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Abstract

A critical foundation to earthquake study and hazard assessment is the understanding of controls on fault rupture, including segmentation. Key challenges to understanding fault rupture segmentation include, but are not limited to: What determines if a fault segment will rupture in a single great event or multiple moderate events? How is slip along a fault partitioned between seismic and aseismic components? How does the seismicity of a fault segment evolve over time? How representative are past events for assessing future seismic hazards? In order to address the difficult questions regarding fault rupture segmentation, new methods must be developed that utilize the information available. Much of the research presented in this study focuses on the development of new methods for attacking the challenges of understanding fault rupture segmentation. Not only do these methods exploit a broader band of information within the waveform than has traditionally been used, but they also lend themselves to the inclusion of even more seismic phases providing deeper understandings. Additionally, these methods are designed to be fast and efficient with large datasets, allowing them to utilize the enormous volume of data available. Key findings from this body of work include demonstration that focus on fundamental earthquake properties on regional scales can provide general understanding of fault rupture segmentation. We present a more modern, waveform-based method that locates events using cross-correlation of the Rayleigh waves. Additionally, cross-correlation values can also be used to calculate precise earthquake magnitudes. Finally, insight regarding earthquake rupture directivity can be easily and quickly exploited using cross-correlation of surface waves.
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### Appendix A
**Seismic Moment Distribution Estimation**

#### A.1 Methods

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**Precise Relative Earthquake Location Using Surface Waves**

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2.1 The outer rise region is located seaward of the subduction zone. Earthquakes occurring in this region are intraplate, rupturing within the oceanic plate. This is opposed to underthrusting interplate earthquakes, which rupture at the interface between the overriding and subducting plate. The outer rise events are commonly interpreted to result from bending forces occurring within the oceanic plate, resulting from coupling within the trench (Figure 2.2). If the trench is strongly coupled, the outer rise is expected to be under a compressive stress regime, potentially resulting in compressive outer rise faulting. Following an underthrust event or if the subduction zone is generally poorly coupled, the outer rise may experience a tensional stress regime, resulting in normal faulting (adapted from Choy and Kirby [1]).

2.2 The occurrence and type of outer rise earthquakes is closely related to the degree of coupling within the subduction zone. In a coupled subduction zone, adjacent to a seismic gap, the outer rise experiences a compressive stress regime, potentially resulting in compressive outer rise events. Seaward of regions of the fault that have recently slipped would be transitioned into a tensional stress regime resulting from the slap pull. This is similarly true to uncoupled subduction zones, where the lack of coupling enables the stress of slab pull to be directly propagated to the outer rise. Both of these tensional stress regimes may result in outer rise normal faulting (adapted from Christensen and Ruff [2]).
2.3 The Vanuatu (formerly New Hebrides) subduction zone is characterized by an abundance of large earthquakes, but no great events. This map displays all seismicity $M_w > 5.0$ recorded from 1973 to 2010 in the NEIC record. Additionally, the focal mechanisms between 1976 to 2012 for events $M_w \geq 7.4$ are also displayed [3]. This includes the 16 May 1995 southern Vanuatu normal outer rise event ($M_w 7.7$). This trench is the interface of the Australian plate (west) subducting beneath the Pacific plate (east). The trench turns sharply in the north transitioning into the Solomon Islands. It also has a sharp corner in the south, where it becomes a diffuse strike-slip plate boundary before making another sharp turn to the south, becoming the Tonga subduction zone.

2.4 Seismicity profile along the strike of the Vanuatu subduction zone. Events are scaled relative to their magnitude. There exists an arching structure of clustered events between $14^\circ$-$17^\circ$S. This arch begins and ends at depth around 200 km and shallows to $\sim 150$ km in the middle. Below this structure there is a dramatic lack of seismicity until a depth of about 600 km. Some have attributed this to slab detachment [4, 5]. The northern and southern limits also lack seismicity $> 200$ km.

2.5 The great 2 March 1933 Sanriku outer rise event ($M_s 8.4$) was a shallow normal faulting earthquake occurring off the coast of Japan. This is the largest event recorded in a region that has experienced multiple other large events over the past 40 years. The 1933 event is also the largest outer rise event to have occurred in this region, the next largest in the modern record occurred on 14 November 2005 ($M_w 7.0$). Besides these two events, this region has experienced a number of considerably smaller outer rise events. The shaded region adjacent to the 1933 epicenter denotes the aftershock region defined by Kanamori [6]. The subduction zone south of the 1933 Sanriku event ruptured in the 11 March 2011 $M_w 9.0$ earthquake.
2.6 The Java subduction zone experienced one of the world’s largest outer rise earthquakes on 19 August 1977 ($M_w 8.3$). This shallow normal faulting event ruptured within the Indian Plate near the termination of the Java Trench at its collision with the Australian Plate. The two-week aftershock activity places the epicenter relatively central within the aftershock zone. Several notable earthquakes that soon followed the great 1977 event include another smaller outer rise event east of the great event on 10 April 1978 ($M_w 6.7$) and a moderate-large strike-slip event north of the trench on 07 October 1977 ($M_w 6.5$). This map displays all seismicity $M_w \geq 5.0$ recorded since 1973 in the NEIC record.

2.7 Southern Vanuatu experiences an abundance of outer rise activity. The largest of these events was the 16 May 1995 ($M_w 7.7$) normal faulting earthquake. The fault plane and slip model (Figure 2.10) of this large event are displayed on the upper map. In addition to multiple large outer rise events, southern Vanuatu has experienced multiple large underthrusting events. Besides the two 2003 events, there appears to be poor correlation between underthrusting and outer rise activity. This may be explained by the complex tectonic setting. The aftershock pattern of the 1995 event suggests most of the slip occurred southeast of the epicenter.

2.8 The Kuril subduction zone provides a good example of a relationship between underthrusting events and outer rise activity. The two most notable events in this region were the 15 November 2006 ($M_w 8.4$) great underthrusting event and 13 January 2007 ($M_w 8.1$) great normal faulting outer rise event. The 2006 event ruptured an identified seismic gap, located north of a region that had last ruptured in 1963. The aftershock patterns of two 2006/2007 great events shows very clear parallel banding to one another, underscoring the relationship between the two events (adapted from [7]).
2.9 Northern Tonga represents a complex tectonic setting, where there is a sharp turn in the trench due to a “corner” in the overriding Australian Plate (Tonga Block). This region experiences abundant outer rise earthquakes, nearly all of which are tensional. Included in this outer rise activity was the great shallow tensional event on 29 September 2009 ($M_w$ 8.1). This event initiated abundant faulting along the megathrust. Displayed is the inferred fault plane of this event by Lay et al. [8]. In addition to the outer rise activity, there is also abundant seismicity along the megathrust and within the Tonga Block. Shallow events larger than $M_w$ 6.8 since 1973 are identified (the focal mechanisms is not available for the 1975 event).

2.10 The 16 May 1995 ($M_w$ 7.7) occurred at 23.01$^\circ$S 169.90$^\circ$E (Figure 2.7). The GCMT solution of the focal mechanisms aligns the fault plane at $\phi=280^\circ$ (110$^\circ$), $\delta=35^\circ$ (56$^\circ$), and $\lambda=-99^\circ$ (-84$^\circ$), with rupture nucleating at 24.7 km depth. NEIC assigns a slightly shallower hypocenter of 20 km. This figure shows the results using the search-based algorithm. This finite fault model of the event assumes a rupture speed of 2.7 km/s. We found the rupture plane was roughly 85 km long and 40 km wide down-dip, however, most of the slip occurred in the southeastern 55 km of the plane. The peak slip was 12.8 m, occurring about 20 km up-dip and southeast of the hypocenter. Our model finds a shallower hypocenter than GCMT and NEIC of 12.604 km, however, the rupture plane focal mechanisms are comparable to those of GCMT. We find the rupture to be oriented at $\phi=115^\circ$, $\delta=35^\circ$, and $\lambda=-90^\circ$. The moment rate function is relatively simple, showing the event was not impulsive at its onset. Most of the energy release occurred between about 4 to 14 seconds of the $\sim$17 second rupture duration.

2.11 The wave fits for the suite of teleseismic P- and SH-waves used in the finite fault inversion of the 16 May 1995 southern Vanuatu outer rise event using the search-based algorithm. The black lines are the observed waveforms; the red are the predicted. This model provides a strong fit to the first $\sim$25 seconds of the waveforms. After 25 seconds, the model has difficulty fitting the second energy arrival.
2.12 These are the model results using the Kikuchi-Kanamori algorithm of the 16 May 1995 (Mw 7.7) outer rise event. Using this model we resolve a slightly slower preferred rupture speed of 2.0 km/s than with the search-based algorithm. The rupture plane was also longer with dimensions of ~100 km long and 40 km wide down-dip. Where in the search-based algorithm we found most of the slip to occur in the southeast, the Kikuchi-Kanamori algorithm produces a largely bilateral slip pattern with large amounts of slip also in the northwest. Both models find significant slip along the up-dip limits of the fault plane. The preferred hypocenter depth of 20 km is close to the depth catalogued by GCMT and NEIC. The focal mechanism in this model, $\phi=116^\circ$, $\delta=48^\circ$, and $\lambda=-68^\circ$, are very similar to the search-based algorithm, but with some variation in the rake. This model is able to pick up the second pulse of energy. This is seen in both the waveforms (Figure 2.13) and the moment rate function.

2.13 The wave fits for the suite of teleseismic P- and SH-waves used in the finite fault inversion of the 16 May 1995 southern Vanuatu outer rise event using the Kikuchi-Kanamori algorithm. SH-waves were significantly down-weighted. The black lines are the observed waveforms; the red are the predicted. This model provides a strong fit to at least the first $\sim$50 seconds of the waveforms.

2.14 Comparing the 16 May 1995 (Mw 7.7) southern Vanuatu outer rise event to similar magnitude local thrust events displays that, unlike other large outer rise events, the Vanuatu event does not appear to be enriched in short-period energy. With the moments scaled to the 1995 event, a seismogram of the 1-second energy (upper) shows the outer rise event had less of this high frequency energy than the 12 February 1994 (Mw 6.9) underthrusting event, and similar to that of the 27 December 2003 (Mw 7.3) underthrust. The 1995 outer rise event displays similar or lower 1-second $m_b$ and 20-second $M_s$ energy as the two local underthrust event (lower).
2.15 The bathymetric map of Vanuatu shows multiple ridges are impinging the trench. Around 15°S, the d’Entrecasteaux ridge is being subducted. Its resistance to subduction is believed to influence not only the local seismicity but also the convergence of the two plates [9]. In the south is the collision between the Loyalty Ridge of the Australian plate with the Vanuatu arc at 22°S. GPS studies note a change in the Australia/Vanuatu relative motion about this collision zone [10], leading some to believe this change is due to the collision [11]. The arrows indicated the motion of the Pacific relative to Australian plate, with the length of the arrow scaled relative to the difference in convergence rate, decreasing from ∼12 cm/yr in the south to ∼4 cm/yr in the north [9] (adapted from [12]).

2.16 The intraplate events (red) display enrichment in high frequency energy (high energy/moment value) relative to interplate (blue) earthquakes. The three Kuril events provide a clear example of the local differences, where both the 2007 and 2009 intraplate events have much higher energy/moment values than that of the 2006 interplate event. This is despite the 2006 event having a higher moment magnitude. (Modified unpublished figure by Lay and Kanamori).

3.1 Left panel: Seismicity of the northern Vanuatu subduction zone, displaying all NEIC earthquake hypocenters since 1973. The Australian plate subducts beneath the Pacific in nearly trench-orthogonal convergence along the Vanuatu subduction zone. Right panel: All GCMT moment tensor solutions and centroids for $M_w \geq 5$ since 1976, scaled with moment. This region experiences abundant moderate and large seismicity events, but lacks any great event ($> M_w 8$) since 1900. The largest events are displayed with dotted outlines of the magnitude-scaled circle. Convergence rates are calculated using the MORVEL model for Australia Plate relative to Pacific Plate [13]. (See Figure B.1 for a color version).

3.2 100-day aftershock maps of all events listed in the ISC catalog for the 1966 sequence and NEIC catalog for the 1980, 1997, 2009, and 2013 northern Vanuatu earthquake sequences (the 2013 event only shows up to the current date, 40-days). The 1966 mainshocks use those listed by Tajima et al. [14]. Events of the 1997 and 2009 sequences were relocated using the double difference method [15] for P-wave first arrival based on EDR picks. The event symbols are scaled to relate the symbol area to the earthquake magnitude based on a method developed by Utsu and Seki [16].
3.3 The 7 October 2009 rupture sequence in northern Vanuatu subduction zone, including the 3-month aftershock sequence (Mw > 4). The sequence began with the Mw 7.7 event (1), followed ∼15 minutes later by the Mw 7.8 event (2). Finally, about an hour later the Mw 7.4 event ruptured (3). The circles located on the NEIC epicenters and the focal mechanisms are from GCMT. Convergence rates are calculated using the MORVEL model [13].

3.4 R1 source time functions (STFs) for the 1997 mainshock earthquake. The rupture appears to propagate in the direction 200°N with an estimated rupture speed of 2.0 km/s and a 75-80 s duration.

3.5 Finite fault model based on teleseismic P- and SH-wave inversion for the 1997 event using a kinematically constrained linear least-squares procedure. Here we show the fit to the steeply dipping nodal plan of the GCMT solution (B.13). All models display a similar characteristic of small slip near the hypocenter and peak slip shallow, correlating with the region of aftershock activity.

3.6 Kikuchi – Kanamori inversion models [17] for a fixed (left) (B.14) and variable mechanism (right) (B.15) of the 1997 mainshock. Both inversion models use a similar depth source as the finite fault model (Figure 3.5) and four subevents. Results for the fixed mechanism show the location of the subevents on the rupture plane. Results for the variable mechanism inversion shows the location of each subevent along strike. Below the subevents is the total mechanism for the entire event as a whole. The contribution of each subevent to the moment rate function is labeled.

3.7 R1 source time functions (STFs) for the first (top) and second (bottom) large 2009 earthquakes. The first rupture shows little evidence of directivity, and appears to have about 40-50 s duration. The second rupture appears to propagate to the north (N340°) with an estimated rupture speed of about 2.1 km/s and an ∼80 s duration.
3.8 Finite fault slip distribution from inversion of teleseismic broadband P- and SH-waves for the first event in the 2009 sequence using a kinematically-constrained linear least-squares procedure. The waveform fits are shown in Figure 3.9. The hypocenter depth is 39 km, rupture speed is 2.5 km/s, and strike = 346°, dip = 45°, and average rake = 86°. This mechanism is similar to the GCMT solution (strike = 344°, dip = 41°, and rake = 87°). The moment rate function indicates steady growth of slip for the first ~20 seconds until the primary asperity, located north of the hypocenter (left star), finally failed. The NEIC hypocenter of the second event in the sequence is located north and slightly up-dip of the first (inverted right star). The arrows indicate the rake and relative slip amplitude at each subfault. The event had a peak slip of around 8 m. A velocity boundary in our layered source structure enhances the along-dip gradient in slip just below the depth of the hypocenter. The white circles within the focal mechanism show locations of stations used in the inversion; the location relative to the center of the mechanism reflect the station azimuth and distance from the source.

3.9 Waveforms fits for the inversion model of the first 2009 event (Figure 3.8). The predicted waveforms (thin) successfully match the major features of the observed waveforms (thick). Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, wave type, and azimuth (degrees). The timescale is in seconds, beginning 10.5 s prior to the P- and SH-wave arrivals.

3.10 Azimuthal plots of the time delays between R1 arrivals for (top) the first and second event, and (bottom) the first and third events. The smooth trends, fit with a cosine curve provide estimates of the relative locations of the event centroids.

3.11 Space-time plot of the 7 October 2009 earthquake sequence. Distance is measured to the south from -11°N, 165.5°E along the strike of the trench. All events with NEIC magnitudes M > 4.5 are displayed. Aftershock activity remained elevated for ~50 days following the mainshock in the north, but quickly dissipated south of this region. The initial rupture appears to have not extended very far south of the third event. However, aftershock activity displays a southern progression, before stopping after about a week.
3.12 Long-period waveform matching of the October 2009 sequence (solid) and July 1980 (dashed). Given the close temporal proximity of the two largest 2009 events, it was not possible to effectively isolate a single event. Consequently, we combined the two 1980 mainshocks with a best-fitting offset to create a composite waveform, simulating the second 1980 event as having a similar time delay as the second 2009 event.

3.13 R1 source time functions (STFs) for the 2013 mainshock earthquake. The rupture appears to propagate to the southeast (N135°) with an estimated rupture speed of about 1.5 km/s and a ~105 s duration.

3.14 Finite fault model based on teleseismic P- and SH-wave inversion for the 2013 event using a kinematically constrained linear least-squares procedure. Here we show the fit to shallow source (18 km) with shallowly dipping plane (dip = 17°) (Appendix B.2, B.4, B.4, B.5, B.6). All models display a similar characteristic of high slip near the asperity, but peak slip shallow to the south-southeast. While peak slip occurs shallow on the fault plane, the peak moment is located near the hypocenter (Appendix B.7). The difference is from the change in shear modulus with depth.

3.15 Kikuchi – Kanamori inversion models [17] for a fixed (left) (B.8) and variable mechanism (right) (B.9) of the 2013 mainshock. Both inversion models use a shallow source (18 km), similar to that used in the finite fault model (Figure 3.14) and four subevents. Results for the fixed mechanism show the location of the subevents on the rupture plane. This model does not resolve the shallow slip observed in the finite fault model. Results for the variable mechanism inversion shows the location of each subevent along strike. Below the subevents is the total mechanism for the entire event as a whole. The contribution of each subevent to the moment rate function is labeled. Models for a deeper source and steeply dipping plane are shown in Appendix B.10, B.11, B.12.

3.16 40-day aftershock pattern of 2013 earthquake, including all events listed in the NEIC catalog. Focal mechanisms are located and scaled to the NEIC catalog information. 19 of the largest events (not including the mainshock) were relocated using surface waves [18] (Table B.1).
3.17 Summary schematic showing the spatial relationships of recent large earthquake ruptures in northern Vanuatu. We interpret the 1997 event as an intraplate rupture based on a collective assessment of broadband seismic observations (ambiguous broadband P waveforms, unusual faulting geometry, deep long period centroid, and unusual aftershock spatial relationships). Due to limited waveform data, we are unable to constrain how much the 1980 and 2009 rupture zones overlapped. However, analysis of available waveforms indicates the 2009 sequence was not an exact repeat of the 1980 (Figure 3.12).

3.18 Space-time plot (right) of all seismicity greater than M 5.0 in northern Vanuatu recorded in the NEIC catalog as a function of distance south of -10°N, 165.25°E. The figure on the left shows the location of the seismicity on a map rotated to make the trench oriented vertically.

3.19 Envelopes of velocity filtered to emphasize the band of 2-4 Hz as observed at MAJO and CHTO stations. The time scale is time from the origin of the respective event. The dotted line shows the relative amplitude, while the solid is absolute. All velocities are normalized for magnitude, depth, and focal mechanism difference, based on the average of modeled P, sP, and pP amplitudes.

4.1 Example waveforms for three events from the Panama Fracture Zone observed at two different stations of significantly different distance (left side panels). The right side panels show the cosine fits to the observed surface-wave time differences between these events. The red dashed curve shows the fit predicted by the original USGS/NEIC location.

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Dedication

To my family and the belief that simply following what interests you will always lead to right place.

xxx
Chapter 1  
Introduction

1.1 Fault Rupture Segmentation

The recent great earthquakes in Tohoku-oki and Sumatra-Andaman-Islands region demonstrated that we continue to be surprised by where the largest earthquakes occur. Both of these regions were known for the large earthquakes they could host. However, traditional seismic hazard perspectives regarding plate-boundary segmentation, plate age, and trench-normal convergence rates led us to believe the Sumatra-Andaman subduction zone was not capable of rupturing in a M9+ event [28]. The Tohoku-oki earthquake redefined our understanding of the seismic hazard that the shallow portions of a subduction megathrust can present [29]. This event also highlighted that even the longest available seismic record might be too short to accurately assess the total seismic potential of a region. Implicit in the lessons from these events is a fundamental challenge in understanding fault rupture segmentation. What determines if a fault segment will rupture in a single great event or many moderate events? How representative are past events for assessing future seismic hazard? How persistent are seismic asperities over time and do they always rupture in a similar manner? Are there fundamental differences in the manner in which subduction, normal, and strike-slip systems respond to elastic strain accumulation?

There exist two end-member cases in the conceptual model of rupture behavior of large faults. The first, the characteristic earthquake model, is based upon observations along the San Andreas and Wasatch fault systems where analysis suggests that a portion of the fault tends to repeatedly rupture over several seismic
cycles in an earthquake of similar magnitude and faulting behavior [30,31]. The model reflects the idea that a fault segment’s earthquake behavior is controlled by the presence of a region of relatively high strength (a seismic asperity) that changes little over a few seismic cycles. The fault principally remains locked until enough slip can accumulate to re-rupture this asperity [30]. Under uniform loading rates, the recurrence interval between re-ruptures would be expected to be similar. Given the periodic nature of these events, this conceptual model provides a general characterization of the expected dominant seismicity for a portion of a fault zone in terms of magnitude and recurrence interval [30]. The second end-member model is the variable slip model, which suggests that the magnitude, location, and timing of events vary along the fault, with no consistent pattern [32]. In some sense at least some complexity is required; the characteristic model causes different amounts of slip on adjacent segments, which would result in an unequal accumulation of slip if an identical rupture pattern were to repeat for more than a few cycles. The reality is probably a mix of behaviors, with some semblance of repeated failures in some areas, but perhaps a more variable behavior over more than a few seismic cycles (e.g. as in Nankai).

1.1.1 Repeated Large Earthquakes

In order to address fault rupture segmentation at a level of deep understanding, one must consider where events start and stop, the nature of slip in the events (including how “repeatable” large earthquakes rupture characteristics are from one sequence to the next), and the relationship between seismicity patterns and the physical characteristics of the plate boundary. Assuming earthquakes nucleate around regions of high stress (asperity) [33], one must consider how the role of an asperity changes over the seismic cycle. Is the seismic behavior of a region modulated by a single asperity, or a collection of interacting asperities? Do clusters of asperities consistently rupture in a similar pattern? A fundamental question regarding the characteristic earthquake model is whether subsequent earthquakes re-rupture the same fault areas as ruptured in earlier sequences or whether there is only a localization of earthquakes along the same portion of the fault. Studies of the 1940 (Mw = 6.4, SCEC) and 1979 (Mw = 6.4, NEIC) Imperial Valley earthquakes, the 1957 (Mw = 8.6), 1986 (Mw = 8.0), and 1996 (Mw = 7.9) Aleutian
Islands earthquakes, and the 1963 \( (M_w = 8.5) \) and 1995 \( (M_w = 7.9) \) Kuril Islands earthquakes are all examples where the subsequent events are believed to re-rupture at least part of the previous asperity [30]. In the Aleutian Islands, while the 1996 and 1986 event ruptures likely did not overlap, both overlap the inferred 1957 rupture to varying degrees [30, 34, 35]. Hwang and Kanamori [36] suggested the presence of two closely spaced strong asperities; each asperity is associated with the epicenter of the 1957 or 1986 event. The 1940 and 1979 Imperial Valley earthquakes have been cited as examples of strike-slip characteristic earthquakes. But the 1979 event re-ruptured only the northern half of the 1940 event. While the more recent earthquake is only half as long as the older rupture, both experienced the same amount of slip on the similarly ruptured portion, providing a good example of a characteristic earthquake [32]. Thus, even the idea of characteristic earthquake is somewhat relative when applied to observed ruptures.

In addition, re-rupture of a similar portion of the fault surface does not automatically support a characteristic earthquake model. Inversion of the 1963 and 1995 Kuril Island earthquakes indicates the epicenter of each event ruptured a common asperity. Comparing the amount of slip at this location between the two events shows that the 1963 event experienced 5-10 times more slip at this location than in 1995 [30, 37]. Additionally, while assessment of seismic risk based on recurrence interval has had limited success along certain plate boundaries, Nishenko [38] had classified this segment of the Kuril fault as low seismic risk prior to the 1995 event. While these events probably re-ruptured the same asperity, the variable slip and inconsistent recurrence led Schwartz [30] to conclude the Kuril events are a direct counter-example to the characteristic earthquake model. More recent examples that illuminate the complexity of earthquake rupture patterns include aspects of the 2011 Maule and 2011 Tohoku-oki earthquakes. The area encompassing the region that failed in the 1985 Chile earthquake sequence also hosted a substantial number of strong aftershocks during the 2010 Maule earthquake sequence. At least part of the down-dip region that ruptured in the 2011 Tohoku-oki earthquake also failed in isolated ruptures during a group of earthquakes in the 1930’s. In chapter 3, I investigate this topic with focus on multiple double sequences in northern Vanuatu. This research exposes that while the behavior of subsequent doublets was similar in event magnitude and rupture geometry, important differences indicated these were not exact repeats.
Despite the numerous studies of specific repeated large-earthquake sequences, these investigations have necessarily been focused on localized regions and small data sets. One way to expand the data set is to look at smaller earthquakes—with the hope that interactions among the smaller events will lead to insights into larger event rupture processes. Working with smaller events reduces the recurrence time for the events but increases the challenges of imaging the earthquake slip characteristics or event centroid location of the events. One aspect that we can improve is the relative locations of earthquakes. For many decades, the use of phase picks to locate earthquakes has ignored the valuable information contained in the seismic waveform. This information can be used in concert with the phase arrivals to improve the locations of moderate-size earthquakes using, for example, waveform correlation methods [18,39]. Chapters 4 and 5 demonstrate an improved method of calculating precise, relative locations for oceanic transform fault earthquakes using surface waves. Chapters 5 extends this method, using cross-correlation methods to calculate accurate relative event magnitudes. As these chapters, as well as chapter 3, demonstrate, some improvement in location and spectral characteristics (when resolvable) may help identify regions of differing physical characteristics that may relate back to the segmentation boundaries, if there is a clear physical control on such features.

1.1.2 Asperity Interaction

In addition to considering the role a single asperity plays throughout the seismic cycle, another important consideration is the relationship between adjacent asperities. As demonstrated in chapters 4 and 5 and in work by VanDeMark [40], seismicity along oceanic transform faults has shown some faults have a tendency to rupture with identifiable spatial patterns; an entire fault length will rupture in numerous events in sequential order from one end of the fault to the other. At the same time, however, nearby faults may display no order in rupture sequence. Asperity interaction also applies to larger scales, such as asking what causes a rupture to continue to propagate 100+ km, while others stop short and fail in multiple smaller events? Another example of earthquake interaction between adjacent asperities is doublets. Chapter 3 displays how observed doublet pattern along the northern Vanuatu subduction zone is quite remarkable because the tendency to
fail in doublet-events seems to occur with overlapping, yet different combinations
of asperities in each sequence.

The occurrence of these doublets raises the question as to why a fault segment
would rupture in two closely spaced (in time and space) earthquakes instead of a
single event. Lay and Kanamori [33] suggested the presence of doublets in a fault
zone may be reflective of uniformity of asperity patches. If so, the asperities must
reach near-failure stress levels at about the same time. Scholz [41] noted that fault
(or asperity) synchronization might be a widespread phenomenon. But Felzer et
al. [42] concluded that statistically the occurrence of doublets is in accord with
a typical physical rupture model based on a single triggering event. Accordingly,
there is no warrant for a unique physical mechanism to generate large doublets.
But in northern Vanuatu, since 1960, five, possibly up to seven, of the eleven events
M < 7.5 ruptured in doublets in northern Vanuatu. This significant proportion of
large events displays this region has a strong proclivity to rupturing in doublets,
suggesting the presence of a synchronizing rupture mechanism. Arguing statistically
with small samples results in unresolved significance (small samples make many
explanations possible, including coincidence, but does not preclude an actual
physical cause).

1.1.3 Seismic Hazard Assessment

The behavior of single asperities and interactions between multiple asperities in
the earthquake cycle coalesce in the characterization of entire fault zones. A classic
approach to seismic hazard is to look for apparent “seismic gaps,” or regions of
an active fault zone that relative to surrounding seismicity display a deficiency
in seismic slip. These regions are defined as having increased potential of future
rupture (e.g. [38,43,44]). This model has provided mixed success, and has been
statistically tested and strongly criticized by some (e.g. [45]). Early results were
based on relatively poorly constrained estimates of earthquake rupture areas (in
particular along the dip direction) and so perhaps the ideas are not as bad as their
application appears in statistical assessments. There remain numerous challenges
to a full assessment or accurate application of the method, and the greatest is our
incomplete historical seismic records. One example of this challenge is assessing the
degree of coupling of a fault zone; specifically, it is unclear how much convergence
across the fault is accounted for in seismic versus aseismic slip. Implicit in this challenge is assessing the extent of the seismogenic zone. Chapter 6 explores this topic through investigation of rupture directivity and how it relates to fault length and rupture speed. In combination with seismic moment, rupture length and speed can be used to estimate rupture zone depth extent.

Assessment of fault zone seismic coupling is additionally challenged if the amount of observed slip during large events differs from that expected based on the elastic strain accumulation along a segment (i.e. variable stress drop) (e.g. [46]). While there were 30-60+ meters of slip during the recent Tohoku-oki event, this amount of slip still does not account for all the slip expected to have accumulated since the last recorded event in AD 1600 if the modern rate of strain accumulation has been uniform since that time [47]. From a seismic hazard perspective, it remains unclear if this “missing” slip will result in another large event in the near future. Cascadia is another example of a fault zone where it remains unclear how much convergence is actually stored as elastic strain energy and how far east the coupling extends. This problem also strikes to the heart of the characteristic earthquake model [48]. The 2004 Sumatra-Andaman earthquake was unexpected because a previous event of this scale had never been observed in this region. We now know most of the slip of this region is accounted for in rare, massive earthquakes, but prior to 2004, the hazard characterization for this region was completely different.

There are two primary ways to approach seismic hazard assessment. First, one could assume a Gutenberg-Richter (G-R) relationship [49], where the frequency and magnitude of seismicity on a fault follows power-law relationship of \( \log N = a - bM \), where \( b \) is roughly 1. This model is based upon a probabilistic growth of rupture size under self-similarity and is universally invariant. One can also assume a characteristic earthquake model, also known as maximum magnitude model (e.g. [31, 50]). This type of model would expect more large earthquakes along a fault than predicted by G-R. Validation of both of these models is heavily influenced by the completeness of the seismic record. These models have been heavily tested in multiple locations, notably in California where a high density of seismic networks affords very detailed and comprehensive (in magnitude range) catalogs. Model validation is a critical focus because the preferred model has large implications for probabilistic seismic hazard analysis (PSHA). PSHA describes the relationship between a metric of ground motion (e.g. peak ground acceleration)
and an events average recurrence interval [51].

A fundamental question to applying PSHA on a global scale is whether similar methods of hazard analysis can be applied to different fault systems (e.g. subduction, normal, and strike-slip). For example, analysis of scaled energy to moment has indicated recent large intraplate slab events display enrichment in high frequency energy as opposed to large interplate earthquakes [8]. High frequency energy is particularly damaging to building structures. This question has begun to be addressed with studies like Schorlemmer et al. [52] looking at variability of $b$ values under different faulting styles and Wiemer and Wyss [53] comparing variability in $b$ values along the San Andreas. The topic of intraplate versus interplate rupture is discussed in chapter 2 through a global survey of outer rise seismicity as well as detailed analysis of the 1995 southern Vanuatu large outer rise event. Discussion in this chapter also addresses another important earthquake hazard question concerning how seismicity behavior may interact with different local fault structures.

