The 2011 Northern Kermadec earthquake doublet and subduction zone faulting interactions

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1. Introduction

[2] Subduction of oceanic lithosphere involves large-scale deformation of the stiff underthrusting plate (slab), elastic and anelastic deformation of the overriding plate, and frictional sliding on the megathrust boundary between the two plates. The complex thermal, petrological, and hydrological environment, and the wide variety of timescales of deformation in a shallow subduction zone make it challenging to disentangle the relative roles played by elastic and anelastic deformations and to quantify the stress distribution in the subduction zone. Earthquakes provide one window into these processes, because their temporal patterns, faulting mechanisms, and magnitude of strain release directly characterize rapid timescale deformations; however, postseismic and interseismic deformation intrinsically involve longer timescale aseismic deformations that appear to involve a mix of poroelastic, anelastic, and/or viscoelastic processes. The relationship between the relatively rapid seismic cycle processes and deformations and the million-year timescale processes by which slabs bend, sink, and unbend, is very difficult to establish.

[3] Intraslab earthquake activity beneath the outer rise and outer trench slope in subduction zones is generally characterized by shallow normal-faulting ruptures with depths less than 30 km into the oceanic lithosphere, with less frequent occurrence of deeper thrust faulting [e.g., Christensen and Ruff, 1988]. The depth variation of faulting mechanism has generally been attributed to elastic bending stresses within the stiffest part of the oceanic lithosphere [e.g., Chapple and Forsyth, 1979].

[4] The relationship between interplate stress cycles and intraslab earthquake activity has been of interest, because it reveals intuitive aspects of stress transfer, faulting interactions, and seismic hazard that may be quantitatively modeled [e.g., Isacks et al., 1968; Stauder, 1968; Abe, 1972; Ruff and Kanamori, 1980; Christensen and Ruff, 1988; Dmowska et al., 1988; Lay et al., 1989; Liu and McNally, 1993; Dmowska et al., 1996; Taylor et al., 1996; Mikumo et al., 1999; King, 2007; Ammon et al., 2008; Lay et al., 2009]. However, it has also been shown that viscoelastic deformation...
can complicate the behavior [e.g., Mueller et al., 1996a, 1996b; Lay et al., 2009].

[5] Intraslab activity can clearly be modulated by the occurrence of great interplate thrust events, which commonly activate extensive normal faulting below the outer trench slope or in the outer rise as part of their aftershock sequence (Figure 1). This activity can persist for decades. Relatively few outer rise normal faulting events occur within the several decades preceding a large interplate thrust event. However, large outer rise thrust-faulting events have been observed in the preceding years to decades before great megathrust failures, typically rupturing at depths greater than 30 km [Christensen and Ruff, 1988; Lay et al., 1989]. This is suggestive of a build-up or depth-shallowing of compressional stresses within the subducting plate during the interseismic elastic strain accumulation interval that reduces shallow horizontal extensional bending stresses while augmenting the deeper horizontal compressional bending stresses. This has been modeled in elastic and viscoelastic calculations as a seismic-cycle modulation of the depth of the neutral stress surface beneath the trench [e.g., Dmowska et al., 1988; Liu and McNally, 1993; Taylor et al., 1996]. Although consistent with seismicity patterns in a general sense, this notion directly juxtaposes processes acting on different timescales; the decades to century timescale of the megathrust seismic cycle and the million-year timescale of plate bending. Some viscoelastic calculations [Mueller et al., 1996a, 1996b] have not confirmed the viability of the intraslab thrust-faulting activity or its temporal variation being a response to the elastic seismic cycle strain fluctuations near the megathrust.

[6] Outer rise faulting also occurs in subduction zones that do not appear to have great thrusting earthquakes, such as the Marianas, Java, and Tonga zones [Lay and Kanamori, 1981]. Great normal-faulting events such as the 1977 Sumba (MW 8.3) [Spence, 1986; Lynnes and Lay, 1988] and 2009 Samoa (MW 8.0) earthquakes [Lay et al., 2010] ruptured the upper oceanic lithosphere seaward of apparently aseismic megathrusts, and likely occur as a result of strong slab pull and plate bending as the subducting plate continuously creeps into the mantle. Intraslab thrust earthquakes seaward of aseismic megathrusts are rare and tend to be small. This is consistent with the notion that without modulation by interplate seismic-cycle stresses, megathrust regions with predominantly aseismic convergence may have deeper elastic neutral bending stress surfaces, and hence less volume for accumulating and releasing elastic compressional stress beneath the trench.

