Body-waveforms and source parameters of some moderate-sized earthquakes near North Island, New Zealand

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ABSTRACT


P-wave first motion and synthetic seismogram analysis of P- and SH-waveforms recorded at teleseismic distances on the WWSSN are used to estimate source parameters of seven of the largest earthquakes (6.1 ≤ mb ≤ 6.3) that occurred in the vicinity of North Island, New Zealand since 1965. The source parameters of three of these (mb ≥ 6.1) events determined outside of this study are included and considered in the final analysis. Four of the earthquakes occurred at shallow depths (< 20 km), of which three were located within and to the north of North Island. Two of the shallow events show strike-slip and normal focal mechanisms with T-axes oriented in a manner consistent with their location in an area of known back-arc extension. One of the shallow events occurred in northern South Island and shows a reverse-type mechanism indicating horizontal contraction of the crust in an easterly azimuth. Six events occurred at intermediate depths (h = 39 to 195 km) of which five exhibit thrust mechanisms with T-axes consistently oriented near vertical. In the light of previously published plate tectonic models, the near vertical orientation of T-axes of the intermediate-depth events may be used to infer that the southern Kermadec plate boundary immediately north of North Island is not strongly coupled, and hence, not likely capable of producing great earthquakes. A similar inference cannot be made for the section of the Hikurangi Margin adjacent to North Island since the intermediate-depth events considered in this study lie to the north of this segment of the plate boundary.

Introduction

The North Island of New Zealand is located on the eastern margin of the Indian–Australian plate (Fig. 1). Relative motion between the Pacific and Indian–Australian plates is primarily convergent to the north Island and becomes increasingly right-lateral in nature to the south, as evidenced by right-lateral offsets along the Wellington, Wairarapa, and Marlborough fault systems (e.g., Lensen, 1983), and by oblique right-lateral strike-slip motion along the Alpine Fault (e.g., Walcott, 1978a). Thus crustal deformation in the vicinity of North Island reflects the transition from a primarily convergent to a largely transform-type plate boundary. The purpose of this paper is to document the source parameters of the largest earthquakes that have occurred in the vicinity of North Island during the last 30 years, as evidenced by synthetic seismogram analysis of P and SH body-waveforms recorded on the Worldwide Standardized Seismograph Network (WWSSN). Motivation for the study resides in our attempt to understand better the state of stress, mechanics of crustal deformation, and the potential for large earthquakes along the northern New Zealand plate boundary.

Earthquakes of magnitude 7 or greater are common to the New Zealand historical record,
with more than a dozen being documented since 1848 (Smith and Barryman, 1986), yet no earthquake of magnitude 7 has occurred since 1960. Hence, the events considered here are generally of moderate magnitude near 6. More specifically, we examine the ten largest earthquakes recorded in the Preliminary Determination of Epicenters (PDE) catalog in the vicinity of North Island since 1960 (Table 1 and Fig. 2). The body-waveforms and source characteristics of two of the events, the 1966 Gisborne (Webb et al., 1985) and the 1987 Edgecumbe (Anderson and Webb, 1989) earthquakes, have been previously studied in detail and the results of those studies are only briefly touched upon in this report. We also draw upon the Harvard centroid-moment tensor catalog (Dzie- wonski et al., 1985) for the source parameters of the December 30, 1984 event for which body-waveforms were not recoverable from the WWSSN library. Additionally, the focal mechanisms of a number of the events in Table 1 have also been determined previously using P-wave first motions recorded at telesismic (e.g., Banghar and Sykes, 1969; Isacks and Molnar, 1971; Johnson and

![Plate tectonic setting, major faults (from Lensen, 1983), and bathymetry near North Island, New Zealand. Overriding plate at the subduction boundary is marked by hachures. Plate motion vectors shown by arrows. Numbers next to arrows indicate rate (mm/yr) and direction of plate convergence, respectively (Minister and Jordan, 1978). Bathymetry contours given in kilometers below mean sea level. The Wairau, Awatere, and Hope faults are members of the right-lateral Marlborough Fault system. Figure adapted from Suggate et al. (1978).](image-url)
TABLE 1