Finally, the recent Tohoku-oki event suggested a possible depth dependent frequency content in large megathrust earthquakes [54]. This depth dependence has major implication for a seismic hazard. As proposed by Lay and Kanamori [29], the 2011 Tohoku-oki event produced both a large tsunami and strong shaking because both the shallow and deep portions of the megathrust ruptured. The 1896 Sanriku-oki event produced a large tsunami, but strong shaking was not felt, suggesting the deeper portions slipped aseismically. This displays that not only the size of events should be of concern, but also which portions of the fault zone rupture affect the hazards. The question of depth dependent frequency content can be explored by comparing the spectral characteristics of events from varying depths along the same subduction zone. The seismicity in Northern Vanuatu discussed in chapter 3 offers a unique perspective at this condition because two very similar sets of earthquakes ruptured at what appears to be differing depths. Analysis of frequency content of these events suggests a possible depth influence proposed described by Koper et al. [54].
1.1.4 Summary

A critical foundation to earthquake study and hazard assessment is the understanding of controls on fault rupture, including segmentation. The following studies explore this topic of fault rupture segmentation through multiple perspectives. Additionally, given the volume and quality of seismic data, as well as the computing capabilities readily available, we find ourselves in an unprecedented era. In the following studies, we also present modern methods that utilize the current data and computing capabilities necessary for answering the difficult questions regarding fault rupture segmentation.

1.1.5 Chapter 2: Uniqueness and Commonality of Outer Rise Earthquakes: Southern Vanuatu, 16 May 1995 (MW 7.7)

This chapter investigates outer-rise seismicity, which are intraplate earthquakes located seaward of the deep-sea trench that marks the plate boundary. A lion’s share of this chapter focuses on a compilation of the state-of-knowledge of global outer-rise seismicity. Particular focus is placed on the largest observed outer-rise events. This chapter also contributes detailed analysis of the rupture behavior of the 16 May 1995 southern Vanuatu MW 7.7 outer-rise earthquake, including finite fault modeling of the slip distribution. Finite fault analysis was performed in collaboration with Charles Ammon. We plan to publish an abbreviated version of this chapter, focusing on an updated analysis of the relationship between outer-rise and megathrust seismicity.

1.1.6 Chapter 3: Large Earthquake Processes in the Northern Vanuatu Subduction Zone

On 07 October 2009 the northern segment of the Vanuatu subduction zone ruptured in three major (MW 7.7, 7.8, and 7.4), shallow thrust earthquakes in close spatial proximity and occurring over about one hour. Prior to this sequence, the same portion of the fault ruptured in a large doublet in 1980 and a large single event in 1997. On 06 February 2013, an MW 8.0 ruptured just north of the 2009 sequence in a region previously ruptured by a 1966 large doublet. This chapter explores the relationship between all of these events through use of numerous analytical
methods. This chapter is co-authored by Charles Ammon and Thorne Lay. Thorne Lay collaborated with some of the slip modeling and interpretations. Charles Ammon contributed the source time function modeling and collaborated with the slip modeling, event relocation, and general interpretations. This chapter is in the final revisions of being converted into a format to be submitted for publication.

1.1.7 Chapter 4: Precise Relative Earthquake Location Using Surface Waves

In isolated continental and offshore areas, earthquake locations must rely heavily on distant observations, often resulting in inaccurate and imprecise locations. This chapter provides thorough development of a method for earthquake location using cross-correlation of Rayleigh (R1) surface waves. We apply this method to relocate vertical strike-slip seismicity along the Panama and Balboa Fracture Zones, located south of Panama. This method was initially developed by Charles Ammon and applied by Ammon and Thomas VanDeMark [40]. In this chapter, we provide a thorough sensitivity analysis of all the model variables and use synthetic seismograms to investigate the influences that differences in faulting parameters and depth have on the relocation procedure. Work on this chapter was completed in collaboration with Charles Ammon. This chapter has been published in Journal of Geophysical Research [18].

1.1.8 Chapter 5: Precise Relative Earthquake Locations and Magnitudes of Seismicity in the Northern Pacific

This chapter extends work by chapter 4, applying the relocation method to seismicity in the northeast Pacific region, including the Blanco Fracture Zone, and Gorda, Juan de Fuca, and Explorer Plates. Application of the relocation method was expanded to also include normal faulting events in addition to strike-slip. Additionally, we demonstrate how cross-correlation coefficient measurements can be used to calculate relative earthquake magnitudes. Work on this chapter was completed in collaboration with Charles Ammon and is in the final revisions to be submitted for publication.
1.1.9 Chapter 6: Strike-Slip Earthquake Rupture Directivity Using Surface Waves

Analysis of measurements of the 27 October 1994 \( (M_w \, 6.3) \) Blanco Fracture Zone earthquake in the previous chapter displayed the ability to use unnormalized correlation coefficient to measure rupture directivity in large earthquakes. In this chapter, we explore application of this method to estimate rupture length and speed for several large strike-slip earthquakes around the world. We also provide detailed analysis of sensitivity of this method to variations in depth and rupture parameters using synthetic seismograms. Analysis in this chapter was performed in collaboration with Charles Ammon. This chapter will be expanded and submitted for publication.
Chapter 2  |
Uniqueness and Commonality of Outer Rise Earthquakes: Southern Vanuatu, 16 May 1995 (Mw 7.7)

2.1 Introduction

Earthquakes have a rich diversity that reflects the wide range of plate interactions and stress transfer processes that ultimately give rise to these powerful events. Most large earthquakes occur near subduction-style plate boundaries, where the lithosphere of one plate thrusts beneath another. Associated with most of these subduction boundaries is another class of events that occur within the subducting oceanic lithosphere. These intraplate earthquakes, located seaward of the deep-sea trench that marks the plate boundary, are generally attributed to the bending of an oceanic tectonic plate as it approaches the plate boundary [2, 55, 56] (Figure 2.1). Seismologists often refer to these events as “outer rise” earthquakes.

Outer rise earthquakes are observed worldwide in most, but not all, of Earth’s subduction zones. These events can help indicate the regional lithospheric characteristics and degree of seismic coupling of a subduction zone megathrust as well help as constrain stress transfer processes in underthrusting oceanic lithosphere during the seismic cycle (i.e. [55], [2], [57]) (Figure 2.2). Outer rise earthquake-related structures are believed to play a significant role in the transport of water
10’s of km down into the subducting plate, where it may play important roles in intermediate-depth earthquakes as well as volcanic processes associated with subduction [58,59]. And finally, although less frequent than large underthrusting events, large outer rise earthquakes may pose significant hazard since they have been found to display higher stress drops and greater enrichment in high frequency shaking than comparable size interplate events [7,8].

The Vanuatu (formerly the New Hebrides) subduction zone is located in the southwest Pacific region, southeast of the Solomon Islands, west of the island nation of Vanuatu, and east-northeast of the Loyalty Islands, between 11°S to 23°S and 165°E to 171°E (Figure 2.3). A unique pattern in the Vanuatu seismicity is a high density of outer rise earthquake activity near the southern extent of the trench before it makes the sharp turn to the east, highlighted by the occurrence of a large tensional outer rise event on 16 May 1995 (Mw 7.7). These events seaward of the subduction zone trench are diffuse and do not appear to be localized on any single structure. The subduction zone’s unique tectonic setting heavily influences southern Vanuatu’s outer rise activity.

The Vanuatu boundary accommodates the subduction of the Australian plate beneath the Pacific plate. The Vanuatu subduction zone is believed to have recently been a part of an ancestral Solomons-Vanuatu-Fiji-Tonga subduction zone. Disruption in the late Miocene caused subduction polarity reversal of the Vanuatu segment. The present day Vanuatu subduction zone subsequently rotated clockwise to the southwest to its present day position, opening the Fiji and Lau basins [60–63]. The NUVEL-1A global plate motion model predicts a present day Australia-beneath-Pacific plate convergence rate of 79-86 mm/yr nearly perpendicular to the trend of the boundary [9]. However, the geodynamics of this plate boundary are unique. At the southernmost extent of the boundary, plate geometry changes rapidly forming a corner (Figure 2.3) and transitions from the nearly orthogonally convergent boundary into an oblique (nearly strike slip) boundary extending to the east-northeast.

Using GPS measurements, Calmant et al. [10] measured convergence rates that ranged from around 30-40 mm/yr in the central segment (~15°S) to over 89-124 mm/yr in the south (~20°S). Taylor et al. [9] suggested that the non-uniform convergence along the boundary is due to the difficulty subducting the relatively buoyant d’Entrecasteaux Ridge (located at ~15°S). Schellart et al. [12] suggested
slab rollback in the south is being resisted by the d’Entrecasteaux Ridge. Govers and Wortel’s [64] STEP model supports this idea, where the Pacific plate tears along the strike-slip fault, allowing rollback of the Australian plate. Consistent with this idea, Millen and Hamburger [65] found evidence of plate tearing in this region.

The Vanuatu subduction zone is characterized by frequent moderately-sized earthquakes ($M_w$ 5-7), but few events larger than $M_w$ 7 [3, 66, 67]. Studying the seismicity in map view, several patterns stand out (Figure 2.3). First, near the latitudes 18.5°S to 20°S, events step away from the trench and toward the east (arc-ward), producing a seismic quiescent zone near the trench. A three-dimensional view of the seismicity of this region reveals an arching structure illuminated by the hypocenters. At around 14.5°S there is a clustering of events at $\sim$200 km depth. This clustering arches up to $\sim$140 km at 15°S, then back down again to $\sim$200 km at 15.5°S (Figure 2.4). The region below this arch is nearly devoid of events until a depth of about 600 km. Several studies have suggested slab detachment in the region to account for the seismicity pattern and the uplift of islands of southern Vanuatu [4, 5]. Louat et al. [68] also noted a lack of deep (>200 km) Benioff Zone in the northern and southern extents of the trench (Figure 2.4). Conrad et al. [69] consider the subducting slab at Vanuatu to be mechanically strong, where by “strength” they mean the fraction of slab pull accounted for in plate convergence rate. A lack of great earthquakes along the boundary suggests that it is seismically weakly coupled [70]. However, this interpretation is contingent upon the completeness of the seismic record, which may not reflect an earthquake cycle longer than the present record.

The magnitude of the southern Vanuatu 16 May 1995 ($M_w$ 7.7) places it among the largest outer rise events recorded. In order to explore the contribution the 1995 Vanuatu earthquake makes to the catalog of outer rise earthquakes knowledge, this study first considers the general characteristics of subduction zones that host outer rise earthquakes. Investigation includes analysis of the interaction between outer rise earthquakes and faulting along the megathrust, in addition to the influence of tectonics and structure on the occurrence of intraplate activity. Highlighting this survey is comparison of notable large and great outer rise events from Samoa, Kuril Islands, Vanuatu, Sanriku, and Sumba over the past 80 years, illuminating common qualities of outer rise faulting. Detailed finite fault analysis of the 1995
Vanuatu event illuminates the slip and moment distribution across the fault plane and the timing of the rupture process. These details of the rupture process provide important insight into the mechanisms that cause outer rise faulting and the hazards associated with these events.

2.2 Global Survey of Outer Rise Earthquakes

2.2.1 Western Pacific, Caribbean, Scotia, and Indian Ocean

The large region including the Western Pacific, Caribbean, Scotia, and Indian Ocean has experienced primarily only tensional outer rise events. Several small, shallow compressional events have occurred very close to the trench zone [66], making it unclear if they are intraplate or interplate. This broad regional grouping includes Java, Sumatra, Philippines, Mariana, Izu Bonin, Japan, Ryukyu, Caribbean, and Scotia. Prior to the 26 December 2004 great Sumatra earthquake (M$_w$ 9.1), the largest recorded event in this large region was an M$_w$ 8.2 on 17 February 1996 in Indonesia [66]. Looking specifically at the Western Pacific and Indian Ocean, there is no apparent spatial-temporal correlation between the tensional outer rise events and the interplate activity. Following the great 2004 megathrust, some faulting occurred oceanward of the hypocenter, however, a linear pattern in the activity oblique to the trench [66] suggests its occurrence is related to an existing structure and not outer rise dynamics.

There have been rare occurrences of extremely large tensional outer rise events. These include the 1933 Sanriku (M$_w$ 8.4) (Figure 2.5) and 1977 Sumba (M$_s$ 8.3) (Figure 2.6). These massive events are inferred to have possibly ruptured through most or the entire lithosphere [6,71]. The stresses for these events may have resulted from variation in the interplate coupling along the subduction zone. In uncoupled zones, tensional stress is continuously transmitted from slab pull. When subduction is temporarily slowed or stopped in a region, this may result in an accumulation of tensional stress in adjacent regions where subduction has not slowed [2]. On 10 August 2009 a large normal faulting event (M$_w$ 7.5) ruptured near the Andaman Islands. It is unclear if this shallow event occurred oceanward or landward of the trench boundary because delineation of the trench in this region is difficult.

The large 1933 Sanriku event occurred off the northern Japan trench. This
event was south of a trench section that later failed in the 1968 $M_s$ 8.1 Tokachi-Oki underthrust. Kawakatsu and Seno [72] note the region of the trench adjacent to where the 1933 event ruptured appears to lack great thrust type events for the past 200 years. This characteristic, while bordered by trench segments that have experienced great thrust events, leads the authors to infer the 1933 event’s region is not strongly coupled. The study suggests the strong coupling in the adjacent fault regions could have resulted in normal component stress concentration prior to the 1933 event. While the 11 March 2011 Tohoku earthquake ($M_w$ 9.0) ruptured in a region identified by Kawakatsu and Seno [72] as a seismic gap, there was no anticipation for the greatness of magnitude as observed in 2011. This underscores the need for caution when assessing seismic hazard based on observed seismicity.

The 1977 Sumba earthquake occurred on the eastern extent of the Sunda trench, south of Sumba Island, near the collision of the Australian continental lithosphere with the Sunda-Banda arc. Similar to Sanriku, lack of great interplate thrust events in the region of the 1977 event is believed to be indicative of decoupling at the plate interface. This, in concert with the Australian continental lithosphere’s resistance to subduction, is suggested to have resulted in concentration of slab pull forces up-dip. These slab pull forces produced the unbending stresses in the outer rise, resulting in the 1977 event [71].

### 2.2.2 Solomon Islands

The Solomon Islands region exhibits abundant outer rise activity, as well as a strong correlation between outer rise and interplate earthquakes. During the 1970s this region experienced six underthrust events, ranging in magnitudes of 7.3 to 8.1. The combined aftershock patterns indicate these events ruptured most of the Solomon Islands subduction zone. These events occurred in three sets of doublets [33]. The 12 and 26 July 1971 underthrust doublet ($M_w$ 8.0, 8.1) was followed by six tensional outer rise events between 1971 and 1984. Two of the largest of these events include the 14 September 1971 ($M_w$ 6.3) and 17 August 1972 ($M_s$ 7.1) earthquakes. The later 31 January and 1 February 1974 ($M_w$ 7.3, 7.4) and 20 July 1975 ($M_w$ 7.7, 7.4) underthrusting doublets were followed by one tensional outer rise event on 1 May 1977. The 1970s underthrust activity was also preceded by a small tensional outer rise event on 28 September 1967. This section of the trench is located between the
1971 and 1975 doublets. The occurrence of the tensional outer rise event suggests this portion of the trench was not in compression while the rest of the trench ruptured in the 1970s [2]. Since the 1970s events, the entire trench has ruptured or re-ruptured in large events, except the 1974 and 1975 doublets regions. Both tensional and a few compressional outer rise events have been associated with these large events [66].

The New Britain region, west of the Solomon Islands has experienced several compressional outer rise events. This region did not rupture in the 1970s, but may have ruptured earlier in a series of large doublets in 1919-1920 and 1945-1946 [33,44]. In both cases, an event in the New Britain trench (1920 and 1946) was triggered by a great event in the Solomon trench [33]. One of the compressional outer rise events occurred on 18 January 1973, which was less than 2° west of the 1971 underthrust aftershock area. Christensen and Ruff [2] believe the New Britain region was accumulating compressional stress, with the adjacent underthrust activity in the Solomon Islands contributing to the accumulated stress. This stress regime theory is complicated by the December 11, 1985 tensional outer rise event located near the January 18, 1973 compressional event. This entire region has nearly entirely ruptured in a suite of large events, including (from west to east): 1999 (Mw 7.0), 1987 (Mw 7.7), 2001 (Mw 7.0), 1985 (Mw 7.3), and 2000 (Mw 7.8); there have been a many compressional and tensional outer rise event associated with each [3,66].

2.2.3 South America (Chile)

The Chile region of the South American subduction zone is a prime example of a strongly coupled region. The plate convergence rate is estimated to be approximately equal to the seismic slip rate [73]. This subduction zone typically releases most of its stress in regular great earthquakes with a large amount of seismic slip [74]. Records suggest the entire shallow subduction zone between 32°S and 46°S ruptures in a series very large earthquakes roughly every century, with evidence going back as far as the sixteenth century [75]. The Chile region has experienced both compressional and tensional outer rise events [2].

A dominant feature of Chile is the great 22 May 1960 (Mw 9.5) underthrust earthquake. This event ruptured the trench from ~37° to 45°S [75]. At least 30 tensional outer rise events have followed the 1960 great earthquake [2,3], five of
which occurred within the first 10 years following the great interplate event [76]. The continued occurrence of tensional outer rise events adjacent to the 1960 underthrust and lack of compressional events, suggests a compressional regime still has not yet been restored.

Multiple outer rise events have been observed in the region adjacent to the segment ruptured by the 19 August 1906 underthrust event ($M_w$ 8.2). The first occurred on 25 September 1971. This tensional event was adjacent to and two months after the large underthrust event occurring on 9 July 1971 ($M_s$ 7.5). This 1971 event ruptured the northern portion of the 1906 region. A large compressional outer rise event later ruptured the same region on 16 October 1981 ($M_s$ 7.5). This shallow event occurred near the trench and just south of the 1971 underthrust aftershock zone. The following year a second smaller compressional outer rise event ruptured in the same region [2].

At the time of rupture, the 1981 earthquake was notable given the seismic condition of the adjacent trench. To the north, the recent 1971 event had a recurrence interval of 25-35 years, and to the south an expected 1906-type event ($M_s$ 8.4) was still 25 years premature [77], suggesting the trench was not in a compressional stress regime. Despite this, the trench subsequently re-ruptured on 3 March 1985 ($M_w$ 7.8). This sequence included the mainshock ($M_s$ 7.8) on 3 March, followed approximately an hour later by an $M_s$ 7.0 aftershock. A little over a month later, this region experienced another large aftershock on 9 April ($M_s$ 7.5). While the juxtaposition of compressional and tensional outer rise events could be problematic, Christensen and Ruff [2] infer the 1971 tensional outer rise event resulted from the adjacent 1971 underthrust event. This subsequently loaded the neighboring locked portion to the south with additional compressional stress, leading to the compressional outer rise events. This series of events ultimately led to the 1985 underthrust event. The region has since experienced several tensional outer rise events, one as large as $M_w$ 6.7 on 9 April 2001 [3, 66].

The Chile region has experienced two more compressional outer rise events associated with a seismic gap following the 11 November 1922 ($M_w$ 8.5) underthrust event. The first of these outer rise events occurred on 18 August 1964 ($M_s$ 6.1). It was located adjacent to the northern portion of the 1922 rupture zone and has since been followed by a moderate underthrust event on 4 October 1983 ($M_s$ 7.4). The second compressional outer rise event ruptured near the center of the 1922
aftershock zone on 13 November 1969 (M<sub>s</sub> 6.0) [2]. Since the 1969 event, there has been some continued moderate compressional outer rise activity between the 1964 and 1983 events, however, a large underthrust event has yet to re-rupture the adjacent trench. Interestingly, a moderate (M<sub>W</sub> 6.2) tensional outer rise event ruptured on 16 July 2006 adjacent to the southern extent of the 1922 aftershock zone. This location is also immediately north of the aftershock zone of the 1943 great interplate event (M<sub>W</sub> 8.2, [73]). This portion of the trench has experienced multiple moderate compressional outer rise events over the past 30 years [3].

2.2.4 South America (Ecuador, Peru, and Colombia)

The seismic character of the Ecuador-Peru subduction zone displays two unique segments. The divide between these segments is the Mendana Fracture Zone, which intersect the trench around 10°S at the site of the 1970 rupture zone. The southern segment includes large underthrust earthquakes [2]. This section (∼8° to 17°S) has ruptured over the past century by large shallow earthquakes [75]. From north to south within this zone, the most recent events occurred in 1966 (M<sub>W</sub> 8.1), 1940 (M<sub>s</sub> 8.4), 1974 (M<sub>W</sub> 8.1) [74], 2007 (M<sub>W</sub> 8.0), 1996 (M<sub>W</sub> 7.7), and 2001 (M<sub>W</sub> 8.4). The 2001 event was locally preceded by multiple compressional outer rise events in the prior 20 years. One tensional event also ruptured farther oceanward than the other compressional events in 1988 (M<sub>W</sub> 5.9) complicating the stress regime. The 1996 event was also preceded and followed by many, shallow compressional events, however, it is difficult to delineate which occurred in the outer rise or along the megathrust.

Contrary to the large underthrusts in the southern segment, the northern segment has been referred to as the “Peru Quiet Zone”, due to its lack of any known large underthrust event [2,66]. This segment stretches from about 0° to 10°S and coincides with the intersection of the Carnegie, Grijalva, and Sarmiento Ridges with the trench. The Nazca plate is believed to subduct at a shallow angle and extend near horizontal, producing a wide plate interface. It remains unclear if the observed quiescence indicates this broad contact is capable of accumulating centuries of strain or is able to subduct aseismically [74]. Despite the lack of underthrust, this region has experienced several outer rise events. There have been at least nine compressional outer rise events since 1967, suggesting compressive stress has been
accumulating in the outer rise. Conflicting this interpretation is the occurrence of two tensional outer rise earthquakes preceding the compressional events in the same region and one that followed [2,3].

Some infer the coexistence of compression and tensional outer rise events suggests the region may be experiencing a transition from a tensional outer rise regime to compressional. Several of the compressional events followed a large underthrust event, which ruptured just south of the Peru Quiet Zone on 17 October 1966 (MW 8.1). It is possible this underthrusting event caused loading in the region to the north. The 3 September 1967 compressional outer rise event was located immediately north of the 1966 rupture zone, suggesting strong correlation to the stress accumulation caused by the 1966 event. A tensional intraplate event occurring on the down-dip edge of the subducting plate on 31 May 1970 (MS 7.6, ISC depth=48km) immediately north of the 1966 interplate event and trenchward of the 1967 compressional outer rise event has also been suggested as evidence of a strongly coupled plate interface [2]. Three additional compressional events have ruptured in the outer rise in 1991 (MW 6.2), 1996 (MW 7.5), and 2005 (MW 5.5), further strengthening evidence of a local compressional stress regime [3,66]. Christensen and Ruff [2] explain despite local stress conditions, given the length of the Peru Quiet Zone, over 1,000 km, correlating all the compressional outer rise events to the 1966 event is a bold assumption. It is unlikely the entire region transitioned to a compressional stress regime due to this single underthrusting event.

In the north, the Colombia-Ecuador subduction zone has a short seismic history. The record is punctuated by the great, shallow 31 January 1906 (MS 8.7) earthquake, thought to have ruptured over 500 km of the trench and produced a large tsunami [74]. Kanamori and McNally [78] recognize the possibility this event was a large normal faulting earthquake, similar to the 1933 Sanriku earthquake. However, the authors also note the impulsive P-wave arrival of the Sanriku event was contrary to the slow buildup in the 1906 event, suggesting it was likely a thrusting mechanism. Much of this fault zone has since re-ruptured in a series of three events in 1942 (MS 7.9), 1958 (MS 7.8), and 1979 (MW 8.1), from south to north. It is believed these three later events have ruptured most of the 1906 fault zone. The northeastern extent may remain unruptured [74]. One tensional outer rise event on 2 January 1981 followed the 12 December 1979 (MW 8.2) underthrust
event [2]. The 1979 earthquake has since been followed by a tight clustering of compressional outer rise events between 2° to 3.5°N. The northern boundary of this clustering marks an apparent break to the seismicity, leading to a quiescent zone where the trench changes strike. The southern extent of this quiescent zone has experienced several tensional outer rise events in 1987 (M_w 5.7) and 1994 (M_w 5.8), and one larger tensional event (M_w 6.7) near the coast [66].

2.2.5 Middle America

The Middle America region, including Mexico and part of Central America, experiences many intermediate size underthrust events. The rupture length and recurrence intervals are typically short in this region [79]. Christensen and Ruff [2] note this short length and recurrence interval cause correlation of outer rise events with specific underthrust earthquakes to be difficult. However, tensional outer rise events have been observed to follow underthrust ruptures in this region. Some of these outer rise earthquakes appear to require multiple adjacent underthrusting events before failing. This may suggest that given the character of the underthrust events in this region, multiple ruptures are required to transfer enough tensional stress to the outer rise to produce faulting [2]. Multibeam bathymetry along this region has found evidence of abundant extensional outer rise faulting. The regions with the most faulting and largest offsets tend to correlate with 5 to 7 km thick crust and pre-existing fabric striking a low angle to the trench [56].

Several outer rise events display very clear correlation to major subduction zone events. Christensen and Ruff [2] correlate 1970, 1972, and 1988 tensional outer rise to specific underthrusting earthquakes. The delay after the interplate events varied from 2 to 11 years. Additionally, the 1970 outer rise events were adjacent to not only a 1968 underthrust, but also interplate events in 1960 and 1957. Other studies have observed outer rise clusters that appear to correlate to the 1990 Nicoya Gulf (M_w 7.0) and 1999 Quepos, Costa Rica (M_w 6.9) underthrusting events [80,81]. Middle America has also experienced one compressional outer rise event on 20 August 1971. This event occurred adjacent to both the 1942 and 1950 underthrusts [2].
2.2.6 Aleutian-Alaska Arc

The Aleutian-Alaska Arc region is a strongly coupled subduction zone, exemplified by the occurrence of three of the largest earthquakes of this past century. Additionally, this region has experienced many tensional outer rise events, most (21 events) of which followed one of the three great underthrusting earthquakes. The first great earthquake on 9 March 1957 ($M_w$ 9.1 [2]; $M_w$ 8.6 [66]) in the Aleutians was followed by 14 outer rise events. Five of these were located on the far eastern region of the aftershock zone, suggesting that they may have been associated with an earlier 1946 underthrust event. The second great earthquake occurred on 28 March 1964 ($M_w$ 9.2). This Alaskan event was followed by only one tensional outer rise event. Finally, the most recent great event on 4 February 1965 ($M_w$ 8.7) at the Rat Islands was followed by six outer rise events [2]. The 1965 great earthquake was also followed on 30 March 1965 by a $M_w$ 7.7 outer rise event [7]. Several outer rise events ruptured in the far western end of the arc near Kamchatka peninsula. One of which occurred only four days after the 1965 Rat Islands event. The close proximity of these events to the Aleutian and Kuril Islands intersection suggests a relationship to more complex tectonics than the rest of the Aleutian-Alaska Arc region. Another outer rise event occurred in 1929 after a series of underthrusting events in the early 1900s [2].

Only one compressional outer rise event has been observed in the Aleutian-Alaska region. The event occurred on July 25, 1990 ($M_w$ 5.3) oceanward of the great 1938 Alaska Peninsula earthquake. This, in conjunction with observed accelerating moment release, has been argued to suggest the Shumagin Islands/Alaska Peninsula region is reaching the end of a seismic cycle, leading to a large or great thrust event [82]. However, a nearby 1992 extensional outer rise event complicates this theory [83]. Considering the size of the underthrust events in the Aleutian-Alaska region, some expect that compressional events have occurred but have not been recorded due to poor station coverage prior to 1962 [2].

2.2.7 Kuril Islands and Kamchatka

The Kuril Islands and Kamchatka region provides an exemplary illustration of the relationship between subduction zone stress state and outer rise activity. Prior to 2006, a 500 km long seismic gap centered in the middle of the Kuril Islands and
Kamchatka region had long been identified [2, 74]. While uncertain, a suspected shallow 1915 underthrust event ($M_s \sim 8$), rupturing 100 km of the trench, was believed to have been the last large event in this gap [7, 44, 74]. Zones of major underthrusting events surrounded this central segment. To the northeast, Kamchatka experienced a great event on 4 January 1952 ($M_w 9.0$), and to the southwest in the Kuril Islands was another great event on 13 October 1963 ($M_w 8.5$). The 1952 event was followed by three tensional outer rise events over the following thirty years, and five tensional events followed the 1963 event over the same duration. Tensional outer rise earthquakes also followed two other underthrust events, the 1952 Hokkaido and 1959 Kamchatka [2].

The Kuril-Kamchatka region has also experienced localized compressional outer rise events since 1963 (Table 1). Christensen and Ruff [2] identified three of these events adjacent to the then Kuril Islands trench gap, the largest of which occurred on 16 March 1963 ($M_s 7.2$) (Figure 2.8). These compressional events were inferred to suggest high seismic potential. In September 2006, the trench of this seismic gap began to experience a swarm of underthrusting events. A great thrust earthquake soon followed this activity on 15 November 2006 ($M_w 8.3$). The aftershocks from this event extended 250 km along the trench. Within minutes of the mainshock, extensional faulting events in the outer rise began rupturing and continued for two months. This outer rise activity created a parallel band of aftershocks extending 200 km along the arc. On 13 January 2007, a second great event ($M_w 8.1$) ruptured in the outer rise. This extensional event was followed by less extensive aftershock activity than the 2006 event, however, roughly matched the limits of the earlier great event [7].

Additional compressional events have occurred north the 2006-2007 doublet (Table 2.1). These may indicate the southern portion of the 1952 Kamchatka rupture zone is beginning to transition from tensional into a compressional regime, having recovered from the 1952 underthrust [2].

Table 2.1: Central Kuril Islands Region Compressional Outer Rise Events Since 1963.

<table>
<thead>
<tr>
<th>Date</th>
<th>Latitude ($^\circ$N)</th>
<th>Longitude ($^\circ$E)</th>
<th>Depth (km)</th>
<th>Magnitude</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>16 Mar 1963</td>
<td>46.79</td>
<td>154.83</td>
<td>10-50</td>
<td>7.7 7.7  7.2</td>
<td>CR</td>
</tr>
</tbody>
</table>
2.2.8 Tonga-Kermadec Islands

The Tonga-Kermadec Islands region has experienced many outer rise earthquakes, including approximately similar numbers of compressional and extensional events. A majority of these events, both compressional and tensional, have occurred between 15°S to 18°S in the northern corner of the trench [2]. Govers and Wortel [64] theorize this corner marks the tip of a tear in the Pacific Plate. Due to the unique geometry of this region, the outer rise events cannot be associated to a simple subduction process. The third largest known outer rise earthquake occurred in this region on 29 September 2009 (Mw 8.1) near Samoa (Figure 2.9). The only larger outer rise events were the 1933 Sanriku (Mw 8.4), the 1977 Sumba earthquake (Mw 8.3), and the 2007 Kuril Islands earthquakes (Mw 8.1). Inversions of this event find a large non-double couple component, underscoring the complex regional tectonic geometry [3]. The 2009 event also produced a tsunami as large as 314 cm in American Samoa [84]. On 30 August 2009, a moderate (Mw 6.6) extensional outer rise event preceded the 2009 great earthquake. Significant seismic activity followed the great event, including back arc faulting and underthrusting events in the trench. Some extensional faulting also occurred in the outer rise.