[7] We note that there are exceptions to these basic temporal patterns, as in the central Kuril Islands where a large, relatively deep intraslab thrust-faulting event in 2009 (MW 7.4) followed the 2006 (MW 8.3) interplate thrust event and the 2007 (MW 8.1) trench slope normal-faulting event. Given that observations of subduction processes span a very limited time interval and the behavior is complex, it is valuable to document and characterize earthquake and plate deformation sequences for many environments. We consider a large intraslab earthquake doublet involving shallow normal faulting on 6 July 2011 (MW 7.6) and deeper thrust faulting on 21 October 2011 (MW 7.4) located below the northern Kermadec trench, evaluating the faulting interactions during the sequence. This 2011 Kermadec doublet, and the earlier 2009 Samoa earthquakes near the northern end of the Tonga trench [Beavan et al., 2010; Lay et al., 2010] indicate that in some instances the roles may be reversed with respect to faulting interactions between the intraslab and interplate regimes.

2. Tectonic Framework

[8] The Tonga-Kermadec trench extends over 2600 km from ~14.5°S, south of Samoa, to ~38°S, northeast of New
Zealand (Figure 2) with the Pacific plate underthrusting the Australian plate along the Tonga and Kermadec Islands, which are backed by the extending Lau Basin and Harve Trough, respectively [e.g., Pelletier and Louat, 1989; Pelletier et al., 1998; Bevis et al., 1995]. Near 26°S, the Louisville Ridge, a chain of seamounts on the Pacific plate, impinges obliquely on the trench, providing a major tectonic segmentation between the Tonga and Kermadec subduction zones [e.g., Bonnardot et al., 2007]. The long arc complex has high levels of low and moderate magnitude shallow earthquakes, but has only a few events with M ~ 8 in the historic record [e.g., McCann et al., 1979; Lay et al., 1982; Nishenko, 1991], and there is large uncertainty in whether realistic potential exists for a megathrust rupture as large as Ms ~ 9.5–9.6 to occur in the region as deemed plausible by overall plate convergence characteristics [Kagan, 1999; Bird and Kagan, 2004; McCaffrey, 2008]. The large shallow interplate and intraslab earthquakes that have occurred in the Tonga-Kermadec region have been the subjects of several seismological studies [e.g., Christensen and Lay, 1988; Eissler and Kanamori, 1982; Houston et al., 1993; Lundgren et al., 1989; Okal et al., 2004; Warren et al., 2007; Lay et al., 2010], and some geodetic analysis [e.g., Beavan et al., 2010; Power et al., 2012].

The long-term plate convergence rate varies rapidly along the Kermadec trench, decreasing southward from 90 mm/yr at 25°S to 53 mm/yr at 35°S [Power et al., 2012].

Over the last century, there have been seven shallow (<60 km deep) earthquakes with Ms ≥ 7.5 along the northern Kermadec trench (Figure 2) between 28°S and 30°S, with this being the primary concentration of large events in the subduction zone. The 1 May 1917 (M ~ 8) event and the 1917 event is generally assumed to be as well [e.g., Nishenko, 1991], suggesting a 59 year recurrence interval, compatible with a high percentage of seismic convergence. An M ~ 7.8 event in 1959 ruptured just to the north of those events, having been preceded in 1955 by an M 7.8 event located seaward of the trench near the outer rise. In 1986, an intraplate Ms 7.7 earthquake occurred somewhat further to the north (28.2°S) below the megathrust, apparently involving a tear in the subducting slab [Lundgren et al., 1989; Houston et al., 1993]. Most recently, the 2011 doublet commenced with an Ms 7.6 normal-faulting event seaward of the 1976 rupture zones, as discussed below.

