Earthquake multiplets in the southeastern Solomon Islands

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The 350-km stretch of the Solomon Islands trench that lies adjacent to the islands of Guadalcanal and San Cristobal has been the site of 10 large shallow earthquakes since 1966: June 15, 1966 doublet ($M_s = 7.7,7.3$), May 20, 21, 1977 quadruplet ($M_s = 6.7,5.7,5.7,5.7$), November 4, 5, 1978 doublet ($M_s = 6.9,7.1$), and two single events on October 23, 1979 ($M_s = 7.1$) and February 7, 1984 ($M_s = 7.7$). Analyses of P-wave first-motions, aftershock distributions, P-waveforms recorded at epicentral distances of 30° to 90° on the WWSSN, and 256 s-period Rayleigh-wave spectra obtained from the IDA and GDSN networks are used to examine the source characteristics of the events. The close temporal association and large size of earthquakes precluded collection of useful seismograms for the latter of the two 1966 events and the second and third events of the 1977 sequence. P-waveforms analyzed show smooth traces that can be modeled with simple trapezoidal source-time functions located at shallow (15–45 km) depths. Focal mechanisms are consistent with oblique subduction of the Indian plate beneath the Pacific plate at about N73°E. Estimates of seismic moment for individual events range from 0.2 to 2.3x10$^{27}$ dyne-cm. Average stress drops of the sequences are estimated to range between 10 and 40 bars. This section of island arc may have ruptured in a similar sequence of earthquakes during the 1930's. Conversion of the cumulative seismic moment to displacement and averaging over 40 years suggests a convergence rate of about 4.5 cm year$^{-1}$, about one half of the rates that have been assessed on the basis of seafloor magnetic lineations.

1. Introduction

The southeastern 350-km stretch of the Solomon Islands trench that lies adjacent to the islands of Guadalcanal and San Cristobal (Fig. 1) has been the site of more than 10 large earthquakes during the last 20 years. An unusual characteristic of large earthquakes in this region is that they commonly occur as multiplets (Lay and Kanamori, 1980): two or more earthquakes of about equal size, spaced closely in space and time (Table 1). The primary purpose of this study is to determine the source characteristics of these large earthquakes, as evidenced from analysis of body and surface waves, and examine the interrelationship

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Fig. 1. Principal plate tectonic features of the Solomon Islands region (adapted from Moore, 1982). The Indian plate underthrusts the Pacific plate in the vicinity of Guadalcanal and San Cristobal at about 10 cm year$^{-1}$ in a northeast direction (Minster et al., 1974). Overriding plates at subduction boundaries marked by hatches. Double lines indicate spreading centers.
of the events as a function of their respective size, time of occurrence, location, and source complexity. The epicentral distribution and aftershock zones of the events are illustrated in Fig. 2.

A number of studies are published concerning the spatial distribution of seismicity (Denham, 1969; Curtis, 1973a; Pascal, 1979) the stress distribution as determined from focal mechanisms (Johnson and Molnar, 1972; Curtis, 1973b; Pas-

![Fig. 2. Epicenters of the large earthquakes examined in this study and the distribution of \( M_b \geq 4.5 \) aftershocks that occurred during the 2 days following each event. Aftershock activity 1 to 2 days subsequent to each sequence decreases rapidly toward background levels of seismicity.](image)

<table>
<thead>
<tr>
<th>Event</th>
<th>Data</th>
<th>Origin time</th>
<th>Location</th>
<th>( M_b )</th>
<th>( \Delta t ) (^a)</th>
<th>( \Delta x ) (^b)</th>
</tr>
</thead>
<tbody>
<tr>
<td>E66-1</td>
<td>June 15, 1966</td>
<td>00 h 59 m 46s</td>
<td>-10.4</td>
<td>160.8</td>
<td>7.5</td>
<td>01 h 33 m 08 s</td>
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<tr>
<td>E66-1</td>
<td>June 15, 1966</td>
<td>01 h 33 m 08 s</td>
<td>-10.2</td>
<td>160.9</td>
<td>7.25</td>
<td></td>
</tr>
<tr>
<td>E77-1</td>
<td>April 20, 1977</td>
<td>23 h 13 m 10 s</td>
<td>-9.81</td>
<td>160.21</td>
<td>6.7</td>
<td>29 m 40 s</td>
</tr>
<tr>
<td>E77-2</td>
<td>April 20, 1977</td>
<td>23 h 42 m 50 s</td>
<td>-9.89</td>
<td>160.35</td>
<td>7.5</td>
<td>6 m 23 s</td>
</tr>
<tr>
<td>E77-3</td>
<td>April 20, 1977</td>
<td>23 h 49 m 13 s</td>
<td>-9.84</td>
<td>160.82</td>
<td>7.5</td>
<td>4 h 34 m 57 s</td>
</tr>
<tr>
<td>E77-4</td>
<td>April 21, 1977</td>
<td>04 h 24 m 10 s</td>
<td>-9.97</td>
<td>160.73</td>
<td>7.5</td>
<td></td>
</tr>
<tr>
<td>E78-1</td>
<td>Nov. 4, 1978</td>
<td>22 h 29 m 22 s</td>
<td>-11.23</td>
<td>162.18</td>
<td>6.9</td>
<td>23 h 32 m 45 s</td>
</tr>
<tr>
<td>E78-2</td>
<td>Nov. 5, 1978</td>
<td>22 h 02 m 07 s</td>
<td>-11.30</td>
<td>162.13</td>
<td>7.1</td>
<td></td>
</tr>
<tr>
<td>E79-1</td>
<td>Oct. 23, 1979</td>
<td>08 h 19 m 01 s</td>
<td>-10.60</td>
<td>161.31</td>
<td>6.2</td>
<td>1 h 32 m 5 s</td>
</tr>
<tr>
<td>E79-2</td>
<td>Oct. 23, 1979</td>
<td>09 h 51 m 07 s</td>
<td>-10.62</td>
<td>161.3</td>
<td>7.1</td>
<td></td>
</tr>
<tr>
<td>E84-1</td>
<td>Feb. 7, 1984</td>
<td>21 h 33 m 21 s</td>
<td>-10.01</td>
<td>160.47</td>
<td>7.7</td>
<td></td>
</tr>
</tbody>
</table>