Of the compressional outer rise events not occurring in the northern corner, several were followed by underthrusting event in the adjacent subduction zone. One of these compressional events occurred on 2 July 1974 (Ms 7.2) and was followed by a large underthrust doublet on 14 January 1976 (Ms 7.7, 8.0). Another compressional event ruptured on 11 October 1975 (Ms 7.8) and was followed on 19 December 1982 by an underthrust event (Ms 7.7) [2]. Not all compressional events
are associated with an underthrusting event. Three of these events have occurred in southern Kermadec (~33.5° to 35°S). These events occurring in 1976 (Mw 6.2), 1985 (Mw 6.9), and 1989 (Mw 6.1) all have depths between ~55 to 63 km, making them deeper than many outer rise events. The remaining compressional events have occurred in central Tonga (~18° to 21°S). This region lies between the 19 March 2009 (Mw 7.6) rupture zone and northern corner region. The lack of adjacent underthrusting events following these compressional events in southern Kermadec and central Tonga leads some to believe that compression is accumulating, possibly leading to a large underthrust rupture [2]. Christensen and Ruff [2] observed the tensional outer rise events in the Tonga-Kermadec region appear to behave most similarly to those of an uncoupled subduction process. While tensional events are located adjacent to regions of the arc that have experienced large underthrusting events, this relationship is not always the case, as some outer rise earthquakes occur in segments lacking large trench events.

2.2.9 Vanuatu

Much of the outer rise activity in the Vanuatu region occurs in the northern or southern extents of the subduction zone, with little in the central region; however, outer rise activity is not exclusive to these segments (Figure 2.3). While most of the outer rise events are tensional, particularly in the south, Liu and McNally [85] find the depths of the tensional and compressional outer rise events overlap and there does not exist a clear depth division between these. A depth division is typically expected in a bending model about a nodal surface within the subducting lithosphere. Above the hypothetical nodal surface the plate is in tension and below it is in compression [55]. Not all of the outer rise activity appears to be correlated with interplate earthquakes, but a significant portion has followed a large underthrust by ten years or less. It is believed the variable correlation with underthrust events suggests the Vanuatu region includes both coupled and uncoupled segments of the subduction zone [2].

Seven tensional outer rise events occurred in a sequence over November and December of 1985. These events were located north of the intersection of the d’Entrecasteaux Fracture Zone with the subduction trench. Two of these events were large (Mw 7.0, 7.1), occurred one hour apart, and were followed by aftershocks.
These events occurred oceanward of a prior moderate interplate underthrust in 1970 and were shortly followed by another moderate interplate underthrust on 21 December 1985. Both underthrust events occurred in the same region [2]. Since this 1985 sequence, this region has continued to experience outer rise activity, most of which has been compressional. This includes moderately-large events in 1990 and 1992 (M$_w$ 6.7, 6.8, respectively) [3].

Christensen and Ruff [2] interpret the collision of the East Rennell Island Ridge and d’Entrecasteaux with the Vanuatu trench to provide a subduction barrier; this is a similar structural setting as the 1977 Sumba great event. Both of the 1985 events occurred directly north of this collision. It is believed the tensional outer rise events were related to subduction being inhibited by the ridge. These outer rise events assist in mitigating the effect of this barrier, enabling some subduction to occur and compressional loading down-dip [2].

One of the region’s few compressional outer rise event occurred in the north on 21 October 1985 in a gap between the 1970/1985 and 1966/1980 events. One interpretation postulates this gap was under compressional stress, being loaded by the 1966 and 1980 events to the north and 1970, 1973, and 1985 events to the south [2]. Wyss et al. [86] suggest the gap had been under compressional loading for some time, leading to a possible future large underthrust. The same gap also experienced a tensional outer rise event in 1964, preceding the compressional outer rise event and the surrounding underthrusting events. The occurrence of a compressional event following a tensional outer rise earthquake indicates the local stress state had changed. A similar example has been seen in the Kuril Islands, where the outer rise stress regime changed from tensional to compressional behind a locked section after becoming loaded by surrounding underthrusting events [2].

On 07 October 2009, the northern Vanuatu gap ruptured in a series of three large underthrusts (M$_w$ 7.7, 7.8, 7.4 [66]). North of the 2009 sequence, a similar region as the 1966 sequence ruptured on 06 February 2013 (M$_w$ 8.0 [66]) (chapter 3). Following the 2013 mainshock, abundant normal faulting occurred in the outer rise, including an M$_w$ 7.1 normal faulting event $\sim$10 minutes after the mainshock. The centroid of this large outer rise event was directly offshore off the centroid of the mainshock along the megathrust.

The remaining compressional outer rise activity occurred near the northern extent of the southern corner of the subduction zone, including a 17 February 1981
(M$_s$ 6.7) event [85]. Tensional outer rise events are observed to the north and south of this compressional event. An abundance of tensional outer rise activity has occurred at the southern extent of the trench, south of 22$^\circ$S [3]. Similar to the northern corner of the Tonga trench, this region is characterized by a complex tectonic setting, with the corner believed to represent the apex of a tear in the Australian plate [64]. This region’s tensional outer rise activity includes several large events occurring in 2003 (M$_w$ 6.8), 2004 (M$_w$ 7.1), and 1995 (M$_w$ 7.7) [3, 66] (Figure 2.7). The magnitude of the 1995 event puts it in the ranks of one of the largest outer rise events recorded.

2.3 Notable Events

2.3.1 Samoa Islands: 29 September 2009 (M$_w$ 8.1)

The 29 September 2009 great Samoa earthquake (15.51$^\circ$S, 172.03$^\circ$W, M$_w$ 8.1) ruptured outboard of the northern region of the Tonga trench where the sharp change in strike of the trench forms a corner (Figure 2.9). Several meter high tsunami waves hit American Samoa 190 km away, with localized tsunami wave heights up to 15 m above sea level [87]. Lack of great underthrusting activity in this region leads Lay et al. [8] to infer this region has weak frictional coupling, similar to the 1933 Sanriku and 1977 Sumba earthquakes.

The faulting mechanics of this event have proven problematic. The global centroid moment tensor [3] solution, based on $\sim$200 s surface wave ground motions, found a mainly normal fault solution with a large non-double couple component [3]. Lay et al. [8] found faulting mechanisms using the W-phase, based on 200 to 1000 s ground motions. These solutions were primarily double-couple normal faulting. While the GCMT and W-phase solutions commonly show good agreement in this region, including aftershocks of the 2009 great event, the solutions for the 29 September 2009 event were significantly different, suggesting source complexities. Inversion of teleseismic P and SH waves found similar faulting solutions as the W-phase [8].

Lay et al. [8] performed inversion modeling of the slip distribution over the fault plane and found from the nucleation point the rupture largely propagated bilaterally across the fault plane with orientation $\phi=144^\circ$, $\delta=65^\circ$, $\lambda=-86^\circ$. As observed in the
waveforms, the event began with low amplitude seismic wave radiation over the first $\sim 12$ s before the main pulses of energy resulting from rupture of two patches of high slip on either side of the hypocenter. With a rupture velocity of 1.5 km/s and rupture plane approximately 150 km long and 45 km wide, the average slip was $\sim 10$ m and the peak slip was over 30 m. The rupture is believed to have propagated into the upper mantle. The calculated depth of the rupture is corroborated by a $\sim 24$ km deep 19 October 2009 ($M_w$ 6.0) aftershock occurring on the mainshock fault. While the rupture plane was $\sim 45$ km wide along the dip, the hypocenter was in the upper 15 km and most of the slip occurred at or above this depth [8].

After speculation of a secondary underthrust event “hidden” within seismic signal of the 29 September great outer rise event, Lay et al. [8] identified two large, shallow underthrusting events ($M_w$ 7.8, 7.8) triggered by the great outer rise event. These two events are believed to have ruptured a shallow portion of the megathrust $\sim 50$ km southwest of the great outer rise event, occurring $\sim 49$ to 89 s and $\sim 90$ to 130 s after the start of the great event; the first of the thrust events likely overlapped a portion of the rupture duration of the normal outer rise mainshock. In addition to the large underthrusting events triggered by the tensional event, much of the aftershock activity was thrust faulting earthquakes occurring within the Tonga block and/or along the plate interface. It is unique for subduction zone thrust faulting to be triggered by intraplate normal faulting [8].

### 2.3.2 Kuril Islands: 13 January 2007 ($M_w$ 8.1)

The 13 January 2007 great outer rise event east of the Kuril Islands is best understood by following the events leading up to it (Figure 2.8). A swarm of thrusting foreshocks began rupturing in the subduction zone east of the Kuril Islands in late September 2006. About six weeks later, a great thrust event ruptured in the subduction zone on 15 November 2006 at 46.592°N 153.266°E ($M_w$ 8.3). Thrusting aftershocks in the plate interface expanded the aftershocks zone up to 250 km along the trench. Additionally, a pattern of aftershocks developed in the outer rise, paralleling the trench. These outer rise aftershocks began minutes after the mainshock and continued for two months, developing a zone expanding a length of 200 km. The faulting mechanisms of the larger outer rise events were normal.

Two months after the November great underthrusting event, a second great
earthquake ruptured. This time the earthquake ruptured in the outer rise parallel to the 15 November fault zone, with an epicenter at 46.243°N 154.524°E. The 13 January 2007 tensional great earthquake (MW 8.1) was followed by aftershock activity spanning several hundred kilometers in the outer rise and along the trench. Only moderate tsunamis were produced from both of these great events; the November event produced a peak height of 1.76 m and the January only 0.37 m. A similar region of the outer rise had previously ruptured on 16 March 1963 (Ms 7.2), in a large thrusting event [7].

Inversion of waveforms found the November event to have thrust faulting mechanisms of φ=266°, δ=15°, and λ=92° [3] and a hypocenter depth of 11 km [7]. While the faulting plane of the November event is parallel to the trench, solutions to the January event are slightly oblique. Studies found the normal faulting mechanisms of the January event to be φ=43°, δ=59°, and λ=-115° or φ=266°, δ=39°, and λ=-54° [3]. Inversion methods loosely resolved a hypocenter depth of 22 km for the January event, but similar depths also worked in the model [7].

The inversion model by Ammon et al. [7] for the 15 November 2006 event found most of the slip expanded to the northeast of the hypocenter. Peak slip of over 5 m occurred in two patches northeast of the hypocenter; the average slip was 4.6 m over the entire fault. The total rupture lasted 120 s. For a rupture velocity of 2 km/s, the rupture plane extended 140 km down-dip and 320 km along the trench. The epicenter was heavily offset to the southern end of the rupture plane [7].

In contrast to the underthrusting November event, the 13 January 2007 event is believed to have ruptured a smaller area but displayed higher slip values. Using a hypocenter depth of 22 km and rupture velocity of 3.5 km/s, Ammon et al. [7] resolved for a rupture plane of about 225 km long and 45 km wide, but most of the slip was included in a 120km long by 20 km down-dip section. Over this smaller region, the average slip was 9.6 m. Most of the energy was released in a main pulse lasting 45 to 50 s. From the nucleation of the rupture, there was relatively low energy release over the first 5 s. Beginning at 5 s and peaking around 10 s, two asperities, bilateral from the hypocenter, began to rupture. The slip expanded shallowly, with peak slips of each rupture patch very near the top of the rupture plane and the surface of the crust. Very little slip occurred below the hypocenter [7].

Between the January 2007 tensional outer rise event and the November 2006 underthrusting event, finite fault inversion show the January event ruptured a
smaller fault plane with almost all of the slip concentrated very shallowly compared to the November event that displayed very broad slip over a wide plane. Comparison of the source time functions display the duration of the January event was much shorter with less complexity than the underthrusting event. In addition to the differences in the slip and time functions, Ammon et al. [7] noted the November 2006 and January 2007 events differed in frequency content. While the January event was smaller than the November earthquake, the outer rise event possessed greater high-frequency amplitudes than the trench event, relating to higher energy release and subsequently greater energy to seismic moment ratio [7].

2.3.3 Southern Vanuatu: 16 May 1995 ($M_w$ 7.7)

The southernmost extent of the Vanuatu subduction zone is a unique tectonic setting (Figure 2.7). Like northern Tonga, the trench makes a sharp turn in this region, forming a corner, where the subduction zone transitions into a strike-slip fault. Govers and Wortel [64] interpret this complex tectonic setting to be the apex of a tear in the Australian plate, forming a STEP fault. The uniqueness of this region is displayed in the abundance of tensional outer rise activity occurring near this southern corner.

The southern Vanuatu region has experienced multiple large tensional outer rise events, the largest of which occurred on 16 May 1995 ($M_w$ 7.7, 23.01°S 169.90°E) [66]. A small tsunami resulted from this event with a maximum water height of 0.50 m. Water run-up was as high as 0.20 m at Port Vila, Vanuatu 606 km away, and experienced as far away as 3,127 km in Rarotonga, Cook Islands with a run-up of only 0.02 m [88]. The GCMT solution of the focal mechanisms aligns the fault plane at $\phi$=280° (110°), $\delta$=35° (56°), and $\lambda$= -99° (-84°), with rupture nucleating at 24.7 km depth. NEIC assigns a slightly shallower hypocenter of 20 km. Based on the W-phase, Hayes (Gavin Hayes, personal communication, 2010) calculates a slightly more oblique focal mechanism of $\phi$=258.9° (132.8°), $\delta$=48.9° (55.9°), and $\lambda$= -132.0° (-52.5°) using the GCMT hypocenter. Using results of W-phase centroid time shift and location searches, Hayes calculated a solution of $\phi$=255.3° (138.1°), $\delta$=51.3° (60.4°), and $\lambda$= -140.7° (-46.0°). The aftershock pattern indicates most of the slip occurred to the south of the hypocenter, with only a slight amount to the north.
Our initial finite fault inversion analysis of this event is conducted using a search-based algorithm (Appendix A.1.1) described in Ammon et al. [7] (Figure 2.10). In this model, we assume a rupture speed of 2.7 km/s, which provides a superior fit to the data over other speeds. With this rupture speed, the slip is well contained within a rupture plane of roughly 85 km long and 40 km wide down-dip, however, most of the slip occurs in the southeastern 55 km of the plane. The best match to the waveforms occurs with the rupture plane oriented at $\phi=115^\circ$, $\delta=35^\circ$, $\lambda=-90^\circ$, and the hypocenter at only 12.6 km depth. This fault plane is similar to the GCMT solution but with a shallower hypocenter depth.

The hypocenter is roughly centered on the broader rupture plane, but offset to the NW in the portion of the plane hosting the higher concentration of slip. While slip extends 40 km down-dip, most of the high slip occurs at similar or shallower depths than the hypocenter. Like the Kuril and Samoa events, the southern Vanuatu outer rise earthquake rupture displayed a slight delay between nucleation and the first pulse of high energy. This is seen in the simple moment-rate function, where the model displays a jump in energy release until around 6 s after the first P-wave arrival, relating to the rupture of the primary asperity. While the aftershock pattern suggests a unilateral rupture to the SE, the finite fault model indicates the rupture had bilateral components; but the slip distribution is asymmetric, favoring the SE. This event includes the rupture of one primary region of high slip to the SE and up-dip of the hypocenter. The peak slip of nearly 13 m occurred very close to the top of the rupture plane and the surface. Stress drop, $\Delta \sigma$, is calculated for a dip slip rectangular fault plate by the relationship,

$$ \Delta \sigma = \left[ \frac{4(\lambda + \mu)}{\pi(\lambda + 2\mu)} \right] \times \left[ \frac{M_o}{w^2L} \right], \quad (2.1) $$

where $M_o$ is the moment, and $w$ and $L$ are the fault dimensions width and length, respectively. If the Lame constant, $\lambda$, and shear modulus, $\mu$, are assumed equal, this equation simplifies to approximately,

$$ \Delta \sigma = \left[ \frac{8}{3\pi} \right] \times \left[ \frac{M_o}{w^2L} \right]. \quad (2.2) $$

Focusing on the region of well-resolved slip (upper 20 km and southeast 55 km...
region), this yields an average stress drop estimate:

\[
\Delta \sigma = \left[ \frac{8}{3\pi} \right] \times \frac{3.9e27 \text{ dyn-cm}}{4.0e12 \text{ cm}^2 \times 5.5e6 \text{ cm}} = 1.5e8 \frac{\text{dyn}}{\text{cm}^2} = 15 \text{ MPa}. \tag{2.3}
\]

While our inversion results provided a strong fit to the first pulse of energy, the second energy pulse is poorly fit (Figure 2.11). To provide a better fit to the waveforms, we used the inversion algorithm of Kikuchi and Kanamori [89] (Figure 2.12) (Appendix A.1.2). Unlike the search-based algorithm, which defines the subfault source time function by a single half-cosine wavelet at each node, the time function in the Kikuchi-Kanamori algorithm is defined by multiple adjacent “boxcar” functions. Results from this model provide significantly improved fit to the waveforms to over 60 s after the first motion (Figure 2.13). Interestingly, the “preferred” rupture parameters of the search-based algorithm do not provide the lowest variance in the Kikuchi-Kanamori algorithm. This model calculates a preferred rupture speed of 2.0 km/s and \( \phi=116^\circ, \delta=48^\circ, \) and \( \lambda=-68^\circ \). The depth of 20 km is also slightly deeper than the search-based algorithm and closer to the catalogued depth by NEIC and GCMT. The difference may be a result of the Green's function.

The slip pattern of the Kikuchi-Kanamori algorithm shows similarity to that of the search-based algorithm (Figures 10 and 12). Both show peak slip up-dip of the hypocenter. The Kikuchi-Kanamori algorithm finds more slip occurring at the hypocenter; the search-based algorithm had very little slip occurring in this region. A large difference between the two models the amount of slip in the northwest region of the fault plane. The search-based algorithm is largely unilateral with no appreciable slip to the northwest. The slip pattern of the Kikuchi-Kanamaori model is bilateral, with significant slip to the northwest and southeast. This pattern is intriguing considering the aftershock pattern is largely unilateral to the southeast. The bilateral pattern of this event does not indicate a significant asymmetry in the amount of slip to explain the aftershock pattern. The disconnect between the mainshock slip and aftershock patterns may be rooted in the complex regional stress regime or distribution of preexisting fault planes. The moment rate functions of the two models show considerable difference. The search-based algorithm was unable to fit the second energy arrival, while the Kikuchi-Kanamori algorithm does. However, the Kikuchi-Kanamori algorithm has more variables, allowing this better
fit. This is seen in both the moment rate functions and waveforms (Figure 2.11 and 2.13). The peak slip of 10.6 m in the Kikuchi-Kanamori algorithm is slightly lower than the 12.8 m in the other model. Additionally, the magnitude of $M_w$ 7.84 ($7.17 \times 10^{20}$ Nm) is higher than that calculated in the search-based algorithm or catalogued by GCMT and NEIC; this difference in moment is typical for these models.

Unlike the Samoa and Kuril events, the 1995 Vanuatu event does not appear to exhibit enrichment in short-period energy relative to local thrust events (Figure 2.14). Both the 1995 event 1-second $m_b$ and 20-second $M_s$ are either similar or smaller amplitudes than comparable local thrust events. Additionally, there was no associated large underthrusting activity linked with the Vanuatu event as there was with the Kuril event.

As mentioned, the tectonics of the southern extents of the Vanuatu subduction zone are inferred to be modulated by a tearing of the subducting Australian plate, developing a ENE striking STEP fault [64]. This region is also just south of a collision between the Loyalty Ridge of the Australian plate with the Vanuatu arc at 22°S. Calmant et al. [10] note a change in the Australia/Vanuatu relative motion about this collision zone. Some studies have suggested the change in relative motion is due to this collision [11] (Figure 2.15). Southern Vanuatu’s apparent lack of correlation between outer rise earthquake activity and seismicity along the megathrust, including a lack of compressional outer rise events, might suggest the subduction zone is weakly coupled. However, unlike a decoupled tectonic setting similar to the great 1977 Sumba event, southern Vanuatu experiences both an abundance of tensional outer rise events and large underthrusting earthquakes.

Not surprisingly, southern Vanuatu seismicity is most similar to northern Tonga, given they both posses a sharp turn in the strike of their trench, which transitions from a subduction plate interface to a transform fault linking the two subduction zones. Both of these regions display abundant tensional outer rise activity, with few compressional outer rise events, while also experiencing abundant underthrusting events along a wide Beniof zone. Govers and Wortel [64] model of lithospheric tearing shows strain rates maximize and rotate in orientation near a STEP. According to this study, roughly a 100 km radius from the STEP would be expected to experience complicated surface deformation.


2.3.4 Sanriku, Japan: 2 March 1933 (M<sub>s</sub> 8.4)

The great Sanriku event was a shallow normal fault rupture occurring on 2 March 1933 (M<sub>s</sub> 8.4) with a hypocenter at 39.2°N 144.5°E and 10 km deep (Figure 2.5). Kanamori [6] believe the earthquake ruptured the entire lithosphere, likely along a healed old fault plane; this study estimated the fault plane was approximately 185 km long by 100 km wide, dipping 45° with a north-south strike, and calculated an average slip of 3.3 m for this event. The fault dimensions and slip correspond well with the aftershock pattern and tsunami data [6]. However, contemporary studies of the 2007 Kuril [7] (section 2.3.2) and 2009 Samoa [8] (section 2.3.1) outer rise earthquakes exemplify how a large rupture area is not necessary to produce the observed large moment. These more recent events are believed to have ruptured regions less than 50 km wide (225 and 150 km long, respectively), however, the average slip is these events (∼10 m) [7,8] is significantly greater than that estimated by Kanamori for the 1933 earthquake.

A large tsunami was produced from this event with peak wave heights of 28.7 m recorded at Ryori Bay, Honshu. Waves as large as 2.9 m were recorded at Napoopoo, Hawaii [84]. North and south of the trench adjacent to the 1933 event later failed in the 1968 Tokachi-Oki (M<sub>s</sub> 8.1) [72] and 2011 Tahoku-Oki (M<sub>w</sub> 9.0) [90] underthrusts, respectively. The seismograms from the great 1933 event show a very impulsive first onset of the event.

Given the size of the event, Kanamori [6] concluded the bending forces alone likely did not produce the event. The author inferred the 1933 event resulted from slab pull. The pull from the sinking slab may result in decoupling along the trench. Weak coupling along the trench is supported by observed lack of seismic activity along the adjacent trench [6]. Kawakatsu and Seno [72] note the region of the trench where the 1933 event occurred appears to lack great thrust type events for the past 200 years. These authors infer the strong coupling in the adjacent fault regions could have resulted in normal component stress concentration prior to the 1933 event.

2.3.5 Sumba, Indonesia: 19 August 1977 (M<sub>w</sub> 8.3)

The 19 August 1977 Sumba earthquake (M<sub>w</sub> 8.3) occurred at 06:08:54.8 UT with an epicenter at 11.16°S 118.41°E [91] (Figure 2.6). This great interplate event
ruptured within the Indian Plate on the eastern extent of the Sunda trench, south of Sumba Island, near the collision of the Australian continental lithosphere with the Sunda-Banda arc [71]. The termination of the Java trench is only 150 km east of the epicenter and several tens of kilometers east of the eastern aftershock zone boundary [92]. While a majority of the one month aftershock activity was concentrated 65 to 115 km east of the epicenter [71], the activity spanned ∼130 km east and ∼110 km west of the epicenter [92].

The depth extents of the 1977 event are unclear. A large tsunami as high as 10 m on Sumbawa suggests significant ocean floor deformation and subsequently shallow fault displacements. The maximum depths of the rupture, however, are problematic. Some have calculated maximum depths of 20 to 25 km (i.e. [93], [94]). Other studies concluded much greater depths of 60 to 90 km depth (i.e. [95], [96]). Lynnes and Lay [92] calculate the event nucleated at 29 km depth and ruptured to a maximum depth of 30 to 50 km, but note the possibility of additional moment release radiating very low frequencies below this depth. As discussed in section 2.3.4, contemporary studies of the 2007 Kuril [7] (section 2.3.2) and 2009 Samoa [8] (section 2.3.1) outer rise earthquakes suggest rupture widths (depths) of these great outer rise events may be narrower and the average slip much greater than earlier studies estimate.

Like the Sanriku great outer rise event, the lack of great interplate thrust events in this region suggest decoupling at the plate interface [71]. This inference is supported by the occurrence of two tensional outer rise events preceding and to the west of the 1977 event in 1967 and 1972. In addition to the apparent aseismic nature, the Java subduction zone’s structure shows similarity to the Marianas and Izu-Bonin subduction zones; the slab has a shallow dip to depths of 50 to 100 km and then quickly steepens to nearly 80°, extending to depths of 650 km [92, 97]. Spence [71] finds the combination of the apparent decoupling of the subduction zone and resistance to subduction of the Australian continental lithosphere has resulted in concentration of slab pull forces up-dip. Subsequently, these slab pull forces produced the unbending stresses in the outer rise, resulted in the 1977 event [71]. Some have interpreted the inferred deeper rupture extents to suggest the earthquake resulted from plate detachment (i.e. [98], [95]). The larger aftershocks were generally all shallow normal faulting events (less than 24 to 31 km depth) [71, 99, 100].

Lynnes and Lay [92] inverted waveforms to study the rupture pattern of this
event. Their model found the event ruptured bilaterally across a 200 km long fault plane. The event displayed significant variability in coseismic displacement across the fault plane. This has been interpreted to suggest the stress drop varied spatially. Additionally, the region of peak displacement occurred 30 to 40 km away from the hypocenter. The foreshock of the event also occurred in a similar location as the point of nucleation of the mainshock. The waveforms display this spatial separation in a delay between the first P-wave arrival and the main pulse of energy, showing the rupture began with a very low level of energy release [92]. Work by Spence [71] showed the region of high displacement experienced a low level of aftershocks. Lynnes and Lay [92] infer the cluster of high aftershock activity to the east correlates with a portion of the fault plane that experienced less displacement during the mainshock. These observations lead Lynnes and Lay [92] to conclude this event resulted from the failure of a large asperity where the peak slip occurred and the rupture nucleated on the perimeter of this asperity.

2.4 Rupture Characteristics of Outer Rise Events

While outer rise earthquakes are rare compared to the frequency of underthrusting events, approximately 20 magnitude 5 to 7 occur annually worldwide [85]. Additionally, outer rise events do not occur adjacent to all subduction zones, but they do rupture along most to varying frequency of occurrence and magnitude. There exist many theories as to what characteristics of a subduction zones qualify it to hosts these events. Following Christensen and Ruff’s [2] outer rise model (Figure 2.2), the degree of coupling within the subduction zone heavily influences the characteristics of outer rise activity. Included in this model, compressional outer rise events preferentially occur adjacent to strongly coupled subduction zones while only tensional events are expected in weakly coupled regions. Plate bending models propose various constraints on lithosphere thickness and rigidity of the subducting plate required to produce outer rise earthquakes (i.e. [55], [57]). Ranero et al. [56] found strong correlation between crustal thickness and tectonic fabric orientation relative to the trench with the occurrence of outer rise events. The same study found the age of the subducting lithosphere only provides second order influences on the occurrence of outer rise earthquakes. Preferential reactivation of preexisting faults has been observed to occur up a 25° angle from the trench.
parallel before new faulting crosscut the tectonic fabric [101].

Convergence direction is a consideration when examining the occurrence of outer rise activity. Tensional outer rise faults typically strike parallel to the trench. This has been attributed to the fact that horizontal tensional stresses generally occur perpendicular to the bending axis of the plate, which aligns parallel to the trench [102]. In addition to being influenced by tectonic fabric [56,101,102], obliquity of plate convergence has also been observed to influence the orientation of outer rise faulting. Mortera-Gutierrez et al. [102] observed in the Aleutians where plate convergence is oblique to tectonic fabric the shallow events preferentially reactivate the preexisting faults, while the plate bending orientation dominated the strike of deeper events. Fromm et al. [103] found a similar example in the 9 April 2001 (Mw 6.7) tensional outer rise earthquake near the Juan Fernandez Ridge. The ~33° strike of this event aligns with preexisting fracture systems associated with bathymetric features. This orientation is subparallel to the trench, but oblique to the 80° azimuth of the Nazca plate motion relative to the South American plate [103].

The traditional outer rise model explains the behavior of outer rise activity is directly related to the degree of coupling within the subduction zone (Figure 2.2). Accordingly, a compressional outer rise event would be expected to occur oceanward of a locked portion of the trench. Conversely, following an underthrusting event, the outer rise would transition into a tensional stress regime resulting in normal faulting outer rise earthquakes. This type of model enables a degree of seismic hazard prediction using either outer rise activity as a gauge of the degree of coupling within the trench or underthrusting events as a precursor for tensional outer rise events. Examples of this are observed along Chile and Peru where compressional outer rise events have been observed adjacent to seismic gaps that later rupture. Similarly, the 2006/2007 great Kuril events exemplify how a tensional outer rise event may follow a large underthrusting earthquake. One caveat to this prospect of seismic prediction is the variable temporal relationship between outer rise earthquakes and underthrusting events. Again, as the 2006/2007 Kuril events displayed, time between the underthrusting and outer rise event can be very short. In other examples, this time between can be many years. This is further complicated in regions like Central America where the high frequency and relatively short fault lengths of underthrusting events make correlation of underthrusting and outer rise
events difficult. Of course, there are also outer rise events, like the great 1933 Sanriku, 1977 Sumba, or 1995 Southern Vanuatu, that are apparently not associated with underthrusting activity. As previously described, these arise from unique tectonic settings; the adjacent trenches of the 1933 and 1977 events have hosted only small underthrusting events, suggesting weak frictional-coupling [8].

Generally, tensional outer rise events are observed to rupture shallowly, while the compressional events from this region are deeper. This relation is believed to support the belief these events are associated with bending forces [76,104]. In addition to depth characteristics, slip models of various large to great outer rise events (i.e. 1977 Sumba, 1995 Southern Vanuatu, 2007 Kuril, and 2009 Samoa) display a common slip pattern. While the rupture areas of these earthquakes varied in dimension, generally the high slip region and a majority of the slip occurred at or above the hypocenter; in many instances this occurred near the upper limits of the fault plane.

Studies have long observed that intraplate earthquakes typically display significantly higher stress drops than interplate events [105–108]. This characteristic is generally consistent with large outer rise earthquakes [8]. Laboratory studies find static friction between two surfaces increases with stationary contact time [109]. It has consequently been suggested the longer recurrence interval between intraplate earthquakes relative to interplate events equates to a longer hold time, possibly explaining the larger stress drop [110]. Scholz et al. [106] has challenged this theory, suggesting the degree of increased contact surface cannot account for the magnitude of stress drop increase. Alternate explanations for the larger stress drop include a relation between fault maturity and amount of fault gouge [106] or a difference in the amount of stress localization between the two types of faults [111]. In addition to a difference in stress drop, many of the largest outer rise events are observed to have enrichment in short-period energy relative to comparable underthrusting earthquakes (Figure 2.16). As the 1995 Southern Vanuatu event displays, this characteristic is not ubiquitous to all outer rise events (Figure 2.14). Compared to local thrust events of similar size, the 1995 Southern Vanuatu event’s 1-second $m_b$ and 20-second $M_s$ was either smaller or similar magnitude.
2.5 Conclusion

Generally, outer rise earthquakes are studied in relation to underthrusting activity. The occurrence of a compressional outer rise event is suggestive of a forthcoming underthrust, and tensional outer rise events commonly follow underthrusting events. Of the four great ($M_w \geq 8$) shallow outer rise earthquakes, the 2007 Kuril events provide an excellent example of the relationship between faulting along the megathrust and outer rise activity. Interestingly, of the largest events, the 2007 Kuril earthquake is anomalous. Weak seismic coupling and a lack of associated large underthrusting earthquakes characterize all other great events (1933 Sanriku, 1977 Sumba, and 2009 Samoa).

The 1995 southern Vanuatu earthquake contributes to the catalog of recorded large outer rise events, being the fifth largest. The 1995 event’s unique tectonic setting and weak seismic coupling highlight several characteristics of most great outer rise earthquakes. The manner in which this event ruptured, with most of the peak slip regions located very shallow and above the hypocenter, is consistent with the other large events. Additionally, like all of the largest outer rise events, the Vanuatu event ruptured as a normal fault. Despite these similarities, the Vanuatu earthquake’s lack of short period energy enrichment relative to local thrust events contrasts an observed trend in many large intraplate earthquakes. The contrast indicates the condition producing this enrichment is not ubiquitous to all outer rise faulting.