Intraslab activity below the Kermadec trench preceded the 1976 megathrust doublet [Christensen and Ruff, 1988], with a thrust-faulting earthquake on 2 July 1974 with Ms 7.2, followed by a smaller normal-faulting earthquake on 3 July 1974 with Ms 6.6 (Figure 3a). Just north of the subducting Louisville Ridge, an outer-trench-slope thrust earthquake in 1975 preceded a large interplate earthquake in 19 December 1982 (Ms 7.5) [Christensen and Ruff, 1988]. These may represent instances of enhanced in-plate horizontal compressional stress prior to megathrust ruptures (Figure 1a).

Power et al. [2012] inferred from westward motion (relative to the stable Australian Plate) of campaign GPS (since the mid-1990s) and continuous GPS measurements (since 2009) at Raoul Island that the northern portion of

Figure 2. (top) Earthquake epicenters in the Tonga-Kermadec subduction zone from the USGS National Earthquake Information Center catalog (1973–2012) are color coded by hypocentral depth. The epicenters of the 2011 Kermadec Doublet (yellow stars) are indicated, along with Raoul Island, approximately 175 km to the west. Velocity of the Pacific plate relative to the Australian plate for model NUVEL-1A [DeMets et al., 1994] is shown by the white arrows. The dashed white box indicates the region for which large shallow events are shown in the timeline below. (bottom) Shallow (h < 70 km) earthquakes with M ≥ 7.0 from 1900 to 2012 located within the white box in the map that are listed in the PAGER-CAT Earthquake Catalog [Allen et al., 2009] are shown in the timeline (the GCMT (www.globalcmt.org/CMTsearch.html) catalog is used to update the catalog from 2007 to 2012). The largest megathrust ruptures appear to be the events in 1917, 1959, and 1976, with the 1959 event locating somewhat north of the other events. The events labeled in red are intraplate events: 1955 was near the outer rise, 1986 was an intraslab rupture, and the 2011 doublet is under the trench slope.
the Kermadec megathrust in the region of the 1976 doublet is frictionally locked at present. As Power et al. [2012] noted, GPS coverage of the Kermadec Trench is restricted to Raoul Island, within the subaerial Kermadec arc, therefore estimation of interseismic locking north or south of Raoul Island is not currently possible. Raoul Island is the largest of the Kermadec islands and is located 175 km west of the 2011 doublet (Figures 2 and 3b). Power et al. [2012] inferred >80% slip deficit in the region from 28°S to 30°S based on block modeling of earthquake slip vectors and GPS velocities. The regionally corrected GPS data indicate that Raoul Island abruptly moved ~20 mm to the west in July 2011, which is consistent with the 6 July 2011 $M_w$ 7.6 outer trench-slope normal-faulting event (Figure 4). Subsequently, the ground deformation resumed its steady rate of westward motion at approximately the same rate as before the earthquake. Approximately 5 mm of abrupt eastward motion that may have resulted from the outer rise thrust faulting in October 2011 is detectable above the background noise, followed by resumed westward motion. This can be contrasted with the GPS observations for the 29 September 2009 Samoa normal-faulting earthquake, where a GPS site in northern Tonga showed unexpected large eastward motion of the upper plate [Beavan et al., 2010], later recognized to represent dynamic triggering of a large ($M_w$ ~ 8.0) megathrust rupture by the outer

![Figure 3](image-url)  
Figure 3. (a) Shallow earthquake focal mechanisms (GCMT lower hemisphere best double-couple solutions from 1 January 1976 to 5 July 2011, plotted at the centroid locations) in the northern Kermadec subduction zone. The large interplate events of 14 January 1976 (15:55:34.9 $M_w$ 7.8 and 16:47:33.5 $M_w$ 7.9), the intraslab rupture of 20 October 1986 ($M_w$ 7.7), and the outer-trench-slope compressional and extensional doublet of 2 July 1974 ($M_w$ 7.3) and 3 July 1974 ($M_s$ 6.6) are shown with large brown solutions. Focal mechanisms for the 1974 doublet are from Chapple and Forsyth [1979]. The trench is shown by the white curve. Raoul Island is highlighted. The subduction zone convergence rate (white arrow) corrects for the back-arc spreading [Power et al., 2012]. (b) The 2011 Kermadec Doublet sequence, with intraslab mainshocks of 6 July 2011 ($M_w$ 7.6) and 21 October 2011 ($M_w$ 7.4) (brown) and smaller interplate and intraslab event (green) GCMT solutions from 6 July 2011 to 31 July 2012 plotted at their GCMT centroid locations. The interplate aftershocks locate up-dip of the second large interplate event in 1976, in an area where there had not been much seismicity for 35 years.