Epicentral data *

<table>
<thead>
<tr>
<th>Event</th>
<th>Origin time</th>
<th>Latitude (°)</th>
<th>Longitude (°)</th>
<th>( m_b )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dec. 8, 1965</td>
<td>18h 05m 25s</td>
<td>37.10S</td>
<td>177.50E</td>
<td>6.2</td>
</tr>
<tr>
<td>Mar. 4, 1966</td>
<td>23h 58m 56s</td>
<td>38.40S</td>
<td>177.90E</td>
<td>6.1</td>
</tr>
<tr>
<td>Mar 23, 1968</td>
<td>17h 24m 15s</td>
<td>41.66S</td>
<td>171.92E</td>
<td>6.1</td>
</tr>
<tr>
<td>Jan. 8, 1970</td>
<td>17h 12m 39s</td>
<td>34.74S</td>
<td>178.56E</td>
<td>6.1</td>
</tr>
<tr>
<td>Jan. 5, 1973</td>
<td>13h 54m 29s</td>
<td>38.99S</td>
<td>175.23E</td>
<td>6.2</td>
</tr>
<tr>
<td>June 29, 1976</td>
<td>18h 30m 09s</td>
<td>33.81S</td>
<td>177.83W</td>
<td>6.1</td>
</tr>
<tr>
<td>Dec. 30, 1984</td>
<td>21h 36m 56s</td>
<td>36.66S</td>
<td>177.51E</td>
<td>6.2</td>
</tr>
<tr>
<td>Sept. 26, 1985</td>
<td>07h 27m 51s</td>
<td>34.69S</td>
<td>178.65W</td>
<td>6.3</td>
</tr>
<tr>
<td>Nov. 7, 1985</td>
<td>19h 12m 31s</td>
<td>35.25S</td>
<td>179.34W</td>
<td>6.2</td>
</tr>
<tr>
<td>Mar. 2, 1987</td>
<td>01h 42m 34s</td>
<td>37.97S</td>
<td>176.76E</td>
<td>6.1</td>
</tr>
</tbody>
</table>

* Data for earthquakes obtained from the Preliminary Determination of Epicenters (PDE) catalog (NEIC, National Geophysical Data Center, Boulder Colorado).

Molnar, 1972), regional (Adams, 1963; Harris, 1982a, b), and local distances (Walcott, 1978a; Reyners, 1980). Similarly, depths for the earthquakes have been determined before on the basis of relative arrival times of body-waves at local to teleseismic distances. This work differs from these latter efforts in that synthetic waveform techniques are used to determine, in addition to the focal mechanism, the depth, duration of faulting, and the seismic moment (\( M_0 \)) for each earthquake. Discussion of our results within the context of regional seismicity and plate tectonic setting will follow a brief description of the data and analysis techniques.

Data and methods

The data analysed in this study are the long-period P- and SH-waves of seven \( m_b \geq 6.1 \) earthquakes (Table 1) recorded by the WWSSN and occurring between 1965 and 1987. Initially the P-wave first motions are used to gauge the orientation of the nodal planes while the S-wave polarization angles are used to infer the location of the pressure axes, thus allowing for an estimate of the focal mechanism for each event. The body-waves of the seven events are then modelled using synthetic seismograms computed according to the technique developed by Langston and Helmberger (1975). Factors controlling the shape and amplitude of the synthetic waveforms are the focal mechanism, source–time function, focal depth, attenuation structure, and the velocity–density structure. A forward modelling approach is used to match the synthetic to the observed WWSSN waveforms to define better the focal mechanism, and to determine the focal depth, source–time history, and seismic moment for each event. Complications due to upper mantle and core structure are avoided by considering waveforms recorded at epicentral distances of about 30° to 90°. The P- and S-wave velocity structures assumed for the source region (used to compute take-off angles) and the surface velocity (used to compute focal depth and reflection coefficients for surface bounces of depth phases pP, sP, and sS) are taken from Jeffreys-Bullen A model (Kanamori and Abe, 1968) and Robinson (1986). The specific velocity–density models (Table 2) used to construct the synthetic seismograms for each event are chosen according to the geographic locations and depths of the earthquakes. Attenuation of the wave along the ray path is accounted for using Futterman’s (1962) operator, with values of 1.0 s
TABLE 2
Velocity and density structures

