1 INTRODUCTION

Oceanic spreading centers impart the primary fabric in oceanic crust that may influence faulting as the plate ages and experiences stresses on its boundaries. Spreading ridge transform faults are commonly seismogenic at shallow depths due to the relative plate motions between offset ridge segments being resisted by friction in the near-surface brittle part of the evolving lithosphere (e.g., Sykes, 1967; Engeln et al., 1986; Bergman and Solomon, 1988; Abercrombie and Ekström, 2001). On the intraplate fracture zone extensions, contrasts in the lithospheric age, differential subsidence rates, thermal contrasts, and bathymetry sustain minor levels of faulting near to the ridge system that drop off rapidly with distance, leaving fracture zones largely aseismic. An exception is the July 15, 2003 $M_w$ 7.6 right-lateral strike-slip rupture that extended about 210 km along a fracture zone in young (<12 Ma) Indian plate lithosphere immediately adjacent to the Central Indian Ridge (e.g., Bohnenstiehl
et al., 2004; Antolik et al., 2006). The corresponding active transform fault on the ridge has left-lateral strike-slip faulting. Strike-slip faulting on fracture zones usually has reversed sense of motion relative to the active transform fault region (i.e., the sense of fracture zone faulting is similar to the ridge segment offset) (e.g., Engeln et al., 1986; Robinson, 2011). Off-ridge normal faulting and strike-slip faulting diminish progressively as lithospheric age increases to about 20 Ma in most plates, while reverse faulting extends across most lithospheric ages (e.g., Wiens and Stein, 1984). The seismogenic expression of fracture zones is thus mostly localized to the near-ridge environment or located far from the ridge. Activity on intraplate fracture zones may be induced by off-ridge volcanism (e.g., Okal and Stewart, 1982), but generally is attributed to large-scale plate stresses and deformation (e.g., Stein and Okal, 1978; Robinson, 2011).

Oceanic lithosphere older than 40 Ma exhibits increased strike-slip activity, including some very large earthquakes and it is commonly assumed that fracture zones may be involved as they represent lithosphere-traversing fault systems that may retain deviatoric stresses from bathymetry and thermal contrasts and may have disrupted mechanical strength due to the presence of gouge and fluid accumulations. However, it actually does not appear that intraplate fracture zones are particularly weak relative to surrounding oceanic lithosphere (e.g., Sandwell and Schubert, 1982; Bergman and Solomon, 1992), so the influence of fracture zones on large intraplate seismicity remains unclear. Here we consider large oceanic intraplate activity at distances >100 km from plate boundaries (including away from the outer-rise region near subduction zones where large normal faulting associated with plate bending occurs), to evaluate the role played by preexisting fracture zones.

2 THE LARGEST OCEANIC INTRAPLATE EARTHQUAKES AWAY FROM BOUNDARIES

Information about global oceanic seismicity is limited to the seismological record, which dates back to about 1900. Restricting our attention to intraplate seismicity more than 100 km from mid-ocean ridges and subduction zones, with long-period seismic magnitude ($M_W$ or $M_S$) $\geq 7.8$, the US Geological Survey National Earthquake Information Center (USGS-NEIC) database contains only 9 events (Table 1) dating back to 1900 (drawing upon the revised magnitudes in the ISC-GEM catalog). The seismic moment tensors and locations of these nine events are shown in Fig. 1. The source mechanisms are all predominantly strike slip, although several of the events have significant non-double-couple components. These largest of oceanic intraplate earthquakes are few in number, but widespread, with events located in the Atlantic, Pacific, Antarctic, and Indo-Australian plates. The May 26, 1975 Azores and December 23, 2004 Macquarie events are relatively close to plate boundary transform faults, but are clearly offset from the main active boundary. Events in the Gulf of Alaska and offshore of Sumatra are relatively close to subduction zone boundaries, but are seaward of the outer rise. We briefly discuss the relationship of these large events to the oceanic fracture zone structure.
### TABLE 1  The Nine Largest Intraplate Strike Slip Events in Oceanic Lithosphere