The rarity of large and great outer rise earthquakes given the relative abundance of moderately sized outer rise events is likely related to the anomalous tectonic settings required to produce these great events. The disconnect these great events display from traditional outer rise faulting is likely inherent in the magnitude of the events. While some of the events discussed (i.e. Sanriku, Sumba) have been inferred to rupture the entire oceanic lithosphere and possibly upper mantle, other events display how the massive moment as these events is possible through observed high average slip values. Regardless of the tectonic settings, the fundamental relationship between the occurrence of outer rise faulting and the degree of coupling in the subduction zone remains. While the tectonic setting is anomalous in these examples of great outer rise earthquakes, these tensional events indicate a weak seismic coupling along the megathrust (except in the Kurils), enabling a greater
transmission of slab pull forces to the outer rise. This highlights an essential condition of all outer rise activity, which is the transmission of slab pull forces into the outer rise resulting in a bending moment. The efficiency in this transmission may play a critical role in deciding the magnitude which outer rise events are able to achieve.
Figure 2.1. The outer rise region is located seaward of the subduction zone. Earthquakes occurring in this region are intraplate, rupturing within the oceanic plate. This is opposed to underthrusting interplate earthquakes, which rupture at the interface between the overriding and subducting plate. The outer rise events are commonly interpreted to result from bending forces occurring within the oceanic plate, resulting from coupling within the trench (Figure 2.2). If the trench is strongly coupled, the outer rise is expected to be under a compressive stress regime, potentially resulting in compressive outer rise faulting. Following an underthrust event or if the subduction zone is generally poorly coupled, the outer rise may experience a tensional stress regime, resulting in normal faulting (adapted from Choy and Kirby [1]).
Figure 2.2. The occurrence and type of outer rise earthquakes is closely related to the degree of coupling within the subduction zone. In a coupled subduction zone, adjacent to a seismic gap, the outer rise experiences a compressive stress regime, potentially resulting in compressive outer rise events. Seaward of regions of the fault that have recently slipped would be transitioned into a tensional stress regime resulting from the slab pull. This is similarly true to uncoupled subduction zones, where the lack of coupling enables the stress of slab pull to be directly propagated to the outer rise. Both of these tensional stress regimes may result in outer rise normal faulting (adapted from Christensen and Ruff [2]).
Figure 2.3. The Vanuatu (formerly New Hebrides) subduction zone is characterized by an abundance of large earthquakes, but no great events. This map displays all seismicity $M_w > 5.0$ recorded from 1973 to 2010 in the NEIC record. Additionally, the focal mechanisms between 1976 to 2012 for events $M_w \geq 7.4$ are also displayed [3]. This includes the 16 May 1995 southern Vanuatu normal outer rise event ($M_w 7.7$). This trench is the interface of the Australian plate (west) subducting beneath the Pacific plate (east). The trench turns sharply in the north transitioning into the Solomon Islands. It also has a sharp corner in the south, where it becomes a diffuse strike-slip plate boundary before making another sharp turn to the south, becoming the Tonga subduction zone.
Figure 2.4. Seismicity profile along the strike of the Vanuatu subduction zone. Events are scaled relative to their magnitude. There exists an arching structure of clustered events between 14°-17°S. This arch begins and ends at depth around 200 km and shallows to ~150 km in the middle. Below this structure there is a dramatic lack of seismicity until a depth of about 600 km. Some have attributed this to slab detachment [4,5]. The northern and southern limits also lack seismicity >200 km.
Figure 2.5. The great 2 March 1933 Sanriku outer rise event (M, 8.4) was a shallow normal faulting earthquake occurring off the coast of Japan. This is the largest event recorded in a region that has experienced multiple other large events over the past 40 years. The 1933 event is also the largest outer rise event to have occurred in this region, the next largest in the modern record occurred on 14 November 2005 (Mw, 7.0). Besides these two events, this region has experienced a number of considerably smaller outer rise events. The shaded region adjacent to the 1933 epicenter denotes the aftershock region defined by Kanamori [6]. The subduction zone south of the 1933 Sanriku event ruptured in the 11 March 2011 Mw 9.0 earthquake.
Figure 2.6. The Java subduction zone experienced one of the world’s largest outer rise earthquakes on 19 August 1977 ($M_w$ 8.3). This shallow normal faulting event ruptured within the Indian Plate near the termination of the Java Trench at its collision with the Australian Plate. The two-week aftershock activity places the epicenter relatively central within the aftershock zone. Several notable earthquakes that soon followed the great 1977 event include another smaller outer rise event east of the great event on 10 April 1978 ($M_w$ 6.7) and a moderate-large strike-slip event north of the trench on 07 October 1977 ($M_w$ 6.5). This map displays all seismicity $M_w \geq 5.0$ recorded since 1973 in the NEIC record.
Figure 2.7. Southern Vanuatu experiences an abundance of outer rise activity. The largest of these events was the 16 May 1995 ($M_w$ 7.7) normal faulting earthquake. The fault plane and slip model (Figure 2.10) of this large event are displayed on the upper map. In addition to multiple large outer rise events, southern Vanuatu has experienced multiple large underthrusting events. Besides the two 2003 events, there appears to be poor correlation between underthrusting and outer rise activity. This may be explained by the complex tectonic setting. The aftershock pattern of the 1995 event suggests most of the slip occurred southeast of the epicenter.
Figure 2.8. The Kuril subduction zone provides a good example of a relationship between underthrusting events and outer rise activity. The two most notable events in this region were the 15 November 2006 ($M_w$ 8.4) great underthrusting event and 13 January 2007 ($M_w$ 8.1) great normal faulting outer rise event. The 2006 event ruptured an identified seismic gap, located north of a region that had last ruptured in 1963. The aftershock patterns of two 2006/2007 great events shows very clear parallel banding to one another, underscoring the relationship between the two events (adapted from [7]).
Northern Tonga represents a complex tectonic setting, where there is a sharp turn in the trench due to a “corner” in the overriding Australian Plate (Tonga Block). This region experiences abundant outer rise earthquakes, nearly all of which are tensional. Included in this outer rise activity was the great shallow tensional event on 29 September 2009 ($M_w$ 8.1). This event initiated abundant faulting along the megathrust. Displayed is the inferred fault plane of this event by Lay et al. [8]. In addition to the outer rise activity, there is also abundant seismicity along the megathrust and within the Tonga Block. Shallow events larger than $M_w$ 6.8 since 1973 are identified (the focal mechanisms is not available for the 1975 event).
The 16 May 1995 ($M_w$ 7.7) occurred at 23.01°S 169.90°E (Figure 2.7). The GCMT solution of the focal mechanisms aligns the fault plane at $\phi=280°$ ($110°$), $\delta=35°$ ($56°$), and $\lambda= -99°$ ($-84°$), with rupture nucleating at 24.7 km depth. NEIC assigns a slightly shallower hypocenter of 20 km. This figure shows the results using the search-based algorithm. This finite fault model of the event assumes a rupture speed of 2.7 km/s. We found the rupture plane was roughly 85 km long and 40 km wide down-dip, however, most of the slip occurred in the southeastern 55 km of the plane. The peak slip was 12.8 m, occurring about 20 km up-dip and southeast of the hypocenter. Our model finds a shallower hypocenter than GCMT and NEIC of 12.604 km, however, the rupture plane focal mechanisms are comparable to those of GCMT. We find the rupture to be oriented at $\phi=115°$, $\delta=35°$, and $\lambda= -90°$. The moment rate function is relatively simple, showing the event was not impulsive at its onset. Most of the energy release occurred between about 4 to 14 seconds of the ~17 second rupture duration.

$\phi = 115°$, $\delta = 35°$, $\lambda = -90°$, $V_R = 2.7$ km/s
Figure 2.11. The wave fits for the suite of teleseismic P- and SH-waves used in the finite fault inversion of the 16 May 1995 southern Vanuatu outer rise event using the search-based algorithm. The black lines are the observed waveforms; the red are the predicted. This model provides a strong fit to the first \(\sim 25\) seconds of the waveforms. After 25 seconds, the model has difficulty fitting the second energy arrival.
Figure 2.12. These are the model results using the Kikuchi-Kanamori algorithm of the 16 May 1995 ($M_w$ 7.7) outer rise event. Using this model we resolve a slightly slower preferred rupture speed of 2.0 km/s than with the search-based algorithm. The rupture plane was also longer with dimensions of $\sim$100 km long and 40 km wide down-dip. Where in the search-based algorithm we found most of the slip to occur in the southeast, the Kikuchi-Kanamori algorithm produces a largely bilateral slip pattern with large amounts of slip also in the northwest. Both models find significant slip along the up-dip limits of the fault plane. The preferred hypocenter depth of 20 km is close to the depth catalogued by GCMT and NEIC. The focal mechanism in this model, $\phi=116^\circ$, $\delta=48^\circ$, and $\lambda=-68^\circ$, are very similar to the search-based algorithm, but with some variation in the rake. This model is able to pick up the second pulse of energy. This is seen in both the waveforms (Figure 2.13) and the moment rate function.
Figure 2.13. The wave fits for the suite of teleseismic P- and SH-waves used in the finite fault inversion of the 16 May 1995 southern Vanuatu outer rise event using the Kikuchi-Kanamori algorithm. SH-waves were significantly down-weighted. The black lines are the observed waveforms; the red are the predicted. This model provides a strong fit to at least the first $\sim$50 seconds of the waveforms.
1-Second Energy of Southern Vanuatu Underthrusting and Tensional Outer Rise Events

Figure 2.14. Comparing the 16 May 1995 ($M_w$ 7.7) southern Vanuatu outer rise event to similar magnitude local thrust events displays that, unlike other large outer rise events, the Vanuatu event does not appear to be enriched in short-period energy. With the moments scaled to the 1995 event, a seismogram of the 1-second energy (upper) shows the outer rise event had less of this high frequency energy than the 12 February 1994 ($M_w$ 6.9) underthrusting event, and similar to that of the 27 December 2003 ($M_w$ 7.3) underthrust. The 1995 outer rise event displays similar or lower 1-second $m_b$ and 20-second $M_s$ energy as the two local underthrust event (lower).
Figure 2.15. The bathymetric map of Vanuatu shows multiple ridges are impinging the trench. Around 15°S, the d’Entrecasteaux ridge is being subducted. Its resistance to subduction is believed to influence not only the local seismicity but also the convergence of the two plates [9]. In the south is the collision between the Loyalty Ridge of the Australian plate with the Vanuatu arc at 22°S. GPS studies note a change in the Australia/Vanuatu relative motion about this collision zone [10], leading some to believe this change is due to the collision [11]. The arrows indicated the motion of the Pacific relative to Australian plate, with the length of the arrow scaled relative to the difference in convergence rate, decreasing from ∼12 cm/yr in the south to ∼4 cm/yr in the north [9] (adapted from [12]).
Figure 2.16. The intraplate events (red) display enrichment in high frequency energy (high energy/moment value) relative to interplate (blue) earthquakes. The three Kuril events provide a clear example of the local differences, where both the 2007 and 2009 intraplate events have much higher energy/moment values than that of the 2006 interplate event. This is despite the 2006 event having a higher moment magnitude. (Modified unpublished figure by Lay and Kanamori)
Chapter 3  
Large Earthquake Processes in the Northern Vanuatu Subduction Zone

3.1 Abstract

On 07 October 2009 the northern segment of the Vanuatu (formerly New Hebrides) subduction zone (11°S to 14°S) ruptured in three major (Mw 7.7, 7.8, and 7.4), shallow thrust earthquakes in close spatial proximity and occurring over about one hour. This region of the plate boundary has previously experienced large (Mw > 7.0), shallow thrust events, including doublets in 1966 and 1980, and large single events in 1997 and 2013. The July 1980 earthquake doublet (Mw 7.5, 7.7) ruptured the plate boundary in approximately the same along-strike location as the 2009 sequence and the focal mechanisms of all of the events are similar. The 2009 sequence involved near-immediate activation of small earthquakes along the entire length of the northern Vanuatu subduction zone, similar to a pattern observed for the 1980 earthquake sequence. We analyze seismic data from the 1980, 1997, 2009, and 2013 sequences to compare and contrast the rupture processes of these large earthquakes. The first large event in the 2009 sequence ruptured to the north, stopping short of the nucleation region of the second, slightly larger event, which initiated about 15 minutes later. These events are consistent with interplate activity. The 1997 earthquake observations suggest strain release within the subducting plate dominated the event, although some triggered slip on the shallow megathrust is likely
to also have occurred. We find that the repeated occurrence of large earthquake
doublets along the northern Vanuatu subduction zone is quite remarkable because
the doublets have overlapping, yet different combinations of asperities in each
sequence.

3.2 Introduction

Northern Vanuatu has experienced an abundance of moderate-to-large earthquakes
since 1973 (Figure 3.1, Appendix B.1); however, no events exceeding $M \sim 8.0$ are
listed in any of the major earthquake catalogs. Of course we have no guarantee
that the maximum expected earthquake size for the region has been observed, we
must exercise caution when interpreting earthquake observations from relatively
short-duration catalogs. The high-level of moderate magnitude seismicity over
the last 40 years has been punctuated by four sequences of major earthquakes
in 1980, 1997, 2009, and 2013. Most recently, a great earthquake ruptured the
northwest extent of the subduction zone on 06 February 2013 (U.S. Geological Survey –
National Earthquake Information Center (NEIC): $M_w 8.0; 10.738^\circ S, 165.138^\circ E$).

Just south of the 2013 rupture, a series of three large, shallow, thrust earthquakes
ruptured on 07 October 2009 (NEIC: $M_w 7.7, 13.052^\circ S 166.187^\circ E$, 22:03:15 UTC;
$M_w 7.8, 12.554^\circ S 166.320^\circ E$, 22:18:26 UTC; $M_w 7.4, 13.145^\circ S 166.297^\circ E$, 23:13:49
UTC). Prior to 2009, a major earthquake on 21 April 1997 (NEIC: $M_w 7.7; 12.58^\circ S$
$166.68^\circ E$) ruptured in the same general area as the 2009 events. The 1980 sequence
included a large underthrusting doublet close to the 2009 sequence (NEIC: 8 July,
$M_w 7.5; 12.92^\circ S 166.21^\circ E$; and 17 July, $M_w 7.7, 12.44^\circ S 165.94^\circ E$). On 31 December
1966, a sequence initiated about 50 km to the northwest of these events, in a
similar location as the 2013 event (Figure 3.2). The 1966 sequence also included
two large underthrusting events (18:23:08.8 GMT, $M_s 7.9, 11.9^\circ S 166.4^\circ E$; and $M_s$
7.3, 22:15:17.1 GMT, $12.1^\circ S 165.7^\circ E$) [14].

The 31 December 1966 earthquake sequence included two large events rupturing
within four hours of each other. The first event occurred at 18:23:08.8 GMT ($M_s$
7.9, $11.9^\circ S 166.4^\circ E$), followed by an $M_s 7.3$ aftershock at 22:15:17.1 GMT ($12.1^\circ S$
$165.7^\circ E$) [14]. While the International Seismological Centre catalog (ISC) lists a
depth of 73 km for the first event, Tajima et al. [14] found the long-period surface-
wave observations were most consistent with a depth of $\sim 40$ km. They were unable
to fully constrain the 1966 faulting geometry using the P-wave first motion data but suggested a similar mechanism to the 17 July 1980 focal mechanisms ($\phi = 347^\circ$, $\delta = 36^\circ$, $\lambda = 91^\circ$). The larger 1966 event was believed to rupture unilaterally to the north along the strike. The peak rate of energy release occurred between 12 to $\sim$50 seconds after the rupture initiation. Additionally, the rupture pattern suggested a distinct barrier blocked rupture propagation to the south [14]. The aftershock pattern propagated almost exclusively to the north and was characterized by rapid expansion in size, reaching a final width of $\sim$200 km. The 100-day aftershock region was more than three times larger than the 1-day, contributing to Tajima and Kanamori [112] noting that the western Pacific characteristically exhibits significant aftershock expansion rates, possibly relating to a lower density of asperities. Smaller events later ruptured in the region south of the 1966 seismicity on 23 January 1972 ($M_s 7.1$, $13.2^\circ$S $166.3^\circ$E) and 06 October 1975 ($M_s 7.0$, $12.5^\circ$S $166.6^\circ$E). The strike of the 1972 event was rotated slightly more to the east of the 1966 events. The aftershock area of the 1972 event did not expand much beyond the 1-day region, remaining under $\sim$125 km in width [112]. As will be discussed later, rupture of the 1966 seismicity region appears to have been reactivated in the 2013 earthquake.

Another sequence of events occurred in July 1980. This sequence began on 08 July 1980 with an $M_w 7.5$ thrust event ($12.92^\circ$S $166.21^\circ$E, 43.6 km) and was followed about a week later by a larger thrust event ($M_w 7.7$) on 17 July 1980 ($12.44^\circ$S $165.94^\circ$E, 34.0 km). Some consider the first event a foreshock, but Tajima and Kanamori [112] noted this event (08 July 1980) produced its own aftershocks, indistinguishable from those of the latter, larger event (17 July 1980). Both events of this sequence displayed similar focal mechanisms, as well as with the 1966 main shock [3,14]. The moment rate function of the 17 July 1980 event was complex with peaks occurring around 30 and 60 seconds. Rupture modeling using body-waveform inversion suggested that the primary rupture occurred in the down-dip portion of the seismogenic zone. In the first $\sim$30 seconds, most of the moment release occurred just south of the epicenter. Following the initial 30 seconds, rupture occurred primarily to the north [14]. Tajima et al. [14] suggest the asperity that served as a barrier to the southern propagation of the 1966 event failed early in the rupture process of the 1980 event.

Like the 1966 event, the aftershock pattern of the 1980 sequence also expanded substantially toward the north. The 100-day aftershock area was more than three...
times the size of the 1-day [112]. The 1-day aftershock area of the 08 July event included part of the aftershock area of the moderately sized 1972 and 1975 events and extended up to the southern boundary of the 1966 1-day aftershock area. The 17 July event ruptured within the aftershock region of the 1980 event. The northern extent of this aftershock area also included the 1966 aftershock region [14]. Estimates by Tajima et al. [14] suggest the source area of the 1980 event was twice as large as the 1966 event, but the seismic slip was only half as great as the former event. There was no appreciable expansion in the aftershock area following the 17 July 1980 event [112].

A single large earthquake ruptured slightly south of the 1980 sequence on 21 April 1997 (M_w 7.7 13.21°S 166.20°E), producing a local tsunami as high as 3 m in the Solomon and Vanuatu islands [113]. The Global Centroid Moment Tensor (GCMT) (e.g. [114]) strike (φ = 301°, δ = 39°, λ = 40°) of this event was slightly more west of north-south than that of the 1966 and 1980 sequences, however the dip is similar to these earlier sequences. Analysis of S and SS waves using an empirical Green's function method, Kaverina et al. [113] suggested the hypocenter was deep but the peak slip of 4.25 m occurred shallow, between depths of 20 to 5 km up-dip and to the south of the origin. The GCMT and W-Phase modeling [Gavin Hayes, personal communication, 19 December 2011] also indicate a deep centroid of 45 km for this event. The aftershock sequence of this event was shallow and near the trench.

On 07 October 2009, ~350 km of the northern extent of the Vanuatu subduction zone ruptured in three major, shallow, thrust earthquakes (Figure 3.3). With similar centroids as the 1980 doublet, the first 07 October 2009 event occurred at 22:03:15 UTC (NEIC: M_w 7.7, 13.052°S 166.187°E), and was followed 15 minutes later by a slightly larger event (NEIC: M_w 7.8, 12.554°S 166.320°E, 22:18:26 UTC). A third major earthquake in the sequence occurred about an hour after the first (NEIC: M_w 7.4, 13.145°S 166.297°E, 23:13:49 UTC). A modestly damaging tsunami with peak water height of 0.1-0.3 m on tide gauges in the Vanuatu Islands (south of the region shown in Figure 3.3) was associated with this sequence [115]. GCMT mechanisms indicate that all of the events in this sequence ruptured similarly oriented faults. The rupture parameters of this sequence are also similar to those of the 1980 sequence.

Most recently, the 06 February 2013 (NEIC: MÂw 8.0; 10.738°S, 165.138°E)
great earthquake ruptured a similar fault segment as the 1966 sequence. The 2013 mainshock hypocenter is over 100 km north of the 1966 hypocenters; however, the aftershocks zones are similar. Like to the 1997 event, the 2013 earthquake produced a local tsunami as high as 3 m on the Nendō Island [115]. The GCMT faulting parameters ($\phi = 314^\circ$, $\delta = 21^\circ$, $\lambda = 74^\circ$) indicate the 2013 event had similar faulting geometry as the 1966, 1980, and 2009 sequences.

Globally, large earthquake doublets have been observed in many regions (e.g. [7, 8, 33, 42, 116]), however, the northern Vanuatu region is similar to the Solomon Islands in having an unusual number of large doublets and triplets (Figure 3.2). In this study, we analyze the body and surface waves generated by the large earthquakes in 1980, 1997, 2009, and 2013. Our primary emphasis is on the 2009 sequence, but we provide analysis of the 2013 event and extend previous work by others on the 1997 and 1980 events as possible (the digital data available for the 1980 sequence are limited) and try to place these recent events in a broader perspective. We find that each sequence appears to have ruptured partially overlapping regions of the plate boundary (and perhaps within the slab in 1997). Although the sequences are complicated, clarifying the relationships of these large events is a central part of the long-term effort to understand the important problem of large earthquake interactions. We begin by describing of our investigation of the 1997 sequence, then analyze the 2009 and 2013 earthquakes, and conclude by discussing the relationship of all five sequences (1966-2013).

3.3 The 1997 Vanuatu Earthquake

We used global broadband short-arc Rayleigh waves (R1) to estimate the overall rupture characteristics of the 1997 event. We accounted for geometrical excitation and propagation effects by using the GCMT point-source solution for the first event in computing fundamental mode synthetics for each observation with corrections for aspherical phase velocities relative to PREM [8, 117]. We performed iterative deconvolutions of the synthetic point-source seismograms from the observed R1 signals and filtered the results with a Gaussian pulse to reduce short-period noise. We then used the resulting R1 wavetrains to estimate the R1 source time functions (STFs) (Figure 3.4). Although the Green’s functions include periods as short as 20 seconds, the STFs contain reliable source information in the period range from

60
about 50 to 200 s. These signals constrain aspects of the overall rupture process, such as rupture direction and duration, but provide only weak resolution of rupture speed.

In our analysis of the R1 STFs, we assumed a point source with the GCMT faulting geometry ($\phi = 301^\circ$, $\delta = 39^\circ$, and $\lambda = 40^\circ$) and a depth of 45 km based on GCMT and W-Phase modeling [Gavin Hayes, personal communication, 19 December 2011]. We examined the pattern of the relatively simple STF pulse-shape duration as a function of assumed rupture direction for the 1997 event at varying azimuths. The most consistent pattern in STF pulse-shape variation was observed for a rupture directed at an azimuth of 200° (Figure 3.4), which is substantially different from the strike of the trench ($\sim 345^\circ$) but not far from orientation of the steeply dipping plane in the GCMT faulting geometry (178°N). We checked STF estimates obtained without using the positivity constraint applied to the signals shown in Figure 3.4 – the results were noisier, but showed no systematic variations that would indicate a complex source [118]. The STF ($\Gamma = 0$) duration of about 75-80 s is somewhat longer than the 60-65 s duration of relative moment-rate functions estimated using body waves for this event by Kaverina et al. [113]. Considering that we have smoothed our estimates with the low-pass Gaussian filter, the difference is not significant. If we interpret the rupture as unilateral toward the 200° azimuth, we infer about 80 s of directivity (120 s – 40 s). Assuming an R1 phase velocity of roughly 4 km/s the 80 s of directivity suggests a rupture length of about 160 km; the estimated rupture speed is about 2.0 km/s, but, again, this estimate might be slightly low as a result of the Gaussian filter broadening of the STFs. The onset of the time functions is quite gradual, so if the Gaussian has introduced a bias, it would be primarily on the end of the rupture – assuming 15 s of Gaussian filter broadening suggests a rupture speed of about 2.5 km/s.

A number of issues are perplexing about the 1997 event. First, the long-period point-source faulting geometry is significantly different from that of the large presumed underthrusting earthquakes in the region (Figure 3.1). Neither of the two planes aligns with the trench and both GCMT and W-Phase inversions have only small non-double-couple components of the moment tensor, so there is not clearly any large-scale geometric complexity during faulting. Both long-period analyses also suggest a relatively deep centroid ($\sim 50$ km). Additionally, a significant number of aftershocks in the sequence are offset to the south of the hypocenter and near
the trench. Perhaps these aftershocks were triggered off-fault, but the R1 STFs are consistent with a rupture direction just west of south, generally towards that aftershock cluster. It is plausible that the event ruptured along a roughly south-striking fault within the slab. Considering the dip of the descending lithosphere, the plane has roughly the orientation expected for a former transform-fault or fracture zone that was initially near vertical in the oceanic lithosphere. Similar intra-slab ruptures have been suggested as part of a complex M$_w$ 7.9 rupture near Sumatra [119] and the 1986, M$_w$ 7.7, earthquake in the Kermadec region [120]. In each of these cases, the intraplate component of the earthquake was believed to be strike-slip in nature, but both show that strain retained within the down-going slab is capable of producing substantial seismic events. On the other hand, Kaverina et al. [113] tested the roughly north-south striking, steeply dipping plane in their analysis of body waves and found that they could fit their observations better by adjusting the dip of the other plane to have a more traditional subduction plate boundary orientation. However, that model does not account for the oblique geometry of the GCMT and W-phase solutions.

To explore this issue further, teleseismic broadband P- and SH-waves were used in a finite-fault inversion to estimate the slip distribution for the 1997 event using a number of rupture plane scenarios. We applied the procedure of M. Kikuchi and H. Kanamori, to invert teleseismic body waves using a kinematic-constrained linear least-squares procedure for a simple layered earth model. To estimate hypocenter depth, dip, and strike, we performed a grid search over these parameters to identify optimal values for these required geometric parameters. We tested five models (Table 3.1), including both nodal planes of the GCMT focal mechanisms for this event, the preferred orientation of Kaverina et al. [113], a 200° striking plane indicated by the R1 STF observations, and a plane with the megathrust geometry of the events in the 2009 sequence. For all models, we used a rupture speed of 2.0 km/s and subfault source time functions were parameterized by 10 half-overlapping triangles with a triangle width of 2.0 seconds. Smoothing was applied to produce the simplest model that produces an acceptable fit to the observations. Prior to inversion, the data (and the Green’s functions) were filtered using a Butterworth filter to emphasize signal periods between 1.11 and 200 seconds. The data distribution was chosen to provide a balanced azimuthal coverage and sample a range of available take-off angles available in teleseismic P waves. Unfortunately,
all of the orientations provide comparable fits to the waveforms, with all geometries accounting for 75-80% of the observed, complicated waveforms (Appendix B.13). Figure 3.5 shows the slip model for the geometry of the steeply dipping nodal plane of the GCMT solution. Like this figure, all models place substantial slip shallow and near the region of dense aftershocks. Our slip distribution is similar to the Kaverina et al. [113] model, only smoother. We are not certain that the 1997 event ruptured a near-vertical fault within the down-going slab, but such a model is compatible with the deep centroid estimates, R1 STFs, and the ambiguous body-wave constraints.

For additional insight into possible rupture complexities, we investigated the rupture process of the 1997 event using the Kikuchi and Kanamori Teleseismic Body-Wave Inversion Program [17] (KK approach). The KK approach is an iterative inversion to model an earthquake source as a series of double-couple point sources. There are two methods of performing this inversion. In one method, the mechanism is fixed and the point sources, subevents, are specified by their location along the fault plane (along strike and dip), seismic moment, and onset time. The inversion can also be run for a variable mechanism, in which case the mechanisms is allowed to vary among subevents. In this method, the sources are identified by their location along strike (location along dip is not calculated), moment, onset time, and focal mechanism. Using a fixed mechanism similar to that used in the finite fault modeling and four subevents, the KK approach’s model shows general agreement with slip patterns from the finite fault model (Figure 3.6, Appendix B.14, B.15). Allowing the focal parameters to vary among subevents, the KK model includes significant variation between the subevents. Strike for the subevents vary from ~90° to 180°. Rake also varies, with one subevent ~40 km south of epicenter having a rake significantly rotated from the rest of the subevents. This type of complexity may indicate multiple rupture planes were active during the propagation of this rupture.

3.4 The 2009 Northern Vanuatu Earthquake Sequence

The 15-minute time separation between the two largest 2009 events complicates our analyses, but we are able to constrain important aspects of each of the ruptures. To estimate the overall rupture characteristics of each event, we used similar analysis
of R1 STFs as used with the 1997 event. Due to the short duration between the
two 2009 events, we deconvolved the synthetic point-source seismograms from
the observed composite R1 signals (including observed R1 signals from both the
first and second event). For stations at closer distances, the two R1 wavetrains
are reasonably well isolated and we can confidently estimate the R1 source time
functions (STFs) for both events. At larger distances, R1 from the earlier event is
contaminated by multi-bounce S waves from the second event, limiting the number
of useful STFs for the first event. R1 for the second event is generally clear of major
interference. The resulting R1 STFs for the two events are shown in Figure 3.7.
We show STFs computed with a positivity constraint in the iterative deconvolution,
but also examined STFs obtained using a water-level deconvolution (which were
consistent, but slightly noisier than those presented here).

The R1 STFs from the first event are displayed in Figure 3.7 as a function of
the directivity parameter \( \Gamma \), computed assuming that the rupture direction is along
the strike of the trench. We only used R1 STFs that produced at least an 80% fit
to the convolutional model. As a result of event interference, only several dozen out
of nearly 300 observed waveforms met the standard, and the results are noisy, but
they provide a sufficient azimuthal sampling. The STFs vary slightly, but show no
consistent directivity effect. The lack of directivity is an indication of a relatively
compact, possibly bilateral rupture. Including the broadening effects from the
low-pass Gaussian filter, the duration of the event at stations nearly perpendicular
to the assumed along-strike rupture direction is, at a maximum, about 50-60 s.
The moment rate function shape is roughly symmetric. The Gaussian filter width
is about 30 seconds (the width where the Gaussian amplitude is equal to 10% of
its peak amplitude), but the amount of measurable broadening produced by the
filter depends on the shape of the signal to which the filter is applied. A smooth
function with a long rise time compared with the Gaussian duration is extended
little by the filter. A sharp, rise or end of an STFs is extended by roughly half
the width of the Gaussian. In this case, our resolution is limited and the duration
could have been up to 30 seconds shorter than the 50-60 s pulse width.

The STFs for the second event have much more variation, including \( \sim 85 \) s of
directivity. From the azimuthal directivity pattern, the second rupture appears to
have propagated along-strike toward the north (N340°). The temporally-stretched
STFs observed to the south (away from the rupture) suggest that the rupture
included at least two, and perhaps three significant sub-events. STFs at azimuths orthogonal to the rupture direction are simple and indicate that the rupture had a total duration of about 80 s. A substantial peak in moment rate occurred at 40 s with a large, slightly asymmetric pulse. The third sub-event apparent in the STFs to the south appears to have occurred close in time to the large pulse. Assuming a simple, unilateral rupture and an average R1 phase velocity of about 4 km/s, the roughly 85 s of directivity indicates a rupture length of about 170 km; the estimated rupture speed is \( \sim 2.1 \) km/s, but this estimate might be biased as a result of the broadening of the STFs by the low-pass Gaussian filter applied to the STF estimates. Assuming a maximum broadening of 30 s would lead to a rupture-speed estimate of 3.6 km/s (the broadening doesn’t affect the length estimate because it is based on a difference in durations). Our resolution is limited, but we believe that the true velocity is closer to the lower rupture speed estimate than the higher.