<table>
<thead>
<tr>
<th>Event</th>
<th>Source</th>
<th>Crust</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\alpha$ (km s$^{-1}$)</td>
<td>$\beta$ (km s$^{-1}$)</td>
</tr>
<tr>
<td>Dec. 8, 1965</td>
<td>8.7</td>
<td>5.0</td>
</tr>
<tr>
<td>May 23, 1968</td>
<td>8.5</td>
<td>3.7</td>
</tr>
<tr>
<td>Jan. 8, 1970</td>
<td>8.7</td>
<td>5.0</td>
</tr>
<tr>
<td>Jan. 5, 1973</td>
<td>8.7</td>
<td>5.0</td>
</tr>
<tr>
<td>June 29, 1976</td>
<td>7.8</td>
<td>4.4</td>
</tr>
<tr>
<td>Sept. 26, 1985</td>
<td>7.8</td>
<td>4.4</td>
</tr>
<tr>
<td>Nov. 7, 1985</td>
<td>7.8</td>
<td>4.4</td>
</tr>
</tbody>
</table>

* Jeffreys-Bullen A velocity model (Kanamori and Abe, 1968), used for events within continental and oceanic lithosphere offshore of North Island.
* From Robinson (1986) for events within crust and subducted slab beneath North Island.

and 4.0 s for $t^*$ of P- and S-waves, respectively, where $t^*$ is proportional to the average attenuation of the wave along the raypath. Forward modelling constrains the duration of the source–time function to within about 2 s (Lay and Kanamori, 1980), though it should be recognized that some trade-off exists between focal depth and the source–time function when computing the shape of the seismograms.

The results from the waveform analysis of the seven events, in addition to the three previously studied earthquakes, are summarized in Table 3 and Fig. 2. For convenience of comparison, when available, the Harvard centroid-moment tensor

TABLE 3
Earthquake source parameters from modelling *

<table>
<thead>
<tr>
<th>Event</th>
<th>$\phi$ (°)</th>
<th>$\delta$ (°)</th>
<th>$\lambda$ (°)</th>
<th>$M_0$</th>
<th>No. Sta.</th>
<th>$h$ (km)</th>
<th>$\tau, \tau_c$ (s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dec. 8, 1965</td>
<td>118</td>
<td>30</td>
<td>56</td>
<td>0.63 ± 0.4</td>
<td>11</td>
<td>155</td>
<td>1, 3.2</td>
</tr>
<tr>
<td>Mar. 4, 1966</td>
<td>249</td>
<td>25</td>
<td>131</td>
<td>0.04</td>
<td>10</td>
<td>18</td>
<td>0.7, 1.7</td>
</tr>
<tr>
<td>May 23, 1968</td>
<td>20</td>
<td>50</td>
<td>120</td>
<td>5.50 ± 3.9</td>
<td>10</td>
<td>10</td>
<td>6.9 b</td>
</tr>
<tr>
<td>Jan. 8, 1970</td>
<td>-20</td>
<td>65</td>
<td>70</td>
<td>5.94 ± 1.9</td>
<td>14</td>
<td>195</td>
<td>1, 4</td>
</tr>
<tr>
<td>Jan. 5, 1973</td>
<td>135</td>
<td>83</td>
<td>50</td>
<td>5.91 ± 2.8</td>
<td>14</td>
<td>158</td>
<td>1, 3</td>
</tr>
<tr>
<td>June 29, 1976</td>
<td>30</td>
<td>45</td>
<td>95</td>
<td>0.38 ± 0.1</td>
<td>13</td>
<td>39</td>
<td>1, 2</td>
</tr>
<tr>
<td>Dec. 30, 1984</td>
<td>86</td>
<td>56</td>
<td>-12</td>
<td>1.80</td>
<td>-</td>
<td>19</td>
<td>-</td>
</tr>
<tr>
<td>Sept. 26, 1985</td>
<td>115</td>
<td>50</td>
<td>125</td>
<td>5.14 ± 1.8</td>
<td>10</td>
<td>52</td>
<td>2, 5</td>
</tr>
<tr>
<td>Nov. 7, 1985</td>
<td>30</td>
<td>50</td>
<td>80</td>
<td>1.30 ± 0.5</td>
<td>12</td>
<td>50</td>
<td>1, 4</td>
</tr>
<tr>
<td>Mar. 2, 1987</td>
<td>225</td>
<td>45</td>
<td>-110</td>
<td>0.43 ± 0.2</td>
<td>15</td>
<td>08</td>
<td>1, 1</td>
</tr>
</tbody>
</table>