<table>
<thead>
<tr>
<th>Date</th>
<th>Origin Time</th>
<th>Region</th>
<th>Lat. (degree)</th>
<th>Lon. (degree)</th>
<th>Mw</th>
</tr>
</thead>
<tbody>
<tr>
<td>April 11, 2012</td>
<td>08:38:36</td>
<td>Wharton Basin</td>
<td>2.327</td>
<td>93.063</td>
<td>8.6</td>
</tr>
<tr>
<td>April 11, 2012</td>
<td>10:43:10</td>
<td>Wharton Basin</td>
<td>0.802</td>
<td>92.463</td>
<td>8.2</td>
</tr>
<tr>
<td>March 25, 1998</td>
<td>03:12:25</td>
<td>Balleny Islands</td>
<td>−62.877</td>
<td>149.527</td>
<td>8.1</td>
</tr>
<tr>
<td>December 23, 2004</td>
<td>14:59:04</td>
<td>Macquarie</td>
<td>−49.312</td>
<td>161.34</td>
<td>8.1</td>
</tr>
<tr>
<td>May 26, 1975</td>
<td>09:11:51</td>
<td>Azores</td>
<td>35.997</td>
<td>−17.649</td>
<td>7.9(Ms)</td>
</tr>
<tr>
<td>November 30, 1987</td>
<td>19:23:19</td>
<td>Gulf of Alaska</td>
<td>58.679</td>
<td>−142.786</td>
<td>7.9</td>
</tr>
<tr>
<td>March 6, 1988</td>
<td>22:35:38</td>
<td>Gulf of Alaska</td>
<td>56.953</td>
<td>−143.032</td>
<td>7.8</td>
</tr>
</tbody>
</table>

**FIG. 1**  Moment-tensor solutions for the nine largest seismologically modeled intraplate ruptures in oceanic lithosphere well-removed from active plate boundaries. The focal mechanisms for all events are from the global centroid-moment tensor catalog with the exception of the Azores earthquake of May 26, 1975, for which the surface wave solution from Lynnes and Ruff (1985) is shown. Each solution is centered at the event hypocenter other than for the 1988 and 2012 events with offset positions. Faulting directions and relationship to nearby fracture zones for each case are discussed in the text.
3 LARGE EVENTS OUTSIDE THE WHARTON BASIN

The May 26, 1975 $M_S$ 7.9 Azores earthquake (Fig. 1) occurred south of the seismicity trend extending from the Azores Islands to Gibraltar that is thought to be the African-Eurasian plate boundary in the eastern Atlantic. The November 25, 1941 $M_S$ ~8.0 earthquake lies on that trend along the Gloria transform fault 200 km NNW of the 1975 hypocenter. The easterly trending fault plane of the 1975 event is quasi-parallel to the plate boundary, but background seismicity in the source region is localized, so the event is best characterized as intraplate. Lynnes and Ruff (1985) analyzed body and surface waves for the event, preferring a right-lateral mechanism with strike 288 degree, dip 65 degree, and rake 205 degree, with a seismic moment of $7 \times 10^{20}$ Nm. The centroid depth is less than 25 km and the fault length is only 50–100 km. Aftershock relocations suggest eastward propagation of the rupture. There is no clear relationship to local structural features within the African plate.

Two of the largest oceanic strike-slip events struck in the Gulf of Alaska, southeast of the 1964 $M_W$ (9.2) subduction zone earthquake, and southwest of the Yakutat terrane. The sequence began with an $M_W$ 7.2 rupture on an east-west trending fault on November 17, 1987, followed by November 30, 1987 $M_W$ 7.9 and March 6, 1988 $M_W$ 7.8 ruptures along a north-south trend (Fig. 1) (Lahr et al., 1988; Pegler and Das, 1996; Quintanar et al., 1995). There had been no prior large seismicity in the region. Hwang and Kanamori (1992) inverted body waves finding that the centroid depths of the largest two events were 20 and 15 km, with faulting extending to at least 25 km. The large-slip rupture lengths of 110 and 40 km, respectively, are shorter than for comparable size continental strike-slip events, suggesting relatively high stress drop. The north-south trend is parallel to a magnetic lineation, so a fracture zone was not involved, and the horizontal orientations of the principle stresses are attributed to pull toward the northwest from the 1964 underthrusting event and compression from the northeast due to the collision of the Yakutat block with the arc. Quintanar et al. (1995) suggest that these events represent a tear in the Pacific plate.