![Figure 4](image-url)  
Figure 4. GeoNet Station RAUL (www.geonet.org.nz), installed on Raoul Island on 16 May 2009, has recorded progressively westward motion, indicating a regionally-locked plate interface. Raoul Island moved ~20 mm to the west following the 6 July 2011 event, as expected for normal faulting within the subducting plate. This relatively abrupt motion is superimposed on the <10 mm/yr movement of the island preceding the 6 July 2011 earthquake. There is a ~5 mm eastward motion for the 21 October 2011 intraslab compressional event.
rise normal faulting earthquake [Beavan et al., 2010; Lay et al., 2010].

3. 2011 Kermadec Doublet

[13] On 6 July 2011 a large normal-faulting earthquake occurred below the Kermadec trench seaward of the 1976 megathrust ruptures (Figure 3). The U.S. Geological Survey (USGS) source parameters (http://earthquake.usgs.gov/earthquakes/) for this earthquake are: 19:03:16.7 UTC, 29.31°S, 176.20°W, body-wave magnitude \( m_b = 7.0 \), and surface-wave magnitude \( M_s = 7.4 \). Global Centroid Moment Tensor (www.globalcmt.org) and USGS W-phase inversions for the event yield \( M_w = 7.6 \), with centroid depth estimates of 22.7 and 15 km, respectively. Modest nondouble couple components are present in both long-period
Figures 6. Observed body wave data (black) and synthetic seismograms (red) for the finite-fault inversion for the 6 July 2011 earthquake in Figure 5a. Waveforms for P and SH (labeled) phases have true relative amplitudes except that SH signals are plotted with 0.2 times the P wave amplitudes, which is the same relative weighting as they were given in the inversion. Each station name and azimuth (φ) is shown.

solutions, with shallow plunging tension axes oriented almost perpendicular to the trench (Figure 3b). The USGS finite-fault inversion indicates a fault plane with strike φ = 170°, and dip δ = 52°. A modest tsunami with peak amplitude above sea level on Raoul Island of 14 cm was reported by NOAA (http://oldwcatwc.arh.noaa.gov/about/tsunamimain.php). This earthquake activated normal-faulting aftershocks distributed 100 km along strike of the trench, and thrust-faulting aftershocks to the west with faulting parameters consistent with plate boundary faulting (Figure 3b) [Hayes et al., 2012].

The second large event of the doublet occurred on 21 October 2011, 50 km to the north of the first event, at greater depth below the outer trench slope. The USGS source parameters are: 17:57:16.9 UTC, 29.00°S, 176.18°W, m0 = 6.9, Mf = 7.6, and Mw = 7.4. The GCMT centroid depth is 48.4 km, and the USGS W-phase centroid depth is 50 km, with minor nondouble couple components and nearly horizontal compressional axes trending almost perpendicular to the trench. The USGS finite-fault inversion indicates a fault plane with φ = 196.1°, and δ = 48.4°. A small tsunami with peak amplitude above sea level on Raoul Island of 14 cm was reported by NOAA (http://oldwcatwc.arh.noaa.gov/about/tsunamimain.php).

The doublet aftershock sequence has continued into 2012 with a mix of thrust-faulting events near the megathrust and normal-faulting events below the trench.

