RUPTURE CHARACTERISTICS OF THE 1983 NIHONKAI-CHUBU (JAPAN SEA) EARTHQUAKE AS INFERRED FROM STRONG MOTION ACCELEROMETERS

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The rupture characteristics of the 1983 Nihonkai-chubu (Japan Sea) earthquake ($M_{JMA} = 7.7$) were investigated using strong motion accelerograms recorded at 10 stations with epicentral distances of between 80 and 280 km. The main shock emitted a large amount of high-frequency seismic energy in two stages, forming two high-amplitude envelopes on the accelerograms. The time difference between S-wave arrival-times of the two events becomes larger as one moves clockwise in azimuth from north to south. Using the time differences, the second event was located 44 km NNE of the first one with the time of origin 26 s after the first event. The azimuthal variation in the amplitude ratio of the two events is consistent with the relative locations. The strong motion accelerograms, combined with the aftershock distribution and the source process time of 63 s obtained from long-period surface waves, suggest the following rupture model. The first event initiated at the southern end of the aftershock area and extended in a direction of N15°E. It stopped near the zone of low after shock activity west of Kyuroku Island. After a pause of about 10 s, the second event started at a place just north of the zone of low aftershock activity and extended in the same direction. A third event initiated near the place where the strike of the aftershock distribution changes from N15°E to N15°W. A low rupture velocity and the paucity of small-scale barriers are possible reasons for the low radiation of high-frequency energy from the third event. The above rupture characteristics appear to be closely correlated with the heterogeneous crustal structure as revealed by other geophysical and geological data in the source region. The observed S-wave accelerations of the first and the second events were jointly inverted for the source acceleration spectra, the attenuation coefficient of $Q$, and the amplification factors of the recording sites. The source spectra of the two events are almost the same, showing a rapid decay of amplitude at frequencies higher than 4 Hz. The $Q$ increases in proportion to the frequency from 66 at 1 Hz to 1,026 at 16 Hz. The local stress drops of subsources in terms of the stochastic source models are estimated to be 380 bars for the first and 340 bars for the second event. The estimates are model-dependent and can vary by as much as a factor of 1.5. The second event being more efficient than the first in radiating high-frequency seismic waves can be interpreted as having smaller subsources.
1. Introduction

Seismic records observed from large earthquakes often show a complicated pattern: a sequence of high-amplitude pulses separated by comparatively long periods of low-amplitude radiation (Wyss and Brune, 1967; Trifunac and Brune, 1970; Fukao and Furumoto, 1975). Such records are thought to represent the complexity of earthquake ruptures. This complexity may be due to geometrical complexities of the fault plane or heterogeneities in the fault strength or tectonic stress (Das and Aki, 1977; Aki, 1979; Lay et al., 1982). An irregular rupture propagation tends to enhance high-frequency waves (Blandford, 1975; Mikumo and Miyatake, 1978; Koyama, 1983). In this sense, strong motion accelerograms are essential in clarifying the nature of the complexities of the earthquake rupture and inferring the heterogeneities in the fault properties (Mom and Shimazaki, 1984).

The purpose of the present paper is to investigate the rupture characteristics of the 1983 Nihonkai-chubu (Japan Sea) earthquake ($M_{\text{JMA}}$ 7.7), principally analyzing strong motion accelerograms recorded at short distances. This earthquake occurred on 26 May 1983 in the Japan Sea about 80 km west of the coast of Aomori and Akita Prefectures in northern Honshu, Japan. Aftershocks at depths shallower than 30 km occurred over an area of 140 km by 40 km (Kosuga et al., 1984; Umino et al., 1985). The depth distribution of aftershocks (Sato et al., 1985) and the first motion distribution of P-waves (Ishikawa et al., 1984; Shimazaki and Mori, 1983; Tohoku University and Hiroasaki University, 1984) suggest that the main shock is a low-angle thrust which dips eastward with an angle of 20°. Long-period body waves recorded at world-wide stations have shown that the rupture was a double event (Shimazaki and Mori, 1983; Ishikawa et al., 1984). But the seismic moment and source duration as estimated from the long-period body waves are much smaller than those from long-period surface waves, suggesting a slow event after the two major events (Mori and Shimazaki, 1983; Ishikawa et al., 1984). The strong motion accelerograms are examined in order to study these details more clearly. It has been shown that the complex rupture revealed by the seismological data correlates well with the heterogeneous crustal structure in the source region (Hydrographic Department, 1984).

We shall also estimate the source acceleration spectra of the first and the second events. The acceleration spectra are interpreted in terms of a stochastic source model to estimate the local stress drops of subsources for the two major events (Hirasawa, 1979; Papageorgiou and Aki, 1983 a). The local stress drops are 2–4 times larger than the global stress drops, in agreement with the result obtained by Izutani (1983). Comparing the source factors, as estimated from the acceleration spectra, with the moments from long-period body waves for the two events, we found that the second event was more efficient in radiating high frequency energy than the first.
2. **Double Envelope Pattern**

Accelerographs installed in northern Japan produced many valuable records with information on the detailed time history of the rupture process of the Japan Sea earthquake. Figure 1 shows the three-component accelerograms recorded at the Earthquake and Volcano Observatory of Hirosaki University (HRD) in Hirosaki City (Fig. 2). The distance is 124 km from the epicenter (40°23.3′N, 138°54.5′E) determined by the microearthquake network of Hirosaki University. We can infer that the instrument started a few seconds after the first P-arrival by comparing the S-P time expected for an epicentral distance of 124 km with the observed time of the first S-arrival which is clearly identified on the horizontal components. Although accelerations exceeding 50 gals last about 50 s on the horizontal components, we have noticed two high-amplitude envelopes each with a duration time of 10–15 s which are separated by a rather long period of low-amplitude radiation. Because the two high-amplitude envelopes are much more conspicuous on the horizontal components, we suppose these wave packets to consist mainly of S-waves.

We used a Husid plot of energy buildup (Husid, 1967) and moving window Fourier spectrum (Trifunac and Brune, 1970) to substantiate the double envelope pattern. The term “energy” is used here for the time integral of the square of the ground acceleration. Figure 3 shows the Husid plot for the ground accelerations...
Fig. 2. Strong motion stations used in the present study (squares) and the epicentral distribution of aftershocks during the period of May 26 to June 8, 1983 (KOSUGA et al., 1984). The solid star denotes the epicenter of the main shock and the open stars denote the epicenters of major aftershocks. Two aftershocks (M 6.0 and M 6.1) in the southern end occurred on June 9 and the largest aftershock (M 7.1) in the northern end occurred on June 21. The fault of the main shock as estimated from the aftershock distribution is represented by a rectangle bent halfway along the fault length. The cross marks the location of Kyuroku Island.

from horizontal components at HRD. The total signal duration can be divided into five sections according to the variation of the average gradient of the Husid plot. Two steep slopes separated by a gentle slope correspond to the two high-amplitude envelopes on the accelerogram. The duration time of the first envelope is about 10 s and that of the second one is about 16 s. The duration separating the two high-amplitude envelopes is about 12 s.

Figure 4 shows the moving window Fourier spectrum of the NS component of acceleration at HRD. A box-car window of 10 s was used. The time window was shifted every one second and the Fourier spectrum was calculated for each window. Spectral amplitudes were averaged over a frequency band-width of 0.8 Hz and normalized by the largest spectral amplitude in the whole signal duration. Then they were subdivided into 20 levels and amplitudes of the same level were connected to each other. The two high-amplitude arrivals, separated by a period of low-amplitude radiation as seen on the record trace and the Husid plot
Fig. 3. Husid plot of the strong motion acceleration recorded at HRD. Solid and dashed lines represent NS and EW components, respectively. Arrows indicate the first and the final S-wave arrivals of the two major events as identified on the Husid plot.