* $\phi$, $\delta$, $\lambda$ = strike, dip, and rake of east nodal plane; $M_0$ = seismic moment of each event, in units of $10^{26}$ dyn cm ($= 10^{19}$ N m); No. Sta. = number of WWSSN stations used in waveform analysis; $h$ = focal depth; $\tau, \tau_c$ = rise time and duration of faulting, respectively.
* Source parameters determined by Webb et al. (1985).
* Complex source–time function, $\tau$ and $\tau_c$ are for main pulse only.
* Centroid-moment tensor solution obtained from the Dziewonski et al. (1985).
* Source parameters determined by Anderson and Webb (1989); two identical sources, separated by 3 s, account for moment release.
TABLE 4
Harvard centroid-moment tensor solutions *

<table>
<thead>
<tr>
<th>Event</th>
<th>$\phi$</th>
<th>$\delta$</th>
<th>$\lambda$</th>
<th>$M_0$ *</th>
<th>$h$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dec. 30, 1984</td>
<td>86</td>
<td>56</td>
<td>-12</td>
<td>1.80</td>
<td>19</td>
</tr>
<tr>
<td>Sept. 26, 1985</td>
<td>196</td>
<td>65</td>
<td>39</td>
<td>2.40</td>
<td>61</td>
</tr>
<tr>
<td>Nov. 7, 1985</td>
<td>28</td>
<td>54</td>
<td>83</td>
<td>0.73</td>
<td>56</td>
</tr>
<tr>
<td>Mar. 2, 1987</td>
<td>36</td>
<td>32</td>
<td>-101</td>
<td>0.64</td>
<td>15</td>
</tr>
</tbody>
</table>


\* In units of $10^{20}$ dyn cm ($=10^{19}$ N m).

solutions are also listed for each event in Table 4. Figures showing the observed versus synthetic waveforms, P-wave first motions, and source parameters determined for each event are placed separately in the Appendix.

Discussion

Discussion of the earthquakes is best placed within the framework of our current understanding of plate tectonics and regional seismicity near North Island. Relative motion between the Indian–Australian and Pacific plates, along the southern Kermadec and Hikurangi trenches, is primarily accommodated by underthrusting of the Pacific plate beneath the Indian–Australian plate at a rate of 5–6 cm/yr (Fig. 1). In contrast, relative plate motion to the south of North Island is taken up largely by right-lateral slip along the Marlborough and Alpine faults in South Island (Fig. 1, e.g., Walcott, 1978a). Epicenters of $m_b \geq 4$ earthquakes, reported by the ISC between 1964 and 1982, are plotted as a function of depth in map view on Fig. 3. Figure 3 and the cross-sections of seismicity illustrated in Fig. 4, show a well-defined Wadati-Benioff zone that dips to the northwest and extends westward beneath North Island. The earthquakes presented in this study, marked in Figs. 3 and 4, are shown to occur both within a shallow band of seismicity west of the trench, and the Wadati-Benioff zone. Initially we will briefly discuss the four shallow events listed in Table 3, then shift attention to the remaining deeper events.

Shallow earthquakes

The southern Kermadec and Hikurangi trenches have not been the site of a great interplate thrust earthquake during historical time. Large thrust ($M \sim 7$) earthquakes occurred near Hawke’s Bay in 1931 ($M = 7.8$) and 1932 (Wairoa, $M = 7$), but it is not certain from geodetic analysis whether the events were associated with slip on the shallowly dipping interplate thrust interface, or solely on more steeply dipping, localized thrust planes (Haines and Darby, 1988). The largest earthquake during the last 30 years to have possibly occurred along the interplate thrust boundary was the Gisborne earthquake ($M \sim 6$) of 1966 (Fig. 2; Webb et al., 1985). Despite the lack of large earthquakes, the source parameters documented here for the moderate-sized earthquakes in Table 3 may add to our insight regarding the state of stress, crustal deformation processes, and the potential for large earthquakes along the plate boundary near North Island, New Zealand. For example, the 1987 Edgecumbe earthquake occurred within the Central Volcanic Region (Figs. 1