The December 23, 2004 $M_W$ 8.1 Macquarie event occurred about 150 km from the Macquarie Ridge Complex plate boundary between the Australian and Pacific plates (Fig. 1). The event is located within a deformation zone called the Puysegur (or Macquarie) Block, which has multiple curved fracture zones within it, now striking at oblique angles next to the plate boundary due to progressive increase in translational motion across the former divergent ridge (e.g., Massell et al., 2000; Mosher and Symons, 2008; Hayes et al., 2009). Historic intraplate activity in the southern Tasman Sea indicates broad intraplate deformation, with large events (up to $M$ ~7.8) dating back to 1924 also having locations off the plate boundary (e.g., Valenzuela and Wysession, 1993). The great May 23, 1989 $M_W$ 8.0 strike-slip event (e.g., Ruff, 1990) occurred directly on the Macquarie Ridge. Das (1992, 1993) suggests that some of the much smaller aftershocks of that event are located on a fracture zone oblique to the ridge, southwest of the 2004 event. Kennett et al. (2014) apply back projection of teleseismic signals for the 2004 event and infer an initial bilateral rupture expansion followed by activation of a parallel rupture to the west. Such back projections are, however, unstable due to the near nodal position of downgoing P waves. The relocated 2004 mainshock and aftershocks generally lie along one of the curved fracture zones (Robinson, 2011). Robinson (2011) models broadband P and SH, finding an along rupture change in strike of about 18
degree with right-lateral strike-slip faulting in the north and oblique thrusting in the south, possibly accommodating motion on the curved fracture zone. For slip models restricted to 18 km depth, and a fault length of 150 km, the estimated stress drop is 20 MPa. As for the previously discussed large events, this stress drop is larger than typical continental strike-slip faulting, mainly due to the shorter rupture length for its size. However, the depth constraint is not strong, and allowing rupture to extend to 28 km depth, at least overlapping the global centroid-moment tensor (GCMT) centroid depth of 27.5 km, lowers the stress drop estimate to 8 MPa.

The largest intraplate earthquake to rupture the Antarctic plate is the March 25, 1998 Balleny Islands $M_W$ 8.1 earthquake (Fig. 1). It struck a region with no prior large events about 250 km west of the Antarctic-Australian plate boundary, with aftershocks and body-waveform inversions indicating predominantly left-lateral strike-slip rupture toward the west, transverse to the regional north-south trending fracture zones (e.g., Kuge et al., 1999; Nettles et al., 1999; Antolik et al., 2000; Henry et al., 2000). The rupture traversed lithosphere of 33–40 Ma age along a nearly 300 km long trend, with the largest aftershock locating 100 km south of the mainshock faulting. The GCMT solution has a significant non-double couple, which Kuge et al. (1999) and Antolik et al. (2000) attribute to secondary oblique normal faulting either between en echelon segments or to the east of the hypocenter, respectively.

Henry et al. (2000) invert broadband body waves, favoring an initial 140-km long westward rupture with 45 s duration with slip bracketed between two fracture zones, and a second, dynamically triggered, slip event initiating 40 s later about 100 km further to the west that extends westward for about 60 km. The mechanisms of the subevents have dips of $\sim-69$ degree and rakes of $\sim-18$ degree; this oblique normal component combines with the space-time extent of the rupture to account for the GCMT non-double-couple moment tensor. Henry et al. (2000) estimated large stress drops of 24 and 21 MPa for the first and second subevents, respectively, assuming a 15-km maximum depth of rupture. Hjörleifsdóttir et al. (2009) analyze long-period signals and demonstrate that observed amplitudes and directivity favor continuous rupture along the westward trend with the long-period slip occurring between the slip patches in the model of Henry et al. (2000), rather than separated dynamic triggering. This complicates the estimation of stress drop, as the slip area may be much larger and may extend deeper. The fracture zones appear to play little, if any, role in the rupture. Tsuboi et al. (2000) propose that this event was driven by postglacial rebound, as the orientation of the principle stress axis is consistent with predictions for Antarctic unloading.

Of these large strike-slip events, only the 2004 Macquarie event is convincingly associated with fossil fracture zone reactivation, and for that event the strong intraplate deformation zone caused by the change of relative plate motion appears to be a controlling factor. Thus fracture zones do not necessarily play a major role in the largest intraplate faulting unless the plate is pervasively deforming. Fig. 1 indicates that the Indo-Australian plate experiences the greatest concentration of intraplate faulting in the Wharton Basin south of Sumatra where there is a broad deformation zone, and we focus on that activity for the rest of this review.