\(^a\) Time to occurrence of subsequent event.

\(^b\) Epicentral offset to subsequent event.

and plate tectonic evolution of the region encompassing the Solomon Islands (Dewey and Bird, 1970; Karig, 1972; Johnson and Molnar, 1972; Curtis, 1973b). The regional plate tectonic setting is illustrated in Fig. 1. Seismicity in the region forms a zone that trends northwest, parallel to the trend of the island arc (e.g., Denham, 1969). Cross-sections of seismicity perpendicular to the trend of the arc show that earthquakes occur to depths of 150 km, defining a Benioff–Wadati zone that dips to the northeast (e.g., Curtis, 1973a). The geometry is consistent with analyses of seafloor magnetic lineations and focal mechanisms that indicate subduction of the Indian plate beneath the Pacific plate has averaged about 10 cm year\(^{-1}\) during the last \( \sim 10 \) Ma, at an azimuth of about N80\(^\circ\)E (e.g., Chase, 1971; Johnson and Molnar, 1972; Minster et al., 1974). The southeastern Solomon Islands thus appear to mark a relatively simple segment of the plate boundary between the Indian and Pacific plates.

2. Data and methodology

The data set used in this study includes earthquake epicenters reported in the EDR catalog, long-period P-waveforms recorded on the Worldwide Standardized Seismograph Network (WWSSN), and vertical long-period seismograms registered by the International Deployment of
Accelerographs (IDA) and the Global Digital Seismograph Network (GDSN). A general approach is taken whereby the epicenters of the mainshocks and aftershocks are used initially to examine the spatial distribution of the earthquake rupture zones. P-wave first-motions and long-period Rayleigh-wave data are then inverted to obtain an estimate of the focal mechanism and seismic moment of each event. Finally, the P-waveforms are modeled with synthetic seismograms to further determine the depth, source-time history, and the seismic moment of each event at shorter periods.

The methods used initially to determine the seismic moment and focal mechanism of each event from the first-motion and Rayleigh-wave data is exactly the inversion technique described by Tanimoto and Kanamori (1986) which, in turn, is based largely on the work of Kanamori and Given (1981). The amplitude and phase spectra of 256 s period Rayleigh wave phases R1 through R3, normalized to the source, are retrieved from the IDA and GDSN seismograms in the manner outlined by Kanamori and Given (1981). Phases contaminated by noise or nonlinear transients are discarded. Each measurement of the surface-wave spectrum $V_r$ is expressed

$$A_1 M_{xx} + A_2 (-M_{xx} + M_{yy}) + A_3 (M_{xx} + M_{yy}) + A_4 M_{yz} + A_5 M_{xz} = V_r$$

(1)

where the matrix $A$ is defined in equation 7 of Kanamori and Given (1981) and $M_{ij}$ are elements of the seismic moment tensor $M$ and defined according to the conventions used by Kanamori and Given (1981). A pure shear dislocation is assumed for the earthquake source and, hence, the trace of the seismic moment tensor is defined to equal zero. For $n$ observations, eq. 1 may be written in matrix form

$$AM = V$$

(2)

where

$$M = \left( M_{xy}, -M_{xx} + M_{yy}, M_{xx} + M_{yy}, M_{yz}, M_{xz} \right) = M_i, i = 1, \ldots, 5$$

(3)

