INVERSION OF TELESEISMIC P WAVES OF
IZU-OSHIMA, JAPAN EARTHQUAKE
OF JANUARY 14, 1978

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Following the method described by KIKUCHI and KANAMORI (1982), we
develop an iterative deconvolution of complex body waves in which subevents
may take two different fault mechanisms. The method is applied to teleseismic
P waves of the Izu-Oshima earthquake of 1978. The result shows that the co-
seismic slip motion occurred not only on a submarine fault but also on an inland fault of the Izu-Peninsula. The seismic moment of the inland faulting is
$4.3 \times 10^{25}$ dyn·cm, which amounts to about one third of that of the submarine
faulting: $1.4 \times 10^{26}$ dyn·cm. Our result is consistent with the fact that aftershock activity increased around the inland and submarine faults almost simultaneously
after the main shock.

1. Introduction

The Izu-Oshima earthquake ($M_s=6.8$) of January 14, 1978 occurred in an area
of intensified observation in Japan. So far, not only the teleseismic waves but
also various kinds of near-field data such as strong motion and strainmeter rec-
dords have been analyzed.

The most characteristic feature of this earthquake is the migration of seismic
activity before and after the main shock. In Fig. 1, the seismicity map obtained
by TSUMURA et al. (1978) is shown schematically. The typical foreshocks and
aftershocks relocated by ISHIBASHI et al. (1978) are also plotted there. As pointed
out by TSUMURA et al. (1978) and YAMAKAWA et al. (1979), roughly three faults
are inferred from this map, and labelled $F_1$, $F_2$, and $F_3$ in Fig. 1. The main feature
of the sequence is as follows: foreshocks occurred near the eastern end of $F_1$; after
the main shock the seismic activity extended over $F_1$ and $F_2$ almost simultaneously;
the largest aftershock ($M=5.8$) occurred on $F_3$ about 19 h after the main shock.

OKADA (1978) analyzed the crustal strain data and obtained a quartet fault
model in which the submarine fault $F_1$ is divided into two subfaults. SACKS et al.
(1981) used the borehole strainmeter records and proposed a slow earthquake model
in which a slow faulting followed the normal event and migrated from the north-
western end of $F_2$ to the entire region of $F_3$. Thus all the static and quasi-static
data suggest the existence of slip motion on more than one subsidiary faults.

It still remains unsolved whether the coseismic rupture was restricted to the submarine fault. In the analysis of near-field strong motion and far-field SH waves, Shimazaki and Somerville (1978, 1979) proposed a single fault model in which the co-seismic slip motion is confined to the submarine fault $F_1$, while Sudo et al. (1978) modelled the strong motion data with the slip motion along $F_1$ and $F_3$. Since then, additional analysis to end the argument has not yet been reported.

In this paper, we will analyze teleseismic P waves recorded at WWSSN stations to examine whether co-seismic slip motion extended to the inland fault or whether it was confined to the submarine fault. To this end, we will first develop an iterative deconvolution technique following the method by Kikuchi and Kanamori (1982).

2. Method

We model an earthquake source as a series of double-couple point sources, each having the same time history. The individual point sources are then characterized by the seismic moment, onset time, location, and focal mechanism. We assume that all the subevents are located on a certain range of fault planes and take
the same focal mechanism on each fault plane.

For the sake of computation, the co-ordinates on the fault planes are discretized and numbered sequentially, so that an integer can identify the location as well as the focal mechanism. Thus each subevent can be specified by three parameters: \((m, s, l)\), where

\[
\begin{align*}
  m & = \text{the seismic moment} \\
  s & = \text{the onset time} \\
  l & = \text{the location number}
\end{align*}
\]

Let \(w_j(t; l)\) denote a synthetic wavelet which is generated by a point source: \((1, 0, l)\) and recorded at \(j\)-th station. Then the wave form generated from a point source of \((m, s, l)\) is given by \(mw_j(t-s; l)\). In our iterative deconvolution, the first solution, \((m_1, s_1, l_1)\), is determined by the criterion that the wavelet \(mw_j(t-s; l)\) best fits the observed wave form. The residual error to be minimized is then defined by

\[
J = \sum_{j=1}^{M} \left[ x_j(t) - mw_j(t-s; l) \right]^2 dt
\]  

