The June 2000 $M_w$ 7.9 earthquakes south of Sumatra: Deformation in the India–Australia Plate

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[1] Two large ($M_w$ 7.9) earthquakes occurred on 4 and 18 June 2000, south of Sumatra, beneath the Indian Ocean. Both earthquakes were predominantly left-lateral strike-slip on vertical N-S trending faults that we interpret to be reactivated fracture zones. The 4 June Enggano earthquake occurred at the edge of the rupture area of the 1833 subduction earthquake. The first strike-slip subevent within the subducting plate triggered a thrust subevent on the plate interface, which comprised at least 35% of the total moment and ruptured SE away from the 1833 earthquake. The 18 June earthquake in the Wharton Basin is one of the largest shallow strike-slip faulting earthquakes ever recorded. A small second subevent with reverse slip is required to fit the body waves. The orientation of both subevents in our preferred model is consistent with the current stress field in the region. Both the June 2000 earthquakes are consistent with recent models of distributed deformation in the India–Australia composite plate. The occurrence of the Enggano earthquake implies that the stress field within the Indian plate continues to a depth of 50 km in the subducting slab. The purely strike-slip source model of the Wharton Basin earthquake obtained by Robinson et al. [2001] matches the $P$ waves very poorly and fits the $S$ waves no better than our preferred model. The strike-slip subevents of both earthquakes had few aftershocks and higher stress drops than the subduction thrust subevent of the Enggano earthquake. This difference is consistent with previous observations of oceanic and subduction earthquakes.

INDEX TERMS: 7209 Seismology: Earthquake dynamics and mechanics; 7215 Seismology: Earthquake parameters; 7230 Seismology: Seismicity and seismotectonics


1. Introduction

[2] On 4 June 2000 a large ($M_w$ 7.9) earthquake occurred near Enggano Island, off the coast of Sumatra, killing 90 people and injuring 2000 more. It was felt as far away as Singapore where many high-rise buildings were shaken [Pan et al., 2001]. This earthquake was followed 2 weeks later by a second earthquake of similar magnitude, approximately 1000 km to the south, beneath the Indian Ocean. Both of these earthquakes were largely unexpected, and the aim of our study is to determine their rupture geometry and to investigate what they reveal about the complex seismotectonics of the Indian Ocean and western Indonesia.

[3] The 4 June Enggano earthquake occurred within the seismically active Sumatra subduction zone. The unusual aspect of this earthquake was its source mechanism. The Harvard catalogue centroid moment tensor (CMT) mechanism [Dziewonski and Woodhouse, 1983; Dziewonski et al., 1981, 2001] is primarily strike-slip. The CMT solution also has a large, but poorly resolved, non-double-couple component. The Indian plate is subducting beneath Sumatra at an oblique angle [DeMets et al., 1994a], and the deformation is divided between predominantly reverse motion at the trench (see Figure 1) [e.g., Fitch, 1972; Newcomb and McCann, 1987; McCaffrey et al., 2002] and right-lateral strike-slip on the Great Sumatra fault [e.g., Sieh and Natawidjaja, 2000; Genrich et al., 2002]. The more recently discovered Mentawai fault in the forearc [Diament et al., 1992] does not appear to be active [Genrich et al., 2000]. Sumatra is often cited as a classic example of a so-called slip-partitioned margin [Fitch, 1972]. The Enggano earthquake was too deep (~50 km) to be on either the Sumatra or the Mentawai faults. The epicenter of the Enggano earthquake is in a region of continuously high seismicity over the last few decades. It is also at the southeastern end of the rupture area of the great (M > 8) 1833 subduction earthquake (Figure 2), as determined from intensity reports [Newcomb and McCann, 1987] and submergence and uplift of corals [Zachariasen et al., 1999]. In the rupture area of the 1833 earthquake, the upper and lower subducting plates are thought to be fully seismically coupled [Zachariasen et al., 2000], but the coupling appears to decrease near the equator [e.g.,
South of Java, where the subducting slab is older and thicker, subduction is nearly aseismic and close to the direction of relative plate motion [e.g., Newcomb and McCann, 1987; Abercrombie et al., 2001].

The 18 June 2000 Wharton Basin earthquake is the largest known earthquake in the central Indian Ocean, and it is one of the largest strike-slip earthquakes ever recorded. There are no previous earthquakes in either the International Seismological Centre (ISC) or NEIC catalogues (since 1970, complete above $M_c \geq 5$) within over 150 km of the hypocenter of the 18 June earthquake. The two earthquakes near its hypocenter in Figure 1 are both aftershocks. Robinson et al. [2001] modeled the $S$ waves and concluded that this earthquake had an unusual rupture geometry. Their preferred model involves simultaneous strike-slip motion on two nearly perpendicular faults, with the slip centroids separated by about 45 km. Such a rupture process is unprecedented and therefore we investigate how well it is resolved.