Finite-fault models provide more information about the rupture characteristics of the first event. To estimate hypocenter depth, dip, and strike, we performed a grid search over these parameters to identify optimal values for these required geometric parameters. Specifically, we varied the strike between \( 342^\circ \) and \( 354^\circ \), and the dip from \( 42^\circ \) and \( 48^\circ \), and tested hypocenter depths between 20 and 50 km. For completeness we also tested the auxiliary focal plane over comparable ranges of dip and depth for strike angles between \( 166^\circ \) and \( 175^\circ \). The waveforms are best fit with a hypocenter depth of 39 km, rupture speed of 2.5 km/s, and focal parameters of strike \( (\phi) = 346^\circ \), dip \( (\delta) = 45^\circ \), and rake \( (\lambda) = 86^\circ \). Subfault source time functions were parameterized by 8 half-overlapping triangles with a triangle width of 1.5 seconds. Like the 1997 event, smoothing was applied to produce the simplest model with an acceptable fit to the observations, and the data (and the Green’s functions) were filtered using a Butterworth filter to emphasize signal periods between 1.11 and 200 seconds. This inversion indicates that a majority of the moment originated from an asymmetric region about \( \sim 60 \) km long by \( \sim 50 \) km wide down-dip (Figure 3.8). The model indicates that the event initiated with a subevent located near the down-dip edge of the eventual rupture zone. The moment rate increased rapidly to a peak at about 15 seconds as the rupture front expanded \( \sim 15 \) km to the north and then expanded in the up-dip direction. The fits to the waveforms are good (Figure 3.9). There is little slip below the hypocenter and the rupture does not appear to extend all the way to the surface, although
the slip 15-50 km from NNW of the hypocenter does proceed substantially up-dip. The peak slip is roughly 3-4 m, and is largest within 10 km of the hypocenter. The model seismic moment is $4.3 \times 10^{20}$ N-m, which is about 30% larger than the GCMT estimate of $3.3 \times 10^{20}$ N-m. The model rupture speed is 2.5 km/s, which is well within the broad range compatible with the R1 STFs. Inversions with the fault model extending further south also place the primary slip north of the hypocenter, so the northward propagation of the rupture is a robust feature constrained by the observations. The source time function duration is about 45 seconds, but moment release was skewed towards earlier times. The hypocenter of the second 2009 event is located just north of the rupture limit of this first event. Unfortunately, we are unable to perform a similar body wave analysis for the second, larger event, because the body waves are overwhelmed by signals from the earlier event.

To better define the relative rupture centroid positions for the three largest earthquakes in the sequence, we used a cross-correlation analysis of the R1 observations from each event. Because the rupture depths and the faulting geometry of all the events are similar, timing relations between the R1 waves can be used to obtain good estimates of the relative centroid locations for each event. Effectively, we cross-correlated signals from the first event with those from the second and third events and then fit a sinusoid to the resulting azimuthal patterns of phase shifts (Figure 3.10). This method cannot provide information on the absolute location of the events; the entire pattern can be shifted without affecting the fit to the relative travel time observations. The later 2009 events were repositioned relative to a location based on results of finite fault analysis (Figure 3.8) for the first 2009 event. The observations suggest that the second event’s temporal centroid was approximately 15.75 minutes later than the first and the third event’s centroid was about 70.50 minutes after the first. The second event’s spatial centroid is located about 70 km due north of the first and the third event’s centroid was about 60 km due south. This timing and locations are generally consistent with the GCMT centroid estimates, except the GCMT centroid of the second event is about 10 km north and slightly to the west of our location and the GCMT centroid of the third event is slightly east of our location. We performed a similar analysis on the 1980 data, and found that the centroid of the 17 July event occurred about 76 km and slightly west of due north ($\phi = 346^\circ$) of that for the 08 July event. Unfortunately, as a result of having very few common stations between the 1980 and 2009 events,
an attempt at computing relative centroid location between these two sequences was inconclusive.

To summarize our findings for the 2009 sequence, the first event was relatively compact and ruptured a distance of about 60 km from south to north along the deeper region of the seismogenic zone. The rupture began with a small, deep subevent and ruptured northward into a larger asperity. About 15 minutes later, the second and larger earthquake occurred at the northern end of the first rupture. This event included at least two and perhaps three subevents, and appears to have ruptured about 180 km to the north with an average rupture speed of about 2.1 km/s. A third event, with much smaller moment than the first two, occurred about 70 minutes later, and was located to the south of the first event (Figure 3.2). The entire length of the northern Vanuatu plate boundary quickly activated with aftershocks. The combined one-day aftershock region extended \( \sim 300 \) km, and over the following week extended to \( \sim 400 \) km along the plate boundary, with the mainshock epicenters located roughly in the middle of this region (Figure 3.11). Aftershocks in the southern half of this region ceased after about 14-days, but those north of the large-event epicenters continued for nearly two months.

### 3.5 Waveform Comparison of the 1980 and 2009 Earthquake Sequences

Comparing the two 1980 and 2009 doublets, the close initial locations, magnitudes, and similar faulting geometries suggest that both of these pairs ruptured similar portions of the plate boundary. Given the close temporal proximity of the two largest 2009 events, it was not possible to effectively isolate a single event. Consequently, we combined the two 1980 mainshocks with a best-fitting offset to create a composite waveform, simulating the second 1980 event as having a similar time delay as the second 2009 event. The composite of the 1980 events was created by determining the best-fitting time offset and relative amplitude that minimizes the L2 difference misfit to the 2009 time series. For completeness, we also investigated all doublet permutations of one or both 1980 events. The best fit was produced with a combination of the 08 July 1980 event waveform followed by the 17 July 1980 waveform; this is what we present here (Figure 3.12). Because the waveforms from
the second 1980 event were clipped at most stations, we included the instrument response of the 1980 earthquake’s waveforms into the 2009 sequence’s waveform and applied a causal low-pass filter to the second 1980 waveform. Also due the waveforms being clipped, we can only reliably compare the waveforms prior to the arrival of the clipped R1 waves of the second 1980 event. This provides comparison of the entire first 1980 event and the body waves of the second with the 2009 sequence.

Limited station similarity, waveform quality, and azimuthal distribution restricted this analysis to only five stations: Albuquerque, NM (φ = 55.5°, Δ = 95.2°, ANMO), Chiang Mai, Thailand (φ = 294.4°, Δ = 73.5°, CHTO), Guam, Mariana Islands (φ = 320.6°, Δ = 33.9°, GUMO), Matsushiro, Japan (φ = 332.9°, Δ = 56.1°, MAJO), and South Karori, New Zealand (φ = 166.7°, Δ = 29.1°, SNZO). Waveforms were low-pass filtered at 50 seconds. ANMO and MAJO stations display very strong fits up to ∼2200 and ∼2500 s, respectively. CHTO also provides a generally good match up to ∼2500 s, however, there is a notable difference in the first P-wave arrival and at ∼1800s. GUMO and SNZO generally display poor correlation between the two sequences, except the R1 waves of the first 1980 event appear to match well with the first 2009 event. Overall, the comparisons suggest that the events are similar but do have some differences in radiation pattern, which suggests the two sequences differ.

### 3.6 The 2013 Vanuatu Earthquake

We analyzed R1 STFs to gain greater insight into the rupture behavior of the 2013 event (Figure 3.13). For the STF modeling, we assumed a point source consistent with the GCMT faulting geometry (φ = 314°, δ = 21°, and λ = 74°) and a depth of 27 km. We aligned the signals by handpicking the onset. The R1 STFs for the 2013 event display pulse durations ranging from ∼75 to 140 s, with an average duration of ∼105 s. The duration did not always varying consistently with azimuth, and some signals display durations shorter than expected. This may indicate complexity in the rupture directivity, propagating in multiple directions. However, the trend of the STFs indicate directivity to φ = 135°. Assuming an R1 phase velocity of roughly 4 km/s, the 62 s of directivity suggests a rupture length of about 125 km; the estimated rupture speed is about 1.5 km/s.
Finite fault modeling provides a more detailed rupture analysis, showing the 2013 event ruptured with a relatively simple slip pattern (Figure 3.14, Appendix B.2). The published GCMT and NEIC centroid moment tensor solutions ($\phi = 311^\circ$, $\delta = 28^\circ$, $\lambda = 68^\circ$, depth = 32 km) are problematic because given the hypocenter at this depth, the dip causes the rupture plane to be $\sim$10 km below the surface at the trench. However, tsunami modeling by Lay et al. [121] indicates rupture extended up to or very near the surface. The NEIC W-phase moment tensor solution is similar to the GCMT solution ($\phi = 309^\circ$, $\delta = 17^\circ$, $\lambda = 61^\circ$), however, the shallower hypocenter depth (depth = 15 km) places the fault plane at a physically reasonable depth at the trench. For our finite fault modeling, we produce a good fit to the P- and SH-waves with a similar faulting geometry as the USGS W-Phase moment tensor solution ($\phi = 309^\circ$, $\delta = 17^\circ$, and $\lambda = 58^\circ$). We fixed the hypocenter to the NEIC location, except decreased the depth to bring the fault plane near the surface at the trench (10.738$^\circ$S, 165.138$^\circ$E, depth = 18 km). We also tested the deeper source of 28.7 km, but with a slightly steeper dip than the GCMT and NEIC centroid moment tensor solutions ($\phi = 309^\circ$, $\delta = 30^\circ$, and $\lambda = 58^\circ$) in order to place the fault plane near the surface at the trench. While both models produced a similar slip pattern, for our discussion, we will focus on the model with a shallower source. The shallower source model provides a moderately improved fit to the waveforms. However, the steeper dip is most similar to the dip of the 1980 and 2009 events to the south. Discussion of the model with a deeper source is provided in the appendix (Appendix B.4, B.4, B.5, B.6).

As indicated by the R1 STF analysis, the observed waveforms are best fit using a relatively slow rupture velocity of 1.5 km/s. Like the 2009 finite fault model, this model’s seismic moment, $2.12 \times 10^{21}$ N-m, is higher than the GCMT estimate of $8.65 \times 10^{20}$ N-m. Because the model is so stable, we applied a modest smoothing constraint. The model shows the hypocenter located in the middle of one patch of high slip. Two other slip patches exist up-dip, with most slip occurring shallow and slightly south of the hypocenter. The model indicates modest slip extending 30 km down-dip and 30 km north of the hypocenter. While peak slip occurs shallow on the fault plane, the peak moment is located near the hypocenter (Appendix B.7). The difference is from the change in shear modulus with depth. Results from this model corroborate rupture length calculations from the STF analysis of $\sim$120 km. Agreement is also observed between the STF pulse shape and the finite
fault model’s moment rate function. The calculated moment rate function of the 2013 event is a relatively simple triangular shape with a duration of $\sim 100$ s. The moment rate function has a $\sim 20$ s rise time to peak slip. Rupture continued for about 70 s following the peak moment, with moment rate remaining nearly constant between 30-60 s after initiation. This model centroid is at 10.88°S, 165.28°E and depth = 17.3 km, which is $\sim 25$ km northeast and shallower than the GCMT centroid (11.08°S, 165.14E, depth = 27.6 km). The centroid time minus hypocenter time for this model is 34.2 seconds, $\sim 6$ seconds greater than the GCMT’s value. STF modeling depicted rupture directivity nearly parallel to the strike of the trench ($\phi = 135^\circ$), while the finite fault model depicts directivity closer to $\sim 180^\circ$. However, it is possible once the rupture reached the top of the fault plane, propagation then proceeded along strike. Low frequency energy from this shallow portion could dominate the R1 waves.

We observe further evidence for depth dependence in the frequency content using the KK inversion approach. Results from the fixed mechanism inversion, using faulting parameters similar to that used in the finite fault modeling, identify subevents near the two down-dip asperities near the hypocenter imaged in the finite fault modeling (Figure 3.15, Appendix B.8, B.9, B.10, B.11, B.12). The KK fixed mechanism inversion does not identify the shallow slip seen in the finite fault modeling. Considering the KK fixed mechanism inversion only utilizes short period energy, the absence of shallow moment in this model may reflect a deficiency in short period energy from shallow slip. For the variable mechanism inversion, we limited the number of subevents to four, any more and the model began to show instability. Results from this modeling included three subevents with focal parameters generally similar to that of the GCMT. The fourth subevent, however, has a substantially different rake, making the event nearly strike-slip. This subevent is located south of the hypocenter, possibly associated with the shallow slip. The finite fault model depicts rake of this shallow slip in accord with the mechanism depicted in the KK inversion.

The 2013 earthquake displayed both distinct differences and similarities to other large events in northern Vanuatu. While the 2013 event lacked a second event of comparable size to classify it as a doublet, it was followed by three shallow $M_w \geq 7.0$ events in the two days after the mainshock. These three events included a normal-faulting $M_w$ 7.1 in the outer-rise (NEIC: 06 Feb 2013 01:23:20 UTC,
11.254°S 164.932°E), and two strike-slip events to the east of the mainshock, near Nendō Island (NEIC: Mw 7.0, 10.479°S 165.772°E, 06 Feb 2013 01:54:15 UTC; Mw 7.1, 10.932°S 166.021°E, 08 Feb 2013 15:26:38 UTC). Rupturing in a similar location as the 1966 sequence, the 2013 mainshock displayed many similarities with this event; however, it also exhibited clear differences from the 1980, 1997, and 2009 events in terms of the foreshock behavior, slip density, and locations of fore-/aftershocks relative to the megathrust.

The 2013 great earthquake was preceded by abundant moderate to large foreshocks; there were seven M ≥ 6.0 foreshocks in the 100 days prior to the 2013 event. Only the 1980 sequence had more than one (total of two) in the same time period prior to the initial mainshock. Based on the ISC catalog, a similar foreshock pattern was also lacking prior to the 1966. For the 2013 event, the 1-month foreshock activity was isolated near the epicenter of the mainshock and in a tight cluster of shallow seismicity ~30 km east of Nendō Island. Following the mainshock, the 1-month aftershock zone indicates the rupture was compact relative to the total moment. The aftershocks suggest the event ruptured the fault zone from the northern “corner” at ~10.5°S to near the northern limit of the 2009 and 1980 ruptures at ~11.5°S. This zone is approximately half of the size of that of the largest 1980 and 2009 events (Figure 3.2), suggesting a considerably higher stress drop. The aftershock zone suggests that rupture of the 2013 event extended up to, but did not overlap the rupture regions of the 1980 and 2009 sequences. Comparison with seismicity of the 1966 event supports the conclusion that the 2013 event re-ruptured the same portion of the megathrust. However, the 1966 mainshock epicenters were located in the southern portion of the rupture region and ruptured to the north. The 2013 event initiated ~150 km north of the 1966 hypocenters, near the middle of the rupture region, and rupture propagated toward the south.

The locations of the fore-/aftershocks for the 2013 event also differ from those in 1980, 1997, and 2009. Based on NEIC hypocenters and GCMT moment tensors, the seismicity associated with the former events appeared to largely occur along the megathrust. However, foreshocks and aftershocks of the 2013 event appear to have occurred along many different structures. In addition to shallow thrusting events near the subduction trench, there was normal faulting in the outer rise, moderate-size shallow strike-slip events in the backarc near Nendō Island, shallow
normal faulting ∼50 km east of Nendō Island, and the plate boundary west of the northern corner was also activated with seismicity of various forms of faulting geometries (Figure 3.16). While the abundance of off-megathrust seismicity was absent from the three prior large sequences, the fore-/aftershock seismicity east of Nendō Island was also observed the 1966 sequence (Figure 3.2).

The complex fore-/aftershock behavior may be due to the unique tectonic setting for the 2013 and 1966 events. At the northern corner of the Vanuatu subduction zone, just north of the 2013 hypocenter, the Australian plate transitions from a strike-slip plate boundary west of the corner, into a convergent boundary south of it. Here the Australian plate subducts beneath the Pacific plate. Consequently, down-dip of the northern limit of the subduction zone is the edge of the Australian plate, with subducted material to the south and none to the north.

3.7 Discussion

The 2009 (M$_w$ 7.7; M$_w$ 7.8; M$_w$ 7.4), 1980 (M$_w$ 7.5; M$_w$ 7.7), and 1966 (M$_s$ 7.9; M$_s$ 7.3) sequences ruptured with multiple large earthquakes; the 1997 (M$_w$ 7.7) sequence had only a single large earthquake and appears to be an intraplate event. Our preferred tectonic interpretation for these events is summarized in Figure 3.17. Analysis of the waveforms supports the strong similarity of the two sequences, but differences in the relative amplitude of different seismic waves indicate that the 2009 sequence was not a precise repeat of the 1980 sequence (Figure 3.12). Estimates of the relative mainshock centroid locations using Rayleigh wave cross-correlation for the largest events (Figure 3.10) corroborates this conclusion (Figure 3.2). The two sequences ruptured overlapping lengths along the trench axis (Figure 3.17). We cannot definitively determine the amount of overlap between these sequences, however, waveform comparisons suggests the 2009 sequence was not an exact repeat of the 1980 doublet. In addition, the 2009 sequence appears to have activated more M$_w$ 4-5.5 aftershock seismicity to the north and south of the hypocenters than the 1980 sequence (Figure 3.18). The low level of 1980 aftershock activity to the north relative to the 2009 sequence may in part be due to a higher event detection threshold in 1980, but also could be reflective of this region still being in “recovery” following the 1966 sequence which produced widespread aftershocks to the north (Figure 3.2).
It is unclear why the 2009 sequence only ruptured a part of the fault zone that failed in the 1980 mainshocks, while also rupturing a region unique from this earlier sequence. It is possible the differences in rupture zones may have later failed in aftershocks activity as opposed to the mainshock ruptures, or each sequence may have not resulted in a complete rupture of all the stored elastic strain energy along the entire width of the seismogenic zone. Either way, the considerable overlap and similarity between the 1980 and 2009 sequences suggests that most of the seismic slip along this fault segment is strongly controlled by several dominant asperities. However, the differences indicate the upper and lower limits of the seismogenic zone may contain second-order asperities that do not always always fail in concert with the rest of the fault segment, allowing for heterogeneous stress levels along the width of the megathrust.

The relationship of the 1980 and 2009 sequences with the 1997 event is complicated, largely due to the uncertain rupture behavior of the 1997 earthquake. Body-wave relocation places the 1997 hypocenter deeper, arc-ward, and between the first and second 2009 events. However, the aftershock activity was densely localized near the trench, south of the southern limit of the 1980 sequence and close to the third 2009 event (Figure 3.2). The GCMT location places the 1997 event’s centroid closer to the aftershock activity near the trench than the body waves indicate (13.21°S, 166.20°E). A tsunami was also reported associated with this event [66], further suggesting, but not requiring, the rupture extended to the trench. Unfortunately, relative centroid locations based on Raleigh wave cross-correlation, linking the 1997 event to the 2009 sequence, proved inconclusive (likely due to differences in faulting geometry). The centroid and body wave locations of the 1997 event suggest the rupture began deep and in the north, then propagated ∼80 km south and towards the trench where much of the slip occurred. The R1-directivity analysis for this event supports this inference, and suggests southward rupture propagation for this event (Figure 3.4). Despite the 1997 event, the segment of the megathrust trenchward of the 1997 hypocenter remained strained, rupturing 12 years later during the 2009 sequence. Interestingly, the rupture region of the third mainshock (Mw 7.4) of the 2009 sequence is in the same region as the dense cluster of 1997 aftershocks. Considering the relatively poor match to the complex waveforms, we cannot with complete confidence choose a preferred rupture geometry for the 1997 event. Despite this, we found, like Kaverina et al. [113], rupture initiated deep
and propagated toward the trench, where the peak slip was shallow and near the region of peak aftershocks (Figure 3.5).

There are systematic differences between the 1980, 1997, and 2009 sequences that suggest each sequence ruptured different regions of the plate boundary. The short period (2-4 Hz) energy radiation provides another metric of comparison of the rupture characteristics of the large northern Vanuatu events (Figure 3.19). The 08 July 1980 and first 2009 events both display narrower rise time to the peak energy release than the respective later doublet events. In contrast, the second 1980 and the 1997 events display either gradual or delayed slopes to the peak energy. Interestingly, despite having greater moment than the other events (Table 3.2), based on the $m_b/M_w$ ratio, the 1980 events display a relative deficiency in short-period energy. This could be attributed to radiation patterns effects given the limited available data, which makes quantifying the deficiency difficult.

A similar deficiency in short-period energy is observed for the 2013 event. Given their moment, both the 1966 and 2013 events appear to have ruptured small regions compared to the 1980, 1997, and 2009 events. Tajima et al. [14] noted the 1980 event appeared to display nearly twice as large of a source area but half as much the seismic slip as the 1966 event. Finite fault modeling suggests a similar disparity between the 2009 and 2013 events (Figures 3.8 and 3.14). While this would typically suggest a higher stress drop in these northern sequences, the 2013 event displays a significant deficiency the short period (2-4 Hz) energy compared to these other events (Figure 3.19). Energy calculations by the USGS, $1.3 \times 10^{15}$ N-m, result in an energy magnitude $M_e = 7.18$ [66] using the formula presented by Choy and Boatwright [122], significantly less than the moment magnitude, $M_w = 8.0$. It is possible this difference in short-period energy is related to the depth at which the slip occurred. The hypocenter of the 2013 event was over 10 km shallower than the 2009; assuming the GCMT centroids, the 1980 was slightly deeper than the 2013 and up-dip of the 2009 earthquakes and 1997 event (assuming a rupture geometry consistent with the megathrust). It is possible the relative short-period energy deficiency may be a result of depth dependence in the frequency content [123].

As we have discussed, the 1966 and 2013 events display characteristics that are different from the other large events discussed in this paper. Aside from common aftershock rupture regions, both of these sequences activated shallow seismicity in the backarc, east of Nendö Island. Locations and focal mechanisms of events
associated with the 2013 mainshock show seismicity occurring along many different nearby structures, including thrust, strike-slip, and normal faulting parameters. A significant difference between the 1966 and 2013 sequences is the hypocenters. The 1966 sequence initiated in the southern portion of the rupture zone, while the 2013 initiated ∼150 km north of this location. It is unclear how to interpret this observation. The amount of aftershock activity near the hypocenters of the 1966 sequence has been minimal. This region experienced significant aftershock activity following the 2009 sequence, suggesting it may be in recovery following the 2009 sequence. It is also possible elastic strain energy remains in this region.

Conversely, while the 1966/2013 region experienced significant aftershock activity following the 1980 and 2009 sequences, the aftershock region of the 2013 event did not extend south into the 1980/2009 ruptures. While highly speculative, it is possible the abundant aftershock activity following the 1980 and 2009 sequences indicates the northern corner of the Vanuatu subduction zone had very high strain during the 1980 and 2009 sequences, leading to the 2013 event. Following the 2013 rupture, there was little strain south of the rupture due to the recent 2009 sequence, resulting in a smaller aftershock region. Observation of this type of aftershock behavior could possibly provide identification of regions with high strain prior to their subsequent rupture. If the entire 1966 rupture zone re-ruptured in 2013, this would indicated rupture of this fault segment is capable of initiating at very different locations. Additionally, this difference in hypocenters may be related to why the 2013 event ruptured as a single event, as opposed to a doublet.

The significance of the repeated occurrence of doublets in the northern Vanuatu region is unclear. Felzer et al. [42] concluded that statistically the occurrence of doublets is in accord with a physical rupture model based on a single triggering event. Accordingly, there may be no need to invoke a unique physical mechanism to generate large doublets. In northern Vanuatu, the NEIC catalog includes 10 events since 1973 greater than magnitude 7.5 (Table 3.3). Four of these events include the 1980 and 2009 doublets/triplets. Two more of these events include another possible doublet in 1985 (28 November, $M_s$ 7.6, $M_w$ 7.0; 21 December, $M_s$ 7.6, $M_w$ 7.1). These events had very different faulting geometries (one strike-slip, the other thrust). Expanding this catalog to 1960 with the United State Earthquake (USE) and California Geologic Survey (CGS) catalogs adds two more events, one of which includes the largest of the 1966 doublet. Overall, since 1960, eight of the twelve
events M > 7.5 ruptured in doublets in northern Vanuatu. This high proportion of large doublet events suggests that effective triggering of nearby asperities plays an important role in seismic strain release along this plate boundary. Interestingly, while high quality data in this region has only been available for the past 15-20 years, comparably large doublets would likely have been detected over the past 100+ years. But doublets have not been identified in this region prior to 1966 [124]. Some doublets with short inter-event times, like for the largest 2009 pair, could be misinterpreted as a single event, but pairs of events even an hour apart would likely have been noted.

3.8 Conclusion

We find the northern Vanuatu 2009 multiplet earthquake sequence ruptured a similar region as 1980 doublet, however, differences in waveforms from these events suggest each sequences ruptured slightly different, albeit overlapping regions. Analysis of the rupture processes shows the first 2009 earthquake rupture propagated towards the second event. The 1997 event has an ambiguous relationship with 1980 and 2009 events. While the 1997 rupture geometry remains uncertain, we find the event nucleated deeper than the earlier large events, rupture propagated obliquely towards the trench, with substantial slip occurring in a region roughly coincident with the third event in the 2009 sequence. Both the 1966 and 2013 events ruptured the fault segment north of the other large events. These two sequences appear to have ruptured very similar segments, however, initiated at very different locations. Given their magnitude, the rupture area of the northern events is significantly smaller than that of the south; however, the 1966 and 2013 events display much higher slip. Despite the higher slip and smaller rupture region, the 2013 event displays a notable deficiency in short period energy relative to the events to the south, possibly relating to depth dependence in the frequency content. Comparing the 1980 and 2009 sequences with the 2013, the aftershock patterns were much broader with the southern 1980 and 2009 events, extending up through the 2013 region. Aftershocks of the 2013 event did not extend south into the regions of the 1980 and 2009. While speculative, this could be reflective of the level of stored elastic strain energy in the surrounding regions following each sequence, possibly providing a tool for assessing future seismic hazard.
The relatively high number of large doublets with respect to all large earthquakes in northern Vanuatu suggest a dynamical asperity failure interaction rather than random triggering of events. The similarity in magnitude and location of the large events adds this region to the model proposed by Lay and Kanamori [33], in which the plate boundary contains large interacting asperities. The overlap in rupture regions between subsequent events indicates that these interactions can be quite complex. Considering the strong presence of large doublet activity in the region, it is probable this behavior is more than mere random clustering. If so, the temporal variation in doublet activity over the last 100 years may reflect inherent variability in the interaction of nearby, similar-sized asperities. This modulation of large seismicity characteristics has broad implications for understanding fault segmentation and characterizing subduction plate boundary hazards.
Figure 3.1. Left panel: Seismicity of the northern Vanuatu subduction zone, displaying all NEIC earthquake hypocenters since 1973. The Australian plate subducts beneath the Pacific in nearly trench-orthogonal convergence along the Vanuatu subduction zone. Right panel: All GCMT moment tensor solutions and centroids for $M_w \geq 5$ since 1976, scaled with moment. This region experiences abundant moderate and large seismicity events, but lacks any great event ($> M_w 8$) since 1900. The largest events are displayed with dotted outlines of the magnitude-scaled circle. Convergence rates are calculated using the MORVEL model for Australia Plate relative to Pacific Plate [13]. (See Figure B.1 for a color version).
Figure 3.2. 100-day aftershock maps of all events listed in the ISC catalog for the 1966 sequence and NEIC catalog for the 1980, 1997, 2009, and 2013 northern Vanuatu earthquake sequences (the 2013 event only shows up to the current date, 40-days). The 1966 mainshocks use those listed by Tajima et al. [14]. Events of the 1997 and 2009 sequences were relocated using the double difference method [15] for P-wave first arrival based on EDR picks. The event symbols are scaled to relate the symbol area to the earthquake magnitude based on a method developed by Utsu and Seki [16].
Figure 3.3. The 7 October 2009 rupture sequence in northern Vanuatu subduction zone, including the 3-month aftershock sequence (Mw > 4). The sequence began with the Mw 7.7 event (1), followed ~15 minutes later by the Mw 7.8 event (2). Finally, about an hour later the Mw 7.4 event ruptured (3). The circles located on the NEIC epicenters and the focal mechanisms are from GCMT. Convergence rates are calculated using the MORVEL model [13].
Figure 3.4. R1 source time functions (STFs) for the 1997 mainshock earthquake. The rupture appears to propagate in the direction $200^\circ$N with an estimated rupture speed of 2.0 km/s and a 75-80 s duration.
Figure 3.5. Finite fault model based on teleseismic P- and SH-wave inversion for the 1997 event using a kinematically constrained linear least-squares procedure. Here we show the fit to the steeply dipping nodal plan of the GCMT solution (B.13). All models display a similar characteristic of small slip near the hypocenter and peak slip shallow, correlating with the region of aftershock activity.
Figure 3.6. Kikuchi – Kanamori inversion models [17] for a fixed (left) (B.14) and variable mechanism (right) (B.15) of the 1997 mainshock. Both inversion models use a similar depth source as the finite fault model (Figure 3.5) and four subevents. Results for the fixed mechanism show the location of the subevents on the rupture plane. Results for the variable mechanism inversion shows the location of each subevent along strike. Below the subevents is the total mechanism for the entire event as a whole. The contribution of each subevent to the moment rate function is labeled.
Figure 3.7. R1 source time functions (STFs) for the first (top) and second (bottom) large 2009 earthquakes. The first rupture shows little evidence of directivity, and appears to have about 40-50 s duration. The second rupture appears to propagate to the north (N340°) with an estimated rupture speed of about 2.1 km/s and an ~80 s duration.
Mo = 0.430 × 10^{21} \text{Nm}
M_w = 7.69
Depth = 39 km
Strike = 346°
Dip = 45°
Rake = 86°
V_{rupture} = 2.5 \text{km/s}

Figure 3.8. Finite fault slip distribution from inversion of teleseismic broadband P- and SH-waves for the first event in the 2009 sequence using a kinematically-constrained linear least-squares procedure. The waveform fits are shown in Figure 3.9. The hypocenter depth is 39 km, rupture speed is 2.5 km/s, and strike = 346°, dip = 45°, and average rake = 86°. This mechanism is similar to the GCMT solution (strike = 344°, dip = 41°, and rake = 87°). The moment rate function indicates steady growth of slip for the first \sim 20 seconds until the primary asperity, located north of the hypocenter (left star), finally failed. The NEIC hypocenter of the second event in the sequence is located north and slightly up-dip of the first (inverted right star). The arrows indicate the rake and relative slip amplitude at each subfault. The event had a peak slip of around 8 m. A velocity boundary in our layered source structure enhances the along-dip gradient in slip just below the depth of the hypocenter. The white circles within the focal mechanism show locations of stations used in the inversion; the location relative to the center of the mechanism reflect the station azimuth and distance from the source.
Figure 3.9. Waveforms fits for the inversion model of the first 2009 event (Figure 3.8). The predicted waveforms (thin) successfully match the major features of the observed waveforms (thick). Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, wave type, and azimuth (degrees). The timescale is in seconds, beginning 10.5 s prior to the P- and SH-wave arrivals.
Figure 3.10. Azimuthal plots of the time delays between R1 arrivals for (top) the first and second event, and (bottom) the first and third events. The smooth trends, fit with a cosine curve provide estimates of the relative locations of the event centroids.
Figure 3.11. Space-time plot of the 7 October 2009 earthquake sequence. Distance is measured to the south from -11°N, 165.5°E along the strike of the trench. All events with NEIC magnitudes $M > 4.5$ are displayed. Aftershock activity remained elevated for $\sim 50$ days following the mainshock in the north, but quickly dissipated south of this region. The initial rupture appears to have not extended very far south of the third event. However, aftershock activity displays a southern progression, before stopping after about a week.
Table 3.12. Long-period waveform matching of the October 2009 sequence (solid) and July 1980 (dashed). Given the close temporal proximity of the two largest 2009 events, it was not possible to effectively isolate a single event. Consequently, we combined the two 1980 mainshocks with a best-fitting offset to create a composite waveform, simulating the second 1980 event as having a similar time delay as the second 2009 event.