Fig. 4. Moving window Fourier spectrum of the strong motion acceleration of NS component recorded at HRD. Domains of amplitudes are progressively stippled in four steps corresponding to the normalized amplitudes, 0.0–0.25, 0.25–0.5, 0.5–0.75, and 0.75–1.0. Domain of amplitude 0.0–0.25 is not stippled at all.

are also demonstrated in this moving window Fourier spectrum. Predominant frequencies contained in the first high-amplitude envelope are 2–3 Hz and those of the second one are around 4 Hz. It appears that the second envelope consists of seismic waves with slightly higher frequencies.

The accelerograms of the largest aftershock ($M_{\text{JMA}}$ 7.1) recorded at the same
station, HRD, do not show such double envelope patterns, although the epicentral
distances from the main shock and the aftershock are almost identical. Therefore,
the double envelope pattern of the main shock accelerogram is not the effect of
local site but is due to the source. Similar patterns of amplitude variation are
recorded at many other stations (Kurata et al., 1983; Sawada et al., 1983).
Hereafter, we call the seismic source for the first envelope the first major event
or E1 and the source for the second envelope the second major event or E2.

3. Locations of Two Major Events

In the following, we shall analyze the accelerograms recorded at the 10 stations
listed in Table 1. Absolute hypocenter parameters of the two major events cannot
be determined from the accelerograms alone because no absolute times are marked
on the records. We shall locate the starting point of E2 relative to that of E1 using
time differences between the first S-arrivals of the two events on the accelerograms.
Correct identification of the first S-arrivals is crucial in precise determination
of the relative locations. Because long-period surface waves generated by E1
may obscure sharp S-arrivals from E2, we applied a high-pass filter to the accel-
erograms. The frequency response of the filter is

$$H(\omega) = \frac{\omega^2}{\omega_0^2 - \omega^2 - 12h\omega^2 \omega},$$

where $\omega_0 = 2\pi/T_0$, $\omega$ is the angular frequency, $h$ is the damping constant of 0.1, and
$T_0$ is the corner period of the filter which is set at 1.0 s. We applied the Husid
plot to the filtered accelerograms and identified the first S-arrivals of the two events.

Table 1. Strong motion stations used in the present study.

<table>
<thead>
<tr>
<th>Station name</th>
<th>Station code</th>
<th>Distance (km)</th>
<th>Azimuth (deg)</th>
<th>$T$ (s)</th>
<th>$a_2/a_1$</th>
<th>Operating Institution</th>
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</thead>
<tbody>
<tr>
<td>Muroran</td>
<td>MUR</td>
<td>275</td>
<td>38</td>
<td>14.5</td>
<td>2.5</td>
<td>PHRI*</td>
</tr>
<tr>
<td>Hakodate</td>
<td>HAK</td>
<td>217</td>
<td>44</td>
<td>18.0</td>
<td>1.7</td>
<td>PHRI</td>
</tr>
<tr>
<td>Shiranuka</td>
<td>SRN</td>
<td>225</td>
<td>67</td>
<td>19.0</td>
<td>1.1</td>
<td>CRIEPI**</td>
</tr>
<tr>
<td>Rokkasho</td>
<td>RKK</td>
<td>210</td>
<td>71</td>
<td>17.5</td>
<td>1.5</td>
<td>MOOSC***</td>
</tr>
<tr>
<td>Furofushi</td>
<td>FRF</td>
<td>85</td>
<td>73</td>
<td>18.5</td>
<td>1.5</td>
<td>CRIEPI</td>
</tr>
<tr>
<td>Aomori</td>
<td>AOM</td>
<td>169</td>
<td>73</td>
<td>17.5</td>
<td>1.3</td>
<td>PHRI</td>
</tr>
<tr>
<td>Namieka</td>
<td>NAM</td>
<td>154</td>
<td>76</td>
<td>19.0</td>
<td>0.97</td>
<td>TAAB****</td>
</tr>
<tr>
<td>Hirosaki</td>
<td>HRD</td>
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<td>80</td>
<td>21.5</td>
<td>0.89</td>
<td>Hirosaki Univ.</td>
</tr>
<tr>
<td>Akita</td>
<td>AKI</td>
<td>127</td>
<td>123</td>
<td>31.0</td>
<td>0.61</td>
<td>PHRI</td>
</tr>
<tr>
<td>Sakata</td>
<td>SAK</td>
<td>183</td>
<td>154</td>
<td>32.0</td>
<td>0.65</td>
<td>PHRI</td>
</tr>
</tbody>
</table>

Epicentral distances and azimuths are calculated for the epicenter of the main shock
determined by the microearthquake network of Hirosaki University. Also listed are the
time differences, $T$ between S-arrival times of the two major events and r.m.s. amplitude
ratios of the two events, $a_2/a_1$. * PHRI, Port and Harbour Research Institute; ** CRIEPI,
Central Research Institute of Electric Power Industry; *** MOOSC, Mutsu-Ogawahara
Oil Storing Company; **** TAAB, Tohoku Agricultural Administration Bureau.
with the times when the energy began to increase sharply. The time precision was estimated from the sharpness of the onsets. The time differences between the two S-arrivals are summarized in Table 1. They were determined with errors of about 1–2 s for most stations except MUR, HAK, and SAK; the first S-arrival of E1 for MUR and HAK and that of E2 for SAK were not clear, so the errors for these stations are rather large, as much as 2–3 s. It is noted that the observed time difference generally gets larger as one moves clockwise in azimuth from MUR to SAK.

We have assumed that a homogeneous half space with a constant S-wave velocity of β exists. Since it is difficult to get any depth resolution on the relative locations using those time differences which contain rather large errors, we assume that the sources are situated on the free surface. Then the time difference at the j-th station can be written as

$$T_j = T_1 + t_1 + (r_1 - r_2)/\beta,$$

where $T$ is the time difference of the origin times of the two events, and $r_1$ and $r_2$ are the epicentral distances of the j-th station from E1 and E2, respectively. The epicenter of E2 relative to that of E1 is given by $l, \theta$, where $l$ is the distance between the two epicenters and $\theta$ is the azimuth of E2 from E1. Using $l, \theta$, we can rewrite (2) as

$$T_j(l, \theta) = T_1 + \sqrt{(x_2 - l \cdot \cos \theta)^2 + (y_2 - l \cdot \sin \theta)^2} - \sqrt{x_1^2 + y_1^2}/\beta,$$

where $(x_j, y_j)$ is the coordinates of the j-th station in the coordinate system with its origin placed at the epicenter of E1. The unknown parameters $(\tau, l, \theta)$ can be determined by minimizing the sum of the squared residuals, i.e.,

$$\sum_{j=1}^{N} (T_j - \hat{T}_j)^2 = \text{min},$$

where $N$ is the total number of stations, $T_j$ is the observed time difference and $\hat{T}_j$ is the predicted time difference given by (3).

Two locations were assumed for the starting point of E1. One is the main-shock epicenter determined by the microearthquake network of Hirosaki University, and the other is a point close (within 5 km) to the main-shock epicenter determined by the Japan Meteorological Agency (JMA). We have assumed that $\beta = 3.7$ km/s based on the average P-wave velocity of the crust estimated by Sato et al. (1985). For the former epicenter of E1, we have determined that $\tau = 27.4 \pm 2.9$ s, $l = 44.3 \pm 5.2$ km and $\theta = N35.0^\circ E \pm 15.3^\circ$, which gives a r.m.s. residual of 1.5 s. For the latter epicenter of E1, we have determined that $\tau = 26.2 \pm 2.5$ s, $l = 44.2 \pm 4.1$ km, and $\theta = N27.2^\circ E \pm 15.0^\circ$, which gives a r.m.s. residual of 1.3 s. These epicenters are shown in Fig. 5 together with the fault area of the main shock which was estimated from the aftershock distribution during the period between May 26 and June 8, 1983 (Fig. 2). This period precedes the time of conspicuous expansion of aftershock activity (Kosuga et al., 1984). At the southern and northern
ends, the estimated fault does not include the rupture, areas of the major aftershocks of $M 6.0$, $M 6.1$, and $M 7.1$. After several tests with different epicenters of $E_1$, we found that the solutions for $(r, l, \theta)$ did not largely depend on the epicenter of $E_1$ as long as it was assumed to be near the main-shock epicenters determined by the microearthquake network and JMA. Judging from the focal depths of aftershocks (SATO et al., 1985), the depths of the two events are probably shallower than 20 km. For the stations with epicentral distances greater than 100 km, i.e., all the stations except for FRF, hypocentral distances do not change significantly as the focal depth changes in ranges shallower than 20 km. Therefore, the solutions for $E_2$ cannot be significantly affected by the assumed depths of the two events.