4. Doublet Rupture Processes

We determine finite-fault slip models for the July and October 2011 doublet events using broadband P and SH wave displacement recordings from teleseismic distances, following the methods of Kikuchi and Kanamori (http://www.eri.u-tokyo.ac.jp/ETAL/KIKUCHI/). We assume the USGS W-phase inversion fault geometry for the July earthquake with φ = 170°, δ = 52°, allowing variable rake. This geometry is slightly preferred relative to the orthogonal plane in the body wave inversions, and is consistent with previous findings of trenchward-dipping normal faults extending to a depth of 25–30 km along the Kermadec trench [Warren et al., 2007]. For the October earthquake, it proves difficult to distinguish between the fault plane and auxiliary fault plane; we ultimately adopt the trenchward-dipping plane from the best double-couple of the GCMT solution with φ = 203°, δ = 38°, and variable rake.

Figure 5 shows the average focal mechanism, source time function, and slip distribution for each doublet event. The July earthquake rupture is modeled as continuing for approximately 65 s, based on coherent secondary arrivals in the P waves (waveform observations and predictions are shown in Figure 6). The subfaults were parameterized with 10 symmetric triangles with 1.5 s rise-time lagged by 1.5 s, yielding 16.5 s subfault rupture durations, and a rupture velocity of 1.5 km/s, with fault length of 150 km along strike and width of 50 km along dip. The P wave data show impulsive early ground motions and this is manifested in the spiky source time function found for the first 15 s of rupture. The inverted slip distribution has estimated slip of up to 9 m distributed across the upper 30 km of the oceanic lithosphere (a water depth of 6 km is included in the depths in Figure 5a) extending about 50 km along strike, with some later slip to the south about 35 s into the rupture. The large shallow slip likely generated the tsunami that was recorded regionally in the Kermadec Islands, New Zealand and Tonga (http://oldwcatwc.arh.noaa.gov/about/tsunamimain.php).
The estimated seismic moment of $7.4 \times 10^{20}$ Nm is larger than the long-period GCMT and W-phase inversions estimates of about $2.8-2.9 \times 10^{20}$ Nm, likely due to limited resolution of the long-period component of the body wave inversion caused by use of simplified Green functions. If we consider only subfaults with at least 20% of the peak subfault moment, we obtain a moment of $6.6 \times 10^{20}$ Nm, an average slip of 2.9 m, and static stress drop of $2.2 \times 10^{16}$ MPa. The USGS estimates a high seismic energy release of $4.0 \times 10^{16}$ Nm, which gives an energy magnitude $M_e$ of 8.2. Using the USGS W-phase inversion seismic moment yields a moment-scaled energy value of $14.3 \times 10^{-5}$. This is a relatively high ratio even for an intraplate event [e.g., Venkataraman and Kanamori, 2004], comparable to the 1993 Kushiro-oki intermediate depth event, and is related to the impulsive arrivals in the ground motions.

[17] Located 50 km northward along trench, the October rupture lasted at least 15 s, possibly with some low-amplitude seismic wave radiation as late as 60 s. We assumed $V_e = 2.0$ km/s and a fault extending 195 km along strike and 75 km along the 38° dipping plane, but the slip remained concentrated within a region extending about 45 km along strike with a centroid depth near 45 km. The subfault source time functions were parameterized with seven symmetric triangles with 1.5 s rise-time lagged by 1.5 s, yielding 11.5 s subfault rupture durations. There are again some secondary arrivals about 30 s into the waveforms (waveform fits are shown in Figure 7), but these may involve scattered energy not accounted for in the Green’s functions. The body wave inversion seismic moment of $3.0 \times 10^{20}$ Nm again exceeds the long-period GCMT and W-phase inversion estimates of about $1.5 \times 10^{20}$ Nm, so the peak slip estimate of 3.7 m may be overestimated by a factor of 2. If we consider only subfaults with at least 10% of the peak subfault moment, the body wave moment estimate is $2.0 \times 10^{20}$ Nm, and the average slip is ~1 m with a static stress drop of $1.1 \times 10^{16}$ MPa. The USGS seismic energy release estimate is $1.4 \times 10^{15}$ Nm ($M_e 7.2$) and using the W-phase inversion seismic moment gives a moment-scaled energy value of $0.9 \times 10^{-5}$. These values are over an order of magnitude lower than for the July earthquake, but the moment-scaled energy is consistent with other intraslab compressional earthquakes such as the 2009 Kuril Islands earthquake [Lay et al., 2009].