In turn, the far-field P-wave displacement may be expressed as $C'M$, where

$$C' = \left( \sin^2 i \sin 2\phi, -\frac{1}{2} \sin^2 i \cos^2 \phi, \frac{1}{2} (1 - 3 \cos^2 i), -\sin 2i \sin \phi, -\sin 2i \cos \phi \right)$$

and $\phi$ and $i$ denote the azimuth and take-off angle of the P wave from the source, respectively (Strelitz, 1978; Fitch et al., 1980; Kanamori, 1983). The product $C'M$ is $>0$, $<0$, and $0$ for compressional, dilatational, and nodal P-wave arrivals, respectively. To simultaneously invert the combined data set of first-motion and Rayleigh-wave spectral observations for each event for focal mechanism and seismic moment, we used the linear programming approach and algorithm presented by Tanimoto and Kanamori (1985) to minimize

$$\sum_i \sum_{j=1}^{5} \left| (A_{ij} M_j - V_i) \right|$$

(4)

with the constraint that

$$\sum_i C_i M_i > 0,$$

(5)

where, for computational purposes, observed dilatational first-motions are multiplied by $-1$. Underlying this method is the assumption that the long-period and short-period mechanism of earthquakes is the same. Similar inversion procedures that utilize only long-period surface-wave data are ill conditioned for shallow earthquakes, due to the indeterminacy of the $M_{xx}$ and $M_{xy}$ components of the moment tensor (e.g., Kanamori and Given, 1981). Incorporation of P-wave first-motion data in this way works to remove that indeterminacy and yields a simple and relatively rapid approach for estimating the seismic moment and focal mechanism of large earthquakes. For convenience in later discussion of individual earthquakes, this method is referred to here as the combined inversion.

Synthetic P-waveforms are computed for each event and matched to the observed waveforms to provide further information concerning the depths, source-time histories, and seismic moments of the earthquakes at shorter periods. The synthetic seismograms are constructed according to the
method detailed in Langston and Helmberger (1975). The primary factors controlling the shape of the synthetic waveforms are focal mechanism, source-time history, focal depth, attenuation, and the velocity–density structure. Complications due to upper mantle structure are avoided by considering only P waves recorded at distances from about 30° to 90°. To calculate earthquake depths and the reflection coefficients for surface bounces of depth phases, we assume that P- and S-wave velocities are equal to 6.4 km s⁻¹ and 3.7 km s⁻¹, respectively. Take-off angles and seismic moments \( M_0 \) are computed assuming that the density \( \rho \) is 3.2 g cm⁻³, and that P- and S-wave velocities in the source region are 7.8 km s⁻¹ and 4.4 km s⁻¹, respectively. The values of 1.0 s and 4.0 s are used for \( t^* \) in Futtermans (1962) attenuation operator for P and S waves, respectively. A forward-modeling approach is used to fit the synthetic seismograms to those observed. The initial mechanism assumed is that found from the initial inversion of first-motion and Rayleigh-wave spectral observations. The source-time function is taken to be of trapezoidal shape, as illustrated in Fig. 3. The forward modeling approach generally allows constraint of the earthquake rupture duration \( \tau_c \) to within about 2 s (Kanamori and Stewart, 1976;)

Fig. 3. Observed and synthetic P-wave seismograms and P-wave first-motions for event E66-1. The synthetics (lower traces) correspond to the focal mechanism and depth shown. The average \( M_0 \) is \( 5.4 \times 10^{26} \) dyn- cm. The source-time function used for synthesis of P-waves for this, and other events in this study, is characterized by the rise time \( \tau \) and rupture duration \( \tau_c \). In this case, \( \tau = 3 \) s and \( \tau_c = 7 \) s. Also, average values of \( M_0 \) are computed with the values of \( M_0 \) obtained for non-nodal stations within a distance range of about 30 to 90°, and the estimates of \( M_0 \) at the individual non-nodal stations are shown adjacent to the seismograms in units of \( 10^{26} \) dyn- cm.

<table>
<thead>
<tr>
<th>Event</th>
<th>( \phi^0 )</th>
<th>( \delta^0 )</th>
<th>( \lambda^0 )</th>
<th>( M_0 ) a ( 10^{26} ) (dyn- cm)</th>
<th>( M_0 ) b ( 10^{26} ) (dyn- cm)</th>
<th>( \sigma )</th>
<th>#</th>
<th>( h ) (km)</th>
<th>( \tau, \tau_c ) (s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>E66-1</td>
<td>223</td>
<td>84</td>
<td>90</td>
<td>–</td>
<td>–</td>
<td>5.4</td>
<td>2.1</td>
<td>4</td>
<td>38</td>
</tr>
<tr>
<td>E66-2</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>E77-1</td>
<td>145</td>
<td>67</td>
<td>90</td>
<td>–</td>
<td>–</td>
<td>2.0</td>
<td>1.2</td>
<td>6</td>
<td>18</td>
</tr>
<tr>
<td>E77-2</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
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<td>–</td>
<td>–</td>
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<tr>
<td>E77-4</td>
<td>160</td>
<td>50</td>
<td>100</td>
<td>23.2</td>
<td>4.2</td>
<td>0.2</td>
<td>2</td>
<td>45</td>
<td>2.8, 5.5</td>
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<tr>
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<td>164</td>
<td>73</td>
<td>139</td>
<td>5.1</td>
<td>6.0</td>
<td>3.4</td>
<td>7</td>
<td>15</td>
<td>4.8 c</td>
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<tr>
<td>E78-2</td>
<td>160</td>
<td>78</td>
<td>118</td>
<td>6.4</td>
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<td>3.8</td>
<td>11</td>
<td>15</td>
<td>5.5, 9</td>
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<tr>
<td>E79-1</td>
<td>165</td>
<td>65</td>
<td>90</td>
<td>–</td>
<td>0.3</td>
<td>0.05</td>
<td>8</td>
<td>15</td>
<td>2.4 d</td>
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<td>75</td>
<td>100</td>
<td>6.2</td>
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<td>–</td>
<td>–</td>
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| a Surface-wave moment. |
| b Body-wave moment. |
| c Complex source-time function. See Fig. 9. |
| d Complex source-time function. See Fig. 12. |
TABLE III