The Indian Ocean is the most seismically active oceanic plate interior on Earth [Bergman and Solomon, 1985]. The distributed deformation is thought to be a consequence of the collision of India with Eurasia impeding the approximately NNE motion of the India–Australia plate in the west and continuing subduction along the Sunda Arc beneath Java and Sumatra. Seismicity is distributed along and around the Ninety East (90°E) ridge, a prominent feature (Figure 1) that is thought to be an old hot spot trace following a fossil transform fault [e.g., Stein and Okal, 1978]. To the west of the 90°E Ridge, long wavelength folds trend E-W [e.g., Weissel et al., 1980], and similarly oriented faults have been identified as currently active [Chamot-Rooke et al., 1993; Bull and Scrutton, 1990]. The seismicity is characterized by reverse faulting earthquakes striking E-W and strike-slip faulting earthquakes striking NW-SE (or NE-SW) (see Figure 1). To the east of the 90°E ridge, however, the earthquakes occur on N-S (or E-W) striking strike-slip faults and a few NE-SW striking reverse faults. The area to the east of the 90°E ridge, including the

**Figure 1.** Location map of the 4 June (Enggano) and 18 June (Wharton Basin) earthquakes. The Harvard CMTs are joined to their NEIC epicenters. The Harvard catalogue CMTs for earthquakes $M_c \geq 6.5$ are shown in midgray. In the Indian Ocean (where the seismicity rate is lower than along the Sumatra and Java trenches), all mechanisms in the Harvard catalogue are shown (pale gray). All mechanisms are plotted as lower hemisphere projections throughout the paper. The arrow indicates the motion of the India–Australian plate relative to the Eurasian plate [DeMets et al., 1994b]. The bathymetry is shaded in this and in subsequent maps [from Smith and Sandwell, 1997].
Wharton Basin, is cut by many N-S trending fossil transform faults, including the Investigator Fracture Zone, clearly seen in the bathymetry (Figure 1). Recent seismic profiles in the Wharton Basin show evidence that these N-S faults are currently being reactivated as left-lateral strike-slip faults [Deplus et al., 1998]. These various observations imply a change in the orientation of the maximum compressive stress from N-S west of the 90°E ridge, to more NW-SE in the east [e.g., Cloetingh and Wortel, 1986; Kreemer et al., 2000].

Studies of plate motion in the Indian Ocean region cannot simultaneously fit data along the three mid-ocean ridge systems that meet at the triple junction between India–Australian, Antarctic, and African plates [Minster and Jordan, 1978; Gordon et al., 1987]. These observations have led to proposals that the India–Australia plate should be considered as two or more separate plates, divided by a distributed zone of deformation. Initially Minster and Jordan [1978] and Okal and Stein [1978] proposed a two-plate model divided by a broad N-S zone of distributed deformation parallel to the 90°E ridge. Wiens et al. [1985], Gordon et al. [1990], and DeMets et al. [1994b] preferred a broad equatorial plate boundary striking roughly E-W. The Wharton Basin earthquake occurred about 5° south of the distributed

Figure 2. (a) Map showing the seismotectonic setting of the Enggano earthquake sequence. The Harvard CMT mechanism for the 4 June main shock is joined to its NEIC epicenter (black star) and CMT centroid (white star). The mechanisms of earthquakes in the Harvard catalogue $M_w \geq 6.5$ are plotted in gray at their centroid locations joined to their NEIC epicenters. All seismicity in the ISC catalogue from 1973–1998, relocated as described in the text, is plotted (small gray circles). All the NEIC located aftershocks from June to October 2000 are plotted as black circles. Note how well the aftershock sequence matches the location of the area of high background seismicity. The approximate rupture area of the 1833 subduction zone earthquake [Newcomb and McCann, 1987; Prawirodirjo et al., 1997] is marked (thick gray box), and the area shown in (b) is outlined by the black rectangle. The bathymetry is shaded as in Figure 1. Enggano Island is where the maximum damage was reported [Pan et al., 2000]. (b) Close-up map of the Enggano earthquake sequence. The mechanisms of all earthquakes in the Harvard CMT catalogue before June 2000 (pale gray) and those of the earthquakes occurring in June–October 2000 (dark gray) are plotted at their relocated epicenters. The mechanisms of two large (M6.8 and M7.3) earthquakes that occurred in January and February 2001, after this work was completed, are labeled 01. All earthquakes located by the NEIC from June to October are shown as black circles.
zone of deformation proposed by Gordon et al. [1990] (Figure 3). More recently, a three-plate model has been proposed to explain the region of complex deformation in the central Indian Ocean (Figure 3) [Royer and Gordon, 1997].

The two large earthquakes that occurred on 4 and 18 June 2000 can provide critical constraints on the nature of present-day deformation in the complex region of the Indian Ocean. We perform waveform modeling and relocate aftershocks to determine the fault planes of the two earthquakes. We then interpret our results in the context of the regional seismotectonics. We also consider what we can learn about the deformation of oceanic lithosphere from the source properties of the two earthquake sequences.