4 WHARTON BASIN ACTIVITY

Fig. 2 displays the intraplate seismicity within the broad deformation zone in the Indo-Australian plate around the Wharton Basin. Locations of all events with $M \geq 6.0$ from the
USGS-NEIC database from 1900 to 1975 are shown, along with focal mechanisms for all events with $M_w \geq 5.0$ from the GCMT catalog from 1976 to 2017. Four of the large events in Fig. 1 are labeled on the map. The positions of fracture zones are indicated by white dashed curves (Matthews et al., 2011). The investigator fracture zone is labeled IFZ. The red curve is the fossil Wharton ridge location with the youngest active age of $\sim 38$ Ma.

The diffuse deformation region between the Indian and Australian plates accommodates $\sim 11$ mm/year of internal deformation based on global plate motion models (e.g., Wiens et al., 1985; Gordon et al., 1990; Royer et al., 1997; Delescluse and Chamot-Rooke, 2007; DeMets et al., 2010). The largest earthquakes in this region involve strike-slip faulting; among them is the largest known intraplate rupture of April 11, 2012 ($M_w 8.6$). Lower magnitude thrust faulting is observed in the southern and western portions of the region, extending off of Fig. 2.

The north-south striking fracture zones east of the Ninetyeast Ridge are associated with left-lateral offsets of the Wharton fossil ridge (dark red in Fig. 2), which was active between 38 and 85 Ma and had very rapid spreading (e.g., Deplus et al., 1998; Matthews et al., 2011; Jacob et al., 2014). Strike-slip faulting near these structures is primarily left lateral, assuming rupture of the north-south striking planes, again displaying the typical reversal of motion relative to faulting that would have occurred on the transform fault segments. It is commonly
asserted that the regional earthquake faulting geometry is controlled by the reactivation of fossil fracture zone by present day deformation within the intraplate oceanic basins and subducting slab of the Indo-Australian plate (e.g., Bull and Scrutton, 1990; Deplus et al., 1998; Abercrombie et al., 2003; Delescluse et al., 2008; Rajendran et al., 2011; Geersen et al., 2015; Aderhold and Abercrombie, 2016). There is also extensive activity along the Ninetyeast Ridge, but the actual fault planes are not known for most events (e.g., Stein and Okal, 1978; Sager et al., 2013). There is indeed a general tendency for spatial concentrations of seismic deformation to be aligned with the fracture zone fabric, as are the focal mechanisms, but the regional stress state must also be considered.

The predominance of intraplate strike-slip deformation across the Wharton Basin is likely due to horizontal principle compressional and tensional stresses in the Indian and Australian plates. This is the expected result from the combination of Himalayan collision along the northern Indian plate and slab pull due to subduction of the Australian plate along the Sunda megathrust (e.g., Cloetingh and Wortel, 1985; Coblentz et al., 1998; Delescluse et al., 2012; Wiseman and Bürgmann, 2012). Close to the trench, there is some outer-rise extensional activity (Fig. 2) typical of slab bending, but strike-slip faulting continues all the way to the trench from 2° to 5°N along the southern portion of the great 2004 Sumatra-Andaman rupture zone. This is a rare predominance of strike-slip deformation over normal faulting seaward of a major underthrusting earthquake. The Gulf of Alaska activity discussed above has some similarity in having an along plate boundary variation from great underthrusting events to terrane collision, but again the scale of continental collision of the Indian plate dwarfs activity elsewhere.

The diffuse Indo-Australian deformation zone may progressively localize faulting, eventually resulting in a well-defined plate boundary along the lines of what has already occurred to segment the Capricorn plate from the Indian plate (e.g., Royer and Chang, 1991; Royer and Gordon, 1997; DeMets and Royer, 2003; DeMets et al., 2005), but this process is greatly complicated by the lateral compression in the plate due to the India-Eurasia collision inhibiting progress in the development of a discrete boundary. At present, a nascent boundary between Indian and Australian plates in the northwestern Wharton Basin would have to primarily be a convergent one, and onset of subduction within old oceanic lithosphere presents a difficult challenge. While the long-term evolution is unclear, the present state of stress in the plate and the influence of fracture zones can be considered relative to the large intraplate earthquakes in the region.