Fig. 3. Epicenters of $m_b \geq 4$ earthquakes in the New Zealand and southern Kermadec regions reported from 1964 to 1982 by the International Seismological Centre (ISC). Epicenters are plotted as a function of depth. Large circles are earthquakes analyzed in this study with either a filled box or star plotted in center to denote focal depth. Lettered brackets show location and zone of seismicity projected onto the cross sections, shown in Fig. 4.
Fig. 4. Focal-depth distribution of seismicity recorded by the ISC, projected onto four cross sections that are normal to the Hikurangi Trench. Triangles represent the approximate location of the volcanic front, downward pointing arrows show approximate location of the trench. Horizontal lines above cross sections are the location of New Zealand Islands. Large circles are the earthquakes analyzed in this study. Projections of axes of least compressive stress (T-axes) onto plane of cross sections are shown by oppositely facing arrows. Tensions axes are plotted only for events with tension axes parallel to the plane of the cross section. The latitude and longitude of the endpoints of the cross sections are given at the bottom edge of each cross section. The figures have no vertical exaggeration.

and 2). Prior microearthquake (Webb et al., 1986), Quaternary fault (e.g., Grindley and Hull, 1986), and geodetic (Walcott, 1978b; Sissons, 1979) studies show the Central Volcanic Region (CVR) is an area of back-arc extension within central North Island. The CVR is thought to be the direct continuation of the Havre trough oceanic back-arc basin into the continental structure of New Zealand (e.g., Stern, 1985). The shallow depth (8 km) and normal mechanism, determined for the Edgecumbe earthquake by Anderson and Webb (1989; Table 3), indicate northwesterly extension
which is consistent with the process of back-arc extension within the CVR. The December 1984 earthquake is also shallow (19 km), and occurred to the north of the Edgecumbe earthquake, within the Bay of Plenty (Fig. 2, Table 3). In contrast to the Edgecumbe earthquake, the mechanism for the 1984 event is strike-slip. The orientation of the 1984 earthquake’s T-axis is similar to that observed for the Edgecumbe earthquake and thus consistent with back-arc extension. However, the presence of strike-slip faulting in this region suggests back-arc extension is not necessarily occurring by simple rifting (Smith and Webb, 1986).

The May 1968 Inangahua earthquake occurred within the thin band of shallow seismicity that extends westward into South Island from the Hikurangi Trench (Figs. 2 and 4). The near horizontal P-axis of the 1968 earthquake has an easterly trend and, hence, is consistent with the results of geodetic studies (Bibby, 1981) that show maximum horizontal contraction of the crust in northern South Island is oriented similarly. The close spatial proximity of the 1968 thrust event to the active, strike-slip Marlborough fault system suggests that the least and intermediate principal stress axes are equivalent in amplitude in the crust of northern South Island.

**Intermediate-depth earthquakes**

The remaining six events are of intermediate depth (39 to 195 km) and are indicated in the cross-sections of Fig. 4. The January, 1973 earthquake, (1) occurred directly beneath North Island, (2) shows primarily strike-slip motion with one nodal plane oriented in the direction of dip of the subducted Pacific plate, and (3) may reflect tear faulting within the subducted slab (Reyners, 1983). The remaining five events are located beneath the Bay of Plenty, to the north of North Island, and show thrust mechanisms with T-axes that are near vertical (Figs. 2 and 5). Although the nodal planes of these events are not particularly well-constrained, variations in the strike and dip of the nodal planes will not significantly affect the orientation of the vertical T-axes. Further, these five events are most likely not a result of slip along the interplate boundary. Stern (1985) inferred from gravity profiles across the Wadati-Benioff zone of North Island that the crustal thickness of the overriding plate, from the southern Kermadec to the southern Hikurangi Trench, is approximately 25 to 30 km. Thus the interplate boundary is inferred here not to extend deeper than 30 km, and hence, the five intermediate-depth events are interpreted to reflect stresses within the subducting plate, and not relative plate motion.