![Figure 3.12](image1)

**Figure 3.12.** Long-period waveform matching of the October 2009 sequence (solid) and July 1980 (dashed). Given the close temporal proximity of the two largest 2009 events, it was not possible to effectively isolate a single event. Consequently, we combined the two 1980 mainshocks with a best-fitting offset to create a composite waveform, simulating the second 1980 event as having a similar time delay as the second 2009 event.

![Figure 3.13](image2)

**Figure 3.13.** R1 source time functions (STFs) for the 2013 mainshock earthquake. The rupture appears to propagate to the southeast (N135°) with an estimated rupture speed of about 1.5 km/s and a ~105 s duration.
Figure 3.14. Finite fault model based on teleseismic P- and SH-wave inversion for the 2013 event using a kinematically constrained linear least-squares procedure. Here we show the fit to shallow source (18 km) with shallowly dipping plane (dip = 17°) (Appendix B.2, B.4, B.4, B.5, B.6). All models display a similar characteristic of high slip near the asperity, but peak slip shallow to the south-southeast. While peak slip occurs shallow on the fault plane, the peak moment is located near the hypocenter (Appendix B.7). The difference is from the change in shear modulus with depth.
Figure 3.15. Kikuchi – Kanamori inversion models [17] for a fixed (left) (B.8) and variable mechanism (right) (B.9) of the 2013 mainshock. Both inversion models use a shallow source (18 km), similar to that used in the finite fault model (Figure 3.14) and four subevents. Results for the fixed mechanism show the location of the subevents on the rupture plane. This model does not resolve the shallow slip observed in the finite fault model. Results for the variable mechanism inversion shows the location of each subevent along strike. Below the subevents is the total mechanism for the entire event as a whole. The contribution of each subevent to the moment rate function is labeled. Models for a deeper source and steeply dipping plane are shown in Appendix B.10, B.11, B.12.
Figure 3.16. 40-day aftershock pattern of 2013 earthquake, including all events listed in the NEIC catalog. Focal mechanisms are located and scaled to the NEIC catalog information. 19 of the largest events (not including the mainshock) were relocated using surface waves [18] (Table B.1).
Figure 3.17. Summary schematic showing the spatial relationships of recent large earthquake ruptures in northern Vanuatu. We interpret the 1997 event as an intraplate rupture based on a collective assessment of broadband seismic observations (ambiguous broadband P waveforms, unusual faulting geometry, deep long period centroid, and unusual aftershock spatial relationships). Due to limited waveform data, we are unable to constrain how much the 1980 and 2009 rupture zones overlapped. However, analysis of available waveforms indicates the 2009 sequence was not an exact repeat of the 1980 (Figure 3.12).
Figure 3.18. Space-time plot (right) of all seismicity greater than M 5.0 in northern Vanuatu recorded in the NEIC catalog as a function of distance south of -10°N, 165.25°E. The figure on the left shows the location of the seismicity on a map rotated to make the trench oriented vertically.
Figure 3.19. Envelopes of velocity filtered to emphasize the band of 2-4 Hz as observed at MAJO and CHTO stations. The time scale is time from the origin of the respective event. The dotted line shows the relative amplitude, while the solid is absolute. All velocities are normalized for magnitude, depth, and focal mechanism difference, based on the average of modeled P, sP, and pP amplitudes.

Table 3.1: 1997 Rupture Models

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<th>Rake (°)</th>
<th>Depth (km)</th>
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<td>33</td>
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<tr>
<td>S200</td>
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<td>66</td>
<td>122</td>
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<tr>
<td>S340</td>
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</table>

1 Global Centroid Moment Tensor catalog
2 Kaverina et al. 1998
Table 3.2: Large Northern Vanuatu Earthquakes Investigated in this Study

<table>
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<th>Focal(^2) (°)</th>
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\(^1\) National Earthquake Information Center catalog

\(^2\) Global Centroid Moment Tensor catalog
Table 3.3: Northern Vanuatu Events M > 7.5

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<sup>1</sup> National Earthquake Information Center catalog, () indicates UK magnitude scale
<sup>2</sup> Global Centroid Moment Tensor catalog
<sup>3</sup> Tajima et al. 1990
<sup>4</sup> United State Earthquake catalog
<sup>5</sup> California Geologic Survey catalog
Chapter 4  |
Precise Relative Earthquake Location Using Surface Waves

4.1 Abstract

Earthquake locations provide a fundamental tool for seismological investigations. While dense seismic networks can provide robust locations, accuracy and precision of these locations suffer outside dense seismic networks. This is particularly true in offshore areas, where location analysis relies heavily on distant seismic observations. This study presents a method for estimating precise relative seismic source epicentroid locations using surface waves. Several reasons, including slower velocities and strength of the signal at distance, make use of surface waves for event location appealing. We focus on the Panama Fracture Zone region and relocate 81 strike-slip earthquakes to produce tectonically consistent epicentroid locations. The resulting pattern of earthquakes more clearly delineates recently active regional structures than original body-wave locations. The mean shift between the USGS NEIC epicenter and our epicentroids is about 14 km (the median is about 11 km), and typical origin time changes are generally less than ±2 s. We find that north of 6.5°N, the plate boundary motion is split across two roughly north-south striking structures, the Panama and Balboa Fracture zones. For the last 36 years, slip along these two structures roughly matches slip along the Panama Fracture Zone to the south (from 4.5°N to 6.25°N), but the Balboa Fracture zone has roughly three times the moment than the northern Panama Fracture Zone. Our analyses show that observed Rayleigh-wave signal-to-noise ratios for moderate-to-large
shallow earthquakes are suitable for applying the procedure and that Rayleigh-wave observations form a self-consistent set of constraints on the relative location of earthquake centroids.

4.2 Introduction

Earthquake locations are fundamental parameters necessary for the in situ study of earthquake physics, faulting, and interaction and are widely used to map and to quantify Earth’s deformation. Within dense seismic networks, earthquakes can be accurately and precisely located. However, for many important tectonic environments, existing catalog locations are neither accurate nor precise. In particular, for isolated continental and offshore areas, earthquake locations are estimated primarily using distant observations, which often results in inaccurate and imprecise locations [125].

Surface waves are sometimes viewed as a complicated component of the seismic wave field because of their dispersive character and sensitivity to faulting geometry and source depth. These complications often preclude their use in routine seismic analyses such as event location. Additionally, phase and group velocity maps are rarely known well enough to enable accurate locations using traditional absolute time methods. However, surface waves are widely observed for moderate-to-large size shallow earthquakes. Utilizing relative surface-wave time shifts allows the estimation of relative event locations; but surface-wave time shifts can be complicated by waveform depth dependence and azimuthal variations in phase produced by differences in faulting geometry. In some environments, particularly oceanic transform fault systems, these problems are substantially diminished, and relative surface-wave time differences can be measured to an accuracy of one to two seconds, sufficient to produce precise relative locations.

Relative to body waves, the low propagation velocities of surface waves provide a substantial sensitivity to location. For example, a teleseismic P wave has a horizontal phase velocity of about 15-20 km/s; a typical intermediate period Rayleigh wave has a phase velocity of about 3.0-4.0 km/s. Because of this difference, a 16 km location difference can produce about a 5.0 second shift in the surface waveform while the P-wave is only shifted about 1.0 second. These characteristics of surface waves have been exploited by a number of researchers to use surface wave observations in
event location [40,126–131] and to provide important constraints on the moment
distribution of large earthquakes [90,132–134].

Provided that the waveform similarity is high, relative surface-wave time shifts
can be measured precisely, and waveform similarity generally translates into locating
events with similar faulting geometry and depth. This is similar to the requirements
necessary for waveform correlation methods applied at the local and regional
scales [15,135]. As we discuss later, tests using synthetic waveforms show the
required degree of similarity in depth and faulting geometry between events is
functionally broad in important tectonic environments. Example short-arc Rayleigh
(R1) waveforms for three events from the Panama Fracture Zone (PFZ) are shown in
Figure 4.1. Each panel contains three vertical component displacement waveforms
filtered to include periods between 80 and 30 seconds. Both groups of waveforms
are from the same three events. The similarity in wave shape is apparent, even
for events located more than 50 km apart. The relative time shifts (meaning
relative to other local events) can be estimated with a precision on the order of
one to two seconds using a relatively unsophisticated peak of the cross-correlation
function. The systematic azimuthal pattern of observed time shifts supports this
claim (Figure 4.1). A difference in event location should produce a systematic
cosine variation in the time shifts [136]. Although the travel time shifts for the event
pairs are shown on the same scale to communicate the variations of the pattern
with inter-event distance, a close view of the nearby event variation (Events 01
and 02 in Figure 4.1) shows that the measurements are consistent with the cosine
variation to within one or two seconds. In the application described below, we
show that the RMS associated with the inversion of such observations is closer to
1.0 than to 2.0 seconds. This demonstrates that intermediate-period surface wave
shifts can be measured at least that well.

The observations in Figure 4.1 illustrate the raw data and the patterns that
can be utilized to relocate events using Rayleigh waveforms excited by nearby
moderate-size events with similar faulting geometry and depths. In what follows,
we show how similar surface-wave time-shift observations can be combined in a
multiple-event, double-difference based inversion to estimate optimal event locations
over a relatively broad and remote region. Our work extends earlier applications of
these ideas by Ammon [137] and VanDeMark [40], and we use synthetic tests to
explore the limitations and advantages of the method. We illustrate the procedure
using a set of vertical strike-slip earthquakes from the PFZ region (Figure 4.2). Eighty-six events were selected based on a search of the Global CMT (GCMT) catalog [99,114] for vertical strike-slip earthquakes. Most of the moment magnitudes for these events were between 5.0 and 6.0, but events as small as Mw 4.7 produced sufficient observations to be included in the analysis.

Before presenting the results for the PFZ region, we describe the data processing procedure, explore the sensitivity of the observations to differences in event depth and faulting parameters using synthetic waveforms, and discuss the inversion procedures. Descriptions of additional sensitivity tests are provided in an electronic supplement. We then conclude with a brief discussion of the results and their implications for the tectonics of the PFZ region.

4.3 Surface-Wave Double-Difference Earthquake Relocation

4.3.1 Seismogram Processing

The seismogram processing procedure is designed to be simple. Only short-arc Rayleigh waves are included, but clearly the ideas will work with short-arc Love waves (and any other easily identifiable phase on the seismogram, such as P, PP, S, SS, etc.). Love waves for vertical strike-slip earthquakes can be more robust to small mechanism variations than their Rayleigh counterparts [138], but we focus on Rayleigh waves because experience indicates that vertical component seismograms more consistently provide good signal-to-noise ratio for the desired band. We acquired long-period seismograms from stations operating at the time of the event from the IRIS data archive. Most temporary network data were excluded due to frequent low signal-to-noise and possible timing issues for some of the older events. We computed displacement seismograms from the original records using a frequency-domain instrument-response deconvolution. Each signal was inspected for quality control; each seismogram was graded from A to F by visual inspection of the signal quality. We only use the A, B, and C, quality signals in our analyses. We isolated Rayleigh waves using a simple, group-velocity window (5.0 to 3.0 km/s) and filtered the signals to enhance the period range from 80 to 30 seconds. We then cross-correlated the signals to estimate a time difference for the surface waveforms.
We chose to use a range of frequencies that produce a characteristic wave shape that reduces the sensitivity to cycle-skipping in a more narrow-band signal. We discuss issues related to dispersion across this period range later.

Cross-correlation provides a waveform similarity measurement and a time shift. We explored the accuracy of the time-shift measurements by adding a varying amount of observed earth noise to two similar signals. As we increased the amplitude of the perturbation noise, the likelihood of cycle skipping increased. For signal-to-noise values lower than about four, occasional cycle skips began to occur depending on the particular characteristics of the noise. As the signal-to-noise ratios decreased towards unity, the likelihood of cycle skipping increased. Because of the bandwidth, cycle-skip produces about a 20 s difference in the observations, so isolated cycle skips are easy to identify. The quality control screening removes most of the signals with low signal-to-noise ratios so we don’t see this as an issue on our analyses. However, even when the signal-to-noise ratio is high, variations in the faulting geometry between events can affect the cross-correlation estimated time shifts. We explore those effects using synthetic tests.

4.3.2 Depth and Faulting-Geometry Sensitivity Tests

We conducted sensitivity tests using fundamental-mode synthetic seismograms [139] computed with a flat-layered, earth model adapted from the 0-20 Myr oceanic lithosphere model of Anderson and Regan [27]. This model includes a 6.2 km thick oceanic crust and a 20 km thick seismic lid (LID). This is consistent with Richardson et al. [140] for a 5-20 Myr crust, a representative age for the lithosphere in the PFZ region [141,142]. We tested the effect of varying lateral offset (inter-event dispersion) and depth between two sources, as well as the sensitivity to variations in fault parameters (i.e. strike, dip, and rake).

The tests indicate that only at very large inter-source distances ($\gg 120$ km) does inter-event dispersion significantly influence the cross-correlation alignments. However, within the 80 to 30 s period band, the model’s group and phase velocities are flat, producing little dispersion sensitivity. We limit the double-difference event linking distance to less than 120 km in location inversion; over this range, the distance between sources does not produce any effect on the waveforms correlation. We note that for shorter periods, which may be necessary to locate smaller events,
inter-event dispersion may be an issue. These tests also provide an opportunity to measure a representative surface wave slowness to use in the relocation inversion. Comparing the Rayleigh wave packet travel time as a function of distance shows that for the assumed mechanism and period range, a near constant slowness of $\sim 0.25 \text{ s/km}$ is appropriate for this model and bandwidth.

4.3.2.1 Depth Effects on Relative R1 Time Shifts

We explored sensitivity of cross-correlation measured R1 time shifts to source depth by comparing waveforms from sources with similar focal parameters but with varying depths of 5, 12, 24, and 36 km. Waveforms from 5, 12, and 24 km all show strong similarity. The azimuthal travel-time distribution of these waveforms produces a smooth cosine pattern representative of what would be expected for their respective source offsets. Waveforms from a source depth of 36 km are significantly different from the shallower sources, but a source at this depth is within the model’s low-velocity zone (LVZ) and the waveform changes because the R1 eigenfunctions change abruptly at the top of the LVZ. A source in the LVZ is physically unrealistic and most observations suggest there are few earthquakes in the oceanic lithosphere at depths greater than 25 km [143,144]. We conclude that the effects of depth are negligible for the likely range of sources and period range we use. Again, we note that for any analysis relying on shorter periods, the issue should be reexamined.

4.3.2.2 Faulting Geometry Effects on Relative R1 Time Shifts

To explore sensitivity of cross-correlation measured R1 time shifts to faulting geometry, we performed a number of tests comparing R1 waveforms produced by different faulting geometries. We first explore some simple cross-correlation sensitivity using analytical expressions on end-member cases. We then illustrate the ideas using synthetic seismograms. Aki and Richards [145] present the modal expansion expression for the displacement produced by a moment-tensor point...
source in a laterally homogeneous medium as:

\[ u_z(k, h, R, \phi) = \sum_n A(k_n, z_0) \frac{e^{ik_n R}}{\sqrt{k_n R}} \]

\[ \times \left\{ (M_{xy} + M_{yx})k_n r_1(k_n, h) \sin \phi \cos \phi 
+ \cos \phi \left[ M_{xz}k_n r_1(k_n, h) \cos \phi + iM_{xz} \frac{r_3(k_n, h)}{\mu(h)} \right] 
+ \sin \phi \left[ M_{yz}k_n r_1(k_n, h) \sin \phi + iM_{yz} \frac{r_3(k_n, h)}{\mu(h)} \right] 
+ \frac{dr_2(k_n, h)}{dz} \right\} \frac{M_{zz}}{h} \]

(4.1)

where \( n \) is the mode number, and \( A \) is an amplitude factor that depends on the earth model, wavenumber, \( k_n \), and receiver depth, \( z_0 \). The exponential describes the propagation phase shift that depends on wavenumber and the distance from source to receiver, \( R \). The term in brackets describes the effects of faulting depth and geometry, where \( M_{ij} \) is the moment tensor, \( r_i \) is the eigenfunction, \( h \) is source depth, \( \phi \) is the azimuth measured clockwise from north, and \( \mu \) is the shear modulus near the source. Assuming a strike-slip earthquake of strike = 0°, dip = 90°, and rake = 0° (generally representative of PFZ events), the fundamental mode (\( n = 0 \)) term simplifies to:

\[ u_z(k, h, R, \phi) = A(k_0, z_0) \frac{e^{ik_0 R}}{\sqrt{k_0 R}} \times 2M_{xy}k_0 r_1(k_0, h) \sin \phi \cos \phi \]

(4.2)

For a second co-located source with a dip \( \neq 90° \), both the amplitude and phase change relative to the first source because \( M_{xz} \) becomes nonzero:

\[ u'_z(k, h, R, \phi) = A(k_0, z_0) \frac{e^{ik_0 R}}{\sqrt{k_0 R}} \times 2M'_{xy}k_0 r'_1(k_0, h) \sin \phi \cos \phi 
+ iM'_{xz} \cos \phi \frac{r'_3(k_0, h)}{\mu(h)} \]

(4.3)

Of particular importance here is the phase shift, which can directly affect the estimated time shift of the two modes. Cross-correlating waveforms described by Eq. 2 with Eq. 3, we find that the phase (for a single frequency) of the
cross-correlation is described by:

$$\theta(\phi, r_1, r_3, k_0, \mu, M) = \arctan \left[ \frac{M'_{xz} r_3(k_0, h)}{M'_{xy} r_1(k_0, h) k_0 \mu(h) \sin \phi} \right]$$

(4.4)

The arctangent of a weighted cosecant of the azimuth produces a phase shift across all azimuths, but is most pronounced near the radiation pattern nodes at 0° and 180° (Figure 4.3a). For a variation in rake, the cosecant term in Eq. 4 simply changes to the secant of the azimuth because $M_{yz}$ becomes nonzero ($M_{xz} = 0$). This moves the strongest phase shifts to the ± 90° nodes, instead of the 0° and 180° nodes. For our specific geometry of interest, the ratio of $M_{xz}$ and $M_{yz}$ moment tensor elements scales the phase shift. For small deviations from dip = 90°, $M_{yz}$ is close to one ($M_{xz}$ = 1 - $M_{yz}$), so the phase shifts are small. If the dip = 90° and rake = 180°, variations in strike only change $M_{xy}$, $M_{xx}$, and $M_{yy}$, and do not produce a phase shift (Eq. 1). However, when the dip or rake deviate from 90° and 180°, respectively, $M_{xz}$ or $M_{yz}$ are nonzero and variations in strike contribute to a phase shift. The phase shift is also sensitive to a ratio of the shear modulus and the eigenfunctions, $r_1$ and $r_3$, suggesting that differences in source depth can affect the phase shift produced by dip and rake. As discussed earlier, the change in this ratio is minor for the likely range in shallow source depths and a period band of 80 to 30 s. The phase shift also depends on the wavenumber, and hence period, so the above analyses are applicable to narrow-band signals. To explore the integrated effects when using a range of periods, we use synthetic tests.

### 4.3.2.3 Synthetic Tests of Faulting Geometry Sensitivity

To illustrate and quantify these effects, we used synthetic waveforms to calculate the cross-correlation time shifts from seismograms computed using different focal parameters (i.e. strike, dip, and rake). We varied the strike, dip, and rake over a range ± 20° from a simple strike-slip fault (strike = 0°, dip = 90°, rake = 180°). These variations produce effects in accord with what is expected from theoretical considerations described in the previous paragraph. Variation in only the strike, with dip and rake fixed at 90° and 180°, respectively, only shift the radiation pattern node without affecting the phase. The effect is to broaden the azimuthal range where waveform correlation may decrease (in the presence of noise).

For the basic PFZ geometry, the waveforms are most sensitive to variations in
dip or rake. A dip \( \neq 90^\circ \) introduces phase shifts around the 0° and 180° radiation pattern nodes (these effects are relative to the strike direction). Variations in dip of 10° have a noticeable affect within 20° of the nodes and introduce a small, asymmetric bias at other azimuths. Similar to phase shifts calculated in the analytic analysis, the result is a “splitting” phase shift effect across radiation nodes at 0° and 180°. The cross-correlation values reflect these effects, dropping below our threshold of 0.9 roughly \( \pm 15^\circ \) from each node (Figure 4.3b). Greater variation in dip, while strike and rake are the same, enhances the phase shift, affecting a wider azimuthal range around the nodes. Variations in rake produced a similar phase shift effect, except the artifacts are centered on the \( \pm 90^\circ \) nodes (relative to strike) (Figure 4.3c). Most important for our analysis is that inter-event differences in dip or rake produce perturbations to the azimuthal travel-time distribution that can map into a cosine pattern. For a dip difference of 10°, this perturbation produces a cosine with amplitude of about one second. This could produce a relative location difference of about 3-4 km. A dip difference of 20° can produce a 4 second artifact that can lead to an uncertainty of up to 10-15 km. Larger differences in dip can lead to larger biases, so careful inspection of the observations is necessary for a successful utilization of R1 cross-correlation time measurements. When the dip is not 90°, small variations in the strike (10°) also produce similar minor cosine pattern perturbation with peak amplitude of less than one second. As the variation in strike increases to 20°, this causes the amplitude of the biased cosine perturbation to increase to nearly 5-second peak amplitude. Additionally, a cycle skip can occur across the 0° and 180° nodal points for each respective strike (Figure 4.3d).

Since variations of focal parameters can produce systematic biases in the relative travel times, we must consider their potential influence on the relative event locations. In all scenarios, the strong nodal effects result in a reduced correlation coefficient value (Figure 4.3b-d and 4.6a). Setting a high correlation threshold greatly reduces the influence of these nodal observations. Also, with good azimuthal coverage, the strong nodal effects are easily identified, and events with significant differences in focal parameters can be removed from the inversion. We discuss relocation sensitivity to the small cosine pattern perturbations at the end of the next section on the inversion procedure (section 4.3.3). We examined all the cross-correlation patterns used in the PFZ relocation. Of the 86 events analyzed in this study, five of the events displayed strong near-nodal effects or scatter in some of the event
pairs due to possible differences in focal parameters. Besides increasing confidence that effects from focal parameter differences have minor influence on our locations, this observation also indicates the Panama and Balboa Fracture zones have strong homogeneity in faulting geometry. Our observations are corroborated by the GCMT results where the majority of data have similar strike and dip; for over 700 km of fault, strike and dip appears to vary by less than ± 20° (Figure 4.3).

Although they complicate the location effort, the phase shifts near the R1 nodes may provide useful information on relative focal parameters. For example, the relative time-shifts between PFZ events occurring on 22 October 2009 and 30 March 2011 include strong near-nodal effects in the azimuth travel-time pattern. The GCMT catalog lists the 2009 and 2011 events as having strikes 354° and 5°, dips 82° and 83°, and rakes 179° and -180°, respectively. However, synthetic modeling of the phase shifts pattern using these focal parameters produces a significantly different azimuthal travel-time distribution than the observed. Analysis of the azimuth travel-time pattern agree with a difference in strikes similar to what is listed in the GCMT catalog, however, the splitting around the 0° node suggests a more significant difference in dip. By adjusting the relative dip of the events, we can roughly model the observed azimuth-travel-time pattern with a difference in dip of 10° (Figure 4.5). The calculation suggests that we may be able to extract more information regarding nearby events than just relative locations. But more work is needed to confirm this because working with near-nodal signals is always a challenge.

4.3.3 The Relocation Inversion Procedure

The inversion for relative event locations is performed using a spherical-earth version of the double-difference equations of Waldhauser and Ellsworth [15]. Double-difference methods were developed to integrate cross-correlation observations of local earthquakes [146] with catalog time picks, and the method we use is a member of a class of differencing algorithms widely employed in relative event location [147–149]. Rayleigh waves in the period range of interest have horizontal wavelengths on the order of 100-300 km, so they constrain the earthquake epicentroid, not the epicenter, and the centroid time [150], not the origin time. We use the term “epicentroid” to indicate the spatial average of seismic moment release in the period range from
about 80 to 30 seconds. In actuality, the difference between the epicenter and the epicentroid is within the uncertainty of the locations for the smaller events, but the two locations need not be the same for the larger events. Similarly, the difference between the centroid time and the origin time should be small for smaller events, but could be several seconds for the larger events. A nonlinear estimation problem is set up by relating the observed and predicted differences in surface-wave delay from all stations and events with the partial derivatives of those delays with respect to position and time. The partial derivatives for surface wave time delay with respect to colatitude, $\theta$, and longitude, $\zeta$, have a very simple form:

$$
\frac{\partial t_i}{\partial \theta} = \frac{\partial T(\Delta_i, z)}{\partial \Delta} \cos \zeta_i
$$

$$
\frac{\partial t_i}{\partial \phi} = -\frac{\partial T(\Delta_i, z)}{\partial \Delta} \sin \theta \sin \zeta_i
$$

where $t_i$ are the arrival times, $T(\Delta_i, z)$ is the travel time as a function of distance from epicenter, $\Delta$, and focal depth, $z$, and $\phi_i$ is the azimuth from the source to receiver [151]. Depth is ignored (as discussed earlier, all the events along the transforms are shallow), but it could be included if we were incorporating body wave times, which have greater relative depth sensitivity in their time shifts. The horizontal slowness is formally frequency dependent (the inverse of group velocity when integrating a range of frequencies); we use an average slowness suitable for the period range. Calculations with reasonable oceanic earth models shows that in the period range of 80 to 30 s the group velocities are relatively flat (and slightly faster than the phase velocity); this is not the case for shorter periods. To first order, we can use equation (4.5) to form a linearized inversion for the unknown centroid location and centroid time shift perturbations. Differencing the linearized equations for nearby events results in a set of equations for the relative positions of the earthquake epicentroids and centroid times as a function of the relative times between the events. We invert the equations with a singular value decomposition using LAPACK routines [152] and applied minimum length constraints on the correction vector [15]. Despite our efforts to include only the highest quality observations, the data set may retain some outliers. Thus, each iteration of the inversion includes three matrix inversions with iteratively reweighted observations to reduce the sensitivity to outliers. The iteratively adapted weights
were computed from the misfits - any misfit larger than 3.0 seconds was assigned a weight equal to \(3.0 / \delta t\) where \(\delta t\) is the time shift residual. Otherwise, the weight was unity. The cutoff at 3.0 seconds was chosen based on the fit of better data in early runs of the inversion. The algorithm produces a result that approaches an L1 optimization of the fit to the observed arrival time differences. The inversions described below all converged within a few iterations and remained stable with further iteration. In addition, we also weight the double-difference observations by the distance between the events so that observations from nearby events are more important in the misfit than observations for events separated by greater distance. The distance-weight decreases from 1 to 0 as distance ranges increases from 0 to 120 km. We start the inversion with the epicenters and origin times reported by the US Geological Survey’s (USGS) National Earthquake Information Center (NEIC) in the Preliminary Determination of Epicenters (PDE) catalog. We investigated using the EHB Catalog of Engdahl and others [125], but not all of the most recent or the older events are included in the EHB catalog. Also, while the EHB depths are more reliable than the NEIC PDE estimates, for this region, the epicenter scatter in the EHB catalog is roughly similar to that in the PDE catalog.

We examined the cross-correlation patterns for 86 events that exhibit similar faulting geometry based on their GCMT mechanism estimates. Most of the events are moderate size; the smallest is \(M_w\) 4.7 and the largest is about an \(M_w\) 6.5. Eighty-one of the events show robust correlation patterns and are included in the location analysis. The 81 events produce 79,080 double differences to constrain 243 unknowns. We explored the sensitivity of the locations to the assumed surface-wave slowness (see the electronic supplement). Our preferred solution corresponds to an average Rayleigh-wave slowness of 0.245 s/km (a 4.08 km/s speed). This speed is roughly consistent with the relatively simple oceanic lithospheric structure used in the numerical tests, but \(\sim\)5% lower than tomography studies in this region [153]. Global shear-wave speed models [154] smooth the structure laterally and vertically, resulting in lower speeds. Our primary preference for the assumed Rayleigh-wave slowness value is rooted in the fact that it minimized the differences in our locations and the independent locations estimates of the USGS.

The results of a three-iteration inversion are shown in Figure 4.6. Following the second iteration the maximum distance any event moved was insignificant. The initial and final double-difference misfit distributions are shown in Figure 4.7. The
inversion results in a reduction in the mean absolute misfit from 4.59 to 0.89 s (the reference is the PDE locations).

In addition to the mean near-source surface wave slowness, other parameters that must be specified to perform a double-difference location include the minimum acceptable cross-correlation value, the minimum number of links to link two events, and the maximum linking distance. We set the correlation coefficient threshold to 0.90 (cross-correlation normalized to 1.00), but analysis showed that the re-locations are very similar down to a coefficient of 0.75 when azimuthal coverage is good (as it is here). In the inversion, events were only linked if they are located within 120 km of each other and there are at least 12 observations satisfying the cross-correlation threshold criteria. A systematic investigation of the effects these assumptions have on the estimated locations is detailed in the electronic supplement. The combined sensitivity tests show that the overall pattern of the epicentroids is robust with respect to our assumptions.

We also tested the effect of differences in faulting geometry (section 4.3.2.3) on the relocation process using synthetic waveforms to locate clusters of events. For the tests, all the events had similar faulting parameters, except one “outlier” that varied in dip (±20°). When an event cluster includes six or more events, the outlier event is mis-located by ~20 km; but all other events are within 5 km accuracy with no uniform shift for all events. The misfit decreases for the similar events as the cluster size increases, but remains roughly the same for the outlier. For smaller clusters (<6 events), the misfit increases for the similar events, but decreases for the outlier (the cluster centroid is shifted). We repeated this test but with variance in the dip (±10°) and strike (±10°) in the outlier; results are similar for outliers with strike varying up to ±20°. The results are similar as earlier tests, except the misfit for the similar events was twice as high. These results indicate that as long as each event is linked to multiple other events, the locations are not strongly influenced by an outlier with modest differences in dip and strike. But the possibility of variations in faulting geometry should be considered for events that move large distances or when locating few events. In an oceanic transform fault setting, it is reasonable to assume the strike and dip would not vary significantly over a small region (Figure 4.3). We note, however, that this could be more problematic in regions of even modest changes in the faulting geometry.

Specific locations are sensitive to the assumed slowness value and to the choices
in linking different events and weighting the data to the point where changes in the epicentroids of a few kilometers are unresolved. We suggest a $\pm 5$ km uncertainty, but in some instances we believe that the relative locations are reliable to within just a few kilometers. This assumes that our screening procedure for substantial mechanism differences removed events that faulting geometry differences would bias relocations by more than a few kilometers, which we believe is true. Again, we credit the high-quality data that we have to explore in this region for the robust results, but we also think this is the case for many of oceanic transform faults.

### 4.4 Relative Relocation Results for the PFZ

Our preferred relocations are shown in Figures 4.3 and 4.8 and listed in Table C.1. Both the original NEIC and the relocated events may be subject to a regional bias (the entire pattern of events can be shifted). Differencing the travel times improves only the relative positions of the events. Improvements in absolute locations require either ground truth information to constrain the inversion, or the use of absolute travel times and knowledge of the regional seismic velocity structure. Unfortunately, we have no ground truth for any of the events in the available catalogs such as that at the International Seismic Centre (http://www.isc.ac.uk). Still, the overall centroid pattern inherited from the NEIC locations appears surprisingly good because the relocations of the linear features lie along the bathymetric signature of the PFZ and Balboa Fracture Zone (east of PFZ), despite including no direct constraints on the absolute locations.