The ratios between r.m.s. amplitudes of the two high-amplitude envelopes are also listed in Table 1. In calculating the r.m.s. amplitude, the time window was chosen to be 10 s after the first S-arrival of each event. We used the average of the two horizontal components. The r.m.s. amplitude ratio of $E_2$ to $E_1$ generally be-
comes smaller as one moves clockwise in azimuth from MUR to SAK. This is consistent with the relative locations of the two events, and suggests that radiation patterns for all the stations do not change considerably between the two events.

A small initial arrival of P-waves was followed 3 s later by a much larger impulsive arrival at most long-period WWSSN stations. The epicenter of the larger event is estimated to be within 5 km of the initial epicenter since there is no resolvable (less than 0.5 s) time variation with an azimuth in the arrival time of the larger event (SHIMAZAKI and MORI, 1983; ISHIKAWA et al., 1984; JAPAN METEOROLOGICAL AGENCY, 1984). On the strong motion accelerograms, we cannot discriminate between the arrivals of the initial minor event and the larger event. But it seems quite reasonable that the first S-arrival identified on the accelerograms was emitted from a source very close to the epicenter of the initial event, which probably corresponds to the epicenter of the main shock as determined by the local networks.

From the synthetics of long-period P-waves, SHIMAZAKI and MORI (1983) inferred that the second major event started 24 s after the initial event at a point 30-35 km N0°-30°E of the initial epicenter. Using an inversion method developed by KIKUCHI and KANAMORI (1982), ISHIKAWA et al. (1984) found that the second major event occurred 26 s after the initial event at a point about 50 km north-north-east of the first major event. It is noted that the relative locations and origin times of the two major events obtained from the strong motion accelerograms roughly agree with those obtained from the long-period body waves. We believe that the sources for long-period body waves were also effective in radiating a large amount of high-frequency seismic energy.

The location of the stopping point of E1 relative to the starting point of E2 can be determined directly by examining the azimuthal variation in the time difference between the final S-arrival of E1 and the first S-arrival of E2. However, the time resolution in picking up the final S-arrivals of E1 was not very good so we could not use this data to determine the relative locations. We believe the aftershock distribution in Fig. 2 presents a clue to this problem. About 30 km NNE of the starting point of E1 and about 10 km SSW of the starting point of E2, we found an elongated zone where aftershocks are sparsely distributed to the west of Kyuroku Island which subsided about 32 cm coseismically (YAMASHINA et al., 1985). This zone suggests the existence of a barrier which has stopped the first rupture.

4. Three-Stage Strain-Release Process

We shall present a rupture model which is consistent with the distinct characteristics of the strong motion accelerograms for the entire length of the record. The rupture area during the main shock is estimated from the aftershock distribution. The assumed fault projected on the free surface is shown in Fig. 5. It is a rectangular fault with a change in strike halfway along the length. The strike of the fault is N15°E on the southern part and N15°W on the northern part. According to the
depth distribution of aftershocks (SATO et al., 1985), the fault is assumed to dip eastward with an angle of $20^\circ$ over the entire fault. A fault width of 35 km is derived from the dip angle and the width of the fault projected on the free surface.

We assume that the rupture initiates at the epicenter of E1, point 0 in Fig. 5, and extends in a direction of N15$^\circ$E. Postulating that all of the high-frequency energy originates at the rupture front (MADARIAGA, 1977), we first calculate the arrival-times of S-waves from the rupture front as the rupture propagates in time. The arrival-times are compared with the accelerograms by correlating the observed first S-arrival with the theoretical arrival-time from the starting point of E1, as shown in Fig. 6. For simplicity, we have considered only the S-arrivals from sources distributed along the center line which bisects the width of the fault. The depth of the center line is 9 km if the top of the fault is assumed to be situated at a depth of 3 km, which is the depth to the sea floor in the aftershock area. The velocity structure of S-waves is assumed to be a three-layered model: the upper layer is 15 km thick with a velocity of 3.57 km/s; the middle layer is 13 km thick with a velocity...
of 3.88 km/s; and the lowest layer is a half space with a velocity of 4.45 km/s. The arrival-time here is understood to be the first arrival from source to receiver. If the stopping point of E1 is assumed to be located at the zone of low aftershock activity, which is 30 km NNE of the initial rupture, the first high-amplitude envelopes of acceleration must be included in the time interval between the arrivals from points 0 and 3. The condition was generally satisfied by assuming a constant rupture velocity of 2 km/s as shown in Fig. 6.

Although the starting point of E2 was determined about 14 km NNE of the stopping point of E1 in the previous section, we assume in the present model that E2 starts at the same point as the stopping point of E1 and propagates unilaterally in the same direction as before. This assumption is made partly for the sake of model simplicity and partly because we infer E2 may have started much closer to the stopping point of E1 than 14 km, noting that the elongated zone of the low aftershock activity has a width of only about 5 km. The starting time of E2 is assumed to be 24 s after the initial rupture, which is about 2 s shorter than the most probable estimate obtained in the previous section. We have taken into account the possibility that the first S-arrivals of E2 have systematically been determined late because of the preceding seismic wave trains of appreciable amplitudes.

The whole rupture time of 63 s is adopted here since the source process time estimated from the long-period surface waves is about 60–63 s (Mori and Shima-Zaki, 1983; Ishikawa et al., 1984). This is considerably longer than the approximate 40 s duration of the main P-waves as seen on the long-period WWSSN records (Shimazaki and Mori, 1983). Given this total source duration, E2, which starts 24 s after the initial rupture would have a duration of 39 s. At ordinary rupture velocities of 2.5 to 3.0 km/s, the rupture length of E2 would extend much

Fig. 7. Mode of rupture propagation assumed in the present study. Lapse time (above) and rupture velocity (below) are both shown against the instantaneous position of the rupture front. This model satisfies the condition that the total rupture over the fault length of 100 km is completed in 63 s.
further than the length of the fault that is estimated from the aftershock distribution. Instead of assuming a constant slower rupture velocity, we assume that the rupture velocity of E2 changes from an ordinary value to a much smaller one at the midway point of the rupture propagation. This can better explain why there was relatively little high-frequency seismic energy emitted at the later stage of E2. The bending corner of the fault is assumed to be the place where the rupture velocity begins to decrease since such a geometrical discontinuity may hinder smooth rupture propagation. The rupture velocity of the earlier stage of E2 is assumed to be 2 km/s, the same as that of E1. The rupture velocity of the later stage is determined such that the rupture reaches the northern end of the fault 63 s after the initial rupture. The rupture model described above is shown in Fig. 7.