The June 18, 2000 $M_W$ 7.9 (Fig. 2) earthquake struck west of the Investigator fracture zone (IFZ) near a parallel fracture zone (F) (Robinson et al., 2001; Abercrombie et al., 2003). The GCMT has a large non-double-couple component indicative of complexity in the rupture, and larger aftershocks included strike-slip and thrust faulting, with an earlier thrust fault having occurred nearby in 2001 (Fig. 3). Robinson et al. (2001) model the source as two strike-slip events with similar mechanisms but orthogonal fault planes, one of which parallels F and the other is orthogonal to IFZ. Abercrombie et al. (2003) also model the event with two subevents, the first of which has 75% of the total moment and is a left-lateral strike-slip event and the second being a less well-resolved oblique reverse slip event (Fig. 3), significantly different from the second event in the model of Robinson et al. (2001). The Abercrombie et al. (2003) model accounts for the non-double-couple moment tensor much better and also predicts the observed P and SH waves much better. This model involves faulting off of both
fracture zones. The stress drop for the primary strike-slip rupture was estimated as 5–10 MPa by Abercrombie et al. (2003) and 20 MPa by Robinson et al. (2001). While fabric of the plate may have influenced the strike-slip geometry, it appears that ruptures did not occur directly on a fracture zone, indicating that F and IFZ are not necessarily weak portions of the plate that localize deformation.

The MW 7.8 earthquake on March 2, 2016 ruptured a region in between well-defined fracture zones in close proximity to the fossil Wharton Ridge (Fig. 2). The failure process of this event is less complicated than for the June 18, 2000 event, with almost pure strike-slip faulting and a simple GCMT with a centroid depth of 34.6 km. Re-located aftershocks define a north-south trend, indicating a left-lateral strike-slip faulting with a strike of about 5 degree (Lay et al., 2016). Analysis of broadband body and surface waves reveals minor overall directivity, indicative of bilateral rupture over limited spatial extent, and Lay et al. (2016) favor a fault very steeply dipping to the east with a rupture expansion speed of 2 km/s with slip of up to 14 m concentrated at depths less than 30 km. Models with a deeper centroid depth are hard to reconcile with the body waves, but the resolution of slip distribution from seismic wave analysis is limited due to the lack of clear directivity.

**FIG. 3** The 18 June 2000 MW 7.9 Wharton Basin rupture sequence. NEIC locations of the sequence are shown by crosses with relocations given by stars. The global centroid-moment tensor solutions for a nearby 2001 event, the 2000 mainshock and two aftershocks are shown by the light gray mechanisms. The dark gray mechanisms are two subevents of the mainshock determined by teleseismic broadband body-wave modeling indicating rupture on separate faults; the first subevent likely ruptured the northwest striking fault, the fault for the second subevent is not resolved. The events are located west of two fracture zones (F and IFZ: investigator fracture zone). Modified from Abercrombie, R.E., Antolík, M., Ekström, G., 2003. The June 2000 MW 7.9 earthquakes south of Sumatra: deformation in the India-Australia Plate. J. Geophys. Res. 108(B1), 2018, https://doi.org/10.1029/2001JB000674.
Although the 2016 event strike-slip faulting generated limited tsunami, there are clear recordings of tsunami signals from this event and Gusman et al. (2017) take advantage of the sensitivity of slow tsunami waves to spatial extent and geometry of faulting to improve constraints on the rupture process. They obtain a model from joint inversion of seismic and tsunami waveforms that has a bilateral rupture with speed of 2 km/s and a steeply dipping fault to the west, with the slip being spread more along strike than in the seismic wave solution of Lay et al. (2016). The tsunami data appear to prefer slip at greater depth than the seismic data, but still within the upper 30 km of the lithosphere. The rupture clearly parallels the nearby fracture zone, but is only directly related to one if there is a very subtle fracture zone not evident as a ridge offset or in the bathymetric structure. Errors in event locations certainly exist, but as in the other cases discussed, they would have to be unexpectedly large to shift the event to the nearest fracture zone.