Astiz et al. (1988) compiled a global data set of intermediate-depth earthquake focal mechanisms, and on the basis of that compilation, showed that intermediate-depth earthquakes within a specific subduction zone generally show either reverse or normal mechanisms, but not both. It was further noted empirically that those subduction zones characterized by normal faulting at intermediate depths commonly show high (6–11 cm/yr) convergence rates, subducting oceanic lithosphere of relatively young age (20–80 Ma), a shallow dipping Wadati-Benioff zone, and the occurrence of the largest historically recorded interplate thrust earthquakes with moment magnitudes exceeding 9 (Ruff and Kanamori, 1983; Astiz et al., 1988). In that regard, Astiz et al. (1988) interpret the normal faulting earthquakes at intermediate depth to reflect down-dip extension within the subducted
slab, caused by gravitational pull of a gently dipp- 
ing subducted plate beneath a strongly coupled 
interplate thrust interface. In contrast, subduction 
zones showing reverse faulting at intermediate 
depths are generally characterized by lesser con- 
vergence rates (2–6 cm/yr), subducting oceanic 
lithosphere of older age (90–150 Ma), Wadati-Beni- 
off zones that dip at relatively steeper angles, 
and interplate thrust earthquakes of only mod- 
erate 6 ≤ M ≤ 8 magnitude. The characteristically 
smaller interplate thrust earthquakes at such 
boundaries are attributed to a smaller coefficient 
of coupling between the overriding and subduct- 
ing plates. The smaller coefficient of coupling 
results from an oceanward migration of the trench 
which is attributed to gravitational pull on the 
steeply dipping, slowly converging, older, and 
therefore, cooler and denser lithosphere (Fujita 
and Kanamori, 1981). Observations that the age of 
the subducting Pacific plate along the Kermadec 
Trench is about 100 ± 20 Ma (Molnar and Atwater, 
1978), the rate of subduction is about 
5–6 cm/yr (Fig. 1), and, as shown in this study 
(Fig. 5), intermediate-depth earthquakes char- 
acterized by vertical tension axes all may be used 
to infer that the southern Kermadec plate 
boundary is not strongly coupled and, therefore, 
the maximum size of interplate thrust events would 
most likely be limited to magnitudes between 6 
and 8. Extension of this interpretation farther 
south along the Hikurangi Margin is not war- 
ranted since only one of the intermediate events in 
this study is located directly beneath North Is- 
land. Some weak support for the contention that 
large interplate thrust events along the Hikurangi 
Trench will be limited to magnitudes 6 ≤ M ≤ 8 
may be found in the historical record and recent 
studies of uplifted wavecut terraces along the east 
coast of North Island. Historically, no interplate 
thrust events of magnitude greater than 8 have 
been unequivocally documented. Furthermore, 
flights of uplifted wavecut terraces along the east 
coast of North Island have recently been analysed 
by Berryman et al. (1989) and tentatively attrib- 
uted to episodic uplifts of the coast during 
thrust-type earthquakes of about magnitude 7. 
However, it is unclear whether or not the 1931 
Hawke’s Bay earthquake or the uplifted terraces 
are directly the result of slip on the main plate 
interface or, rather, the result of more localized 
thrust faults along the coast. Moreover, the sub- 
ducting Pacific lithosphere along the Hikurangi 
Trench may be significantly thicker (B. Davies, 
pers. commun., 1989) and more buoyant than 
found along the Kermadec Trench to the north 
(Robinson, 1986; Smith et al., 1989). Such dif- 
fferences in the character of the subducting litho- 
sphere that exist between the Kermadec and 
Hikurangi trenches may well result in significantly 
different seismic behavior along the two trenches 
(Smith et al., 1989). Indeed, in contrast to the 
presence of intermediate-depth thrust faulting 
within the subducted slab beneath the Bay of 
Plenty, Robinson’s (1986) composite focal mecha- 
nism study of microearthquakes directly beneath 
the thrust interface and southern North Island 
show nearly pure normal faulting mechanisms, 
consistent with the presence of a strongly locked 
plate interface. Available observations thus remain 
rather equivocal regarding the potential for great 
interplate thrust earthquakes along the east coast 
of North Island, New Zealand.

Appendix

Figures A1–A7 present, in chronological order, 
the P- and S-wave first motions, focal mechanisms, 
source parameters, and synthetic and observed P- 
and SH-waveforms of the earthquakes analysed in 
this study as listed in Table 3 and shown in Fig. 2. 
Comparison of these results to prior studies and 
any information regarding special features of the 
events are discussed in the figure captions. The 
format of the figures is as discussed below. For 
each seismogram pair, the upper and lower seis- 
mograms correspond to the synthetic and ob- 
served long-period P- and SH-waveforms, respec- 
tively. SH-waves are annotated by an SH shown 
next to the station name. Heavy lines within the 
mechanism are P-wave nodal planes; dashed lines 
are SH-wave nodal planes. Bars bisecting the sta- 
tion symbol represent the S-wave polarization an- 
gle determined, when available, for that station 
using the technique of Dillinger (1975). Open and 
filled stars labeled by T and P respectively, repre- 
sent the T- and P-axes. The synthetic seismo-
grams correspond to the focal mechanism, source–time function, and depth shown. Strike, dip, and rake of the east nodal plane are $\phi$, $\delta$, and $\lambda$, respectively. Source-time functions are assumed to be trapezoidal in shape and characterized by a rise time $\tau$ and a rupture duration $\tau_c$. The seismic moment $M_o$ determined from individual waveforms is shown beside each station name and is in units of $10^{26}$ dyne cm ($10^{26}$ dyn cm $= 10^{19}$ N m). The seismic moment for each event was obtained by averaging the specific values of seismic moment determined at non-nodal stations, and is shown in the header of each figure.