<table>
<thead>
<tr>
<th>Event</th>
<th>$\phi^0$</th>
<th>$\delta^0$</th>
<th>$\lambda^0$</th>
<th>$M_0 \times 10^{26}$ (dyne-cm)</th>
<th>$h$ (km)</th>
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<td>E77-1</td>
<td>146</td>
<td>71</td>
<td>116</td>
<td>1.9</td>
<td>11</td>
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<tr>
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<td>67</td>
<td>102</td>
<td>9.0</td>
<td>16</td>
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<td>77</td>
<td>5.0</td>
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<td>159</td>
<td>55</td>
<td>97</td>
<td>12.0</td>
<td>40</td>
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<td>73</td>
<td>119</td>
<td>3.6</td>
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<td>E84-1</td>
<td>182</td>
<td>72</td>
<td>124</td>
<td>25.2</td>
<td>22</td>
</tr>
</tbody>
</table>

* Dziewonski et al. (1985); Giardini et al. (1985).

and, hence, surface-wave data for these two events are not analyzed. Also, P-waveforms of the latter event of the doublet are lost in the coda of the initial event. Hence, analysis of the doublet is limited to study of the first-motions and P-waveforms resulting from the first event of this sequence.

A thrust-type focal mechanism is indicated for event E66-1 by the distribution of P-wave first-motions, though only the orientation of one of the nodal planes is constrained well (Fig. 3). Complete P-waveforms were recovered from the WWSSN library for five stations. Synthetic seismograms computed at those five stations are compared to the observed waveforms in Fig. 3. The synthetics are calculated for a simple source time function situated at 38 km depth and characterized by a $M_0 = 5.4 \times 10^{26}$ dyne-cm. The first 20–25 s of the observed waveforms are generally matched well by the synthetics. The waveforms observed for this and other events in this study, however, also commonly show significant energy arriving during the subsequent 60 s of the record. Lay and Okal (1983) discussed similar P-waveforms produced by earthquakes in the Gilbert Islands. They showed that the later arriving energy likely results from reverberations in a layered crust and overlying water layer. Figure 4 is reproduced from Lay and

Lay and Kanamori, 1980). Although a trade-off exists between the effect of assumed focal depth and source-time function on the shape of synthetic seismograms, the ambiguity is reduced significantly by modeling waveforms at varying azimuths for each event (Astiz and Kanamori, 1984).

The results of the data analysis are summarized in Table II for each event, and further discussed below. Centroid-moment solutions (e.g., Dziewonski et al., 1981) are now routinely reported in the Monthly Listings of the National Earthquake Information Service (NEIS) and the *Physics of the Earth and Planetary Interniors* (e.g., Dziewonski et al., 1985). Previously published centroid-moment solutions for the events in this study are also, for reference, listed in Table III.

3. Earthquake sequences

3.1. 1966 doublet

The earliest earthquake sequence considered is the doublet that occurred June 15, 1966; the two shocks were separated in time and distance by about 1.5 h and 25 km, respectively (events E66-1 and E66-2 in Table I). Aftershocks, during the 2 days subsequent to the latter shock, span an area of about $65 \times 40$ km in the region offshore and between the islands of Guadalupe and San Cristobal Islands (Fig. 2). Aftershocks are generally absent within the region directly adjacent to the mainshock epicenters. The doublet occurred prior to installation of the IDA and GDSN network.

Fig. 4. Comparison from Lay and Okal (1983) of WWSSN P-waveforms (top trace), synthetics including reverberations in a layered crust and water layer (middle trace), and half-space synthetics (bottom trace) for two Gilbert Island earthquakes that occurred in 1982. The layered structure calculations provide a more detailed match to the entire waveform, but the first 15 s are well matched by the half-space synthetics as well.
Okal (1983), and compares P-waveforms of two Gilbert Island earthquakes to synthetics computed for a half-space and layered model, respectively. Although inclusion of crustal and water-layer reverberations yields a superior overall fit to the observed waveforms in Fig. 4, there is little difference in the first 15 s or so of the observed waveforms, which indicates the adequacy of half-space modeling for constraining earthquake mechanisms. With respect to event E66-1 then, the synthetic waveforms are not sensitive to changes in rakes of the assumed mechanism. Thus, neither the first-motions nor modeling of P-waveforms eliminate a possibly large component of strike-slip movement for this event.

