( P-O-\Delta/6 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>FO</td>
<td>39.85</td>
<td>0.52</td>
</tr>
<tr>
<td>MZ</td>
<td>71.36</td>
<td>0.33</td>
</tr>
<tr>
<td>OY</td>
<td>75.28</td>
<td>0.53</td>
</tr>
<tr>
<td><strong>1st Kurayosi Explosion</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>1st Hanabusa Explosion</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>HM</td>
<td>143.00</td>
<td>0.35</td>
</tr>
</tbody>
</table>

Table 5

a) Number of earthquakes of which hypocenters are determined within a certain value of standard deviation \( \delta \).

<table>
<thead>
<tr>
<th>( \delta )</th>
<th>0.05≤( \delta )&lt;0.10</th>
<th>0.10≤( \delta )&lt;0.15</th>
<th>0.15≤( \delta )</th>
</tr>
</thead>
<tbody>
<tr>
<td>941</td>
<td>438</td>
<td>132</td>
<td>19</td>
</tr>
</tbody>
</table>

Total 1530

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

b) Number of earthquakes observed at each station.

<table>
<thead>
<tr>
<th>Station</th>
<th>MZ</th>
<th>FO</th>
<th>OY</th>
<th>HM</th>
<th>IZ</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1346</td>
<td>722</td>
<td>1309</td>
<td>1260</td>
<td>1384</td>
</tr>
</tbody>
</table>

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

<table>
<thead>
<tr>
<th>Earthquake</th>
<th>3</th>
<th>4</th>
<th>5</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>503</td>
<td>623</td>
<td>404</td>
</tr>
</tbody>
</table>

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 observation 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 microearthquake⁵,

\[
M = 1.25 \log(V) + 2.50 \log(R) - 6.50
\]

(3.1)

where \( V \) is the maximum amplitude in \( \mu \) 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,

\[ E = 11.8 + 1.5M \]  \hspace{1cm} (3.2)

to our network.

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

\[ \Delta M_j = M_j - M \]  \hspace{1cm} (3.3)

In Table 6 the mean value of \( \Delta M \) and mean epicentral distance \( R \) is shown for 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 \( \Delta M_j \) 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.

\[ M = 1.25 \log(V) + 2.90 \log(R) - \text{const.} \]  \hspace{1cm} (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 \( \mu \)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\(^a\). 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 \( \mu \)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.
In Fig. 6 the magnitude \( (M) \)-accumulated frequency \( (N) \) relation is shown. At a glance on this figure we find Gutenberg—Richter's relation,

\[
\log n(M) = a - bM \quad (3.6)
\]

which does not hold in this district. If we manage to fit straight lines, their coefficients \( b \) are calculated by the method of Utsu\(^n\),

\[
1.2 < M < 2.2; \quad b = 0.89
\]

\[
2.2 \leq M; \quad b = 0.52
\]

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\(^n\).
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.

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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 April 1964 to June 1966. In this period seismicity is underestimated.

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 $3.0 \leq M$. 
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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 $0<d\leq5$, $5<d\leq10$, $10<d\leq15$, $15<d\leq20$, $20<d\leq30$, and $30<d\leq40$ respectively.

The details of seismicity and quaternary geotectonics and related problems will be discussed in succeeding papers.


Fig. 12 (a) Seismicity map of the focal depth $d$ as $0<d\leq 5$.

Fig. 12 (b) Seismicity map of the focal depth $d$ as $5<d\leq 10$. 
Fig. 12 (c) Seismicity map of the focal depth $d$ as $10 < d \leq 15$.

Fig. 12 (d) Seismicity map of the focal depth $d$ as $15 < d \leq 20$. 
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 $30 < d \leq 40$. 
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- Seismogram interpreters: Mrs. M. KOGA
- Station keepers: Mr. S. YABE, Mr. R. OGURA, Mr. T. SHIKATA, Mr. G. HOSOMI, Mr. Y. UENAKA

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References