In Fig. 6, only the arrival-times of S-waves from points 0, 3, 6, 7, and 10 are plotted and connected by lines between the stations, which are arranged from the top in order of northern to southern stations. There are two arrivals from point 3: the earlier corresponds to the stopping phase of E1 and the later to the starting phase of E2. Generally, the first S-arrival of E2 appears to coincide with the starting phase from point 3 and the second high-amplitude envelope seems to be included in the time interval between the arrivals from points 3 and 6. Although we cannot refute the view that appreciable energy is arriving in the time interval

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**Fig. 8.** Source areas of the three major events in the present rupture model. The shaded areas in the southern and central parts correspond to the source areas of the first and the second events, respectively. The unshaded area in the northern part indicates the source area of the third event which radiated relatively little high-frequency energy. The star symbol indicates the epicenter of the main shock.
between the arrivals from points 6 and 7, very little can be seen in the time interval
in which the S-waves from points 7 and 10 arrive. Hereafter, we call this part of
rupture the third major event or E3, in order to discriminate it from E2.

Figure 8 shows the source areas of the three major events. The source area of
E1 is represented by the darkly shaded area between points 0 and 3. The south-
ernmost area between points —1 and 0 is lightly shaded since it might possibly
be ruptured during the first stage. Thus the fault length of E1 is estimated to be
30–40 km. The source area of E2 is the darkly shaded area between points 3 and
6. E2 might have extended past the change in orientation of the fault, although it
probably stopped around point 7. Therefore, the area between points 6 and 7 is
lightly shaded. Thus the fault length of E2 is estimated to be 30–40 km, the same
as that of E1. The source area of E3 is the unshaded area in the northern part of the
fault. The fault length of E3 is also estimated to be 30–40 km.

It has been suggested that a source process time of 60–63 s and a total
moment of $7.5-8 \times 10^{27}$ dyn-cm, which are derived from the long-period surface
waves, require that the second event be followed by a slow event to make up for a
source duration of about 40 s and a moment of $3.5-5 \times 10^{27}$ dyn-cm as estimated
from the long-period body waves (MORI and SHIMAZAKI, 1983; ISHIKAWA et al,
1984). This theory agrees with the data of the strong motion accelerograms in
that little high-frequency seismic energy is observed in the time interval correspond-
ing to the third event.

In order for the first high-amplitude envelopes of accelerations to be included
in the time interval between the arrivals from points 0 and 3, we had to choose a
lower rupture velocity of about 2 km/s. The rupture velocity of E2 was assumed
to be the same as that of E1. Therefore, the change in the rupture velocity at
the change in orientation of the fault is rather small, as shown in Fig. 7. It only
changes from 2 km/s to 1.67 km/s. In comparing the accelerograms with theoret-
ical S-arrivals, we considered only the contribution from sources on the center line
bisecting the width of the fault. If we consider a more realistic two-dimensional
source of the same fault length, we will have to choose a rupture velocity greater
than 2 km/s in order to include the two high-amplitude envelopes in the same re-
spective time intervals. If the rupture velocity of E2 is 2.5–3.0 km/s, the rupture
velocity of E3 becomes 1.5–1.4 km/s, showing a sharp decrease in rupture velocity
at the change of fault orientation. A lower rupture velocity for E3 can better ex-
plain why it did not radiate abundant high-frequency seismic energy.

Assuming that the slip over each segment conforms to the static displacement
expected for the crack model (Model A), we shall estimate the average slip and the
stress drop of each segment. Here we assume two sets of seismic moments in
Table 2: One (MO.SM) is based on estimates given by SHIMAZAKI and MORI
(1983) and MORI and SHIMAZAKI (1983), and the other (MO.I) by ISHIKAWA et al.
(1984). The moment caused by E3 was obtained by subtracting the total moment
estimated from long-period body waves from the total moment estimated from
long-period surface waves. For MO.I, the total moment from long-period body
Table 2. Two sets of seismic moments ($\times 10^{27}$ dyn-cm) estimated for the three major events.

<table>
<thead>
<tr>
<th></th>
<th>E1</th>
<th>E2</th>
<th>E3</th>
<th>Total</th>
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<tbody>
<tr>
<td>MO.SM</td>
<td>3</td>
<td>2</td>
<td>3</td>
<td>8</td>
</tr>
<tr>
<td>MO.I</td>
<td>1.5</td>
<td>0.75</td>
<td>4</td>
<td>7.5</td>
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MO.SM is based on the estimates by Shimazaki and Mori (1983) and Mori and Shimazaki (1983). MO.I is based on the estimates by Ishikawa et al. (1984).

Table 3. Two sets of fault lengths (km) estimated for the three major events.

<table>
<thead>
<tr>
<th></th>
<th>E1</th>
<th>E2</th>
<th>E3</th>
<th>Total</th>
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<tr>
<td>FL.A</td>
<td>30</td>
<td>30</td>
<td>40</td>
<td>100</td>
</tr>
<tr>
<td>FL.B</td>
<td>40</td>
<td>40</td>
<td>30</td>
<td>110</td>
</tr>
</tbody>
</table>

The average slip, $D$, and the stress drop, $\Delta \sigma$, of each segment are listed in Table 4 for every combination of the seismic moment in Table 2 and the fault length in Table 3. The fault width, $W$ is assumed to be 35 km for all three segments. The average slips are given by

$$D = \frac{M_0}{\mu L W}, \quad (5)$$

where $M_0$ is the seismic moment, and $\mu$ is the rigidity and is assumed to be $3.7 \times 10^{11}$ dyn/cm². The stress drops are calculated for the circular crack model (Eschelby, 1957),

$$\Delta \sigma = \frac{7}{16} \frac{M_0}{R^8}, \quad (6)$$

where $R$ denotes the radius of the circular crack which was given by the equation, $R = (L W/\pi)^{1/2}$. The stress drops in round brackets in Table 4 are calculated for the infinite two-dimensional fault of a pure dip-slip type (Starr, 1928). The formula is given by

$$\Delta \sigma = \frac{16}{3\pi} \frac{M_0}{L W^2}. \quad (7)$$

In Table 4, the average slips of E1 and E2 are much smaller than that of E3 for the moments estimated by Ishikawa et al. (1984) (MO.I). This does not agree with the tsunami heights observed along the coast adjacent to the source region: the tsunami heights on the northern source region were roughly the same as those on the southern region (Aida, 1984; Tsujii et al., 1984). In terms of Model A, MO.SM is more consistent with the observation of tsunami heights. Using MO.SM, we obtained a stress drop in the range of 100–200 bars for all three segments. The stress drops of intra-plate earthquakes agree with these values better than those of...
Table 4. Stress drops $\Delta \sigma$ (bars) and average displacements $D$ (m) for the three major events.

<table>
<thead>
<tr>
<th></th>
<th>E1</th>
<th></th>
<th>E2</th>
<th></th>
<th>E3</th>
<th></th>
<th>Total</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>$\Delta \sigma$</td>
<td>$D$</td>
<td>$\Delta \sigma$</td>
<td>$D$</td>
<td>$\Delta \sigma$</td>
<td>$D$</td>
<td>$\Delta \sigma$</td>
<td>$D$</td>
<td></td>
</tr>
<tr>
<td>MO.SM+FL.A</td>
<td>220</td>
<td>7.7</td>
<td>140</td>
<td>5.2</td>
<td>140</td>
<td>5.8</td>
<td>94 (110)</td>
<td>6.2</td>
</tr>
<tr>
<td>MO.SM+FL.B</td>
<td>140</td>
<td>5.8</td>
<td>93</td>
<td>3.9</td>
<td>210</td>
<td>7.7</td>
<td>82 (100)</td>
<td>5.6</td>
</tr>
<tr>
<td>MO.I+FL.A</td>
<td>110</td>
<td>3.9</td>
<td>54</td>
<td>1.9</td>
<td>190</td>
<td>7.7</td>
<td>88 (100)</td>
<td>5.8</td>
</tr>
<tr>
<td>MO.I+FL.B</td>
<td>70</td>
<td>2.9</td>
<td>35</td>
<td>1.5</td>
<td>290</td>
<td>10.0</td>
<td>77 (94)</td>
<td>5.3</td>
</tr>
</tbody>
</table>

These are calculated for combinations of the seismic moments in Table 2 and the fault lengths in Table 3, according to Model A.

inter-plate earthquakes (KANAMORI and ANDERSON, 1975). We call the stress drops listed in Table 4 “global stress drops” in order to discriminate them from “local stress drops” which represent the stress drops over localized areas of the source (PAPAGEORGIOU and AKI, 1983 a).