3.2. 1977 quadruplet

The 1977 sequence occurred during a span of about 5 h (Table I). Surface-wave magnitudes for the events range from 6.7 to 7.5. Aftershocks during the 2 days following the last shock form a distinct two-lobed pattern along the coast of eastern and central Guadalcanal (Fig. 2). Epicenters of the first two shocks (E77-1 and E77-2) are located 15 km apart, and each is situated at the eastern edge of the western lobe of aftershock activity. The time separating the occurrence of events E77-1 and E77-2 was about 30 min. The latter two shocks of the sequence (E77-3 and E77-4) are also located about 15 km apart, but are separated to the east by about 45 km from the first two events of the sequence. Epicenters of shocks E77-3 and E77-4 locate within and at the northern edge of the eastern lobe of aftershock activity. The time-delay between events E77-3 and E77-2 was only about 6 min as compared to the 4.5 h separating event E77-4 and E77-3. Hence, the shortest delay in time between consecutive events in this sequence is associated with the largest spatial jump between consecutive epicenters.

The eastern and western lobes of aftershock activity are about 55 x 25 km² and 50 x 50 km², respectively. If it is assumed that the aftershock zones represent the approximate extent of rupture during this sequence, it appears that the region between the two lobes of aftershock activity re-
mained unbroken until 1984 (Fig. 2). One may also infer that events E77-1 and E77-2 triggered a westward and unilateral rupture of the zone marked by the western lobe of aftershock activity, based on the relative locations of the two shocks with respect to the aftershock distribution. Likewise, the latter two shocks may have triggered a southward and unilateral rupture of the zone marked by the eastern lobe of aftershock activity. It is convenient to discuss each event separately.

3.2.1. E77-1

Rayleigh-wave phases subsequent to R1 for event E77-1 are contaminated by the occurrence 30 min later of event E77-2. The limited set of R1 spectra are also insufficient to yield a meaningful inversion of the event mechanism and moment. Thus, analysis of the event is limited to study of P-wave first-motions and P-waveform analysis. Numerous first-motions recorded on the WWSSN yield reasonable limits on the orientations of a southwest dipping nodal plane, and indicate that the event produced thrust motion (Fig. 5). P-waveforms suitable for modeling were recovered for six stations and are compared to synthetic waveforms in Fig. 5. The synthetics are computed.

Fig. 5. Observed (upper) and synthetic (lower) P-wave seismograms and P-wave first-motions for event E77-1. The synthetics correspond to the focal mechanism, depth and source-time characteristics shown. The average M₀ is 2.0 x 10^26 dyne-cm.
for the pure thrust mechanism illustrated in Fig. 5, and a simple source-time function characterized by \( \tau_c = 6 \) s, located at 18 km depth with a \( M_0 = 2.0 \times 10^{26} \) dyne-cm. The waveforms examined are generally insensitive to changes in rake of the assumed mechanism. Thus, neither the first-motion nor waveform data exclude potentially large amounts of strike-slip movement.

3.2.2. E77-2 and E77-3

An interval of about 6 min separates the second and third events of the quadruplet. The combination of the large size resulting in driving signals off scale and signal interference resulting from preceding events prevented the recovery of either first-motions or P-waveforms from the WWSSN library. Also, interference between surface waves from events E77-2 and E77-3 prevented retrieval of phase and amplitude spectra that are useful to invert for source parameters. Maximum amplitudes of seismograms for the earthquakes are generally similar to amplitudes registered for the subsequent event E77-4, which is in accord with the observation that \( M_s \) for each event is reported to equal 7.5 (Table I). A sample seismogram for this sequence recorded at station ANMO is shown in Fig. 6.

![ANMO.LPZ seismogram](image)

**Fig. 6.** Vertical long-period seismograms at ANMO (\( \Delta = 98^\circ \), Az = 56°) for the 1977 quadruplet. \( R_1^3, R_2^3, R_3^3, \) and \( R_4^3 \) indicate location of \( R_1 \) phases for events E77-1 to E77-4, respectively. \( O_2 \) and \( O_4 \) are the origin times of the second and fourth events of the sequence (Table I), respectively. Interference between \( R_1^3 \) and \( R_3^3 \) phases is registered in upper trace. Maximum amplitudes of the \( R_4^4 \) and the \( R_1^3 \) and \( R_3^3 \) phases are generally similar, but vary in relative amplitude by a factor of ±2 from station to station.

3.2.3. E77-4

The combined inversion of first-motion and Rayleigh-wave spectra, and modeling of P-waveforms indicate motion during this event was thrusting, as shown by the focal mechanism in Fig. 7a. The theoretical amplitude and phase spectra for 256-s period Rayleigh waves for this mechanism are compared to the observations in Fig. 7b. The phase data show the event is predominantly thrust, although scatter in the amplitude spectra introduces an uncertainty of 10° or more to the strike of the nodal planes. P-waveforms recorded on-scale at two WWSSN stations are compared to synthetics in Fig. 7a. The two ob-

![Observed and synthetic P-wave seismograms and P-wave first-motions for event E77-4](image)

**Fig. 7.** Observed and synthetic P-wave seismograms and P-wave first-motions for event E77-4 are shown in (a). The synthetics (lower traces) correspond to the focal mechanism, depth and source-time function shown. The average \( M_0 \) is \( 4.2 \times 10^{26} \) dyne-cm. Observed 256 s period Rayleigh-wave amplitude (circles) and phase (triangles) spectra are compared to theoretical values in (b) for a source with the mechanism shown in (a) and a \( M_0 = 2.3 \times 10^{27} \) dyne-cm.
served waveforms are reasonably reproduced by a source located at 45 km depth that is characterized by \( \tau_c = 5.5 \) s and a \( M_0 = 4.2 \times 10^{26} \) dyne-cm. Inversion of the surface-wave data yields a significantly larger estimate of seismic moment equal to \( 2.3 \times 10^{27} \) dyne-cm.