5. Aftershock Sequence

[18] The 6 July 2011 earthquake triggered shallow intraslab faulting within minutes reported in the USGS catalog, and activity near the megathrust updip of the 1976 rupture area within hours. For the events in the sequence large
enough to obtain GCMT solutions \((M_w > 4.5)\), most of the intraslab events involve normal faulting, and most of the megathrust activity is thrust faulting (Figures 3b and 8a). The normal-faulting aftershocks may be located on the fault plane that ruptured in the mainshock or on adjacent faults of similar geometry; the source depth estimates are imprecise and no clear alignment is apparent in the cross-section in Figure 8a. The July earthquake also triggered two deep (40–45 km) intraslab thrust-faulting aftershocks, including the 21 October 2011 earthquake. Figure 8b shows that normal-faulting and thrust-faulting aftershocks were occurring simultaneously, indicating a stress increase on faults of similar geometry to the mainshock as well as on the megathrust interface.

The 21 October 2011 earthquake generated its own aftershock sequence of normal-faulting intraslab earthquakes and underthrusting earthquakes. As shown in Figure 8b, the number of thrust fault aftershocks large enough for GCMT solutions to be determined decreased for a few months after the October earthquake, possibly due to a static stress change shadow \([e.g., \text{Toda et al.}, 2012]\), but picks up again in early 2012. The brief reversal of upper plate motion at the time of the 21 October 2011 event (Figure 4) was quickly undone and compression of the upper plate resumed as the shallow interplate thrusting picked up again.

The GCMT depths and focal mechanisms of the thrust-faulting aftershocks west of the trench indicate that the earthquakes are likely located along the plate interface (Figure 8a), although there is some scatter. \textit{Hayes et al.} [2012] estimate an approximate dip of 22° for the megathrust along the Kermadec Trench and most of the thrust faulting aftershocks west of the trench have fault planes dipping between 20° and 25°. There appears to be a secondary alignment of thrust events about 10 km deeper than the primary alignment, with some of the events having slightly greater dip than the shallower events. This may indicate intraslab activity, but the events are relatively small and the depth estimates have at least 10 km uncertainty. We sought to confirm validity of the GCMT centroid depths in this region, but only a few events produced teleseismic \(P\) wave signals with high enough signal-to-noise ratios for reliable modeling. For the few events with adequate data we found good consistency between our depth estimates from \(P\) wave modeling and the GCMT depths (Appendix A), so we have no indication of a regional bias in GCMT centroid depth estimates for northern Kermadec. We focus on the likely megathrust activity, noting that there are several events near the megathrust that do clearly have mechanisms indicating intraslab faulting (Figure 8a), and it is possible that there are multiple activated fault structures in the region.

6. Stress Transfer Cycle

We compute stress perturbations caused by the 6 July 2011 and 21 October 2011 earthquakes for both the intra-plate environment and the plate boundary. The calculations are made with Coulomb 3 software produced by S. Toda, R. Stein, J. Lin, and V. Sevilgen. A coefficient of friction of 0.4 is used in the calculations, with the basic patterns not being strongly affected by this particular choice. We
use the slip distributions from our finite-fault rupture models in computing Coulomb stress changes on the primary fault geometries activated during the sequence. The 6 July normal-faulting earthquake produced Coulomb stress increases of ~1–2 bars on faults with shallow thrust fault geometries in the vicinity of the megathrust to the west (Figure 9b). In this respect, triggering of interplate thrust aftershocks is a reasonable outcome; just as the more common reverse situation of interplate thrusting triggering outer rise normal-faulting is consistent with corresponding Coulomb stress changes. The normal-fault slip predicts larger Coulomb stress increases of ~3–5 bars on 30–50 km deep intraslab thrust-faults below the normal fault with the geometry and approximate location of the 21 October event (Figure 9d). The 21 October faulting produced a small reduction (<1 bar) of driving stress (a static stress change shadow) for shallow thrusting near the megathrust (Figure 9c), consistent with the ephemeral reduction of thrusting activity, and modest

Figure 9
increases of driving stress on overlying normal fault geometry near the 6 July event (Figure 9e), reconciling the occurrence of some normal-faulting aftershocks.