The two largest intraplate events shown in Fig. 1 ruptured in the Cocos Basin (north Wharton Basin) on April 11, 2012, in a region of intense intraplate activity that extends all the way to the trench (Fig. 2). The great earthquakes were preceded by the January 10, 2012 $M_{W}$ 7.2 strike-slip earthquake to the north, which was located near, but not on a fracture zone. The focal mechanism for this event has a north-south plane paralleling the fracture zone, but the aftershocks do not clearly define the fault plane and body-wave inversion slightly favors rupture on an east-west plane (Aderhold and Abercrombie, 2016). The $M_{W}$ 8.6 and 8.2 great strike-slip earthquakes also have moment tensors and locations possibly consistent with rupture of either a well-defined fracture zone (Fig. 2), or on a more subtle parallel structure to the west (Singh et al., 2011). Numerous investigations have examined the faulting complexity for these events, although the $M_{W}$ 8.2 aftershock signals are partially obscured by signals from the mainshock (e.g., Duputel et al., 2012; Satriano et al., 2012; Meng et al., 2012; Yue et al., 2012; Ishii et al., 2013; Wei et al., 2013; Hill et al., 2015; Wang et al., 2012; Zhang et al., 2012; Yadav et al., 2017).

The $M_{W}$ 8.6 earthquake is the largest recorded intraplate event and also the largest recorded strike-slip event. It is also one of the most complex events ever, in that it involved failure of at least four near-orthogonal fault segments spread over a region ~350 km across. The faulting complexity is apparent in the spatial distribution of aftershocks, in the loci of locations of coherent short-period and long-period seismic radiation imaged by $P$-wave back projection (e.g., Meng et al., 2012; Satriano et al., 2012; Wang et al., 2012; Yue et al., 2012; Zhang et al., 2012), and in finite-source modeling based on regional and teleseismic body waves and regional geodetic data (Yue et al., 2012; Wei et al., 2013; Hill et al., 2015). Satriano et al. (2012) interpret back-projection results as favoring a westward migration of rupture of three en echelon parallel NNE trending fracture zones. In contrast, the back-projection interpretations from Meng et al. (2012), Wang et al. (2012), Yue et al. (2012), and Zhang et al. (2012) favor initial rupture on a WNW-ESE plane, followed by rupture on an orthogonal NNE-SSW plane that crosses the initial fault, and then rupture on another WNW-ESE plane to the south, with further faulting on a separate plane either parallel to or perpendicular to the Ninetyeast Ridge (Zhang et al., 2012 only image the first three faults). Satriano et al. (2012) recognize the possibility of faulting in the WNW-ESE direction, but invoke the common, but casual assumption that most strike-slip faulting in the Wharton basin is along fracture zones. As described before this is actually not the case for most large events globally.

The slip distribution of the multiple faults for the $M_{W}$ 8.6 event was obtained by finite-fault inversions constrained by back projections and aftershock patterns. The solution from Hill et al. (2015) shown in Fig. 4 is based on joint analysis of teleseismic body waves, surface waves
and regional high-rate and campaign GPS measurements along Indonesia and the Nicobar Islands. The inclusion of the geodetic observations improves resolution of the fault geometries and orientations relative to earlier seismic wave only inversions. Six faults rupture in the numbered sequence in Fig. 4, with time lags indicated by the superimposed moment-rate functions and with relative position indicated by the hypocenters (stars). Slip is largest at shallow depths (the fault models have four 20-km wide segments on steeply dipping planes), but slip of about 5 m does extend down to 60 km in limited portions of some of the faults. This is also found in the models of Yue et al. (2012) and Wei et al. (2013). This is of particular relevance given that the GCMT centroid depth is 45.6 km, while it is 54.7 km for the \( M_W 8.2 \) aftershock. Centroid depths of 30–40 km were estimated using long-period signals for both events by Duputel et al. (2012), with a two point-source model being used for the \( M_W 8.6 \) event. There is limited resolution of the spatial extent and magnitude of slip in the finite-fault models, but Hill et al. (2015) estimate an average stress drop of 17 MPa for the \( M_W 8.6 \) event overall, and \( \sim 25 \) MPa for the first fault, which has the largest slip. There is uncertainty in the westernmost fault orientation. Meng et al. (2012) discuss multibeam bathymetry showing numerous WNW-ESE fault scarps along the ridge, but Hill et al. (2015) present evidence from the geodetic fitting for the fault plane being parallel to the Ninetyeast Ridge. It is possible that