(1)

where \(x_j(t)\) is the wave form observed at \(j\)-th station and \(M\) is the number of stations. After simple calculation we modify (1) as

\[
J = r_s - 2mr_w(s; l) + m^2 r_w(l)
\]

(2)

where

\[
\begin{align*}
  r_s & = \sum_{j=1}^{M} \int [x_j(t)]^2 dt \\
  r_w(s; l) & = \sum_{j=1}^{M} \int [w_j(t; l)x_j(t+s)] dt \\
  r_w(l) & = \sum_{j=1}^{M} \int [w_j(t; l)]^2 dt
\end{align*}
\]

(3)

For \(s\) and \(l\) being fixed, the residual error \(J\) is minimized if

\[
m = r_w(s; l) / r_w(l).
\]

(4)

For this value of \(m\), the residual error becomes

\[
J = r_s - [r_w(s; l)]^2 / r_w(l).
\]

Thus, we find that the first solution \((m_1, s_1, l_1)\) is obtained from the criterion

\[
r_w(s_1; l_1)^2 / r_w(l_1) = \text{maximum}
\]

(5)

and

\[
m_1 = r_w(s_1; l_1) / r_w(l_1).
\]

(6)

Next the observed wave form \(x_j(t)\) is replaced by the residual wave form given by

\[
x_j'(t) = x_j(t) - m_1w_j(t-s_1; l_1)
\]

(7)
and the same procedure is repeated to find the second and third solutions. After \( N \) iterations we obtain a sequence of \( N \) subevents: \((m_i, s_i, l_i), i=1, 2, \ldots, N\). Then the moment rate function is given by

\[
\dot{M}_i(t) = \sum_{i=1}^{N} m_i u(t-s_i)
\]

where \( u(t) \) represents the time history of subevent with the unit seismic moment.

The resulting synthetic wave form at \( j \)-th station becomes

\[
y_j(t) = \sum_{i=1}^{N} m_i w_j(t-s_i; l_i)
\]

The final residual error is given by

\[
\delta = r_{\text{res}} - \sum_{i=1}^{N} r_{w}(l_i)m_i^2
\]

or in the normalized form,

\[
\delta = \frac{\delta}{r_{\text{res}}} = 1 - \frac{\sum_{i=1}^{N} r_{w}(l_i)m_i^2}{r_{\text{res}}}
\]

3. Analysis

Figure 2 shows the vertical components of long period P wave forms which were digitized with a sampling interval of 0.4 s. The stations and the fault plane solution obtained later are also plotted onto an equal area projection. The wave forms are very complex, especially at northwestern stations (HLW, KJF). At these stations even the initial phase cannot clearly be identified. Hence we estimate the P arrival from the sP phase by taking the delay time into account.

The complexity of the observed wave forms suggests that the earthquake source consists of several subevents. Actually these subevents may have individual time history and source depth. However, since our knowledge about structure effects and noise contained in the observed records is poor, the estimation of so many parameters would not provide us with significant results. Moreover, our primary concern here is restricted to seeing whether a co-seismic slip motion occurs only along a single submarine fault or extends to an inland fault. Thus we assume that all subevents have the same time history of a trapezoidal function and the same focal depth at 7 km.

We take 10 discrete points, each being 5 km apart along the adjacent two fault strikes (Fig. 3). The first 7 points are located on the submarine fault and the remaining points are located on the inland fault. In the choice of focal mechanisms we fix the strikes at N95W and N55W for the submarine and inland faults, respectively. These values are inferred from the aftershock activity (see Fig. 1). The dip and slip angles are varied by trial and error. For given source mechanisms, we place double-couple point sources at a hypocenter in a homogeneous
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Fig. 2. Teleseismic long-period P waves used in the present analysis. The stations (WWSSN) and the focal mechanism of submarine fault are plotted onto an equal area projection.