2. 4 June 2000 Enggano Earthquake Sequence
2.1. Broadband Body Wave Modeling of the Main Shock

[8] We model teleseismically recorded broadband body waves to investigate the rupture geometry of the Enggano earthquake. The seismograms are relatively long and complex (Figure 4), even for an earthquake of this size. We use all the available recordings of both $P$ and $SH$ waves at Global Seismic Network (GSN) stations. We correct for the instrument responses to obtain displacement seismograms filtered between 1 and 100 s. We then invert for the moment tensor, source time function, depth, and rupture directivity using the method developed by Ekström [1989]. We correct for attenuation using a $t^*$ (travel time divided by average $Q$) of 0.6 s for $P$ waves and 3.0 s for $S$ waves and a layered velocity model over a half-space, with a 23 km thick crust (essentially PREM) and 2 km of water. We pick $P$ and $S$ arrivals for the inversion, but fix the source at the NEIC epicenter. Initially, we invert only $P$ waves and predict the $S$ waves to assist in the correct picking of the $S$ wave arrivals. The method allows us to include the CMT solution as a constraint in the body wave inversion, but we do not use this option. If we allow the moment tensor to include a non-double-couple component, we can fit the waveforms with a single mechanism similar to that of the CMT solution.
When we constrain the source to be a double couple, however, we cannot fit the waveforms with a single mechanism. A close look at the P wave onsets shows that they require a strike-slip solution (Figure 4). For example, note the change from positive first motion in the NE (e.g., TLY round to PMG) to negative in the SE (CAN, SNZO, and VNDA). The later phases (24–75 s after the P onset) are the same polarity in the NE and SE, implying a change of mechanism. We thus try inverting for two separate mechanisms. We fix the first mechanism at strike-slip solutions consistent with the polarity and amplitude of the first motions (which provide tight constraints of ±2°C) (Figure 4) and invert for the mechanism and source parameters of the second subevent. We vary the onset time of the second subevent and the duration of both subevents. Initial inversions demonstrate that the second subevent has to be predominantly reverse faulting, striking approximately NW-SE. When we find a solution that fits the P waves well, we can use the S waves predicted by the P wave model to assist in picking the S arrivals in the data. We can then include the S waves in the inversion. They provide sufficient additional constraints to enable us to invert for all the source parameters of both subevents and obtain a stable solution.

We also consider the possibility that the second, reverse faulting subevent is simply an artifact produced by using a simple 1-D velocity model to match waveforms that were excited in a dipping structure. Wiens [1989] showed that dipping interfaces between sediments, crust and mantle have a relatively small effect on waveforms, but that reverberations in a deep water layer above a dipping seafloor produce late arriving pulses, which are not present in a horizontal geometry and that can easily be mistaken for source complexity. Such water multiples are only observed in the P waves, and the SH waves are unaffected. Both P
Figure 4. Broadband body wave modeling of the Enggano (4 June) earthquake, showing the observed $P$ and $SH$ seismograms (solid) and synthetics (dash) for our preferred model. The vertical lines on each seismogram indicate the time window used in the inversion. (Those seismograms without such lines were not used to constrain the inversion and are simply forward modeled because of their epicentral distance and high noise level or to balance the azimuthal distribution in the inversion.) The vertical arrows are the picked onset times, and the numbers to the right of the seismograms are the maximum amplitudes in microns. The stations are located on the focal mechanisms of the first of the two subevents, and the moment rate function of the first subevent is shown beneath the $P$ wave mechanism. Both subevent mechanisms are shown at the bottom. The strike, dip, and rake of the two mechanisms are $194^\circ$, $82^\circ$, and $-1^\circ$, and $330^\circ$, $37^\circ$, and $72^\circ$, respectively. The arrows on the focal mechanisms indicate the direction of rupture propagation (at velocity $V_R$). The moment rate functions of the two subevents are also shown, separately and combined.
and SH seismograms of the Enggano earthquake include the second subevent. Also, Wiens [1989] found that such reverberations are insignificant for earthquakes at the relatively large depth and shallow water layer of the Enggano earthquake. Complex reverberations in the dipping sedimentary layers could produce both complex P and SH waveforms, but they should exhibit strong azimuthal variation not observed in the waveforms of the Enggano earthquake. In addition, no such phases are observed in the P and SH seismograms of the largest aftershocks. Therefore, we conclude that the reverse faulting subevent is real and not an artifact of an oversimplified model velocity structure. It is probable that structural heterogeneity is contributing to the smaller-scale complexity of the waveforms, especially the later arrivals, and so we do not try a more complicated source model that attempts to match every detail.

[15] We estimate the uncertainties in our preferred 2 subevent solution using the measured RMS misfit between observed and synthetic seismograms and also noticeable deterioration in the visual fit, especially the polarity of the first motions. We find that there is a relatively broad range of solutions that can fit the data adequately, with little change in RMS. The first motions constrain the mechanism of subevent 1 to within about 2°, but the SH waves are slightly better fit by a mechanism rotated about 10° anticlockwise. This discrepancy may represent a curved fault plane, but we do not have enough resolution to confirm this. The depth of subevent 1 can vary ±10 km with no significant change in RMS. The first subevent is a single pulse of slip lasting about 30 s and includes up to two-thirds of the moment. The second subevent starts 13 s after the first; the onset is well constrained by arrivals in the seismograms not generated by the strike-slip source. The second subevent involves three well-resolved pulses spread out over almost 100 s. The fourth small pulse at ∼100 s (Figure 4) is poorly resolved; shortening the solution to exclude this fourth pulse has a negligible effect on the RMS error, whereas shortening it to only the first two pulses (50 s) increases the RMS error by over 10%. The depth of the second subevent can vary by ±10 km with negligible effect on the waveform fits. The west dipping plane of the mechanism is the better constrained. The orientation of the null axis can vary by about 15° and the strike and dip of the east dipping plane between 300° and 315° and 28° and 40°, respectively, without significant increase in RMS. The moment of the second subevent varies from about a third to a half of the total, depending on its orientation. Both subevents show significant directivity. The first subevent ruptured to the north, which implies that the NS plane is the fault plane. The second subevent ruptured to the SE, along the strike of the subduction zone. Fixing either source as a point source increases the RMS error by about 5% and results in an obvious decrease in visual fit. We thus believe the directivity in both subevents to be resolvable and the rupture propagation velocity to be ∼2 km/s. The sum of the moment tensors of the two subevents is close to the Harvard catalogue CMT solution derived from long period body and mantle waves.