[22] These calculations are made without any prestress in the medium, so the Coulomb stress perturbations would be superimposed on any ambient stresses from bending and flexure of the slab or interplate stress accumulation. The deeper compressional event of the Kermadec doublet is likely a consequence of combined transient stresses and longer-term bending stresses, but these calculations cannot determine the relative importance of the two. The close spatial proximity of the July normal-faulting and October thrust-faulting events, separated vertically by only ~20 km, is comparable to the separation seen between intraplate normal-faulting and thrust-faulting in other environments [e.g., Christensen and Ruff, 1988; Lay et al., 2009]. At face value, an elastic bending interpretation would suggest that a bending stress neutral surface exists around 30–35 km deep in this region within the elastic core of the plate lithosphere. This implies a relatively thick elastic lithosphere of perhaps 70 km thickness, for which only the upper 2/3 or so is sufficiently brittle to experience rapid faulting. Billen and Gurnis [2005] analyzed topography and gravity profiles parallel to the Kermadec trench and infer a dramatic decrease in plate strength due to either a decrease in flexural rigidity by 3–5 orders of magnitude or a decrease in the effective elastic thickness by more than 15 km. The latter may be consistent with a relatively shallow neutral surface in the plate, and the occurrence of large-scale faulting within the plate may provide one mechanism for weakening the subducting lithosphere. However, Billen and Gurnis [2005] favor such pronounced weakening of the lithosphere that it is unclear that significant elastic stress interactions could even occur between the interplate and intraslab environment. It is possible that the interactions that are observed are influenced by dynamic stress changes that induce triggered aftershock sequences. The problem at hand is again one of evaluating stress transfer for processes operating on vastly different timescales.

[23] The relative roles of elastic versus anelastic processes in faulting interactions in subduction zones is not resolved, but the 2011 Kermadec doublet sequence provides additional support for importance of elastic interactions, given that the elastic Coulomb stress calculations are consistent with the observed faulting patterns and plausible long-term bending and plate interaction stresses. The Kermadec sequence is, however, distinct from typical faulting patterns that have been observed in regions with larger interplate and intraslab events, as summarized in Figure 10. Large, shallow outer-trench-slope normal-faulting events are commonly observed following great underthrusting events in regions with strong and/or shallow interseismic coupling, as in the 1896/1933 Sanriku-oki (Figure 10a) and 2006/2007 Kuril Island (Figure 10b) sequences. Based on these and other observations, the time delay between the great interplate rupture and the great outer-rise extensional faulting can range from minutes, as seen in the 2011 M, 9.0 great Tohoku-oki earthquake [Lay et al., 2011], to months for the 2006/2007 Kuril sequence [Lay et al., 2009] to decades, as for the 1896/1933 Sanriku-oki sequence [Kanamori 1971, 1972]. The varying delay time is plausibly attributable to viscoelastic and poroelastic processes, although in the Kuril sequence small intraslab activity was activated very quickly after the thrusting [Lay et al., 2009], so there is some role of dynamic triggering or elastic stress change even if the main faulting is delayed.

[24] The 2009 Samoa normal-faulting event (Figure 10c) dynamically triggered thrust-faulting on an adjacent stretch of the megathrust [Lay et al., 2010], and produced far more extensive thrust-faulting aftershock activity than intraslab activity. As found here, the occurrence of megathrust faulting is a reasonable consequence of trench slope normal-faulting, although the mechanism by which large extensional stress builds up and releases if the megathrust is locked is not clear (for the Samoa event, the thrusting occurred to the south of where the normal fault ruptured, so there may be a creeping region immediately down-dip of the normal-fault that allowed slab pull stress and/or slow slip events on the megathrust to load the normal fault). The triggered thrusting may have occurred in a region of conditional stability, which was undergoing stable sliding that accelerated into earthquake faulting when the stresses abruptly increased. Much of the triggered megathrust activity in the Kermadec sequence is located in a region where few GCMT solutions were previously found (Figure 3). The 6 July 2011 Kermadec normal-fault earthquake (Figure 10d) is either an example of very delayed response to the 1976 interplate events similar to the Sanriku events or an example of extensional stress accumulation and release up-dip of a coupled megathrust region. Like the Samoa sequence (Figure 10c), the trench slope faulting activated thrusting on the megathrust, but in this case, it also activated deeper intraslab compressional activity, similar to the 2009 Kuril event.