5. Local Stress Drops

We shall determine the source acceleration spectra of the first and the second major events and then estimate the local stress drops of the subsources. According to the results in the previous section, the direct S-waves from the first major event (E1) are assumed to be within the time of the arrivals from points 0 and 3. Similarly, the direct S-waves from the second major event (E2) are assumed to be within the time of the arrivals from points 3 and 6. Applying a sine-tapered window (5 percent at each end of the time window) to that portion of the direct S-waves, we obtain the Fourier spectrum $F(f)$. We averaged the spectral amplitudes over a frequency bandwidth of 0.8 Hz following BERRILL (1975), who concluded that a frequency band width of $\sim 1$ Hz was appropriate for this kind of analysis. The power spectral density $P(f)$ is estimated from the square of the Fourier spectrum divided by the duration of the direct S-waves $T_d$, i.e.,

$$P(f) = \frac{|F(f)|^2}{T_d}. \tag{8}$$

We assume that the earthquake source is a point source with the source power spectrum $G(f)$. In order to estimate the spectrum from the observed power spectra at various stations, we assume that the predicted value of a power spectrum of frequency $f_i$ recorded on a single horizontal component at distance $r_j$ is given by

$$\hat{P}(f_i, r_j) = \left(\frac{1}{\sqrt{2}}\right)^2 (\alpha_j)^2 \frac{G(f_i)}{(4\pi \rho \beta^2)^2} \frac{(F_j)^2}{r_j^2} \cdot \exp\left\{ -\frac{2\pi f_i r_j}{Q_\beta(f_i) \beta}\right\}, \tag{9}$$

where $\rho$ is the density, $\beta$ is the shear wave velocity, the factor $(1/\sqrt{2})^2$ accounts for vector partition into two horizontal components of equal magnitude, the factor
accounts for local site effects, the factor \((F_0^s/r_j)^2\) accounts for the radiation pattern and geometrical attenuation of the direct S-waves, and the factor \(\exp \{ -2\pi f_i r_j/Q_\beta(f_i) \beta \}\) accounts for attenuation due to material anelasticities and/or scattering. Introducing new variables, we rewrite (9) as

\[
P(f_i, r_j) = c_j \frac{S(f_i)}{r_j^2} \exp \{ -2\pi f_i r_j/Q_\beta(f_i) \beta \},
\]

where

\[
c_j = (\alpha_0 F_0^s/\alpha_0 F_0^s)^2,
\]

and

\[
S(f_i) = \frac{1}{2} (\alpha_0 F_0^s)^2 \frac{G(f_i)}{(4\pi \rho^s)^2},
\]

\(\alpha_0\) and \(F_0^s\) denote the factors of local site effects and radiation pattern at a reference station, respectively.

Let \(P_{ij}\) denote an observed value of a power spectrum at frequency \(f_i\) and a hypocentral distance of \(r_j\). Let us define factor \(k_{ij}^p\) as follows:

\[
k_{ij}^p = \frac{P_{ij}}{P(f_i, r_j)}.
\]

Following PAPAGEORGIOU and AKI (1983 b), we assume that the distribution of \(\ln (k_{ij}^p)\) is approximately Gaussian. Therefore, for a set of \(m\) discrete sampling frequencies and a set of \(n\) stations (i.e., \(2n\) records, since each station provides two horizontal components of motion), the best estimates of the parameters \(S(f_i), Q_\beta(f_i)\) and \(c_j\) are those that minimize

\[
\Phi = \sum_{i=1}^{m} \sum_{j=1}^{2n} (\ln k_{ij}^p)^2
\]

over the data set, where \((2mn)\) is the number of data entries.

In the present case, we have jointly used the two major events which occurred in close proximity to solve the unknown parameters by assuming that \(Q_\beta(f_i)\) and \(c_j\) are common to these two events. In the joint determination, (14) is replaced by

\[
\Phi = \sum_{i=1}^{m} \sum_{j=1}^{2n} (\ln k_{ij1}^p)^2 + \sum_{i=1}^{m} \sum_{j=1}^{2n} (\ln k_{ij2}^p)^2,
\]

where \(k_{ij1}^p\) and \(k_{ij2}^p\) denote the factors \(k_{ij}^p\) of E1 and E2, respectively, and \((4mn)\) is the number of data entries. The data are given equal weight assuming that the scatter remains approximately the same with respect to variations in distance and frequency. Unknown parameters are \(Q_\beta(f_i), c_j\), and the source factors \(S_1(f_i)\) and \(S_2(f_i)\) of E1 and E2. From (10), it is clear that \(c_j\) can only be determined in relation to the fixed value of a reference station. Therefore, the number of unknown parameters is \(3m+n-1\) for the model of frequency-dependent \(Q_\beta\). Since the station FRF is the nearest to the hypocentral region and the accelerograph there
was installed on a hard rock site, we choose FRF as the reference station and had $c_j$ of FRF fixed at unity. We selected the sampling frequencies of 1, 2, 4, 8, and 16 Hz. Since there are 10 stations and 5 sampling frequencies, there are 200 data entries, and 24 unknown parameters. Distance $r_j$ in (10) for E1 is assumed to be an average of the epicentral distances from points 0 and 3, and the distance $r_j$ for E2 is assumed to be an average of the epicentral distances from points 3 and 6.

First, we shall assume a model of constant $Q_\beta$ and test the accuracy of the data. YAMADA (1984) estimated a $Q_\beta$ of 500–600, using the spectral analysis of aftershocks of the 1983 Japan Sea earthquake. We assumed that $Q_\beta$ was constant and independent of frequency. We assume that $Q_\beta$ is 600 in order to determine

![Fig. 9 (a)](image-url)
Fig. 9. (a) Power spectral densities of strong motion accelerations corrected for the inferred $c_j$ and the fitted curves for a model of constant $Q_p=600$ for the first event (E1). (b) Power spectral densities of strong motion accelerations corrected for the inferred $c_j$ and the fitted curves for a model of constant $Q_p=600$ for the second event (E2).

Fig. 9 (b)

Figures 9 (a) and (b) shows the fitted curves and the observed values of the power spectra which are corrected for the inferred factor $c_j$. Similar to the results obtained by Papageorgiou and Aki (1983 b), the fitted curve overestimates the spectral amplitudes at 1 and 2 Hz and underestimates those at 16 Hz, when referring to large distances. In contrast, at short distances the fitted curve underestimates the spectral amplitudes at 1 and 2 Hz and overestimates...
those at 16 Hz. This implies that $Q_\phi=600$ is too large for frequencies of 1 and 2 Hz and too small for a frequency of 16 Hz, which in turn implies that $Q_\phi$ is frequency-dependent.

Next, we allow $Q_\phi$ to vary freely at each sampling frequency and apply the least-squares method to determine $c_j$, $Q_\phi(f_i)$, $S_1(f_i)$, and $S_2(f_i)$. In Figs. 10 (a) and (b), the observed power spectra corrected for the inferred factor $c_j$ is demonstrated together with the fitted curves. It is clearly shown that the model of frequency-dependent $Q_\phi$ matches well with the observations although there is a large scatter around the fitted curves at 1 Hz for E1 and at 8 Hz for E2. The determined $Q_\phi$ increases roughly in proportion with the frequency, as shown in Fig. 11. In the same figure, we show the $Q$-values estimated from strong motion S-waves by
CONSOLE and ROVELLI (1981) and PAPAGEORGIOU and AKI (1983 b). They show almost the same dependence on frequency, though the coverages of the epicentral distances in the three data sets are different.