2.2. Aftershocks and Background Seismicity

[16] The 4 June Enggano earthquake had an active aftershock sequence, including 51 earthquakes M ≥ 5 in the first month (Figure 2). The epicentral distribution is close to the pattern of background seismicity in this active region. This seismicity pattern has been stable over the last few decades, with about 8 earthquakes M ≥ 5 each year in the area of Figure 2b. The aftershock distribution is also consistent with our preferred rupture geometry of the Enggano earthquake. Most aftershocks are located to the SE of the main shock epicenter, extending 150 km along the subduction interface, coinciding with the inferred rupture area of the second subevent. The aftershocks also extend over 50 km north of the main shock epicenter, along the preferred fault plane of the first subevent. The Harvard CMT mechanisms of the larger aftershocks exhibit considerable diversity (Figure 2). Some are consistent with reverse slip on the plate interface, and the others are mostly strike-slip with nodal planes in various orientations. To confirm the reliability of the CMT mechanisms, we model the body waves of the three largest aftershocks with very different CMT mechanisms. Our results are consistent with the CMT catalogue solutions, and thus we think that the catalogue mechanisms are reliable and that the diversity is real.

[17] In order to investigate further the rupture geometry of the sequence we relocate the aftershocks recorded by the NEIC from June to October 2000, using the hypocentral decomposition method of relative relocation developed by Jordan and Sverdrup [1981] and Bergman and Solomon [1990]. We use direct and reflected P and S phases reported by the NEIC in the inversions. The largest earthquakes are usually the best recorded, so we start with those aftershocks with M ≥ 5. We discard the ones with the smallest number of recorded phases to enable the inversion to converge. We then add the smaller earthquakes with the largest number of reported phases, resulting in a catalogue of 64 earthquakes with hypocentral relocation uncertainties of less than 15 km (all but 8 are under 10 km), and these are plotted in Figure 5. The epicentral distribution differs insignificantly from that of the NEIC, but our relocations have well resolved depths, whereas all but the three deepest aftershocks are fixed at 33 km depth by the NEIC.

[18] We then apply the same relocation technique to the earthquakes recorded by the ISC between 1973 and 1998 along the Sumatran subduction zone. We divide the region from 0°S to 7°S into overlapping rectangles, allowing comparison of hypocenter relocations derived for earthquakes included in more than one inversion. In this relocation, we use direct and reflected P and S phases reported by the ISC. Our relocated catalogue includes many earthquakes that have larger location errors than the aftershock sequence, with depth errors of up to 25 km. These relocations significantly improve our resolution of the subducting slab compared to the original ISC locations, which show only a very diffuse dipping structure over 100 km thick. Our relocations are sufficiently accurate to provide the context for the Enggano earthquake sequence that is necessary here. Further work to improve these locations and investigate the spatiotemporal patterns of seismicity along the Sunda Arc is planned.

[19] The main shock is included in the relocation of the aftershock sequence and has a hypocentral depth of 38 ± 6 km (95% confidence ellipse), shallower than the broadband centroid depths but within the error bounds. It is plotted
separately in Figure 5b. The relative depths of the hypocenter and centroids suggest downward rupture. The body wave inversion results predict more horizontal rupture directions, but the resolution of the data cannot rule out a component of downward rupture.

[15] The rupture propagation in the second subevent is about 2 km/s to the SE, and the rupture lasted about 90 s. Thus, the halfway point is approximately 100 km SE of the hypocenter, and it is at this point that we plot the mechanism of the second subevent in Figure 5. This location does not imply any preference for the actual slip distribution, but is simply chosen to represent the preferred geometry from the broadband modeling.

[16] The maximum depth of the aftershocks decreases to the SE (Figure 5, cross section A), and several aftershocks are at similar depths to the centroid of subevent 1. The thickness of the subducting slab defined by our relocated background seismicity is poorly constrained, due to the relatively large uncertainties in the locations, but subevent 2 and the related thrust aftershocks are consistent with slip at the slab interface (Figure 5, cross section B). The dips of the nodal planes of these earthquakes are also in good agreement with the dip of the subducting oceanic lithosphere. Subevent 1 is deeper and, within the error range can confidently be considered within the subducting slab. Cross section C (Figure 5) is through the main shock hypocenter, approximately perpendicular to our preferred N-S fault plane for subevent 1. Three aftershocks lying to the south of subevent 1 (Figure 5a) have similar strike-slip mechanisms to subevent 1 and so are consistent with our interpretation of a N-S fault plane within the subducting slab.

2.3. Summary

[17] Our preferred source model for the Enggano earthquake consists of two separate subevents. The first (∼65% of the moment) involves strike-slip motion on a near vertical,
approximately N-S trending fault within the subducting slab, lasting ~30 s. The rupture propagated toward the north from the hypocenter. This first subevent then triggered reverse slip on the subduction interface, rupturing toward the SE. This second subevent consisted of at least three pulses of slip, lasting a total of ~90 s. The locations of the aftershocks provide further support for this interpretation. The orientation of the strike-slip subevent would lead to increased static shear stress on the plate interface to the SE, consistent with triggering slip there in the second subevent.