Of the 81 relocated events, the distances between the initial epicenter and final epicentroid locations shown in Figure 4.6 range in size from about 0.3 to 66 km (Figure 4.9). The median distance shift is about 12 km. Absolute centroid time shifts ranged from 0.1 to 9.8 s with 0.1 s mean and zero median centroid time shift (a consequence of the relative nature of the observations). Five events analyzed in this study were not used in the relocation; we plot these five events (Figure 4.6) at the NEIC locations (open circles). Four of these events are the northernmost earthquakes, of which the waveforms were poorly correlated with all other events. These northern earthquakes occurred onshore and, according to the NEIC and GCMT catalogs, are likely deeper than the offshore events in this study ($> 27$ km depth). Tests with synthetic waveforms (section 4.3.2.1) suggest
that sources located in significantly different seismic velocity structures (affecting Rayleigh-wave excitation) can affect the estimated time shifts. The remaining event not included in the inversion showed significant scatter in correlations with other earthquakes, suggesting that this event is possibly affected by differences in faulting parameters. Figure 4.6 demonstrates that the structures illuminated by the relocated events are more spatially consistent and fluctuate less from a linear trend in their positions than the original NEIC locations. One area that does not collapse to a nearly perfect linear trend is that along the southern PFZ, between 4.5°-5°N. The improved locations suggest a curved structure, but whether this is a region of slightly more complex fault patterns or simply small variations in fault dip and or rake is hard to discern. We can only conclude that the events are located along a less uniform structure than the events to the north and the few located to the south.

Several events in the inversion moved tens of kilometers. The three events farthest east shifted to the linear trend of events immediately to the west of their original locations and likely associated with the Balboa Fracture Zone (BFZ). One of these two events was immediately preceded by an earlier event from the BFZ, which may have biased the original NEIC location (interference in seismic waves of the two events). We checked the event-to-event sinusoidal patterns for these two events that shifted the large distance with events along the BFZ. When the correlations were high, the patterns indicated that the true locations were in fact within the linear trend of the BFZ, which is where they were relocated. The 05 May 2005, Mw 6.5 earthquake originally located east of the linear trend around 5.75°N, shifted nearly 66 km to the southeast and had nearly a 10 s origin/centroid time shift. The GCMT solution for this event also contains a centroid to hypocenter time difference of 15.4 s and our final epicentroid location is quite close to that in the GCMT catalog. Finally, a review of the observed event-to-event cross-correlation patterns for this event confirms the large time shift and estimated relative epicentroid location.

The PFZ is better outlined as a relatively continuous feature containing a westward bend (or step) beginning near 6.25°N. Perhaps more impressive is the illumination of the BFZ located east of the PFZ between 6.5°N and 7.5°N. The structure is suggested by the NEIC locations, but less well defined (Figure 4.1); it becomes active near the westward bend in the PFZ. GCMT mechanisms (Figure 4.8) confirm that the BFZ is a right-lateral fault (as is the PFZ). The BFZ appears
to coalesce into the PFZ around 6.25°N. This geometry could be interpreted as a three-plate system with a triple junction at the intersection of the PFZ and BFZ, similar to the San Juan Bautista Crustal Triple Junction [155]. In this system, slip across the PFZ south of the junction is divided by the two northern faults. Asymmetric distribution of slip between these two faults would cause the triple junction to propagate through time. The occurrence of an \( M_w 5.8 \), east-west oriented normal faulting earthquake at this triple junction on June 26, 1992 (GCMT Catalog) suggests unequal slip rates between these two northern faults. Based on the GCMT catalog, the total seismic moment distribution along the faults since 1976 indicates that in recent decades, the BFZ is about three times more seismically active than the PFZ (over the latitude range from 6.25° to 8.0° N). This relationship remains the same if we exclude five non-strike-slip events (all of which are < \( M_w 6.5 \)) from the calculation. After removing the two largest events from each fracture zone, the BFZ is still about 2.5 times more active. The BFZ has also hosted about twice as many moderate-size events than the PFZ. Keeping in mind these values only represent relatively short catalog duration, it appears the BFZ may be accommodating substantially more slip than the PFZ. The sum of seismic moments along a similar length of PFZ south of the PFZ-BFZ split (4.5° to 6.25° N) is roughly the same as the sum of the two northern faults. The seismic coupling, the ratio of the seismic slip and the amount of slip from tectonic rate (\( \sim 62 \text{ mm/yr} \) [13]), for both the northern and southern regions is approximately 15%, assuming a 15 km seismogenic zone thickness and a shear modulus of 44 GPa. Much is unknown about the geometry of the seismogenic zone along transform faults, but 15% coupling is consistent with the global mean value calculated for oceanic transform faults [156]. Assessing the tectonics significance of these inferences requires consideration of the brevity of the data used to calculate the moment distribution along the structures. One or two large earthquakes along any boundary could substantially change the moment distributions. Combined, these structures have experienced five events \( M \geq 7.0 \) since 1900. The largest was \( M_s 7.4 \) on July 18, 1934 (8.14°N, -82.38°E), roughly north of the BFZ trend, and the last was \( M_s 7.1 \) on July 26, 1962 (7.50°N, -82.80°E) [19], which may have propagated southward into the region of little moderate-size activity along the northern segment of the PFZ.
4.5 Conclusions

Earthquake locations are fundamental to seismic and tectonic studies. Events that occur within dense seismic networks can commonly be precisely and accurately located. This scenario is typically limited to many continental earthquakes; events that occur offshore suffer from lack of local seismic stations, requiring location to be performed with distant stations. In this study, we have presented an alternative procedure using surface waves. Although they are slightly complicated by faulting geometry effects, displacements from surface waves are widely observed for moderate-to-large size shallow earthquakes and they have a significantly lower propagation speed, minimizing the effect of measurement error on location estimation. Our focus on strike slip events along the Panama and Balboa Fracture Zones displays the significant improvement in event location this method can provide. The initial locations for the region depicted diffuse seismicity, broadening to the north with no clear delineation between the PFZ and other parallel fracture systems. The improved relative locations provide a better outline of the PFZ as a relatively continuous linear feature, as well as delineate the BFZ, a parallel fault to the east of the PFZ between 6.5°N and 7.5°N. North of 6.5°N, the plate boundary motion is split across two roughly north-south PFZ and BFZ. For the last 36 years, slip along these two structures roughly matches the slip along the PFZ to the south (from 4.5°N to 6.25°N), but the BFZ has roughly three times the moment than the northern PFZ. In regions with good coherence in focal mechanism of seismicity, like oceanic fracture zones, this method provides a simple and efficient method of estimating precise centroid locations.
Figure 4.1. Example waveforms for three events from the Panama Fracture Zone observed at two different stations of significantly different distance (left side panels). The right side panels show the cosine fits to the observed surface-wave time differences between these events. The red dashed curve shows the fit predicted by the original USGS/NEIC location.
Figure 4.2. Seismicity along the Panama and Balboa fracture zones. The boxed region in the inset shows the study region relative to Central America. This figure shows epicenters of events $M > 4.5$ since 1973 listed in the NEIC catalog, scaled to magnitude. White stars identify the locations of all events $M \geq 7.0$ between 1900-1973 [19].
Figure 4.3. (a) The cross-correlation of vertical-component Rayleigh waves from a strike slip earthquake of strike = 0°, dip = 90°, and rake = 180° with a second co-located source, but with a dip ≠ 90° has phase described by the arctangent of cosecant as a function of azimuth (Eq. 4). (b) The effect of this phase distortion is observed using synthetic waveforms for two co-located events of similar faulting parameters in subplot (a) (strike = 0°, dip = 80° and 90°, and rake = 180°). Because the events are co-located, the time difference should be zero at all azimuths, but the phase distortion produced by the difference in dip produces time shifts most noticeable at 0 and π nodes. (c) Similar to subplot (b), except with similar dip (90°) and varying rake (170° and 180°), the strong phase effects occur at ± 90°. (d) While varying strike (dip = 90°, and rake = 180°) does not produce noticeable effects, when the dip ≠ 90° phase perturbations are produced. In all cases, the nodal effects are reflected by diminished cross-correlation coefficients.
Figure 4.4. Distribution of faulting parameters listed in the GCMT catalog for events used in this study. Number of observations are grouped into 10° bins.
a) Observed

Figure 4.5. (a)(top) Observed time shifts for two events (2009-10-22-00-51 and 2011-04-30-08-19) as a function of azimuth and (bottom) corresponding cross-correlation measurements. The dashed lines in the lower panel identify values of 0.75 and 0.90. The solid line shows a cosine fit to the observations. (b)(top) Time shifts using synthetic waveforms for two events with faulting parameters as listed in the GCMT catalog. The difference in dip of 1° does not account for the magnitude of phase effect observed (a). (bottom) Increasing the difference to 10° provides a closer representation of the observed data.
Figure 4.6. Red circles show the final epicentroids calculated using surface waves. Gray circles mark the original NEIC epicenters. Black lines show how each event location changed in the location process. Open circles mark the NEIC locations for events that were not located using surface waves due to inconsistent azimuth travel-time distributions.
Figure 4.7. (Initial and final misfits (observed double difference less the predicted double difference) for the inversion of observations from 81 vertical strike-slip earthquakes in the Panama Fracture zone region (Figure 4.6). Initial fit is based on USGS/NEIC epicenter locations and origin times.)
Figure 4.8. Global Centroid Moment Tensor (GCMT) focal mechanisms for this study’s events at relocated epicentroids. Symbol area is proportional to seismic moment.
Figure 4.9. Comparison of the epicentroid locations with the original US Geological Survey NEIC locations.
Chapter 5  
Precise Relative Earthquake Locations and Magnitudes of Seismicity in the Northeast Pacific

5.1 Abstract

Interplate seismicity offshore of southwestern Canada and the U.S. Pacific Northwest occurs along the boundaries of the Pacific, Juan de Fuca, and Gorda oceanic plates. The Blanco, Mendocino, Nootka, and Sovanco fracture zones host the majority of this seismicity, which largely produce strike-slip earthquakes. The Explorer, Juan de Fuca, and Gorda spreading ridges join these fracture zones, producing normal faulting earthquakes. Double difference methods applied to cross-correlation measured Rayleigh wave time shifts have been shown to be an effective tool at providing improved epicentroid locations and relative origin-time shifts in remote, oceanic transform regions. In this study, we investigate the possibility of extending the correlation of R1 waveforms from vertical strike-slip transform-fault earthquakes with those produced by normal faulting events from nearby ridges, providing improved relative locations of events along the entire plate-boundary system. We also demonstrate how cross-correlation coefficient values can be used to calculate precise relative event magnitudes. The cross-correlation coefficient may be used to model rupture directivity. The relatively dense seismic networks along the western North America coastal region enable precise location of smaller oceanic events than would otherwise be possible. Precise relative location
of events is vital to interpreting earthquake interactions and characterization of fault properties/behavior.

5.2 Introduction

Oceanic transform faults (OTF) constitute a large portion of the total global plate boundaries. Compared to subduction zones and continental strike-slip faults, OTFs generally do not pose significant seismogenic hazard; their maximum earthquake size is typically too small and the location is generally far enough removed with little tsunami hazard to greatly affect humans. However, compared to other tectonic boundary types, there are better constraints on OTF tectonic parameters (e.g. fault length, slip rate, and thermal structure) [156]. OTF surfaces have more homogeneous compositional structure and the thermal structure is more accurately predicted by plate kinematics [157–159].

Considering the length and linearity, the observed seismicity along oceanic transform faults (OTF) is relatively modest. Between 1976 and 2004, Boettcher and Jordan [156] identified only one earthquake definitely associated with an OTF greater than $M_w$ 7.0. For events with null axis between 70° and 90° and located near an identifiable oceanic transform fault, the Global Centroid Moment Tensor (GCMT) catalog [114] includes nine events $M_w \geq 7.0$ between 1976 and 2013, only 5 of which are $M_w > 7.0$. In their study on seismicity scaling relations for OTF, Boettcher and Jordan [156] found that the transform seismicity is on average well described by a truncated Gutenberg-Richter distribution with a self-similar slope of $\beta = 2/3$, which is in agreement with Bird et al. [160] and is consistent with global averages for all earthquakes [151].

Many studies have identified strong thermal controls on the maximum seismogenic zone depth extent for OTFs [24,144,161–164]). Specifically, seismicity along OTFs appears limited to temperatures less than 600°C (e.g. [24,144,164]). Modeling by Roland et al. [159] suggests that hydrothermal circulation can heavily influence OTF thermal structure and rheology. In their models, hydrothermal cooling causes thermal boundaries of the seismogenic zone to deepen more quickly near ridge-transform boundaries than previous half-space models suggested, and the seismogenic width remains fairly constant over the length of the fault. This thermal structure causes the seismogenic zone to deepen as isotherms are shifted downward.
This is particularly true for longer faults and slower slip rates. Alteration of mantle due to hydration may also extend the depth of brittle deformation [159].

The relative deficiency in cumulative seismic moment along OTFs has led many to conclude that OTFs generally have low seismic coupling (e.g. [165–168]). Bird et al. [160] noted the difficulty in identifying fault segments fully coupled because significant slip occurs aseismically along these faults. On average, OTF seismic coupling varies between 10% to 30%, however, specific fault zones vary significantly [156]. Studies find a decrease in coupling with spreading rate (e.g. [160, 169–171]). While the largest earthquake size and total moment release are functions of fault length and slip rates, they do not directly scale with total area above the 600°C isotherm. Possible models to explain this behavior include a thin seismogenic zone with uniform coupling along strike or a wide seismogenic zone with variable coupling along strike [156]. Boettcher and McGuire [172] suggest that the Blanco Ridge and Gofar transform fault display variable coupling along strike, producing multiple “patches” that only slip in single-mode seismic rupture. They conclude that the Blanco Ridge (the eastern portion of the Blanco Fracture Zone) includes two patches (centered 128°W and 127.7°W), which rupture independently with similar recurrence periods of 13.5 years.

Many have noted that slow earthquakes are common along OTFs (e.g. [173–176]). Some have observed similar OTF segments to experience multiple modes of rupture, including both infraseismic and ordinary (fast) rupture. Some studies contend to observe slow rupture initiating fast (e.g. [176–179]). Studies have also observed adjacent OTFs to display coupled seismic slip [178–180]. In one case, strike-slip swarms were observed nearly simultaneously on two parallel transform faults separated by ~25 km on the western boundary of the Easter microplate. In this instance, Forsyth et al. [180] determined neither static nor dynamic triggering could be occurring, suggesting the swarms were part of a larger, common creeping event. However, the idea of slow rupture initiating fast rupture remains controversial [144]. Additionally, results of McGuire [181] suggest rupture in the largest earthquakes only occurs seismically, lacking any slow slip processes.

In this study, we provide improved relative epicentroid locations and magnitudes for select oceanic transform fault earthquakes in the northeast Pacific region using relative Rayleigh wave time shifts. Multiple studies have relocated events in this region using different methodologies (e.g. [20, 22, 23, 25, 26]). The relocation method
we employ provides particular advantage in OTF settings over body-wave-based location methods. Surface waves are commonly more clearly observed than body-waves at stations of greater distance, increasing the number of usable stations for relocation. Additionally, surface waves exhibit reduced sensitivities to small changes in earth structure and measurement uncertainty. Interpreting the observed patterns and characteristics of these improved locations and magnitudes provides insight about the regional tectonics of the northeast Pacific region. In the Gorda and Explorer Plates, precise locations are important for identifying and analyzing internal deformation. Along the Blanco Fracture Zone, the improved locations assist in interpreting if slip is distributed along a single or multiple active faults. As will be discussed later for the Blanco Fracture Zone, the improved locations and magnitudes also enable investigation of fault properties; this includes analysis of how seismicity changes along the length the fault, as well as estimation of the width of the seismogenic zone.

5.3 Northeast Pacific OTF

This study focuses on OTF seismicity in the northeast Pacific, west of Cascadia subduction zone (Figure 5.1). This region includes seismicity associated with the interaction of the Pacific, Gorda, Juan de Fuca, Explorer, and North American Plates. The Gorda, Juan de Fuca, and Explorer are all remnants of the Farallon Plate, which began breaking up in the Tertiary [182–184]. Around 30 Ma, the Farallon-Pacific spreading center reached the Farallon-North America subduction zone, around which time the Farallon Plate split into the Juan de Fuca Plate in the north and Nazca-Cocos in the south [185]. As the North America-Juan de Fuca-Pacific triple junction propagated north over time, the Juan de Fuca Plate continued to decrease in size and fragment [186]. In the last 5 Ma, two segments of the Juan de Fuca are believed to have begun to move independently, forming the Gorda deformation zone in the south [187–189] and Explorer Plate in the north [190–193].
5.3.1 Gorda Plate

The Gorda Plate is young oceanic lithosphere (≤ 5 My) [25] that subducts beneath Oregon and California and is positioned south of the Juan de Fuca Plate. The Mendocino Fracture Zone (MFZ) forms the southern border and the Gorda Ridge (GR) the western. To the north is the Blanco Fracture Zone (BFZ). The GR and MFZ separate the Pacific Plate to the west and south from the Gorda Plate. Riddihough [187] describes the GR as including three sections, each divided by fracture zones. Spreading rates of these sections slow toward the south, with rates ranging from 5.5 to 2.5 cm/yr [187]. Observation of regional magnetic anomalies shows that while anomalies in the Pacific Plate are linear and align parallel with the Gorda Ridge, those in the Gorda Plate curve, “fanning” to the northeast, with anomalies parallel to the ridge in the north and trending northeast of the ridge in the south. Some [194,195] have noted the anomalies of the Gorda being shorter in length than those of the Pacific. Studies (e.g. [187,196]) have interpreted the differences in the anomalies to indicate internal deformation of the Gorda Plate. Carlson and Stoddard [197] attributes deformation of the Gorda Plate to plate reorganization at 5 Ma. Menard and Atwater [198] describes a clockwise rotation of the Gorda Plate relative to the Pacific Plate over the last 10 My.

Multiple models exist to explain the deformation of the Gorda Plate. In one model, the Plate behaves rigidly in the north and south but experiences right-lateral shear along NW-SE faults [199], causing deformation to be localized to the center of the Gorda Plate. In a different perspective, studies [195,197] propose the Gorda Plate experiences continuous deformation, described by Stoddard and Woods [25] as “flexural-slip buckling.” While this model would see the least amount of internal deformation in the north and south and most in the center, seismicity would be expected throughout the entire block, not only localized to the center and along discrete fault surfaces.

There exist differing perspectives on the adequacy of modeling the Gorda Plate as a single rigid body. Based on PDE and ISC earthquake locations, in addition to earthquakes occurring along the Gorda Ridge, there is significant seismicity within the Gorda Plate, further supporting interpretations of internal deformation. Sverdrup [200] contends mislocation of events in these catalogs makes the intraplate seismicity appear more active than it is in actuality. Sverdrup [200] relocated events
following methodology of Jordan and Sverdrup [148], finding much of the initial “internal” earthquake locations moved to the southwest, closer to the ridge. After relocation, the authors associate earthquakes whose locations remain within the plate to faults scarps and possibly rift mountains, interpreting this to support non-rigid Gorda Plate models with deformation occurring along pre-existing crustal weaknesses. Based on focal mechanisms [188], single-channel seismic data [195], surface expressions [188,201], and the aftershock sequence of the 08 November 1980 M 7.2 Eureka earthquake [189], studies identify northeast-southwest trending faults in the southeast corner of the Gorda Plate near the Mendocino triple junction believed to be resulting from the clockwise rotation of the southern part of the Gorda Plate [202].

5.3.2 Explorer Plate

The Explorer Plate is located north of the Juan de Fuca Plate, separated by the Nootka Fault Zone. The Dellwood Knolls, Revere-Dellwood Transform Fault, Explorer Ridge, and Sovanco Transform Fault all separate the Explorer from the Pacific Plate [203]. Like the Gorda Plate, and different from the Juan de Fuca, the Explorer Plate appears to experience significant intraplate seismicity. However, unlike the Gorda, magnetic lineations do not indicate internal deformation in the Explorer [203]. Studies (i.e. [192,193,204]) propose the Explorer only recently became an independent plate, separating from the Juan de Fuca up to 4 Ma in the creation of the left-lateral Nootka Transform Fault; an independent Explorer Plate is necessary since 4 Ma to account for the differences in orientation between the Explorer and Juan de Fuca [192,193].

Braunmiller and Nábělek [20] believe that the Explorer Plate is rotating clockwise, reducing convergence with North America. Prior to the Explorer-Juan de Fuca split, the Juan de Fuca is estimated to have been subducting beneath North America at >5 cm/yr at ~105° azimuth (at 50°N, -130°E). Following the breakup, the Explorer’s motion is believed to have slowed to 3-4 cm/yr and rotated to 120° azimuth [192]. Since that time, the Explorer’s convergence with North America has continued to slow and rotate to a present ≤2 cm/yr at 145° azimuth [20]. Riddihough [192] proposes the young age of the Explorer, <10 Ma after the breakup, may have made it more buoyant and consequently resistant to subduction, resulting
in the slowing and rotation of convergence. *Braunmüller and Nábělek* [20] consider reduction in slab pull forces, due to limited volume of subducted material, may have also contributed to the change in convergence rate. When the Explorer Plate became independent of the Juan de Fuca, spreading along the Explorer ridge was asymmetric, with more material accreted to the Explore than the Pacific [187,205]. *Braunmüller and Nábělek* [20] explain the asymmetric spreading accounts for the observed distance between Tuzo Wilson seamounts and Dellwood Knolls in the north. *Braunmüller and Nábělek* [20] calculated that the Explorer-Pacific spreading was about 4.5-6.0 cm/yr.

Due to the broad distribution of seismicity, the Sovanco Fracture Zone is a poorly delineated region. *Braunmüller and Nábělek* [20] describes the eastern Sovanco Fracture Zone extending far enough south to include the Heck seamount chain. *Dziak* [206] extends this zone further, proposing the Sovanco Fracture Zone is possibly associated with the formation of the series of seamounts south of this fracture zone. The region is currently composed of multiple rhombus shaped blocks bounded by faults [207,208]. *Botros and Johnson* [193] suggest the southern narrowing of the Brunhes anomaly along the Explorer ridges results from southward propagation of the Explorer ridge during this anomaly’s time. *Braunmüller and Nábělek* [20] point out this would require a period of counter-clockwise rotation of the Sovanco Fracture Zone, possibly resulting in the broken-up character we observe today.

The current tectonic setting of the Explorer Plate remains controversial. In one model, the Explorer is an independent plate bordered by the Pacific, Juan de Fuca, and North American Plates [191,192]. Based on analysis of seismicity and detailed bathymetry, *Rohr and Furlong* [204] proposed a different model, in which the eastern portion of the Explorer Plate is becoming coupled with North America, whereas the western is accreting to the Pacific. This model identifies the formation of a new transform zone representing the creation of a new Pacific-North American Plate boundary splitting the Explorer Plate. *Kreemer et al.* [209] further supports this interpretation, showing the slip vector orientations of abundant Explorer intraplate seismicity aligning closely with to the relative motion of North America and Pacific plates [210]. Additionally, this study refers to GPS measurements in northern Vancouver Island [211,212], which show crustal deformation similar to that expected by North American and Pacific plate relative motion. Inversion of
seismic strain distribution within the Explorer Plate also supports this model [209]. *Rohr and Furlong* [204] infers the new fault is being formed as a result of a southern lengthening of Queen Charlotte Fault. This would result in the extinction of the Explorer Ridge as the Pacific Plate captures it.

*Braunmiller and Nábělek* [20] interpret the seismicity patterns differently than *Rohr and Furlong* [204]. *Braunmiller and Nábělek* [20] believe that the lack of seismicity near the Southern Explorer Ridge and western Sovanco fracture zone is an indication that this region has become part of the Pacific plate. In the south, this model is very similar to that of *Rohr and Furlong* [204]; *Braunmiller and Nábělek* believe a new transform fault is forming, the Southwest Explorer transform boundary, separating the southwest corner from the Explorer Plate. In the north, however, *Braunmiller and Nábělek* do not recognize the formation of a new fault boundary and believe the Explorer continues to move independently of North America and the Pacific north of the Southwest Explorer transform boundary; the former study views it as being captured completely by the two adjacent plates. *Braunmiller and Nábělek* [20] points out that much of the seismicity considered to be intraplate by *Rohr and Furlong* [204] is actually an artifact of systematic mislocation to the northeast of regional seismicity. Earthquake relocation by *Braunmiller and Nábělek* shifted the location of much of the seismicity to the expected plate boundaries. In their study of earthquake slip vectors, *Braunmiller and Nábělek* found azimuths largely ranging between *Riddihough’s* [192] Pacific-Explorer motion (310°) and Pacific-North America motion (340°), with an average value of 323°. *Braunmiller and Nábělek* recognized their slip vector differ significantly from *Kreemer et al.* [209] and attribute this difference to mistakes in data selection made in the earlier study. This study also finds evidence for convergence between the Explorer and North American Plates in the seismicity beneath the Brooks peninsula [213] and margin morphology offshore and northwest of this peninsula.

### 5.3.3 Blanco Transform Fault Zone

The Blanco Transform Fault Zone (BTFZ) is a linear structure that extends ∼360km between the Gorda Ridge to the east and Juan de Fuca Ridge to the west. The BTFZ is commonly described as being composed of five transform fault segments, separated by extension basins. These basins include East and
West Blanco Depressions, Surveyor Depression, Cascadia Depression, and Gorda Depression [22, 214, 215]. Ibach [216] describes the BTFZ as a transpressional wrench fault.

The easternmost portion of the BTFZ is the Gorda Depression Segment. Located ∼20 km west of the Gorda-Blanco intersection, the Gorda Depression is 11 km wide (NE-SW) by 18 km long (NW-SE) and is 4400 m deep at the center. The depression formed as a result of extension between two strike-slip fault strands extending between the southern border and northern Gorda Ridge and the northern border and Cascadia Depression [217]. The earthquakes in this region are generally shallow (4-6 km) and small (M\(_w\) ≤ 5.3). Normal faulting is observed at the northern Gorda Ridge and within the Gorda Depression. Braunmiller and Nábělek [24] calculated 15% seismic coupling for this region. West of the Gorda Depression is the Cascadia Segment, including the Blanco Ridge and Cascadia Depression. In its eastern limit, the Blanco Ridge is the narrow, steep northern border of the Gorda Depression. Trending 111°, this ridge is composed of multiple diamond-shaped highs ranging in width of 3.5 km north of the Gorda Depression to 7 km between the Gorda and Cascadia Depressions. At ∼15 km, the Blanco Ridge is the longest uninterrupted ridge segment of the BTFZ. The geology of the Blanco Ridge is poorly understood and lacks volcanic features [217]. The age is believed to be 5 Ma [24].

Braunmiller and Nábělek [24] described the peak seismic slip rates of the BTFZ as occurring along the Blanco Ridge. The location of the largest recorded earthquake of the BTFZ occurred along the Blanco Ridge [218]. Multiple studies have estimated average seismic coupling value along the entire BTFZ (e.g. [24, 156, 219, 220]). Values range considerably, including 10% [156] to 36% [24] seismic coupling. These differences reflect this calculation’s dependence upon assumed seismic width (vertical extent of seismogenic region) and the assumed value of the shear rigidity. Braunmiller and Nábělek [24] identified a notable change in seismic behavior near 127.9°W. East of this longitude, the ridge is very seismically active, including six of the eight M\(_w\) ≥ 6 earthquakes recorded along the BTFZ [3, 66] and Braunmiller and Nábělek described this region as having full seismic coupling. West of 127.9°W, the former study described the seismicity rate and earthquake sizes as lower and calculate the seismic coupling as less than 50%. They associated the transition in seismicity at 127.9°W as aligning with the western edge of the abyssal hill that runs parallel to the Gorda Ridge and believe all large Blanco Ridge
earthquakes nucleate in this region. North of the Blanco Ridge is the Cascadia Channel. The Blanco Ridge terminates in the west along the southern border of the Cascadia Depression.

Normal faulting seismicity is observed between 128.5° to 129.2° W in the Cascadia and Surveyor Depressions [24]. The Surveyor Depression is located ~10 km northwest of the northwest corner of the Cascadia Depression. This western depression is a double basin, split by a small ridge, and is the shallowest and smallest of all BTFZ basins [217]. The seismicity along the Cascadia and Surveyor Segments is generally deep (6-9 km) and small (M_w ≤ 5.6). 

Braunmiller and Nábělek [24] calculated 15% seismic coupling in these regions. The T-axis of the faults and geologic features suggest the Surveyor Depression is a pull-apart deformation and the Cascadia Depression is a spreading center [24,217]. The Cascadia Depression is believed to have formed 5 Ma when plate motion changed from 100° to 110° E. Comparatively, the Surveyor Depression is significantly younger, 0.35 to 0.4 Ma [217].

The western portion of the BTFZ, the Western Blanco Segment, is structurally more complex than the eastern segments. Like the Surveyor Depression, this segment is believed to be younger, with reorganization 0.4 Ma [217]. Embley [217] included the Surveyor Depression in the Western Blanco Segment. In addition to the Surveyor Depression, this segment includes two basins, East Blanco Depression (EBD) and West Blanco Depression (WBD). The EBD is pull-apart, but the WBD is believed to be neither pull-apart nor a spreading ridge. Seismicity is largely strike-slip and shallow (4-6 km) [24]. 

Braunmiller and Nábělek [24] calculated 33% coupling in the Western Blanco Segment. In addition to the main fault along the north wall of the Western Blanco Depression, multiple transform fault locations are proposed for this region (e.g. [217,221–223]). Due to how close they are, Braunmiller and Nábělek [24] could not delineate these strands based on earthquake locations, however, noted that slip vectors support the presence of three strands.

5.4 Methods

Multiple studies have utilized surface wave observations for earthquake location (e.g. [18,40,126–131]). Having lower propagation velocities than body waves, surface waves can provide enhanced sensitivity to event location. Precise measurement of
relative surface-wave time shifts benefits from high waveform similarity. *Cleveland and Ammon* [18] showed that the necessary degree of similarity in faulting geometry and depth between events for oceanic transform fault tectonic settings is functionally broad. In this study we apply methods developed earlier (i.e. [18,40]), combining surface-wave time-shift observations in a multiple-event, double-difference based inversion, to estimate optimal oceanic transform fault earthquake epicentroid locations in the northeast Pacific region. We selected 125 out of 154 earthquakes based on a search of the Global CMT (GCMT) catalog (e.g. [99,114]) for vertical strike-slip earthquakes $M_w \geq 4.5$ since 1990. The remaining events were either located too far away from surrounding events or lacked a sufficient number of viable waveforms. We also supplemented this set with 36 normal faulting events. Most events in this selection have magnitudes between 4.5 and 6.5, with a mean of about $M_w$ 5.3; however, we include events as large as $M_w$ 7.2.

Following similar data processing procedures as described by *Cleveland and Ammon* [18], we focused only on short-arc Rayleigh waves. After acquiring long-period seismograms from operational stations at the time of each event from the IRIS data archive (Figure 5.2), we computed displacement seismograms using frequency-domain instrument-response deconvolution. We excluded most temporary networks, however, we included US Transportable Array data. Seismograms were graded from A to F by visual inspection based on the signal quality; only signals with quality C and better were used for our analysis. A group-velocity window (5.0 to 3.0 km/s) was used to isolate Rayleigh waves and we filtered signals to enhance the period range 80 to 30 seconds. We used cross-correlation to calculate the time difference for the surface waveforms between all signals. Using these time differences, we are able to invert for relative locations of the events using a spherical-earth version of the double-difference equations of *Waldhauser and Ellsworth* [15]. Further details of the inversion procedure can be found in *Cleveland and Ammon* [18].

Calculation of the locations requires definition of several parameters. Specifically, criteria must be defined to determine cross-correlation significance, event linking parameters, and average surface-wave slowness used to convert time-shifts to distance. The linking parameters include the maximum distance allowed between two linked events and the number of double-difference observations necessary to link two events. *Cleveland and Ammon* [18] provide sensitivity tests of these parameters; with the exception of the slowness value, we use the same parameter values as this
earlier study.