### 3.3. 1978 doublet

The aftershocks of the 1978 doublet span a region of about \( 60 \times 40 \) km, and represent the most southeastern of the rupture sequences considered in this study (Fig. 2). The two main shocks occurred almost exactly 1 day apart (Table I) and were located within about 10 km of each other, near the western edge of the aftershock distribution.

#### 3.3.1. E78-1

The P-wave first-motion observations and the mechanism determined from the combined inversion of first-motion data and 256 s period Rayleigh-wave spectra are displayed in Fig. 8. The theoretical amplitude and phase spectra for this mechanism and a \( M_0 = 5.1 \times 10^{26} \) dyne-cm are compared to the observed data in Fig. 9a. The large component of strike slip associated with the thrust mechanism is indicated by the azimuthal pattern of the surface-wave spectra. P-waveforms observed for the event are complex and are modeled with a complex source-time function that consists of two subevents (Fig. 8). The seismic moment of the first subevent is 10% that of the second. The waveforms are best reproduced when the source is 15 km deep and assigned a \( M_0 = 6.0 \times 10^{26} \) dyne-cm. The synthetics best match the data at stations eastward of the epicenter. The observed waveforms at the non-nodal stations TAU and MUN located south and westward of the epicenter show a shorter period character than is observed for stations to the east. This discrepancy may reflect directivity during the rupture process.

#### 3.3.2. E78-2

The P-wave first-motions and the focal mechanism determined from the combined inversion of the first-motion data and 256 s period Rayleigh-wave spectra are shown in Fig. 10. The theoretical amplitude and phase spectra for 256 s period Rayleigh waves for this mechanism and a \( M_0 = 6.4 \times 10^{26} \) dyne-cm are compared to the observed data in Fig. 9b. The large component of strike-slip characterizing the thrust mechanism is required by the surface-wave observations. In contrast to the initial event of this sequence, observed P-waveforms are reproduced with a simple source-time function characterized by a duration \( \tau_c = 9 \) s, \( M_0 = 8.5 \times 10^{26} \) dyne-cm, and located at 15 km depth.

### 3.4. 1979 doublet

The 1979 sequence is not a doublet in the true sense, since the magnitudes of the events differ by an order of magnitude (Tables I and II). The first
shock E79-1 may be considered a foreshock to the subsequent larger event E79-2. Both events are considered here for completeness. The mainshock epicenters and consequent aftershocks are located offshore of northwestern San Cristobal. Aftershocks of the sequence are spread over a region encompassing 100 × 100 km², although most are concentrated in a smaller area of 40 × 40 km² that abuts and partially overlaps the 1966 aftershock sequence (Fig. 2).

3.4.1. E79-1

The size of this shock was insufficient to produce a useful set of Rayleigh phases from the IDA and GDSN networks. A thrust mechanism is indicated for the event by the distribution of P-wave first-motions. The observed waveforms are reproduced well by a source characterized by the focal mechanism in Fig. 11, located at 15 km depth, with a $M_0 = 0.3 \times 10^{26}$ and a complex source-time function consisting of two triangular sources separated by 4 s. Neither the first-motions nor waveform data are sufficient to rule out a significant strike-slip component of movement like that found for the following event E79-2.

3.4.2. E79-2

The combined inversion of first-motion data
and 256-s Rayleigh-wave spectra for event E79-2 indicate the event $M_0 = 6.2 \times 10^{26}$ dyne-cm, and a fault strike, dip, and rake equal to $159^\circ$, $96^\circ$, and $67^\circ$, respectively. The theoretical Rayleigh-wave amplitude and phase spectra for this mechanism are compared to available observations in Fig. 12. However, a slightly different mechanism is required to best match synthetic P-waveforms to observed data (Fig. 13). Thus, synthetics in Fig. 13 are computed for the mechanism shown, a seismic moment $M_0 = 5.6 \times 10^{26}$ dyne-cm, and a simple source-time function of duration $\tau_c = 10$ s located at 25 km depth.

Fig. 11. Observed and synthetic P-wave seismograms and P-wave first-motion data for event E79-1. The synthetics (lower traces) correspond to the focal mechanism, depth and source-time function shown. The average $M_0$ is $0.3 \times 10^{26}$ dyne-cm.