Fig. 3. Discretized source points, each 5 km apart. The locations numbered 1 to 7 belong to submarine fault F_1 and those numbered 8 to 10 belong to inland fault F_2. The point numbered 2 is an assumed epicenter.

half space and calculate the wavelets following the method by Kanamori and Stewart (1978). We used 6.0 km/s and 3.5 km/s for the P and S wave velocities and 2.7 g/cm³ for the density. Then the wavelets generated from 10 discrete
points are given by taking the time shifts into account. Once the wavelets are calculated, the observed P wave forms are iteratively deconvolved into a subevent sequence: \((m_i, s_i, l_i)\).

We performed 10 iterations while no significant decrease in the residual error was found after 5 iterations. We also repeated the above procedure, varying the dip and slip angles by trial and error, and found the best fit solution of the fault mechanisms. The rise and rupture times were also determined in the least square sense.

The source parameters thus obtained are given in Table 1. The mechanism solution of submarine fault is in good agreement with the fault mechanism adopted by SHIMAZAKI and SOMERVILLE (1978) as their starting model which was determined from S-wave polarization data. The moment rate functions for two fault mechanisms are individually shown in Fig. 4. The location of subevents is also shown there. The spatio-temporal distribution indicates three clusters of the

<table>
<thead>
<tr>
<th>Fault</th>
<th>Strike (assumed)</th>
<th>Dip</th>
<th>Slip</th>
<th>Depth</th>
<th>Rise time</th>
<th>Rupture time</th>
</tr>
</thead>
<tbody>
<tr>
<td>(F_1)</td>
<td>N95W</td>
<td>77</td>
<td>190</td>
<td>7 km</td>
<td>1.2 s</td>
<td>2.0 s</td>
</tr>
<tr>
<td>(F_2)</td>
<td>N55W</td>
<td>95</td>
<td>195</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Fig. 4. (a) Moment rate function for submarine fault. (b) Moment rate function for inland fault. (c) Location of subevents.
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The first cluster consists of subevents on the submarine fault during the first ten seconds. The total seismic moment is \(8.8 \times 10^{25}\) dyn·cm. The second occurs on the inland fault about 9 s after the initial break. The seismic moment is \(4.3 \times 10^{25}\) dyn·cm. The third occurs again on the main fault. The seismic moment is \(5.6 \times 10^{25}\) dyn·cm. Consequently the total seismic moment of the submarine events amounts to \(1.4 \times 10^{26}\) dyn·cm. The details of these subevents are given in Table 2. In Fig. 5, the resulting synthetic wave forms are compared with the observed ones. The wave form matching is satisfactory at all stations.

The above result supports the possibility of co-seismic inland faulting about 10 s after the initial break. However, before justifying this, we have to see if a single fault model cannot explain the observed wave forms. For this purpose, we confine the range of the parameter \(l\) to \(1 \leq l \leq 7\) and perform the same procedure as before.

The results are shown in Fig. 6. We see that both the moment rate function and the location of subevents are basically the same as those of the submarine fault obtained before. Figure 7 shows the resultant synthetic wave in comparison with the observed ones.

Fig. 5. Synthetic and observed wave forms for a double fault model.
Table 2. List of subevents.

<table>
<thead>
<tr>
<th>No.</th>
<th>Onset time (s)</th>
<th>Location number</th>
<th>$M_s$ (10^8 dyn cm)</th>
<th>Subtotal of $M_o$ (10^8 dyn cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Submarine events</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>0</td>
<td>2</td>
<td>1.4</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>1.6</td>
<td>4</td>
<td>2.4</td>
<td>8.8</td>
</tr>
<tr>
<td>3</td>
<td>4.0</td>
<td>5</td>
<td>4.0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>6.0</td>
<td>4</td>
<td>1.0</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>15.2</td>
<td>7</td>
<td>0.9</td>
<td></td>
</tr>
<tr>
<td>21.2</td>
<td>6</td>
<td>2.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>22.8</td>
<td>7</td>
<td>1.1</td>
<td></td>
<td>5.6</td>
</tr>
<tr>
<td>31.2</td>
<td>1</td>
<td>0.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Inland events</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>9.2</td>
<td>8</td>
<td>3.1</td>
<td>4.3</td>
</tr>
<tr>
<td>30.0</td>
<td>8</td>
<td>1.2</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Although the overall agreement of the wave forms is good, a relatively large discrepancy is seen between 10 to 20 s after the initial onset. The discrepancy is remarkable especially at northeastern stations (GDH, COL, DUG). Such misfit at the particular azimuth suggests the change in fault mechanism during the co-seismic rupture process. Combining this with the above analysis, it is reasonable to say that the co-seismic slip motion on the inland fault is necessary for explaining observed P wave forms.
4. Discussion

