total scalar moment determined at low frequencies? and (2) Does the source time function predict the phase or the centroid time shift observed at low frequencies? The source time function shown in Figure 4 satisfies the first test, and the inversion algorithm was indeed designed to en-
sure this. Figure 5 shows a comparison between observed phase shift and the source time function at low frequencies, obtained from narrow frequency bands of long-period Rayleigh waves, and the same quantity calculated from the Fourier spectrum of the source time function in Fig-
ure 4. The comparison shows that the agreement at very low frequencies is good, with an average centroid time of 20-30 s. The discrepancy at the high end of the spectrum (at 8 Hz) may be due to complexities in the source time function that we are not modeling, or the increasing ex-
traneous noise in the long-period measurements due to lateral variations in phase velocity. Thus, while the source time function is not rigorously accurate, it is useful in making relative comparisons from one time to another.

Fig. 5. Phase shift at low frequencies expressed as centroid time shifts 1c. The solid curve shows the time shifts calculated from the Fourier spectrum of the source time function shown in Figure 6. The open hexagons show the time shifts obtained at different frequencies from the Rayleigh wave observations.

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tion 4650 and IPGP contribution 1269.

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Ruff et al. [1980]. They found that the mechanism of large events had bimodal distribution; either strike-slip along the ridge or subduction along the Australia plate subducting beneath the Pacific plate. Ruff et al. [1980] interpreted this complexity as an indication of the subduction process in the region.

In this paper, we determine the fault parameters of the 1980 Macquarie Ridge earthquake, using seismological surface waves and body waves and compute tsunamis using fault models derived from the seismic data.

Table 1. Centroid moment tensor solution

<table>
<thead>
<tr>
<th>Moment tensor elements (10^12 Nm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mxx</td>
</tr>
<tr>
<td>Myy</td>
</tr>
<tr>
<td>Mzz</td>
</tr>
<tr>
<td>Mxy</td>
</tr>
<tr>
<td>Myz</td>
</tr>
<tr>
<td>Mxz</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>best double couple</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mxx</td>
</tr>
<tr>
<td>ε</td>
</tr>
<tr>
<td>σ (strike)</td>
</tr>
<tr>
<td>σ (dip)</td>
</tr>
<tr>
<td>σ (rake)</td>
</tr>
</tbody>
</table>

Copyright 1990 by the American Geophysical Union.
The motion on this plate is right-lateral. The centroid time is 30 sec, very small for an earthquake of this size, 0.3%. This is not similar to that obtained by Dielwowski et al. [1990].

Body Wave Analysis

The GDSN stations usable for body wave analysis (i.e., distance range of 30° to 100°) are very few, especially in some subducts. We used the records from the GEOSCOPE [Romanowicz, 1984] and IDA [Jafow et al., 1980] networks in conjunction with the GDSN records. By combining the above three network stations, the station coverage becomes very good. We use P and S waves recorded on a broadband channel (one for BHR, VBB, BB, or MI). All the records are first deconvolved with their individual response, convolved with the WWSSN (LP) response, and deconvolved at 1 sec intervals. Thus the waveforms from different networks are directly compared to each other. Figure 2 shows the P and SH waveforms with station locations projected on a map of the United States.

We applied a multideconvolution method of P and SH waves [Kikuchi and Kanamori, 1988]. In this method, the observed seismograms are matched by synthetic seismograms computed for a sequence of subevents distributed on the fault plane. We compute Green's functions for five independent moment tensor elements, and represent the waveform by a linear combination of the Green's functions from subevents. By minimizing the difference between the observed and synthetic seismograms, we determine the moment tensor or mechanism of each subevent as well as their spatial and temporal location.

The point sources are distributed in the strike direction (58°) determined from the CMT solution with an interval of 20 km. The Green's functions are computed from each point source for a simple earth model consisting of a water layer (2 km thick), crust, and mantle. We fix the depth at 10 km, since the true depth resolution of body wave is generally poor and there is a trade-off between depth and moment estimates. The source time function for each subevent is assumed to be a triangle with rise and fall times of 5 sec respectively.