If $F^2$ is assumed to be the same for all the stations, the square root of $c_j$ expresses the site amplification coefficient relative to the reference station FRF. The square root of the inferred $c_j$ ranges between 2.4–6.6, as shown in Fig. 12.

Fig. 10. (a) Power spectral densities of strong motion accelerations corrected for the inferred $c_j$ and the fitted curves for a model of frequency-dependent $Q_\beta$ for the first event (E1). (b) Power spectral densities of strong motion accelerations corrected for the inferred $c_j$ and the fitted curves for a model of frequency-dependent $Q_\beta$ for the second event (E2).
Fig. 11. Inverted quality factor of $Q_\beta$ plotted against frequency. Bars indicate the standard errors of the estimates. Also shown are the $Q$-values obtained by Papa-Georgiou and Aki (1983 b) for the San Fernando earthquake of 1971. The straight line indicates the dependence of $Q_\beta$ on the frequency obtained from strong motion records in the Friuli region, Italy (Console and Rovelli, 1981).

Fig. 12. Square root of the inverted $c_j$. Shaded bands indicate the standard deviations of the square root of $c_j$ over the northern stations (MUR-AOM) and the southern stations (NAM-SAK), respectively. FRF is shaded to indicate it is a fixed value.

This range of values is consistent with the scatter of amplification factors for various types of ground obtained by Takemura and Ohta (1983). Therefore, we believe that the inferred factor $c_j$ primarily reflects the effects of site amplification at each station. However, the square roots of $c_j$ at the northern stations are systematically greater than those of the southern stations. This might be explained by the same directivity effect as demonstrated by the long-period body and surface waves (Ben-Menahem, 1961; Hirasawa and Stauder, 1965). The rupture of the 1983 Japan Sea earthquake propagated almost unilaterally from south to north, and thus is expected to enhance the seismic amplitudes at northern stations. Since the focal mechanisms of the two events are estimated to be pure dip-slip with strikes in a N15°-20°E direction (Shimazaki and Mori, 1983; Ishikawa et al., 1984; Sato et al., 1985), it is difficult to explain the regional difference in $c_j$ with the effect of the radiation pattern. The radiation patterns coefficients calculated for the S-wave velocity in Section 4 are on an average almost the same.
between the northern and southern stations. The observed data in (10) showed a much large scatter around the fitted curves for a preliminary inversion when factor $c_j$ was neglected.

The inferred source power spectra $S(f_i)$ is shown in Fig. 13. The source spectra of the two major events are quite similar to each other. This suggests that the local magnitude ($M_L$) and the body-wave magnitude ($m_b$) are probably the same for the two events. The source spectral amplitude is nearly flat at frequencies of 1–4 Hz and decreases sharply beyond a frequency of 4 Hz. Since the effects of attenuation have been removed, the cut-off frequency $f_{\text{max}}$ around 4 Hz can be regarded as a property of the source. However, it remains problematic as to whether this is really a property of source or a property of local site conditions (Hanks, 1982).

The dotted flat lines in the same figure indicate the levels of spectral amplitudes expected for the $\omega$-square source model (Aki, 1967; Brune, 1970). Given the observed moment $M_0^2$, the corner frequency $f_0$ and the duration time of the seismic signal $T_d$, we can express $G(f)$ in (12) in terms of the $\omega$-square model as

$$G(f) = \frac{M_0^2}{T_d} \left\{ \frac{(2\pi f_0)^2}{1 + (f_0 f)^2} \right\}^2. \tag{16}$$

![Figure 13. Inverted source factors $S(f)$ for the first and the second events plotted against frequency. Two flat levels denoted by dashed lines show the power spectral amplitudes expected for the $\omega$-square model.](image-url)
Then the amplitude of the flat part at frequencies higher than the corner frequency is given by

\[ S_0 = \lim_{f \to \infty} S(f) = \frac{1}{2} \left( \frac{\alpha_s F_0^s}{\rho \beta^2} \right)^2 \frac{M_0}{T_a} (2\pi f_0)^4. \]  \hspace{1cm} (17)

Here the moment \( M_0 \) is assumed to be an average value of those estimated by SHIMAZAKI and MORI (1983) and ISHIKAWA et al. (1984). The average moment of \( E_1 \) is about twice as large as that of \( E_2 \). We assume that \( f_0 \approx 1/T_a \) still holds. The values of \( T_a \) are assumed to be 12.2 s for \( E_1 \) and 13.4 s for \( E_2 \). These values are obtained at the stations by averaging the time between the theoretical S-arrivals from points 0 and 3 for \( E_1 \) and from points 3 and 6 for \( E_2 \), respectively. In (17), we assume \( \alpha_s = 2.0 \) and \( (F_0^s)^2 = \langle (F_0^s)^2 \rangle = 0.4 \), where \( \langle (F_0^s)^2 \rangle \) is the expected value of the square of the factor accounting for the S-wave radiation patterns over the focal sphere.

The observed amplitudes of the flat part are greater than the amplitudes predicted for the \( \omega \)-square model. We consider that the spectrum around 1–4 Hz corresponds to the spectral bump observed by GUSEV (1983) for global earthquakes. The ratio of the observed to the predicted amplitudes for \( E_1 \) is about 4 and that for \( E_2 \) is about 10, showing a large difference in the ratios between the two events. This is very important when we compare the source properties of the two events. In spite of the fact that the seismic moment of \( E_2 \) is about half the moment of \( E_1 \), \( E_2 \) radiated almost the same amount of high-frequency seismic energy as \( E_1 \). \( E_2 \) was more efficient than \( E_1 \) in radiating seismic energy at higher frequencies.

According to the HIRASAWA model (1979), the source spectral density \( G(f) \) in (12) is predicted to be

\[ G = (17.8)^2 \cdot (RV) \cdot \psi \cdot E(\tau^2), \] \hspace{1cm} (18)

where \( R \) is the radius of the circular fault, \( V \) is the sweeping velocity, \( \psi \) is the spreading velocity and \( E(\tau^2) \) is the mean-square of the local stress drop. Similarly, the Papageorgiou and Aki model (1983) predicts

\[ G = (17.7)^2 \cdot (WV) \cdot \psi \cdot \Delta \sigma_L^2, \] \hspace{1cm} (19)

where \( W \) is the width of the rectangular fault and \( \Delta \sigma_L \) is the local stress drop which is assumed to be constant, irrespective of different subsources. In both (18) and (19), \( \psi/\beta = 0.75 \) is assumed. We believe that the flat part of the inferred source spectra in Fig. 13 corresponds to the constant amplitude predicted by (18) and (19). Therefore, we take the average of \( G(f_i) \) over the frequencies of 1, 2, and 4 Hz and substitute the average value into (18) and (19) to estimate the local stress drop. \( G(f_i) \) is obtained through (12) from the inferred \( S(f_i) \). We assume that \( \rho = 2.7 \text{ g/cm}^3, \beta = 3.7 \text{ km/s}, V = 2.0 \text{ km/s}, \alpha_s = 2.0 \) and \( (F_0^s)^2 = 0.4 \). The radius \( R \) in (18) is assumed to be given by the equation, \( R = (LW/\pi)^{1/2} \), where \( L \) is assumed to be 30 km and \( W \) 35 km for the two events. Estimated values of the local stress drop are listed in Table 5. In terms of the Hirasawa model, it is 380 bars.
Table 5. Local stress drops $E(\sigma^2)^{1/2}$ and $\sigma_{E_L}$ (bars), radii of subsources $\rho_0$ (km), global stress drops $\sigma$ (bars) and ratios of local to global stress drops for the two major events.