FIG. 4 Six-fault rupture model for the April 11, 2012 \( M_W 8.6 \) earthquake, showing the configuration of faults, numbered chronologically, along with the individual fault moment-rate functions and slip distributions determined from the joint analysis of seismic and geodetic observations. The moment-rate functions are overlapped at lower left with true relative timing and peak amplitudes. The fault grids all extend from the surface down to 60 km and have common length scales corresponding to the red line segments for each fault. Modified from Hill, E.M., Yue, H., Barbot, S., Lay, T., Tapponnier, P., Hermawan, I., Hubbard, J., Banerjee, P., Feng, L., Natawidjaja, D., Sieh, K., 2015. The 2012 \( M_W 8.6 \) Wharton Basin sequence: a cascade of great earthquakes generated by near-orthogonal, young, oceanic-mantle faults. J. Geophys. Res. 120, https://doi.org/10.1002/2014JB011703.
faults 3 and 4 in Fig. 4 locate along fracture zone F7b (Singh et al., 2017), with some lateral offset accounting for an apparent oblique trend relative to the fracture zone.

Low-resolution bathymetry does not reveal clear WNW-ESE structures along fault planes 1, 2, and 5 in Fig. 4, and there is substantial sediment cover on the seafloor. However, Singh et al. (2017) report evidence from high-resolution bathymetry and seismic reflection images for multiple parallel shear zones striking N294°E at high angle to the fracture zone fabric in the plate from 0.5° to 1°N both west and east of fracture zones F6ab, F7ab. These shear zones are defined by sets of normal faults striking N335°E, along the principal compressional stress direction in the lithosphere. The trend of the shear zones is consistent with the major WNW-ESE fault 5 (Fig. 4), supporting the faulting model, but also indicating that the fault is juvenile in that it has not localized into a single through-going structure.

Ishii et al. (2013) argue that the WNW-ESE ruptures are the result of slip partitioning of the oblique convergence of the Indian plate relative to Sumatra. Typically, slip partitioning occurs on through-going strike-slip faults in the upper plate for highly oblique convergent zones. Together with obliquity of thrusting on the megathrust (accounting for about 40% of the trench-parallel component), nearly the full budget of oblique plate convergence is accommodated, leaving little role for the intraplate deformation in the Indian plate in the overall slip partitioning. Thus, the presence of the partitioned Sumatran fault in the upper plate and the lack of a single continuous fault in the Indian plate make it more appropriate to attribute the strike-slip faulting to intraplate deformation just as for other strike-slip events in the deformation zone far from the subduction zone which are not directly manifestations of slip partitioning.

The $M_W$ 8.2 event appears to have ruptured along the NNE-SSW plane, in the vicinity of fracture zone F7b (Yue et al., 2012; Wei et al., 2013). Wei et al. (2013) use regional waveforms to image slip extending down to at least 50 km with a rupture length of about 150 km. The large rupture depth is consistent with the centroid depth estimate of Duputel et al. (2012). Precise location of the rupture is not known due to high noise levels produced by the mainshock, which occurred 2 h earlier. This may be the largest earthquake to rupture through the lithosphere on a fossil fracture zone.

## 5 DISCUSSION AND CONCLUSION

The broad deformation zone in the Indo-Australian plate involves numerous events with ruptures that extend to large depths into the lithosphere. One of the deepest of these is the May 21, 2014 $M_W$ 6.0 Bay of Bengal strike-slip earthquake located below the edge of the thickest part of the Bengal Fan (at latitude 18.2°N, longitude 88.0°E). The NEIC reports a hypocentral depth of 47 km and a W-phase centroid depth of 60.5 km. Rao et al. (2015) estimate a source depth of 50 km by modeling broadband recordings in India, and Aderhold and Abercrombie (2016) estimate a centroid depth of 53 km from teleseismic P-wave modeling. The lithospheric age in this region is 110 Ma, placing the source depth near the 600°C isotherm for half-space and plate-cooling models (Aderhold and Abercrombie, 2016).
In general, the moderate size intraplate seismicity in the Wharton Basin also lies above the 600°C isotherm as shown in Fig. 5. Estimates of earthquake centroid depth and vertical rupture extent for events in the Basin are summarized, extending slightly the compilation of Aderhold and Abercrombie (2016). The Bay of Bengal event plots on the far right. The great ruptures in 2012 have GCMT centroid depths that locate deeper than the 600°C isotherm, but the 30–40km centroid depth estimates of Duputel et al. (2012) indicate that the GCMT estimates are too deep for these events. Allowing for that upward shift, the 600 degree isotherm bounds most of the activity, although it does appear that for the 2012 events rupture may extend to greater depth. This is a strong indication of lithosphere transecting ruptures in the largest strike-slip events.