The primary concern of this paper is to examine whether or not a co-seismic slip motion extended to the inland fault. For this purpose, we have analysed teleseismic long-period P waves using an iterative deconvolution technique. In a strict sense, the solution is not the best fit one for a given number of unknown sources. Nevertheless, the above method has great advantages in the stability of solution and in the saving of computational time.

Suppose that we attempt to determine simultaneously $N$ point sources best fit to the observed records. The number of all possible combinations of the location and the onset time of subevents is then given as $(n_t n_g)^N$, where $n_t$ and $n_g$ are the numbers of the sampling points in time and the grid points in space, respectively. To get the best fit solution, we have to solve the $N$ normal equations for the individual seismic moments in each of the $(n_t n_g)^N$ cases, and compare the residual errors. Thus the computational time and cost would be unbearable even for several sources. In additions, as a more serious problem, the solution might be
unstable because of the ill conditions in the normal equations.

Our result indicates the existence of co-seismic inland faulting. The spatio-temporal distribution of subevents suggests that the rupture almost continuously extended beyond the coast line of the Izu Peninsula. It should be noted, however, that resolution of source location is poor because the relative time shift of P arrivals for varying source locations ranges only within a few seconds. Hence, we cannot clearly say what portion of the inland fault ruptured in a co-seismic way.

The total seismic moment of submarine events is $1.4 \times 10^{26} \text{dyn} \cdot \text{cm}$, while the seismic moment of inland events is $4.3 \times 10^{25} \text{dyn} \cdot \text{cm}$. These values are slightly larger than those obtained in previous works. For example, Shimazaki and Somerville (1978, 1979) obtained $1.1 \times 10^{26} \text{dyn} \cdot \text{cm}$ for the submarine event. Using the static parameters estimated by Okada (1978) with the rigidity of $3.5 \times 10^{11} \text{dyn/cm}^2$, we have the seismic moment of $2.1 \times 10^{25} \text{dyn} \cdot \text{cm}$ for the inland event. However, such differences may be due to experimental errors. Since the amplitude of synthetic body waves is inversely proportional to the cube of body wave velocity, only a $10\%$ decrease of the assumed body wave velocity results in a $30\%$ decrease of the seismic moment to be estimated.

Shimazaki and Somerville did not regard the inland faulting as a seismic wave source, although they did not reject the possibility of aseismic or very slow faulting there. In this respect, they pointed out that any distinct later phase corresponding to the inland faulting cannot be found in the near-field accelerogram. In fact, the accelerogram as well as the strong motion observed at near-field JMA stations are so complex that any discrete subevents cannot be identified. We think, however, such complexity may rather suggest that co-seismic slipping occurs in a relatively large area, or at least that it does not indicate the absence of inland faulting.

In addition, our model seems to be more consistent with the sequence of seismic activity (Fig. 1). Note that the aftershock activity increased almost simultaneously over $F_1$ and $F_2$ without any noticeable migration front, and note that a relatively large aftershock ($M=5.1$) occurred at the northwestern end of $F_2$ only about one hour later than the main shock. Hence, it is natural to think that the co-seismic rupture rapidly extended to the inland fault.

A cluster of retarded subevents was obtained on the submarine fault. Although the resolution of the source location is poor, it is interesting to note that the subevents are located near both ends of the submarine fault. We can regard them as the beginning of aftershocks. We also obtained the value of $2 \text{s}$ for the apparent rupture time of individual subevents. This suggests that a typical length of coherent faults in this area is 4 to 6 km, for the rupture velocity assumed to be 2 to 3 km/s.

REFERENCES

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