3. 18 June Wharton Basin Earthquake

We perform a similar analysis of the 18 June Wharton Basin earthquake, combining broadband body wave modeling and aftershock relocation to investigate the rupture geometry.

Robinson et al. [2001] analyzed the Wharton Basin earthquake and obtained an unusual rupture geometry involving slip on two perpendicular strike-slip faults. They recalculated the long period moment tensor and determined that the large non-double-couple component in the Harvard CMT solution (35%) is not well resolved. We investigate the uncertainties in the Harvard CMT solution and also find that the best double couple solution (strike, dip, and rake are 168°, 53°, and 11°, respectively) has an RMS error insignificantly higher than that of the solution in the catalogue. Thus the non-double-couple component cannot be resolved from the long period waves alone. We do not use the Harvard CMT as a constraint on the body wave modeling.

3.1. Broadband Body Wave Modeling

We model the teleseismically recorded body waves of the Wharton Basin earthquake following the same procedure described for the Enggano earthquake. We use both P and S waves, unlike Robinson et al. [2001] who only attempt to model the S waves. The waveforms (Figure 6) are significantly shorter and less complex than those of the Enggano earthquake. We use a velocity model with 5 km of water and a 6 km crust over a half-space. We can model the broadband body waves with a similar non-double-couple mechanism to the Harvard CMT. We then attempt P wave inversions constraining the mechanism to be a pure double couple. Like the Enggano earthquake the P wave first motions tightly constrain the first subevent to be strike-slip (Figure 6), with negative polarity onsets in the NW and SE and positive in the other two quadrants. Such a mechanism can fit many of the seismograms but it cannot fit the P waves in the NW, nor the SH waves in the ENE. The P wave onsets in the NW are negative polarity but a larger, positive polarity phase that arrives within 5 s after the onset cannot be modeled with a pure strike-slip mechanism. This arrival is too early and too large to be any kind of water reverberation. We thus try adding a second subevent and invert for two subevents with different mechanisms, varying the time delay between them from 0 to 20 s.
Our preferred model is shown in Figure 6. The sum of the two subevents is close to the Harvard CMT mechanism, including the non-double-couple component. Again we estimate the uncertainties in our solution using the calculated RMS error, the visual fit to the waveforms, and the consistency of results of inversions with different starting parameters and constraints. The mechanism of the first subevent is tightly constrained by the first motions to ±2°. Varying the depth of the first subevent by ±5 km increases the RMS error by 4%. The directivity is toward the north and the visual fit to the waveforms decreases significantly (the RMS error increases by 20%) if the solution is fixed as a point source. Thus the N-S plane is the likely fault plane. The location of the Harvard catalogue centroid to the north of the hypocenter is consistent with this interpretation (Figure 7). The second subevent is less well resolved. Varying the depth by ±10 km gives only a 4% increase in RMS error. The mechanism is predominantly reverse, with the null axis to the SW. Varying the strike by 10° increases the RMS error by 5%. The poorer constraints result from the relatively small size of this subevent (25% of the moment) and the fact that it occurs entirely within the rupture duration of the larger, first subevent. Including directivity in the second subevent gives only a negligible improvement to the fit, and the orientation is not consistent from inversion to inversion. Thus we do not believe that the rupture

Figure 6. Broadband body wave modeling of the 18 June Wharton Basin earthquake. The layout and symbols are the same as in Figure 3. A number of stations are omitted from the inversion to give a more balanced azimuthal distribution. The strike, dip, and rake of the two mechanisms are 337°, 77°, and 17° and 206°, 40°, and 30°, respectively.
direction can be resolved or used to distinguish the fault plane for this second subevent. The onset time and orientation of the second subevent are constrained by the need to fit the $P$ waves to the NW and SW. The orientation of the second subevent is consistent with that of the regional stress field, which has maximum compression in the NW-SE direction. It is also similar to that of a late aftershock that occurred in September 2001 (Figure 7), just to the NE of the main shock.

### 3.2. Relocation of Aftershocks

[22] The Wharton Basin earthquake had only 21 aftershocks located by the NEIC during 2000, of which two were large enough to be included in the Harvard CMT catalogue (Figure 7). The ISC and NEIC catalogues (since 1970, complete above M ~ 5) contain no previous earthquakes in the area of Figure 7. The depths of the aftershocks as reported by the NEIC are all fixed at 10 km. We attempt to relocate all of the aftershocks recorded by the NEIC between June and September 2000, but had to discard poorly recorded earthquakes to enable the inversion to converge. Inverting for depth proved unstable, and so we fix all the depths at 10 km. The relocated epicenters of the main shock and the twelve aftershocks with uncertainties <10 km are shown in Figure 7. The relocated epicenters are close to their corresponding NEIC locations, but the NNW-SSE alignment is a little more tightly defined after relocation. This agrees well with the preferred fault plane for subevent 1. The other aftershocks could be interpreted as off-fault events. Off-fault aftershocks are common following strike-slip earthquakes as a result of the static stress changes induced by the main shock [e.g., Das and Scholz, 1981]. Alternatively, those aftershocks to the NE could be related to the second subevent. In September 2001, a M6 reverse faulting earthquake occurred NE of the Wharton Basin earthquake, with a mechanism similar to that of subevent 2 (Figure 7).