The depth extent of the intraplate strike-slip faulting is substantially greater than found in continental strike-slip systems where coseismic slip is limited to the crust, and comparable moment earthquakes tend to have very different rupture lengths in the two environments. In addition, the stress drop estimates for the large intraplate events discussed above, while very uncertain, tend to be in the 15–25MPa range, higher than for continental strike-slip events. Thus, slip can be large in the intraplate events, with estimates of localized slip greater than 25m for the first faults to rupture in the MW 8.6 event in 2012 (e.g., Yue et al., 2012; Wei et al., 2013; Hill et al., 2015). The oceanic intraplate events clearly involve rupture in the mantle as well as the crust, and apparently the high strength of oceanic mantle plays a major role in producing the high stress-drop fractures.

Considering all of the large events discussed here, the commonly invoked notion that fracture zones intrinsically define weak regions within oceanic plates prone to reactivation by intraplate stresses is not very well supported. The strongly deforming regions near the...
Macquarie ridge and in the northern Wharton Basin do provide a few cases where large ruptures appear to locate on fracture zones, but there are many events that rupture on faults laterally offset from fracture zones or on orthogonal structures, even though the focal mechanism is generically compatible with reactivation of a fracture zone. Event location uncertainties in the remote areas are significant, but probably cannot place the major events in 2003 and 2016 on the prominent nearby fracture zones. As mentioned before, it could be that there are very subtle fracture zone features that are not identified in the ridge offsets or bathymetry, or it may be that some form of fracture zone parallel fabric, perhaps associated with lateral thermal gradients, plays a role where these ruptures occur. However, if the more prominent nearby fracture zones are truly very weak structures, it is unclear why they were not preferentially activated.

The NW-SE principle stress direction in the Indian plate arising from the configuration of the India-Eurasia continental collision combines with the oblique orientation of underthrusting and NE-SW slab pull in the Sunda trench to produce intraplate stresses favoring strike-slip faulting that happens to align generally with fossil fracture zones and the Ninetyeast Ridge. It is not clear that the fracture zones or the ridge serve as weak zones that activate preferentially in this stress environment, although some activity does appear to occur on fracture zones (as for the 2012 $M_W$ 8.2 event and possibly the N-S fault activated in the $M_W$ 8.6 mainshock). With the primary moment release being on the orthogonal planes for the largest event in the plate, and with other large events locating off of nearby well-defined fracture zones, there is not compelling evidence that either the preexisting transform fault or ridge parallel fabric play controlling roles in the intraplate faulting, despite the intuitive appeal of this idea derived from the general consistency of focal mechanisms (Fig. 2).

The fast-spreading Wharton Ridge produced thin crust and the associated transform faults likely had only shallow activity when the system was active. In the 38 million years since spreading ended, cooling has thickened the lithosphere progressively across the region, possibly with relatively small “memory” of the fracture zone genesis. In that perspective, the occurrence of large strike-slip events offset from fracture zones or rupturing lithosphere in orthogonal directions is most likely a manifestation of the plate failure being primarily controlled by the stress state and boundary conditions on the plate rather than by any internal fabric. This argument can also be applied to the large earthquakes in the Antarctic, Pacific, and Atlantic plates, which are also not results of fracture zones serving as zones of weakness. That intuitive ‘expectation’ should perhaps be downgraded.

Note Added in Proof

An Mw 7.9 strike-slip earthquake struck on January 23, 2018 under the Gulf of Alaska seaward of Kodiak Islands. Preliminary analysis indicates rupture of multiple, quasi-orthogonal faults, with primary slip on planes parallel to magnetic lineations rather than on fossil transforms.

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References


4. REACTIVATION OF OCEANIC FRACTURE ZONES IN LARGE INTRAPLATE EARTHQUAKES?