[25] These diverse sequences indicate a suite of possible faulting interactions with the intraplate environment responding to changes in the interplate environment and vice versa. It is rare to observe a large normal-fault event seaward.

Figure 9. (a) Schematic diagram indicating the relative locations of the 6 July 2011 and 21 October 2011 earthquakes, with cross-sections perpendicular to their fault planes. (b) Map view of Coulomb stress change on shallow dipping thrust faults consistent with the megathrust geometry [Hayes et al., 2012] at a depth of 24 km for the finite-fault model of the 6 July 2011 event, and a vertical cross-section (along AB) showing corresponding Coulomb stress changes. (c) Map view of Coulomb stress change on shallow dipping thrust faults consistent with the megathrust geometry [Hayes et al., 2012] at a depth of 45 km for the finite-fault model of the 21 October 2011 event, and a vertical cross section (along A'B') showing corresponding Coulomb stress changes. (d) Map view of Coulomb stress change on deep trench slope compressional faults consistent with deeper compressional faulting at a depth of 45 km for the finite-fault model of the 6 July 2011 event, and a vertical cross section (along CD, which intersects the hypocenter of the 21 October event) showing corresponding Coulomb stress changes. (e) Map view of Coulomb stress change on shallow trench slope extensional faults at a depth of 24 km for the finite-fault model of the 21 October 2011 event, and a vertical cross section (along C'D', which intersects the hypocenter of the 6 July event) showing corresponding Coulomb stress changes.
of a locked megathrust region, but the 2011 Kermadec doublet demonstrates that this can occur and appears to load the megathrust driving stresses, likely advancing the clock for the next large megathrust failure to come. The activation of shallow megathrust faulting by the normal fault will contribute further to stress concentration on the deeper, locked megathrust. For the northern Kermadec region we have very limited knowledge of the stress state, so it is unclear what the timing or total size of the next large underthrusting event will be.

7. Conclusion

[26] The 6 July 2011 ($M_w$ 7.6) and 21 October 2011 ($M_w$ 7.4) intraslab earthquake doublet and the accompanying aftershock sequence along the Kermadec trench allow exploration of intraslab and interplate faulting interactions and stress changes in this region. The sequence began with large shallow normal faulting on 6 July 2011 that increased driving stress on both the megathrust environment and intraslab environment, resulting in thrust-faulting and normal-faulting activity in both regions, including the 21 October 2011 thrust-faulting event ~20 km below the northern portion of the earlier normal fault rupture. Coulomb stress change calculations are consistent with the observed faulting patterns in a general sense, although the precise contribution of background plate bending and plate coupling stresses is not known quantitatively. The apparent elastic neutral stress surface depth of about 30–35 km favors a relatively thick elastic lithosphere in the region, in contrast to estimates of low effective elastic thickness from flexural studies at the Kermadec Trench [Billen and Gurnis, 2005]. The 2011 Kermadec Doublet sequence is distinctive in that the previous megathrust rupture was 36 years earlier in 1976 and the GPS data indicate that the megathrust is currently locked. The 6 July 2011 normal-faulting earthquake triggered aftershocks on the megathrust in addition to normal-faulting earthquakes below the trench, with a deeper, large thrust-faulting event on 21 October 2011. Stress should have increased on the 1976 rupture zone, which previously failed in 1917.

Appendix A

[27] We determined finite-source rupture models using $P$ and $SH$ waves recorded at teleseismic distances to constrain the depths of two moderate, $M_w$ 5.9 and 6.0, underthrusting
aftershocks that occurred 6 h apart on 9 July 2011. The inversion indicates consistent centroid depths with the GCMT estimates, suggesting that there is no regional bias in GCMT depth estimates. Selected waveform fits and residual error versus assumed hypocentral depth curves are shown in Figure A1.

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