Fig. 12. Observed 256 s period Rayleigh-wave amplitude (circles) and phase spectra (triangles) compared to the theoretical predictions (lines) for event E79-2. Theoretical values are computed for $M_0 = 6.2 \times 10^{26}$ dyne-cm and fault strike, dip, and rake = $159^\circ$, $67^\circ$, and $97^\circ$, respectively.

Fig. 13. Observed (upper) and synthetic (lower) P-wave seismograms and P-wave first-motion for event E79-2. The synthetics correspond to the focal mechanism, depth and source-time function shown. The average $M_0$ is $5.6 \times 10^{26}$ dyne-cm.
3.5. 1984 event

The epicenter of the 1984 event is located directly between the two lobes of aftershock activity produced by the 1977 quadruplet (Fig. 2). The shock may represent rupture of a patch of plate boundary within the 1977 rupture zone that did not rupture in 1977. The distribution of aftershocks is consistent with this interpretation, although the number of aftershocks are too few to draw any strong conclusions.

The WWSSN film chip library is now incomplete for this event, due to the relatively recent origin time. The few seismograms collected provided P-wave first-motion information, but waveforms were driven off scale and were not suitable for modeling. Hence, analysis of the event is limited to study of P-wave first-motions and observations of Rayleigh-wave spectra. Results of the combined inversion of first-motions and 256 s period Rayleigh-wave spectra are illustrated in Fig. 14. The abundant spectral observations indicate that displacement during the event was thrusting with a large component of strike-slip motion. The seismic moment determined from the surface-wave spectra is $3.8 \times 10^{27}$ dyne-cm.

4. Discussion

The P-waveforms of large southeastern Solomon Islands earthquakes generally show smooth traces that can be modeled with simple source-time functions, ranging in duration $\tau_c$ from about 6 to 10 s, located at shallow depths between 15 and 45 km. Similar P-waveforms have been produced by shallow subduction zone earthquakes in other parts of the world which, in contrast, do not necessarily occur as doublets or multiplets (e.g., Chael and Stewart, 1982; Astiz et al., 1986). Thus, P-waveforms register no obvious clue to the doublet behavior, at least at the longer periods considered here.

Estimates of coseismic slip $D$, stress drop $\Delta \sigma$, and fault dimensions $A$ for individual events that occur in multiplets are problematic. Seismograms for events are commonly contaminated by the coda of a preceding event. Similarly, one is generally limited to examining the aftershock distribution of the sequence, rather than the individual earthquakes. We assume here that the source dimensions $A$ of the earthquake sequences are equal to the mapped dimensions of the aftershock zones shown in Fig. 2. For those sequences where analysis of seismograms was limited to a single event, it is assumed that the seismic moment $M_0$ of the event not studied is equal to the event that was examined. This is reasonable in light of the similar values of $M_s$ reported for events in the respective
multiplets (Table I). The average coseismic slip $\bar{D}$ resulting from an earthquake multiplet can then be defined to equal $\Sigma M_0/\mu A$, where $\Sigma M_0$ is the moment sum of the sequence, $A$ is the fault area, and $\mu$, the crustal rigidity, is here assigned the value of $5 \times 10^{11}$ dyne cm$^{-2}$. Average stress drops $\Delta \sigma$ for earthquakes are commonly estimated with expressions for a circular crack to equal $(7\pi \mu/16)/(\bar{D}/r)$ (Kanamori and Anderson, 1975).

We here define $r$ to equal the radius of a circle of area equal to the aftershock zone area. Seismic moments, both determined and assumed, of the earthquakes that compose each sequence and the areal dimensions of aftershock zones are listed in Table IV. Calculations indicate average coseismic displacements range from 0.8 to 2.0 m and average stress drops between 18 to 50 bars for the separate sequences (Table IV), except for the E77-3 and E77-4 event pair where the assumptions result in sharply greater estimates of coseismic offset and stress drop equal to 6.7 m and 220 bars, respectively. It is of interest to note that epicenters of events E77-3 and E77-4 are located most distant from the trench axis (Figs. 1 and 2), and that body-wave analysis shows event E77-4 to be the deepest (45 km) studied. Related to this observation, Seno et al. (1980), observed that the 1978 Miyagi-Oki, Japan earthquake of 1978 occurred relatively deeper along the plate interface of the Japan subduction zone, and that the event was characterized by a relatively high stress drop over 100 bars. Similar observations were made by Fukao and Furumoto (1979) for the 1958 Etorofu earthquake with respect to other large and adjacent subduction zone earthquakes along the Kurile trench. Withstanding the many assumptions and uncertainties in stress drop estimates, the few observations provide some suggestion that stress drop for subduction zone earthquakes correlates with depth of the events along the plate interface.