The result of the deconvolution is summarized in Figure 3 and Table 2. Only four major subevents are shown. All the subevents show a strike-slip mechanism similar to that obtained from surface waves. The four subevents are located successively from south to north at 40 km interval and the total extent is about 120 km. Deviation from a double couple is fairly small for all subevents except No. 2. The total time duration for the four subevents is about 30 sec. Apparently the rupture propagated towards the north with a rupture velocity of about 4 km/sec. The sum of the seismic moments of the four subevents is 1.1 x 10^22 Nm, which is slightly smaller than that from surface waves. The difference is probably due to the difference in period between surface and body waves.

Fault Parameters

To estimate the fault parameters such as length L, width W, the average slip D, and static stress drop \( \Delta \sigma \), we need seismological or geodetic information other than the seismic moment, because seismic moment merely gives a product of L, W, D, and the rigidity \( \mu \). The fault length can be estimated to be about 120 km from the subevent distribution obtained in the last section. The aftershock distribution seems to support this estimate too. The aftershocks located by PDE are shown in Figure 4. The epicenters are very scattered off the ridge axis, partly because of the poorly determined location for this remote earthquake. Four larger events are located right on the ridge over a distance of about 100 km. Two events just off the ridge are at the northern and southern extension. If we include them, the aftershock zone may be near 200 km long. In any case, the aftershock zone is abnormally short for the earthquake size.

The fault width is difficult to estimate particularly for a remote earthquake such as this event. Empirical relationships from many large earthquakes show that the aspect ratio (L/W) is roughly constant and approximately 2. If we apply this to the present case, the width is estimated to be 60 km. Since the dip angle is almost 90°, this requires that the fault extends as far deep as 60 km. It is very unlikely that this event cuts the upper mantle at such a depth. The depths of all the aftershocks shown in Figure 1 are constrained at 10 km but they are probably shallow. For large strike-slip earthquakes, such as the 1906 San Francisco earthquake and the 1976 Guatemala earthquake, the aspect ratio is much larger, about 20. Here we assume W=30 km. This depth range appears more reasonable for the depth extent of faulting.

If the fault area is 120 x 10 to 120 x 30 km², the average slip, computed from the seismic moment assuming a rigidity of 4x10¹¹ dynes/cm² (10¹¹ Pa), is 27 to 9 m. The static stress drop for a strike-slip fault can be computed as \( \Delta \sigma = (2\mu\Delta D/LW) \), which gives an estimate of 850 to 50 bars (0.8 to 5.0 x10¹⁰ Pa), significantly higher than the average for intra-plate earthquakes. A strike-slip fault produces vertical uplift and subsidence near the end of the fault. The maximum uplift and subsidence for a fault with D=30 m and W=30 km are about 50 cm.

There seems to be a common misconception that tsunamis are not generated by a strike-slip earthquake. Both observation and theory showed that this is not so. The North Atlantic earthquake of 1969, a strike-slip event, generated observable tsunami [Lynes and Ruff, 1981]. Tsunamis were also recorded from the 1968 San Francisco earthquake. The tide-gauge record at Fort Point (San Francisco) shows a very distinct arrival of tsunami with an amplitude of 10 cm. Ward [1980] and Okal [1980] computed excitation of tsunamis theoretically for several types of mechanisms, using normal mode theory and assuming that the ocean depth is uniform. They showed that the tsunami amplitude for a strike-slip source is about a factor of 3 to 4 smaller than for a dip-slip source.

We modeled and computed the generation and propagation of tsunamis from this earthquake. Using the near-source displacement field computed from the fault parameters as the initial condition, tsunamis are computed by a finite-difference method for the assumed bathymetry. The grid size of the bathymetric data is 5 min. Tsunami waveforms are computed at several positions. Figure 5 shows the computed tsunami waveforms at Sydney, Australia and Wellington, New Zealand. Computed tsunamis arrive at these stations at
The motion on this plane is right-lateral. The centroid time is 30 sec, very small for an earthquake of this size. The mechanism obtained from surface waves is shown. Contour interval for the ocean depth is 1000 m.

The point sources are distributed in the strike direction (S) determined from the CMT solution with an interval of 20 km. The Green's functions are computed from each point source for a simple earth model consisting of a water layer (2 km thick), crust, and mantle. We fit the depth at 10 km to the true depth resolution of body wave is generally poor and there is a trade-off between depth and moment estimates. The source time function for each subevent is assumed to be a triangle with rise and fall times of 5 sec respectively.