The body-waves for the 1966 Gisborne earthquake and the 1987 Edgecumbe earthquake were previously studied in detail by Webb et al. (1985) and Anderson and Webb (1989), respectively. The waveforms for the December 30, 1984 event were insufficient for modelling, however, a centroid-moment tensor solution was obtained from Dziewonski et al. (1985). Hence, we have used the results from previous analyses of these three events in our study (Table 3) and do not show the waveform for the events here.

Fig. A1. December 8, 1965 earthquake. The mechanism determined here is similar to solutions previously published by Isacks and Molnar (1971) (e.g., $\phi = 95^\circ$, $\delta = 50^\circ$, $\lambda = 55^\circ$) and Chung and Kanamori (1980) (e.g., $\phi = 117.5^\circ$, $\delta = 29.7^\circ$, $\lambda = 56.2^\circ$). A depth phase observed about 45 s after the first arrival is interpreted to be sP and places the event at about 155 km depth. See Appendix text for explanation of symbols in this and subsequent events.
Fig. A2. May 23, 1968, the Inangahua earthquake. The mechanism indicates predominantly thrust motion with a lesser component of strike-slip movement, in accord with a mechanism (ϕ = 60°, δ = 50°, λ = 120°) presented previously by Rial (1978). The P-waveforms for this event are complex, suggesting a multiple source. The source–time function used is characterized by a small event that precedes the main event by about 1 s. The N20°E nodal plane is parallel to the trend of observed surface rupture associated with the event (Adams and Lowry, 1971).

Fig. A3. January 8, 1970 earthquake. The discrepancy between the S–sS time predicted for the SH-waveform at GUA and the P–pP and p–sP time recorded at the other stations likely reflects the inaccuracy of the average velocity structure assumed for the region.


P and SH - WAVES JANUARY 5, 1973

\[ \phi = 135^\circ, \ delta = 83^\circ, \ \lambda = 50^\circ \]
\[ h = 158 \text{ km}, \ M_o = 5.91 \times 10^{26} \text{ dyne-cm} \]
\[ \tau = 1.0\text{s}, \ \tau_c = 3.0\text{s} \]

Fig. A4. January 5, 1973 earthquake. The deep nature of this event is evidenced by the clear presence of sS phases on the SH-waves recorded at stations MUN and LEM. Harris (1982b) previously reported a mechanism of \((\phi = -20^\circ, \delta = 75^\circ, \lambda = 130^\circ)\) for this event based upon local and regional first motion observations.

P - WAVES JUNE 29, 1976

\[ \phi = 30^\circ, \ delta = 45^\circ, \ \lambda = 95^\circ \]
\[ h = 39 \text{ km}, \ M_o = 3.8 \times 10^{26} \text{ dyne-cm} \]
\[ \tau = 1.0\text{s}, \ \tau_c = 2.0\text{s} \]

Fig. A5. June 29, 1976 earthquake. A thrust mechanism is indicated for this event, but the orientations of the nodal planes are not strongly limited by either first motion or waveform observations.
$\phi = 115^\circ$  $\delta = 50^\circ$  $\lambda = 125^\circ$

$h = 52 \text{ km} \quad M_O = 5.14 \times 10^{26} \text{ dyne-cm}$

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**Fig. A6. September 26, 1985 earthquake.**

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$\phi = 30^\circ$  $\delta = 50^\circ$  $\lambda = 80^\circ$

$h = 50 \text{ km} \quad M_O = 1.3 \times 10^{26} \text{ dyne-cm}$

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**Fig. A7. November 7, 1985 earthquake.**
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