<table>
<thead>
<tr>
<th></th>
<th>$\sqrt{E(\sigma^2)}$</th>
<th>$\sigma_{E_L}$</th>
<th>$\rho_0$</th>
<th>$\sigma$</th>
<th>$\sqrt{E(\sigma^2)/\sigma}$</th>
<th>$\sigma_{E_L}/\sigma$</th>
</tr>
</thead>
<tbody>
<tr>
<td>E1</td>
<td>380</td>
<td>550</td>
<td>9.1</td>
<td>220</td>
<td>1.7</td>
<td>2.5</td>
</tr>
<tr>
<td>E2</td>
<td>340</td>
<td>500</td>
<td>6.7</td>
<td>140</td>
<td>2.4</td>
<td>3.6</td>
</tr>
</tbody>
</table>

$\sigma$ and $\rho_0$ are calculated for the combination of MO.SM and FL.A.

for E1 and 340 bars for E2. In terms of the Papageorgiou and Aki model, it is 550 bars for E1 and 500 bars for E2. These values fall near the upper bound of the local stress drops estimated by AKI (1984) for various earthquakes. If the sweeping velocity is assumed to be greater than 2.0 km/s, for example, 2.5–3.0 km/s, the local stress drops become 0.8–0.9 times smaller than those listed in Table 5.

The estimated values of $E(\sigma^2)^{1/2}$ and $\sigma_{E_L}$ are almost identical for the two events. This agreement results from the same source area and almost the same observed spectral densities for the two major events. The values of $\sigma_{E_L}$ are about 1.4 times larger than those of $E(\sigma^2)^{1/2}$, which is attributed to the different assumptions on the rupture mode between the two stochastic models (SATO, 1985).

According to the Papageorgiou and Aki model, the radius $\rho_0$ of the subsource is obtained from

$$\rho_0 = \frac{7}{4} \frac{1}{\Delta \sigma_L} \frac{M_0}{L W}.$$  \hspace{1cm} (20)

The subsource radii listed in Table 5 are obtained from a combination of the seismic moments, MO.SM in Table 2 and the fault lengths, FL.A in Table 3. They are 9.1 km and 6.7 km for E1 and E2, respectively. The radius for E1 is 1.4 times larger than that for E2, which results from the fact that the moment of E1 is 1.4 times larger than that of E2 while the local stress drops and the source areas are almost the same for the two events. The difference in subsource size obtained from the Papageorgiou and Aki model predicts a difference in the source spectra between the two events at frequencies of around 1.0 to 0.1 Hz. After comparing the source spectra in the frequency range, we could not conclude that there was any such difference as had been predicted from the difference in subsource size. This is due to a large scattering of data. Surface waves generated by E1 probably contaminate the spectra of E2 at lower frequencies and make the discrimination very difficult.

The ratios of the local stress drops to the global stress drops are also listed in Table 5. Here we use the global stress drops estimated for the combination of MO.SM and FL.A. The values range from 1.7 to 3.6. According to IZUTANI (1983), the ratios are between 1 to 5.3 for the Matsusiro swarms and major earthquakes in Japan with a seismic moment of from $10^{28}$ to $2 \times 10^{28}$ dyn-cm. Independent of the stochastic model, the ratio for E2 is 1.4 times greater than that.
for E1, again suggesting E2 was more efficient in radiating high-frequency energy than E1.

Little radiation of high-frequency seismic energy from E3 can be interpreted in terms of the stochastic source models as being due to the paucity of small-scale barriers in the source area of E3. The space-time distribution of aftershocks (TOHOKU UNIVERSITY and HIROSAKI UNIVERSITY, 1984) shows that the aftershock activity in the source area of E3 was much lower than that in the source areas of E1 and E2. According to the stochastic barrier models, high-frequency seismic waves are generated by the presence of barriers, which subsequently act as stress concentrators responsible for aftershocks. This process may explain the interesting correlation between the aftershock activity and the radiation of high-frequency energy during the main shock. Both the low rupture velocity and the paucity of small-scale barriers can be considered causes for the low level of radiation of high-frequency seismic energy from E3.

6. Discussion

If the fault of the main shock has a dimension mapped by the aftershock distribution, E3 might be due to a separate rupture in the area to the north of the change in fault orientation, which was not ruptured during the preceding two events (Model A). This assumption was taken for granted in Section 4. However, there are some other possible processes involved in generating E3. In the following, we shall discuss the plausibility of three models which are schematically illustrated in Fig. 14.

In Model A, the slip distribution of the northern segment was assumed to con-
form independently to the crack model. However, it is possible that the rupture in the northern segment was coalesced with the slip zone of E2 to form a slip distribution expected for a single crack with length equal to the sum of the central and the northern segments (Model B). Although the aftershock activity in the northern segment manifested itself soon after the main shock, it was very low in comparison to those in the southern and central segments (Tohoku University and Hirosaki University, 1984). This suggests that the northern segment had not ruptured during the main shock. If the fault of the main shock is limited to the aftershock area south of the bend in the aftershock distribution, E3 might be explained by the failure of the barrier between the two slip zones which were ruptured during the preceding two major events (Model C). One other explanation for E3 is a mixture of Model A and Model C; E3 is due in part to a new slip created in the northern segment and due in part to the failure of the barrier between the two slip zones for E1 and E2 (Model D).

Let us determine the increase of the moment due to E3 in terms of Model B. Using (6), we then calculate the moment due to the slip which has adapted to a single crack with the combined area of the central and northern segments. Here $R$ is given by $(LW/\pi)^{1/2}$ and the stress drop is assumed to be the same as E2 in Table 4. The moment of E2 in Table 2 is then subtracted from the calculated moment to obtain the moment due to E3, which is listed in Table 6 together with the average slip over the combined area of the central and northern segments. Only the moment predicted for a combination of MO.SM and FL.B agrees with the observed one. For the other cases, the predicted moments largely differ from the observed values.

The increase of the moment due to E3 in terms of Model C can be predicted following Rudnicki et al. (1984), who theoretically evaluated the increase of the moment caused by the failure of the barrier for a two dimensional case. On the assumption that the stress drop is constant over the whole rupture area, the ratio between the sum of moments of two slip zones separated by a barrier and the moment due to the failure of the barrier is determined from two parameters: one is the ratio between the lengths of the two slip zones and the other is the ratio of the barrier interval to the length of either one of the two slip zones. For the two sets of fault length listed in Table 3, the ratio between the lengths of E1 and E2

<table>
<thead>
<tr>
<th>Model</th>
<th>$M_0$ ($\times 10^{27}$ dyn-cm)</th>
<th>$D$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MO.SM+FL.A</td>
<td>5.1</td>
<td>7.9</td>
</tr>
<tr>
<td>MO.SM+FL.B</td>
<td>2.6</td>
<td>5.1</td>
</tr>
<tr>
<td>MO.I+FL.A</td>
<td>1.9</td>
<td>3.0</td>
</tr>
<tr>
<td>MO.I+FL.B</td>
<td>1.0</td>
<td>1.4</td>
</tr>
</tbody>
</table>

These are calculated for combinations of the seismic moments in Table 2 and the fault lengths in Table 3.
Table 7. Seismic moments \( \times 10^{27} \text{ dyn-cm} \) due to the third event predicted for Model C.

<table>
<thead>
<tr>
<th></th>
<th>( e = 1 \text{ km} )</th>
<th>( e = 5 \text{ km} )</th>
<th>( e = 14 \text{ km} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>MO.SM (+FL.A)</td>
<td>3.6</td>
<td>9.3</td>
<td>20</td>
</tr>
<tr>
<td>MO.SM (+FL.B)</td>
<td>3.3</td>
<td>7.5</td>
<td>12</td>
</tr>
<tr>
<td>MO.I (+FL.A)</td>
<td>2.5</td>
<td>6.5</td>
<td>14</td>
</tr>
<tr>
<td>MO.I (+FL.B)</td>
<td>2.3</td>
<td>5.3</td>
<td>8.2</td>
</tr>
</tbody>
</table>

These are calculated for three different barrier length \( e \), for combinations of the seismic moments in Table 2 and the fault lengths in Table 3.

is unity. For MO.SM, the sum of moments of E1 and E2 is \( 5 \times 10^{27} \text{ dyn-cm} \). For MO.I, however, a value of \( 3.5 \times 10^{27} \text{ dyn-cm} \) is used as the sum of moments of E1 and E2 for the reason mentioned earlier, in relation to the estimation of moment due to E3. Table 7 shows the predicted moments due to E3 for three barrier intervals of 1, 5, and 14 km. We see that MO.SM requires that the barrier interval be about 1 km if the observed moment due to E3 is to be interpreted in terms of Model C. Similarly, MO.I requires that the barrier interval be 2-4 km. A barrier interval greater than 5 km predicts a much larger moment due to E3 than the observed one for any combination of moments and fault lengths, and thus cannot be accepted in terms of Model C.