The results of the deconvolution is summarized in Figure 3 and Table 2. Only four major subevents are shown. All the subevents show a strike-slip mechanism similar to that obtained from surface waves. The four subevents are located successively from south to north at 40 km interval and the total extent is about 120 km. Deviation from a double couple is fairly small for all subevents except No. 2. The total time duration for the four subevents is about 30 sec. Apparently the rupture propagated towards the north with a rupture velocity of about 4 km/sec. The sum of the seismic moments of the four subevents is $1.1 \times 10^{11}$ Nm, which is slightly smaller than that from surface waves. The difference is probably due to the difference in period between surface and body waves.
Satake and Kanamori: Mechanism and tsunami of 1985 Macquarie earthquake

Reference


RUPTURE PROCESS AND STRESS-DROP OF THE GREAT 1989 MACQUARIE RIDGE EARTHQUAKE

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Abstract. On May 23, 1989, a great (Mw=8.1) earthquake occurred in the Macquarie Ridge complex, south of New Zealand. The earthquake is one of the largest in the Macquarie Ridge since this century, and represents the right-lateral strike-slip component of the motion between the Pacific plate and the Australian plate. Subduction initiation appears to be presently occurring in the Macquarie Ridge complex, and it has been suggested that plate boundary strike-slip earthquakes in a transitional tectonic environment have a high stress-drop. We have investigated the source rupture process of the 1989 earthquake, using teleseismic P and SfS waves. The best point source depth is 12 km below the ocean bottom. The deconvolved source time function is dominated by a single pulse with a large moment release (1.8x1022 Nm) and a short duration (20 s). The moment rate increases slowly in the first 10 s of the rupture process, and is suddenly truncated at 20 s. There is no resolvable directivity to this sharp truncation, which gives the temporal resolution of the data set, means that the spatial location of the sharp truncation is within approximately 50 km of the epicenter. The stress-drop and average displacement may be as high as 370 bars and 36 m. These values are unusually high, and strongly support the suggestion that high stress-drop earthquakes are characteristic for transitional tectonic environments.

Introduction

The Macquarie Ridge complex forms the boundary between the Pacific plate and the Australian plate, and runs from New Zealand southward to the Pacific-Australia-Antarctica triple junction. The deformation is characterized by rapid evolution, and has changed from a tensional environment in the Oligocene, to strike-slip with a compressional component since the Miocene (Melgar et al., 1975). This rapid evolution is the result of the southwest migration of the Pacific-Australia rotation pole, which is located close to the Macquarie Ridge complex (Walker, 1978). The current tectonic setting is oblique convergence with an increasing thrust component going northward from the triple junction, and the topographic expression in an alternation of troughs and rises (Figure 1). Subduction initiation appears to be presently occurring (Ruff et al., 1989), which is reflected by the seismicity in the region. The seismicity in the Macquarie Ridge consists of numerous low-angle thrust earthquakes, but most of the seismic motion is released in a few great strike-slip earthquakes (Figure 1). This combination of thrust earthquakes with a few great strike-slip events is characteristic for the transitional tectonic environment of the Macquarie Ridge complex (see Ruff et al., 1989).

The great Macquarie Ridge earthquake (Mw=8.1) of 23 May 1989 occurred in the central section of the Macquarie Ridge Complex. It is the largest earthquake recorded anywhere in the last decade, and the largest well-recorded earthquake in the Macquarie Ridge region. Its focal mechanism (Figure 2) represents both lateral strike-slip on a plane that strikes in the local direction of the plate boundary, and thrusting on the Australian plate west of a small trench (Figure 1). Two smaller thrust earthquakes (Mw=5.5) have occurred in the same section of the ridge since 1966 (Figure 2). The thrust earthquakes are consistent with westward subduction of the Pacific plate beneath Australia, which agrees with the location of the oceanic trench relative to a bathymetric high in the ocean floor. About 500 km farther to the north, the great 25 May 1981 Mw=7.7 right lateral strike-slip earthquake (Ruff et al., 1989) occurred in the northern section of the ridge. This region is similar to the 1989 region,