[23] The Wharton Basin earthquake occurred near to the Investigator Fracture Zone but, although the fault orientation is similar, the locations are too far to the west (~75 km) for the earthquake to have ruptured the ridge identified as the Investigator Fracture Zone, and it must have occurred on a subparallel feature. Robinson et al. [2001] proposed that a trough just to the west of the Investigator Fracture Zone (fracture F, Figure 7) may have been the main fault plane. Accordingly, they shift the relocations of all the earthquakes in the sequence 50 km to the east. The orientation of fracture F is more N-S than that of both the main shock fault plane and the alignment of the majority of the aftershocks contrary to their conclusion. In addition, a systematic shift of 50 km is large, especially in a region where there is no strong lateral heterogeneity in velocity structure as, for example, in a subduction zone. Antolik et al. [2001] and Smith and Ekström [1996] performed location tests with earthquakes M5.5 and above and observed RMS misfits of 12–14 km using the same velocity models as the NEIC and ISC, and so it is unlikely that this entire set of earthquakes could be 50 km or more mislocated by the NEIC or ISC. Therefore, we do not prefer any particular feature in the bathymetry as the main shock fault plane.

### 3.3. Summary and Comparison With the Study of Robinson et al. [2001]

[24] Our preferred model of the Wharton Basin earthquake involves two subevents. The first (75% of the moment) is left-lateral strike-slip on a near vertical, almost NS plane, lasting just under 30 s. Our model for the first subevent is very similar in moment, slip orientation, and dimension to that of Robinson et al. [2001]. The rupture directivity and hypocenter, centroid and aftershock locations support this interpretation and also imply rupture to the north. The fault plane is subparallel to the Investigator Fracture Zone, but too far away to have occurred on this fracture. The second subevent is much smaller and relatively poorly resolved as it occurred contemporaneously with the first subevent. It was predominantly reverse slip, with the null axis dipping to the SW and a fault plane striking either SW-NE or E-W. We cannot resolve which nodal plane is the fault plane.

[25] Our preferred model for the second subevent differs significantly from that of Robinson et al. [2001]. The main reason for the difference between the two models is that we model both the $P$ and $S$ waves, whereas they only consider the $S$ waves. To compare the two models, we fix our inversion parameters to their preferred solution and obtain
the model shown in Figure 8. We use their mechanisms, depths, rupture velocities, and temporal and spatial separation between the two subevents. We invert only for moment rate functions. The moment of the first subevent is close to their published model, but that of the second is smaller. Our preferred model is able to fit both $P$ and $S$ waves from all quadrants (Figure 6). The model of Robinson et al. [2001] does not fit the $S$ waves any better than ours (Figure 8). The poor fit to stations SSE and YSS (Figure 8) would perhaps be improved if the second subevent were larger, as in their published model. The location of these stations near the $SH$ node for both subevent mechanisms suggests that the poor fit is more likely the result of incorrect model source orientation. More importantly, the model of Robinson et al. [2001] does not fit the $P$ waves to the NW and SW. Robinson et al. [2001] claimed that the $P$ waves cannot be modeled because of reverberations in the water layer. We do not see any indication of such reverberations in the part of the waveforms that we model. Modeling strike-slip earthquakes in the Atlantic Ocean with and without a water layer made a negligible difference to the shape of the synthetics, suggesting that such reverberations are small for strike-slip mechanisms in relatively flat-lying strata [Abercrombie and Ekström, 2001]. The $P$ phase that Robinson et al. [2001] failed to model is the phase with the largest amplitude recorded at stations to the NW and starts only about 5 s after the onset. It is the occurrence of this phase that corresponds to the onset of our second subevent and requires it to have a primarily reverse slip orientation. Simply increasing the moment of the second subevent in Robinson et al.’s model cannot produce this phase, even if it did start significantly before the 12 s delay time that they propose. The relatively poor fit of the rupture geometry proposed by Robinson et al. [2001] to the larger data set used here suggests that their results for the location, timing and orientation of slip are not well resolved.

4. Results and Discussion

4.1. Seismotectonics

[26] Our preferred interpretation of both the 4 and 18 June 2000 earthquakes is that they primarily involve left-lateral slip on N-S trending near-vertical faults (Figure 1). Although the faults are not well known, the Wharton Basin is cut by many parallel N-S fossil fracture zones. The 18 June earthquake is close to (though not on) the Investigator Fracture Zone, which is the most prominent fracture zone east of the 90°E Ridge. Newcomb and McCann [1987] cite evidence for fracture zones at 102.5° and 104.5°, based on age offsets and sediment filled troughs. The westerly of these two is a likely candidate for the fault that ruptured in the 4 June Enggano earthquake. Previous seismicity east of the 90°E Ridge suggests that some of these fossil fracture zones are currently active, and DePluss et al. [1998] find clear evidence for reactivation of N-S fractures in a left-lateral sense on seismic profiles through the Wharton Basin. The two June earthquakes thus confirm that these N-S fractures are active and capable of moving in large earthquakes.