If the moment due to the failure of the barrier equals the observed moment due to E3, the slip over the northern area is not necessary. Therefore, Table 7 shows that Model D is allowed if the barrier interval is much less than 1 km for MO.SM and less than a few kilometers for MO.I.

Although the above arguments narrow the range of possible mechanisms for generating E3, it is difficult to determine the final model for lack of precise estimates of the seismic moments, fault areas and the barrier interval. If we could directly observe the displacement near the boundaries of fault segments as in the case of the El Asnam earthquake of 1980 (YIELDING et al., 1981), we might better determine the best model.

We suppose that the zone separating the two source areas of E1 and E2 was not fractured during the main shock because the aftershock activity in the zone was very low, as shown in Fig. 2. Even if the zone was ruptured, the width of the fault may be small there. In this respect, Models A and B are more likely than Models C and D. Moreover, the fact that the largest aftershock occurred north of the northern segment and no large aftershocks with magnitude greater than 6.0 had occurred in the northern segment as of November, 1985, suggests that the northern segment ruptured during the main shock releasing most of the strain energy in this area.

At present, there is no strong evidence for differentiating between Models A and B. For the El Asnam earthquake, the displacement near the bend of the fault strike appears to be compatible with Model A (YIELDING et al., 1981). It is uncertain whether such a change in fault strike always causes a similar slip.
distribution. In Models A and B, the rupture velocity must decrease someplace near the bend of the fault strike to make up for the larger source process time. Past the bend of the fault strike, the stress concentration at the crack tip for Model B may be larger than that for Model A since the fault length associated with the stress concentration is larger for Model B in the northern segment. A larger stress concentration suggests a higher rupture velocity, so Model A is more consistent with the probable decrease in average rupture velocity beyond the bend of the fault strike than is Model B.

The HYDROGRAPHIC DEPARTMENT (1984) investigated bathymetry, geological structure, geomagnetic total intensity force, and free-air gravity anomaly in the

Fig. 15. Geological structures in the source area (HYDROGRAPHIC DEPARTMENT, 1984) superimposed upon the estimated fault of the main shock. The shaded patches near Kyuroku Island indicate the Tertiary to Quarternary volcanic rocks (GEOLOGICAL SURVEY OF JAPAN, 1981). The solid circle indicates the place where yellow-colored materials were witnessed on the sea floor by Hotta et al. (1984).
source area soon after the occurrence of the main shock (May 27–June 15, 1983). Figure 15 shows the geological structures superimposed on the estimated fault of Fig. 8. The general trend of geological structures change around 40°48'N in correspondence with the bend of the estimated fault. The general trend is perpendicularly intersected by an anticlinal axis to the west of Kyuroku Island. This tectonic line is situated just north of the zone of low aftershock activity and is very close to the epicenter of the second event determined from the strong motion accelerograms.

There is evidence of high volcanic activity in the past and considerable thermal activity at present in the region surrounding Kyuroku Island. Several concentrations of Tertiary and Quaternary volcanic rocks are found around the island, which itself is composed of volcanic rocks (GEOLOGICAL SURVEY OF JAPAN, 1981). Hot springs and sulfur-like materials were found on the sea floor a few kilometers west of Kyuroku Island (FUKUDOME et al., 1984). Also observed on the sea floor about 40 km west-north-west of Kyuroku Island were yellow-colored materials which seem to have originated from hydrothermal activity (HOTTA et al., 1984). Therefore, we infer that the temperature is high in the zone of low aftershock activity and the inactivity of aftershocks might be caused by the ductility of rock materials. The same zone of low aftershock activity probably acted as a barrier during the main shock. For this reason, the barrier is classified as a "ductile" barrier.

The source area of E1 shows positive free-air gravity anomalies of greater than 20 mgals, whereas the source area of E2 shows negative anomalies of less than —10 mgals. The contour line of geomagnetic total intensity force in the southern segment generally extends in a N-S direction parallel with the trend of topography, whereas the contour line in the central segment runs perpendicular to the general trend of topography. The boundary seems to coincide with the zone of low aftershock activity. In the northern segment, the contour line appears to extend again in a N-S direction. The change in direction of the contour line suggests a lateral variation of crustal structure in the source area. Although we cannot fully explain the cause, it is very interesting that the anomalous patterns of the free-air gravity and geomagnetic total intensity force appear to correlate well with the fault segmentation.

Rupture complexity may be due to geometrical complexities of the fault plane and heterogeneities in the fault strength and tectonic stress. These fault properties may be closely related to the crustal structure in the source area. In the case of the Japan Sea earthquake, the geological and geophysical data in the source area seem to coincide with the three-stage strain-release process derived from the seismological data.

7. Conclusions

1. From analysis of strong motion accelerograms, the 1983 Nihonkai-chubu
earthquake was found to have emitted a large amount of high-frequency seismic energy in two stages, forming two high-amplitude envelopes on the accelerograms. As one moves clockwise in azimuth from north to south, the time difference between S-wave arrival-times of the two events becomes larger and the amplitude ratio of the second to the first event becomes smaller. Using the time differences, the second event was located 44 km NNE of the first one with a delay of 26 s in the origin time. The strong motion accelerograms, combined with the aftershock distribution and the source process time of 63 s obtained from long-period surface waves, suggest the following rupture model. The first event initiated at the southern end of the aftershock area and extended in a N15°E direction. It stopped near the zone of low aftershock activity to the west of Kyuroku Island. This zone seems to have acted as a ductile barrier since the temperature in the zone obtained from geological data is estimated to be high. After a pause of about 10 s, the second event started at a place just north of the zone and extended in the same direction. A third event followed the second near the bend of the aftershock distribution and extended in a N15°W direction with a lower rupture velocity. The slow rupture and the paucity of small-scale barriers are both considered causes for the low level of high-frequency radiation from E3.

2. The observed accelerations of the first and the second events were jointly inverted for the source acceleration spectra, the attenuation coefficient of $Q_\text{eff}$, and the amplification factors of the recording sites. The source spectra of the two events are almost the same, both in amplitude and in shape, showing a rapid decay of amplitude at frequencies higher than 4 Hz. The $Q_\text{eff}$ is strongly frequency-dependent and increases approximately in proportion to the frequency of 66 at 1 Hz to 1,026 at 16 Hz. Interpreting the acceleration spectra in terms of the stochastic source models, we estimated the local stress drops of subsources: the r.m.s. local stress drop in terms of the Hirasawa model is 380 bars for the first and 340 bars for the second event: the local stress drop in terms of the Papageorgiou and Aki model is 550 bars for the first and 500 bars for the second event. In terms of the Papageorgiou and Aki model, the second event being more efficient than the first in radiating high-frequency seismic energy can be interpreted as being due to smaller subsource sizes for the second event compared with those of the first event.

3. The three-stage strain-release process derived from the seismological data appears to be supported by other geological and geophysical data in the source area, suggesting that the rupture characteristics during the main shock were closely related to the heterogeneous crustal structure. More observational and theoretical studies along this line may allow us to predict a complex rupture during a large future earthquake.

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