Depths and focal mechanisms of the events analyzed are summarized in Fig. 15. Depths of the initial events in the 1977 and 1979 multiplets are located shallower than subsequent events in the respective sequences, and the 1978 earthquakes appear to have initiated at the same depth. Source depths range from 15 to 45 km. Focal mechanisms of events E66-1, E77-1, E77-4, and E79-1 show predominantly thrust displacement, although the easterly nodal plane is constrained by data only for event E77-4. Data place limits on the orientation of both nodal planes for events E78-1, E78-2,

<table>
<thead>
<tr>
<th>Event</th>
<th>$M_0$ $10^{26}$ (dyne-cm)</th>
<th>Length (km)</th>
<th>Width (km)</th>
<th>$r$ (km)</th>
<th>$\bar{D}$ (m)</th>
<th>$\Delta \sigma$ (bar)</th>
</tr>
</thead>
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<tr>
<td>E66-1</td>
<td>5.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E66-2</td>
<td>(5.4)</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td></td>
<td>60</td>
<td>45</td>
<td>30</td>
<td>0.8</td>
<td>18</td>
</tr>
<tr>
<td>E77-1</td>
<td>2.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E77-2</td>
<td>(23.2)</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\Sigma M_0 = 25.2$</td>
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<td>50</td>
<td>28</td>
<td>2.0</td>
<td>49</td>
</tr>
<tr>
<td>E77-3</td>
<td>(23.2)</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>23.2</td>
<td></td>
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<td>$\Sigma M_0 = 46.4$</td>
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<td>55</td>
<td>25</td>
<td>21</td>
<td>6.7</td>
<td>219</td>
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</tr>
<tr>
<td>E78-2</td>
<td>6.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>$\Sigma M_0 = 11.5$</td>
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<td>60</td>
<td>40</td>
<td>28</td>
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<td>25</td>
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<tr>
<td>E79-1</td>
<td>0.3</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>E79-2</td>
<td>6.2</td>
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</tr>
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<td>40</td>
<td>23</td>
<td>0.8</td>
<td>24</td>
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<tr>
<td>E84-1</td>
<td>37.5</td>
<td>?</td>
<td>?</td>
<td>?</td>
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</tr>
</tbody>
</table>

* Determined from Rayleigh waves, except for events E66-1 and E77-1, where values are from body waves. Values in parentheses are assumed (see text for discussion).
Fig. 15. Lower-hemisphere focal mechanisms and depths (in parentheses) of earthquakes determined in this study. Depths are determined from P-waveform analysis. Compressional quadrants are stippled or solid depending on whether data provide limits on orientation of a single or both nodal planes, respectively.

E79-1, E79-2, and E84-1. The mechanisms of these latter earthquakes and the mechanism for E77-4 are consistent with thrust motion with a large component of left-lateral slip along fault planes that dip to the north. Slip vectors on the northerly dipping planes range in azimuth from N66°E to N85°E with an average of N73°W. Estimates of relative plate motion that are based largely on seafloor magnetic lineations, in contrast, are closer to N80°E (Chase, 1971; Minster et al., 1974).

The southeastern Solomon Islands may have ruptured in a similar sequence of earthquakes during the 1930's (Lay and Kanamori, 1980; McCann, 1981; Lay et al., 1982). The observation allows a crude approximation of the seismic slip rate along this portion of the island arc. Assuming that all the events produced displacement along a north-dipping thrust at about N80°E, the rate of displacement averaged over the T = 40 years since the last sequence of events may be written to equal \( \bar{D} = \Sigma M_0/\mu A T \), where in this case \( \Sigma M_0 \) is the cumulative moment of all earthquakes during the time \( T \). Taking \( A \) to equal 300 × 50 km², and the cumulative seismic moment sum listed in Table IV, we found the seismic slip rate averaged over the last 40 years to equal about 4.6 cm year⁻¹, a rate about one-half that of the 10 cm year⁻¹ indicated by analysis of magnetic lineations on the seafloor (e.g., Chase, 1971; Minster et al., 1974). This discrepancy in rates, if real, implies that either plate subduction is accommodated in a large part by aseismic processes, or that contemporary rates are less than those rates averaged over many millions of years.
The distribution of aftershocks suggests that earthquakes since 1966 have released accumulated strain along most of the length of the Solomon Island trench that extends along the coasts of Guadalcanal and San Cristobal. Only the segments along the western end of Guadalcanal and southeastern coast of San Cristobal have not ruptured in large earthquakes during the last 20 years (Fig. 2). Assuming a constant rate of strain accumulation along this portion of the arc, it is likely that either of these two sites will be the locus of the next large shock along this plate boundary.

The earthquakes within each multiplet are summarized in Fig. 16 as a function of size, time of occurrence, and epicentral location. Figure 16a is a plot of event magnitude $M_s$ versus the epicentral offset to the subsequent event in each sequence. Figure 16b is similar, except that the number of hours to the subsequent event is plotted against $M_s$. The number of kilometers between epicenters is plotted against the number of hours between events in Fig. 16c. A trend is observed only in Fig. 16a. The trend shows epicentral offsets between sequential shocks within a multiplet is proportional to the size of the preceding event. In general, however, there is no distinct pattern that sheds a conclusive light on the process of stress transfer and the initiation or triggering of the earthquake rupture process.

Note in Addendum

A $M_s = 6.9$ earthquake occurred west of the 1977 earthquake sequence at 9.8°S, 159.2°E on September 27, 1985. No aftershocks of $M_b > 4.5$ are reported for the 2 days subsequent to the event.

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