[27] The direction of strike-slip motion in both of these earthquakes is consistent with recent deformation models of the Indian Ocean (Figure 3) [Roy et al., 1997]. The location of the Wharton Basin earthquake confirms that the diffuse boundary extends over a wider region than initially defined by Gordon et al. [1990]. The orientation of the earthquakes is also consistent with the present-day orientation of stress and strain in this part of the India–Australia plate [Cloetingh and Wortel, 1986; Kreemer et al., 2000].

The occurrence of the 4 June earthquake implies that this deviatoric stress field exists in the oceanic lithosphere that has been subducted to a depth of at least 50 km. This interpretation is consistent with a study of stress in the Sunda arc by Slancová et al. [2000]. They used CMT solutions of earthquakes prior to 1997 to identify domains with similar stress orientations within the subducting slab and observed a domain with compression parallel to the trench in the depth range 25–225 km along the length of Sumatra. They suggested that this might represent stresses in the oceanic plate continuing into the subduction zone. The continuation of the deviatoric stress field in the oceanic lithosphere into the subducting slab may also explain the seismicity on the subducting Investigator Fracture Zone, recorded by a local network [Faure et al., 1996]. A band of earthquakes, consistent with faulting on a near vertical N-S plane follows the subducting Investigator fracture zone to a depth of about 150 km. Faure et al. [1996] noted that such seismicity is unusual as most subducting fracture zones are aseismic. Their preferred explanation is that the seismicity is related to lateral compression in the slab caused by its local geometry. They did not observe a tear in the slab, but were unable to resolve a vertical offset in hypocenters of less than 20 km.

[28] The 4 June Enggano earthquake is quite similar to the 1986 (Mw 7.7) Kermadec earthquake in many ways. Houston et al. [1993] showed that the Kermadec earthquake involved primarily strike-slip faulting in the subducting slab at a depth of about 45 km. Thus it represents another example of tearing of the subducting slab. The Louisville Seamount Chain is subducting just north of the location of the Kermadec earthquake and so stresses involved with this could have produced the tearing in 1986. The Louisville Seamount Chain follows the trace of the fossil Eultan fracture zone, similar to our present understanding of the 90°E Ridge [Herzer et al., 1987]. Thus it is possible that both the 4 June 2000 and the 1986 Kermadec earthquakes ruptured fossil fracture zones in the subducting slab. The 1986 Kermadec earthquake did not include reverse slip on the plate interface, like that observed in the Enggano earthquake, but most of the aftershocks, including the largest (Mw 6.1), were interplate thrusts [Houston et al., 1993]. The 1994 Kurile Islands earthquake (Mw 8.3) has also been interpreted as a tear within the subducting slab, but it did not trigger slip at the plate interface [Tanioka et al., 1995]. The more recent December 1999 Kodiak Island earthquake was also intraplate with an oblique strike-slip mechanism, but aftershock relocations imply that it ruptured a steeply dipping fault plane parallel to the trench and was driven by downdip tension in the slab [Ratchkovski and Hansen, 2001]. The Enggano earthquake is also consistent with downdip tension in the slab, but our analysis implies that the trench-parallel nodal plane is not the fault plane.

[29] The strike-slip subevent of the 4 June Enggano earthquake triggered a second subevent that ruptured the plate interface to the SE. The hypocenter and strike-slip subevent are near the southern extent of the great 1833
rupture (Figure 3). The second subevent thus ruptured away from the 1833 rupture zone. Simple static stress modeling suggests that subevent 1 increased the stress on the plate interface to the SE, in the region of the second subevent. The uncertainty in the distribution of slip in the subevents of the Enggano earthquake prevents us from calculating a meaningful model of the static stress change induced by this earthquake in the rupture area of the 1833 earthquake. The slip on the plate interface would have increased the shear stress on the neighboring parts of the interface, including the 1833 rupture area. The aftershocks are all in an area of continued high activity over recent decades. The degree of seismic coupling in this region is not well known. Prawirodirdjo et al. [1997] show that Enggano Island is

![P Waves](image)

Subevent 1
Depth = 15 km
$M_0 = 4.66\times10^{20}$ Nm
$V_R = 3.1$ km/s

Subevent 2
Depth = 15 km
$M_0 = 2.83\times10^{18}$ Nm
$V_R = 4.0$ km/s

![SH Waves](image)

Figure 8. Broadband body wave modeling of the 18 June Wharton Basin earthquake [after Robinson et al., 2001]. The layout and symbols are the same as in Figure 4, and the same stations are included in the inversion as in Figure 6. We fix the orientation, depth, and rupture velocities at their published values and invert for the moment rate function. The strike, dip, and rake of the two mechanisms are 165°, 87°, and −2° and 75°, 82°, and 173°, respectively. The moment of the first subevent is close to the published model but that of the second subevent is smaller in our inversion.
moving with the forearc and the Indian plate, suggesting that the subducting plate was coupled to the overlying plate in this region prior to the 4 June earthquake. It will be interesting to see what motions are observed in the region after the Enggano earthquake.

4.2. Source Properties

[30] The two earthquakes considered in this study are among the largest and best-recorded oceanic events. They can thus provide information about the nature of fracture in the oceanic lithosphere. They are both mainly intraplate strike-slip earthquakes, although the Enggano earthquake has a substantial subduction component. We can estimate the rupture areas of the various subevents, using the durations and rupture velocities determined from the body wave modeling and also from the aftershock areas. We can then calculate the stress drops, following the study of Eshelby [1957]. Both earthquakes occurred in ~70 Ma lithosphere and so can be assumed to have rupture widths of 30–40 km (corresponding to a brittle–ductile transition at 600–700°C) [Wiens and Stein, 1983; Abercrombie and Ekström, 2001]. The length of rupture in the strike-slip subevent of the Enggano earthquake was ~50–100 km, and in the Wharton Basin earthquake, ~80–120 km. These rupture areas imply stress drops of 10–30 and 5–10 MPa for the earthquakes, respectively. Robinson et al. [2001] also estimated a relatively high stress drop for the Wharton Basin earthquake. These values are similar to the stress drop of 16 MPa estimated by Antolik et al. [2000] for the 1998 Antarctic earthquake (Mw 8.1) which was the largest intraplate strike-slip earthquake recorded to date. The 1986 Kermadec earthquake appears to have been similar to the Enggano earthquake and that had a stress drop (dominated by the strike-slip component) of 10–20 MPa [Houston et al., 1993]. In contrast, assuming a downdip width of 50–70 km and length of 150–200 km for the second, subduction subevent of the Enggano earthquake gives a lower stress drop of only ~1 MPa. This is significantly lower, despite the wide range of possible rupture areas considered in the calculations to reflect the resolution of the parameters. Choy and Boatwright [1995] performed a global study of earthquake apparent stress, an alternative estimate of stress drop. They found that oceanic intraplate and oceanic transform strike-slip earthquakes have the highest apparent stress and that the apparent stress of subduction zone earthquakes is significantly lower. Thus our observations are consistent with this global trend. In the case of the Enggano earthquake, the low stress drop could be related to the degree of seismic coupling. The plate interface may be slipping in part aseismically. The overall rupture area of the earthquake, therefore, could include large regions that slipped aseismically during and following the seismic slip. Estimating the stress drop of subduction earthquakes in such regions from the total dimensions and the seismic radiation could therefore produce an underestimate of the stress drop.

[31] The number of aftershocks is also different for the strike-slip and plate interface subevents of the Enggano earthquake. Typically continental earthquakes have one after- shock within an order of magnitude of the main shock, 10 within 2 orders of magnitude and so on (Bath’s Law, first cited by Richter [1958]). Subduction zone earthquakes have similarly active aftershock sequences. Both the strike-slip subevents of the Enggano and Wharton Basin earthquakes have few aftershocks. The Enggano earthquake had only three strike-slip aftershocks within 2 orders of magnitude and the Wharton Basin had only one. Houston et al. [1993] noted only one strike-slip aftershock within 2 orders of magnitude of the 1986 Kermadec earthquake, and the December 1999 Kodiak Island earthquake had only two such aftershocks. Other examples of oceanic strike-slip events with few aftershocks include the 1994 Romanche transform earthquake (Mw 7.1) which had no recorded aftershocks at all (NEIC and ISC, complete above about the M4.5) and the 1994 (Mw 6.5) earthquake on the Blanco transform which had only two aftershocks within 2 orders of magnitude [Dziak et al., 2000]. The 1998 Antarctic earthquake (Mw 8.1) had only 26 aftershocks recorded by the ISC, of which only two were within 2 orders of magnitude of the main shock. A recent global study by Boettcher and Jordan [2001] also finds that oceanic transform earthquakes typically have few aftershocks. By contrast, both the Enggano and Kermadec earthquakes had large numbers of plate interface aftershocks, suggesting that the subduction zone earthquakes are more like continental earthquakes in their ability to trigger aftershocks.

[32] Both the stress drop and number of aftershocks are of interest as they are first-order observations as to the differences in earthquake rupture in continental and oceanic lithosphere. Kanamori and Anderson [1975] observed a difference in stress drop between interplate and intraplate earthquakes and interpreted it as a dependence on strain rate and recurrence time. It is hard to define whether the seismicity in the Indian Ocean is strictly intraplate or interplate, however, since it is considered part of a broad zone of deformation.

5. Conclusions

[33] The 4 June Enggano and 18 June Wharton Basin earthquakes were both primarily left-lateral strike-slip on N-S, vertical faults, one within the subducting slab and the other beneath the Indian Ocean. Both earthquakes had a second, reverse slip subevent. The purely strike-slip ruptures on conjugate planes proposed for the Wharton Basin earthquake [Robinson et al., 2001] are inconsistent with the P waves and so can be discounted. In the Enggano earthquake the reverse slip subevent was a plate interface thrust comprising at least 35% of the total moment and rupturing to the SE away from the rupture area of the great 1833 earthquake.

[34] The occurrence of these earthquakes indicates that fossil fracture zones in the Indian Ocean are being reactivated in the current stress field as strike-slip faults and that they can rupture in large earthquakes. The stress field causing this reactivation is present to depths of at least 50 km in the subducting slab.

[35] Both strike-slip subevents have high stress drops and few aftershocks, consistent with previous observations of transform and oceanic strike-slip earthquakes. In contrast the subduction thrust subevent had an active aftershock sequence and a lower stress drop, typical of subduction earthquakes. These differences between earthquakes in oceanic and continental lithosphere and at subduction zones are consistent and reliable. They should assist us in under-
standing the state of stress in plates and the role of rheology in governing the nature of earthquake rupture.

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