The misfit to the double-difference times is only slightly sensitive to average slowness values (within a range of reasonable values). We determined an optimal surface-wave slowness by using the value that minimized the average misfit between the revised and original, NEIC locations. For the Explorer and Gorda regions, a slowness of 0.26 s/km (3.85 km/s) provided the minimum difference from the original locations. Along the Blanco Fracture Zone, the optimal slowness was slightly higher, 0.28 s/km (3.57 km/s). In this region, the difference between slownesses of 0.26 and 0.28 s/km expanded/contracted the event locations along strike from the centroid of the cluster, with a maximum relocation difference of 7.9 km; this difference did not change the event distances normal to the fault trace. Compared with PREM and GDM52 [224], a slowness of 0.26 s/km is appropriate for all three regions. For these reasons, we use a slowness of 0.26 s/km for the entire dataset.

We also introduce a linking criteria based on the azimuth distribution of stations (something not used in Cleveland and Ammon [18]). Previously, we required at least 12 common stations with good correlation for two events to link; but if these stations are all located at a narrow azimuth region from both events, they will not provide enough information to sufficiently identify the variation of travel time with azimuth. In this study, we require the largest gap in station coverage is no larger than 310°, or in other words, we require at least 50° of more or less dense observations. Application of this criteria unlinked 8 events, which had very narrow azimuthal control (and thus very little control on the relative epicentroid locations of the two events). The specific value is imprecise; we selected the largest gap that would still provide enough data to reasonably constrain the expected azimuth travel-time cosine pattern. We find this value can be changed without significant influence on event relocations. Comparing improved locations calculated with a 50° limit and without any azimuth limits, the maximum difference in location is 0.78 km and the mean and median differences are 0.06 and 0.00 km.

Previous work by Cleveland and Ammon [18] focused only on improved locations of near vertical strike-slip earthquakes. As an extension to development of the relocation methods, we explore application of the method to linking strike-slip and normal faulting events and use of unnormalized cross-correlation coefficient to calculate relative event magnitudes.
5.4.1 Linking strike-slip and normal faulting events

Cleveland and Ammon [18] showed that cross-correlation measured R1 time shifts are sensitive to roughly $> 10^6$ differences in faulting geometry between two events. At smaller differences, the measured time shifts are uncontaminated by excitation differences resulting from faulting geometry variations. In larger oceanic ridge-transform fault systems, it would be useful to link successive transform fault strike-slip events through the use of the ridge, normal faulting earthquake. The strikes of these normal faulting events are generally nearly perpendicular to the slip events occurring along the fracture zones.

In a similar manner as Cleveland and Ammon [18], we analyzed the sensitivity of cross-correlation measured R1 time shifts to linking strike-slip and normal faulting events. We begin with the modal expansion expression for the displacement produced by a moment-tensor point source in a laterally homogeneous medium, presented by Aki and Richards [145] as:

$$u_z(k, h, R, \phi) = \sum_n A(k_n, z_0) \frac{e^{ik_n R}}{\sqrt{k_n R}}$$

$$\times \left\{ (M_{xy} + M_{yx})k_n r_1(k_n, h) \sin \phi \cos \phi 
+ \cos \phi \left[ M_{xx} k_n r_1(k_n, h) \cos \phi + i M_{xz} \frac{r_3(k_n, h)}{\mu(h)} \right]
+ \sin \phi \left[ M_{yy} k_n r_1(k_n, h) \sin \phi + i M_{yz} \frac{r_3(k_n, h)}{\mu(h)} \right]
+ \frac{dr_2(k_n, h)}{dz}\bigg|_h M_{zz} \right\}$$ (5.1)

which is expressed in terms of an amplitude factor that depends on the earth model, $A$, the mode number, $n$, wavenumber, $k_n$, and receiver depth, $z_0$. The propagation phase shift is described by the exponential term and depends on wavenumber and the distance from source to receiver, $R$. The bracketed term describes the effects of faulting depth and geometry. In this term, $M_{ij}$ is the moment tensor, $r_i$ is the eigenfunction, $h$ is source depth, $\phi$ is the azimuth measured clockwise from north, and $\mu$ is the shear modulus near the source. For a strike-slip earthquake of strike = $0^\circ$, dip = $90^\circ$, and rake = $0^\circ$, the fundamental mode ($n = 0$) term
simplifies to:

\[ u_z(k, h, R, \phi) = A(k_0, z_0) \frac{e^{ik_0R}}{\sqrt{k_0R}} \times 2M_{xy}k_0r_1(k, h) \sin \phi \cos \phi \quad (5.2) \]

For a second co-located normal faulting earthquake of strike = 90°, dip = 45°, and rake = -90°, the amplitude, but not the phase, changes relative to the strike-slip source. Moment tensor elements \( M_{xx} \) and \( M_{zz} \) become nonzero and \( M_{xy} \) becomes zero, but \( M_{xz} \) and \( M_{yz} \), which affect the phase, remain zero:

\[ u'_z(k, h, R, \phi) = A(k_0, z_0) \frac{e^{ik_0R}}{\sqrt{k_0R}} \times M_{xx}k_0r_1(k, h) \cos^2 \phi + \frac{dr_2(k_n, h)}{dz} \bigg|_{h} M_{zz} \quad (5.3) \]

Given these faulting parameters for the dips and rakes, only the amplitude (not the phase) is affected by variations in strike. As with comparing two strike-slip events, a phase shift is introduced if the dip or rake varies from 90° and 0° for the strike-slip event and 45° and -90° for the normal faulting because the \( M_{xz} \) and \( M_{yz} \) become nonzero. However, as discussed in Cleveland and Ammon [18], the introduced phase shift is generally isolated near the radiation pattern nodes and easily recognized when the differences are minor.

**5.4.2 Relative event magnitude**

When calculating the relative locations, we use the normalized cross-correlation coefficient as a criterion to determine if two events should be linked [18]. We can also use the unnormalized cross-correlation coefficient to calculate relative amplitude. Schaff and Richards [225] describe use of this methodology for two events. If we assume two identical events of different magnitude, \( x_1 \) and \( x_2 \), we can relate their signal linearly: \( x_1 = \alpha x_2 + n \), where \( n \) is uncorrelated noise. Consequently, we estimate a least-squares optimal value for the amplitude scaling factor, \( \alpha \), as:

\[ \alpha = \frac{x_1 \cdot x_2}{x_1 \cdot x_1} \quad (5.4) \]

where \( x_1 \) and \( x_2 \) are seismograms of two events. By inspection, we see \( \alpha \) is the unnormalized cross-correlation coefficient scaled by the dot product of the \( x_1 \) event.
As Schaff and Richards show, we can then define a relative magnitude as:

\[ \delta \text{mag}(CC) = \log \alpha \]  

(5.5)

\[ \log CC = \log x_1 - \log x_2 \]  

(5.6)

We can extend this procedure for a set of events observed at many stations. For example, for four events, \( x_i \), we can write this relationship as:

\[
\begin{bmatrix}
    y_{(1,2),1} \\
    y_{(1,2),2} \\
    y_{(1,3),1} \\
    y_{(1,4),1} \\
    y_{(1,4),2} \\
    y_{(2,3),1} \\
    y_{(2,3),2} \\
    y_{(2,3),3} \\
    y_{(2,4),1} \\
    y_{(3,4),1} \\
    y_{(3,4),2} \\
    \log(M_{\text{tot}})
\end{bmatrix} = \begin{bmatrix}
    -1 & 1 & 0 & 0 \\
    -1 & 1 & 0 & 0 \\
    -1 & 0 & 1 & 0 \\
    -1 & 0 & 0 & 1 \\
    -1 & 0 & 0 & 1 \\
    0 & -1 & 1 & 0 \\
    0 & -1 & 1 & 0 \\
    0 & -1 & 1 & 0 \\
    0 & -1 & 0 & 1 \\
    0 & 0 & -1 & 1 \\
    0 & 0 & -1 & 1 \\
    1 & 1 & 1 & 1
\end{bmatrix} \begin{bmatrix}
    \log(x_1) \\
    \log(x_2) \\
    \log(x_3) \\
    \log(x_4)
\end{bmatrix}
\]

(5.7)

where \( y_{(i,j),k} \) is the logarithm of the unnormalized cross-correlation coefficient for events \( i \) and \( j \) observed at station \( k \). We then invert for \( \log(x_i) \). We stabilize the inversion with \( M_{\text{tot}} \), by making the total moment equal to the total moment listed in the GCMT catalog for these events. While this makes the average relative moment equal to the average GCMT moment, this simply introduces a linear shift and the relative magnitudes are preserved. It is unlikely to have of uniform azimuthal coverage. In order to minimize possible azimuth-dependent factors that could influence the measurements, we weight the observations as a function of azimuth. We do this by sorting the measurements into 5° bins according to azimuth. We weight each bin evenly, consequently down-weighting individual measurements that fall into a bin with more measurements. While small, the azimuth bin weighting has an influence on the calculated values. Comparing relative magnitudes for a catalog with both strike-slip and normal faulting events calculated with a 5° bin width versus 10° width bins, the maximum magnitude difference is 0.08, but about
68% of the events differed by less than 0.01 magnitude units. In the binning tests, there is more scatter in difference for magnitudes smaller than $M_w 5.5$.

5.5 Relative Relocation and Magnitude Results for the Northeast Pacific

5.5.1 Earthquake relocations

Our improved locations are displayed in Figure 5.3 and listed in Table 5.1. Recall the nature of this relocation method is such that it provides relative locations. The relationship of event locations within the cluster of earthquakes is preserved, however, the entire cluster could be shifted. Consequently, if there is a regional bias in original event locations, as has been noted for events in this study region (e.g. [22, 24, 218]), the relocated events may inherit this static shift, however, the relationship between individual events is improved. Ground truth information could be used to acquire absolute locations. While there have been ocean bottom seismometer studies in this region (i.e. [226]), none are found including common events with this study.

As described earlier, in this study we link both strike-slip and normal faulting events together. Considering the acceptable range of focal parameters, we use the GCMT catalog to limit the included strike-slip events to those with null axis of $70^\circ$ to $90^\circ$ and the normal faulting events with null and tension axis of $0^\circ$ to $20^\circ$. One must also be concerned about the strike, particularly when linking strike-slip and normal faulting events. Inspection of the included events displays the normal faulting events have strikes nearly parallel or normal to the nearby strike-slip events. We inverted for locations using both a data set with only strike-slip events and a set supplemented with normal faulting events. There is negligible difference in the improved locations of the strike-slip events in each set, providing confidence that linking with normal faulting events does not adversely effect event locations. Additionally, we analyzed the azimuth travel-time pattern between normal fault and strike-slip events, and we do not observe appreciable azimuthal effects that would produce erroneous effect on the relocations. Azimuthal effects would be reflective of phase shifts resulting from moment tensor differences (Equation 5.1) that inaccurately influence event locations.
The distance between the initial epicenter and final epicentroid locations for all 161 events range from about 0 to 69 km, with a median shift of about 12.9 km (Figure 5.4). The absolute centroid time shifts for all events range from about 0.01 to 9.37 s; the mean and median are about 0.00 s and 0.05 s, respectively. Part of absolute centroid time shifts value reflects the difference between time of initiation of rupture and the centroid time, which can be large for larger events. For most of the events, distance and time changes do not correlate. However, the events with the largest distance offsets also had large time changes, but not necessarily vice versa. Separating the Explorer events from the remaining events south of 46° N, the northern events have a median distance shift of about 19.6 km and mean and median time shifts of 0.00 s and about 0.36 s. Similarly, the southern events have a median distance shift of about 11.2 km and mean and median time shifts of about 0.00 s and 0.03 s. As mentioned earlier, analysis of the azimuth travel-time patterns did not indicate any nodal effects large enough to influence the relocations. All the patterns were also generally very smooth. Because we limited the catalog to only include events with null and tension axis within a certain range as listed in the GCMT catalog, the lack of nodal effects largely indicates the strike of the events does not vary significantly on a 120 km scale.

As a whole, relocated events in the north moved a greater distance than those in the south. This difference is reflected in the two median distance shifts (∼11.2 km in the south and ∼19.6 km in the north). In the southern section, six events moved more than 25 km; twenty-eight move more than this distance in the north. The three events in the south that moved the farthest were all located along the Blanco (2012-08-08 17:45:33, Mw 4.9, moved 69.2 km; 2008-01-10 01:37:19, Mw 6.3, moved 54.6 km; and 1994-10-27 17:45:58, Mw 6.3, moved 47.6 km). The difference between epicenter and epicentroid could partly contribute to the distance moved in the two larger events, however, both the 2012 and 2008 events moved more inline with the rest of the events from their original locations. The azimuth travel-time patterns for all of these events are smooth and without inconsistencies that would suggest erroneous relocations. While the northern events had a higher median shift, only two moved more than 40 km (2008-09-11 23:26:19, Mw 5.2, moved 56.3 km; and 2001-09-14 04:45:08, Mw 6.0, moved 40.4 km). The 2008 events moved to where a cluster of seismicity was consolidated. The 2001 event is located in the region of diffuse seismicity of the Sovanco Fracture Zone. Like the events in the
south, the azimuth travel-time patterns support the accuracy of the relocations for these events.

Generally, the improved locations collapse seismicity into tighter clusters or more linear features. The improved locations for events along the Blanco generally concentrate the set of events into a more compact linear feature, however, there remains width to the linear trend; all of the events do not arrange into a single line. The western half of the Blanco appears more diffuse than the eastern. Normal faulting in the middle of the fault zone coalesces into a tighter cluster. Our catalog does not include any normal faulting events in the eastern half, while the western half has multiple. As mentioned earlier, several events moved large distances, however, this resulted in bringing these events more in-line with the linear trend. Our catalog of the Gorda seismicity is relatively sparse, and the relocations do not significantly influence the seismicity patterns, other than collapsing several of the events into a tighter cluster along the northern Gorda Ridge. The relocations of the northern events display both clustering and linear features. South of the Sovanco, two tight clusters of seismicity form around 47.75°N, -128.75°E and 48.25°N, -129°E. Through the middle of the Explorer, from around 49°N, -129°E to 49.5°N, -128.75°E, the initially diffuse seismicity coalesces into an arching linear feature. Southeast of this linear trend, seismicity remains diffuse. North of 49.5°N, seismicity becomes generally more linear, however, remains diffuse in a cluster centered around 51°N. North of this region, the seismicity appears to bulge to the west before bending back east and following the coast of Graham Island, British Columbia, Canada. North of 52°N, all the events in our catalog are normal faulting.

For those events in common, we compare our improved locations to those of other studies (Figures 5.5, 5.6, 5.7). In attempt to remove possible regional bias and achieve values closer to absolute locations, we can apply static shifts to the relative locations. Given the paucity of seismicity along the Juan de Fuca Ridge, seismicity around the Explorer Plate does not link with any Blanco or Gorda seismicity; as a result there are no constraints on the relative locations between these regions. In the north, we follow the method of Braunmüller and Nábělek [20] and fix the 06 April 1992 Mw 6.7 earthquake to the location carefully calculated by Cassidy and Rogers [21] (50.55°N, -130.46°E); Cassidy and Rogers located the 1992 event using P- and S-wave travel-time residuals and corrected for regional bias using Wahlström and Rogers [227]. Comparing our locations to those of Braunmüller
and Nábělek [20], we relocated 8 similar events with this earlier study (Figure 5.5, Table 5.2). Excluding the 06 April 1992 event, the seven remaining events differ in location between the two catalogs by about 3.6 to 37.5 km, with a median difference of 10.4 km; there is no systematic difference in azimuth. The biggest difference between the two catalogs (37.5 km) was the 07 April 1992 00:42:21 earthquake. The relocation of Braunmiller and Nábělek moved this event over 77 km southwest the NEIC location and over 30 km anomalously southwest of the closest relocated events. On the contrary, our location aligns the 1992 event with the pattern of the surrounding events, making us confident of our location. Other than the 1992-04-07 00:42:21, which Braunmiller and Nábělek [20] located well west of the other events, Braunmiller and Nábělek’s locations are generally in agreement with the trends that ours display.

In the south, we do not have a viable single event to use as an absolute location. Instead, we shift the Blanco events 30 km at 200° azimuth, a minimal amount that appears to best correlate with the bathymetry (Figure 5.6). Other studies have used locations calculated using T-phase data as absolute location. Considering the influence bathymetry local to the source can influence these locations, we do not feel this method is necessarily any more accurate. For this region, we include four other catalogs with relocations of common events as ours. Differences between catalogs vary; with the static shift applied, we find our locations provide the best agreement with the local bathymetry. The general patterns are similar as those represented in Dziak et al. [23] and Braunmiller and Nábělek [24]; however, our results locate the events along a narrower band of seismicity. The locations of Cronin and Sverdrup [22] appear biased to the northeast and again display broader scatter in locations without any strong agreement with local bathymetry. We do not apply a static shift to the events in and around the Gorda Plate because the locations appear to correlate well with the Mendocinco Fracture Zone and the Gorda Rise (Figure 5.7). The most comprehensive catalogs of earthquake relocations in this region predate the period of our catalog [25,26], so we cannot compare similar events, however, the general seismicity patterns appear to agree with our catalog.
5.5.2 Earthquake relative magnitudes

Our intermediate-period R1 magnitude estimates are compared with the GCMT values in Figure 5.8 and Table 5.1. Like the locations, because these are relative magnitudes, a static shift can be applied to the entire set. For these results, we set the mean magnitude to be equal to the mean magnitude of the GCMT catalog. Our magnitudes display a strong similarity with the GCMT’s at all magnitudes. The absolute differences between the two magnitudes range from 0.00 to 0.37 M$_w$ units. In the north (Explorer Plate region), the mean and median absolute differences are 0.06 and 0.05, respectively, and standard deviation of 0.01. At magnitudes above M$_w$ $\sim$ 5.4, the GCMT values generally lower than our relative magnitudes. Below M$_w$ $\sim$ 5.1, the opposite is true. In the south (Blanco and Gorda region), these values are 0.08 and 0.06 and standard deviation of 0.10. While not as apparent as the northern events, GCMT magnitudes for events M$_w$ $<$ 5.5 are generally slightly higher than our relative magnitudes. For all events, we do not observe any strong spatial correlation of magnitude difference between the GCMT and our relative magnitudes.

The use of correlation coefficients causes the calculation of relative magnitudes to be sensitive to differences in focal parameters. Similar to the azimuth travel-time patterns, the effect from these differences is easily identified in the azimuth correlation value patterns. In some instances, analysis of the azimuth correlation value pattern is more effective at identifying focal parameter differences than that of the azimuth travel-time patterns. The events with the largest differences were 28 October 2012 23:26:50 north of the Explorer Plate, and 15 June 2005 02:50:54 in the Gorda Plate. While we observe some nodal effects for the 2012 event with the 30 October 2012 02:49:02 earthquake, we do not observe factors that would erroneously influence the relative magnitudes of these events. Comparing a catalog with both strike-slip and normal faulting events with one of only strike-slip, the maximum relative magnitude difference is 0.09 magnitude units and the mean and median absolute differences are 0.02 and 0.01, respectively. The two events with the largest differences in magnitude were 2006-03-25 20:14:06 ($\delta$M = 0.09; GCMT: M$_w$ 5.0; 41.73°N, -126.30°E) and 2007-07-28 00:20:38 ($\delta$M = 0.08; GCMT: M$_w$ 5.0; 44.22°N, -130.07°E).
5.5.3 Tectonic Implications

The precise earthquake locations we present provide new insight into the tectonic settings of the northeast Pacific region. Specifically, we focus on seismicity in and around the Gorda and Explorer Plates and along the Blanco Fracture Zone. The improved locations display abundant intraplate seismicity in the Gorda and southern Explorer Plates, suggesting internal deformation. The band of seismicity along the BFZ narrows with our improved locations, however, indicates multiple, parallel, active faults. These improved locations provide detailed comparison between western and eastern BFZ and estimation of fault rupture dimension.

For the Gorda Plate seismicity, our improved locations do not alter the first-order patterns presented by the NEIC locations; while some locations move over 20 km, these do not significantly affect the seismicity patterns. The fact that some locations remain within the Gorda Plate supports interpretations of active internal deformation. While these improved locations cannot determine how the plate is deforming (i.e. rigidly or continuously), they do contribute to a growing catalog of improved locations necessary to make this interpretation.

Similar to the Gorda, the seismicity of the Explorer region remains complex with the improved locations. Between of 48.25°N to 49.5°N, the seismicity appears very diffuse, not aligning a single fault structure. This is not particularly surprising, as deformation of the Sovanco Fracture Zone has been interpreted as being very diffuse. The linear feature of seismicity in this region is not easily explained in any of the current tectonic models, suggesting northeast-southwest internal deformation of the Explorer Plate. North of this region, we observe a strong linear feature extending toward the Queen Charlotte Islands. These observations are consistent with models by both Braunmiller and Nábělek [20] and Rohr and Furlong [204]; however, they do not resolve any conflict between these two models. We also note the curving seismicity pattern just south of the Graham Island in the north. While the locations do not provide tectonic explanation for this pattern, they do appear to correlate with a notable change in the morphology of the continental margin.

The improved locations of seismicity narrows the width of seismicity along the BFZ. Introducing a static shift to the Blanco relocations, we observe a strong correlation between the improved locations and observed bathymetry. Additionally, these locations agree well with current interpretations of this well-studied fault zone.
While the width of the seismicity delineated fault zone is reduced, not all the events collapse to a single line, supporting interpretations of multiple active faults in the east as opposed to a single fault. However, seismicity of the western extent appears more scattered than the east, again agreeing with previous studies suggesting more active faults than the eastern half of the Blanco. We lack the precision to delineate specific fault traces, however, careful analysis of the measurements supports the observed fault zone width. Normal faulting events are tightly clustered in the middle of the fault, likely marking the Cascadia Depression. It is also interesting to note the relative abundance of normal faulting events in the west relative to the east.

While this is not a complete catalog, the stark difference may be indicative of the increased faulting complexity of the west relative to the east. Studies believe the western half of the Blanco is younger than the east (i.e. [217]). The difference in age and faulting complexity may indicate that as the fault ages, the fault zone narrows to fewer active parallel faults. Observing the seismicity through time (Figure 5.9), there is a noticeable difference in event magnitude and rupture behavior between western and eastern Blanco. The western half has ruptured in three \( M \geq 6.0 \) since 1965 while the east has experienced seven (NEIC). The east tends to rupture in single events, with little or no aftershock sequences. The west displays possibly repeating swarm-type rupture events between \(-130^\circ\)E to \(-129^\circ\)E. These differences in rupture behavior may also result from maturing of a single dominant fault surface, enabling larger single ruptures.

Observing original NEIC locations, there is a noticeable gap east and west of the Cascadia Depression (Figures 5.6 and 5.9). The relocation of the 27 October 1994 (\( M_w \) 6.27; 43.75°N, -127.92°E) and 10 January 2008 (\( M_w \) 6.22; 43.77°N, -127.94°E) epicentroids place them over 40 km west of their original location and significantly closer to the Cascadia Depression. The similarity in locations, rupture parameters, and magnitudes suggest the 2008 event may be a repeat of the 1994 event. Prior to 1994, there are two local \( M_w \) 6.4 (GCMT) events in 1981 and 1985 that may also be related. Two other large events occurred in 2000 (\( M_w \) 6.03) and 2012 (GCMT: \( M_w \) 6.0). This study relocated the 2000 event \( \sim 30 \) km to the east (43.65°N, -127.61°E). We did not relocate the 2012 event because the GCMT dip did not meet our criteria, however, preliminary relocation places it very near the 2000 event, suggesting it may be a repeat of the 2000 event. Based on the relocations, if the 1994/2008 events
ruptured the entire space between the Cascadia Depression and the 2000/2012 epicentroid, they could have ruptured up to \( \sim 60 \) km of the fault. Without our improved locations, the fault segment between -128.5°E to -127.75 would appear to not experience any large seismicity since 1981.

To test the estimation of rupture length, we use the azimuth correlation value plot of the 1994 event. In this plot, the measurements display a cosine pattern (Figure 5.10). This type of pattern is not representative of differences in focal parameters. We interpret this pattern to result from rupture directivity. To test the influence of directivity, we use fundamental-mode synthetic seismograms [139] of vertical strike-slip sources computed with a flat-layered, earth model adapted from the 0-20 Myr oceanic lithosphere model of Anderson and Regan [27]. While the synthetics are generated for a point source, we model rupture length by adding synthetic waveforms from a series of subfault sources spaced 1 km apart and shifted in time depending upon the rupture speed. Using these waveforms, we are able to model the observed cosine pattern.

The cosine pattern reflects several aspects of the rupture behavior. The phase of the cosine pattern is related to the strike of the source and direction of rupture; the maximum value aligns to the direction of rupture. Either increasing the length of the fault or decreasing the rupture speed increases the amplitude of the cosine pattern. Based on the observed pattern, we see the 1994 event ruptured to the northwest (Figure 5.10). The best fit to the data is an azimuth of \( \sim 282^\circ \); the GCMT calculates a strike of 296° and the very clear bathymetric features suggest a strike of \( \sim 300^\circ \). There is considerable scatter in the observed data, but the amplitude of the pattern appears to vary by \( \sim 30\text{-}40\% \); the L1 fit is 34%. We are confident the rupture was at the least very strongly unilateral; bilateral rupture results in a very unique pattern, with two downward dipping curves reflected about the zero azimuth. Using the synthetic waveforms, we are unable to model amplitudes close to what is observed for unilateral rupture lengths below 30 km. A length of 30 km requires a rupture speed near or less than 1 km/s, while a rupture length of 60 km, the entire width of the seismic "gap", requires speed of over 4 km/s.

Using a simple approximation, we can estimate the dimensions of the 1994 earthquake rupture plane. We calculate a relative moment of \( \sim 3.2\text{e}25 \) dyne-cm for this event. If we assume the 2008 event is a repeat of the 1994 and there was no slip between events, based on MORVEL [13], we can approximate the seismic slip is
∼64 cm (∼49 mm/yr). Using a shear modulus of 44.1 GPa [156], we approximate the rupture area to be ∼115 km². Assuming a unilateral rupture of 60 km causes the width of the fault to be ∼2 km; 30 km unilateral rupture makes the width ∼4 km. Studies suggest 600°C is an upper limit of the seismogenic zone in oceanic lithosphere (e.g. [144,159]). Based on mineral properties, Roland et al. [159] believe the 350°C is a lower temperature limit of the seismogenic zone. Thermal modeling by Roland et al. [159], predicts a near constant seismogenic width of ∼4 to 5 km along the portion of the fault ruptured by the 1994 event. Comparatively, a half-space cooling model would predict a seismogenic width that varies from nearly zero in the west to over 5 km in the east. Our calculations indicate the rupture width may be even narrower than these models predict. These results provide two important suggestions. First, it is possible to model rupture directivity using unnormalized correlation coefficient values. Additionally, oceanic transform fault earthquake rupture width appears to be very narrow. These ideas require further testing with more events at different transform faults.

5.6 Conclusions

Precise earthquake locations and magnitudes are important to our understanding of fault rupture behavior and geodynamic interpretations. We utilize cross-correlation of surface waves to calculate precise relative earthquake epicentroid locations in the northeast Pacific region. As an extension to previous research of this relocation method by Cleveland and Ammon [18], we demonstrate how this method can be used to jointly relocate strike-slip and normal faulting events; however, one must remain careful of possible influences from variations in rupture parameters.

Patterns observed in the distribution of seismicity can be used to interpret tectonic processes. Our relocations in and along the Explorer Plate demonstrate seismicity north of ∼49.5°N falls within a narrow band, as opposed to the initial more scattered locations. South of this latitude, seismicity within the Explorer remains scattered prior to our improved locations, however, our locations suggest a possible SSW-NNE linear structure of seismicity through the middle of the plate. Similarly, relocations within the Gorda Plate support observations indicating internal deformation.

Improved locations can also be used to understand rupture behavior and how slip
is distributed along a plate boundary. Our improved Blanco locations demonstrate the width of seismicity is more compact than initial locations. However, seismicity does not collapse to a single line, suggesting the presence of multiple active parallel faults along the fracture zone. There is also a noticeable difference between western and eastern Blanco; seismicity is more scattered and generally smaller magnitude (based on our catalog) in the west, possibly relating to this portion of the fault being younger.

Finally, we demonstrate how the data can be used for more than just calculating improved locations; we show how the unnormalized correlation coefficient may be used to calculate relative earthquake magnitudes. Our improved relative magnitudes show strong correlation with the GCMT magnitudes. For some larger events, we observe cosine patterns in the correlation coefficients values as a function of azimuth and demonstrate how these patterns may result from rupture directivity. Modeling rupture directivity to fit the observe patterns, we are able to approximate rupture length and speed; these parameters can be combined with other information to estimate the width of the seismogenic zone. Modeling directivity of the 27 October 1994 ($M_w$ 6.27; 43.75°N, -127.92°E) earthquake along the Blanco, we estimate a long rupture (30-60 km) with a narrow width, between 2-4 km wide.
Figure 5.1. Regional seismicity of the northeast Pacific. Shown are locations and magnitudes of all events $M \geq 4.0$ in the NEIC catalog (since 1973). Focal mechanisms from the GCMT catalog are also shown for all events available (also located and scaled according to NEIC values). Plate motion directions are calculated using MORVEL [13].
Figure 5.2. Distribution of stations from which observations were used in earthquake locations. Each station location is scaled to the number of observations used from that station. Most temporary networks were excluded, however, the US Transportable Array was used.
Figure 5.3. Relocated epicentroids from this study (red circle) compared to original NEIC locations (gray) labeled by the respective GCMT focal mechanism.
Figure 5.4. Comparison of the epicentroid locations with the original NEIC locations (subplots a and b) and origin time (subplots c and d) for events north (subplots a and c) and south (subplots b and d) of 46°N.
Figure 5.5. Explorer Plate relocated epicentroids from this study (red circle) compared to original NEIC locations (gray) and relocations from Braunmiller and Nábělek [20] (smaller, blue). Following the method of Braunmiller and Nábělek [20], we shift our relative relocations to fix the 06 April 1992 Mw 6.7 earthquake to the location carefully calculated by Cassidy and Rogers [21] (50.55°N, -130.46°E).
Figure 5.6. Blanco Fracture Zone relocated epicentroids from this study (red circle) compared to original NEIC locations (gray) and relocations from Cronin and Sverdrup [22] (blue), Dziak et al. [23] (green), Braunmiller and Nábělek [24] (purple), and Stoddard and Woods [25] (yellow). We shift our relative relocations a minimal amount that appears to best correlate with the bathymetry (30 km at 200° azimuth).
Figure 5.7. Gorda Plate relocated epicentroids from this study (red circle) compared to original NEIC locations (gray) and relocations from Cronin and Sverdrup [22] (blue), Dziak et al. [23] (green), Braunmiller and Nábělek [24] (purple), Stoddard and Woods [25] (yellow), and Sverdrup [26] (light blue). There is no shift in our relative relocations.
<table>
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Figure 5.8. Comparison of GCMT magnitude with relative magnitudes calculated in the study for events north (left) and south (right) of 46°N. Calculation using only vertical strike slip events (gray) and both strike slip and normal faulting events (red) show minor difference for common events, but 68% of the common events varied by less than 0.01 magnitude units.
Figure 5.9. Timeline plot of Blanco Fracture Zone seismicity since 1992. Shown are both relocated epicentroids from this study (red) and all NEIC epicenters M ≥ 4.0 (gray). Note differences in event locations are apparent, particularly for the large 1994, 2000, and 2008 events.
Figure 5.10. Unnormalized cross-correlation coefficient values (divided by the mean value) between the 27 October 1994 Mw 6.3 Blanco event and other surrounding events (labeled by color) at varying event to station azimuths. The observed cosine pattern (estimated with the gray, dashed line) results from rupture directivity of the 1994 event.
Table 5.1: Event relative relocations, shifted relocations and relative magnitudes.

<table>
<thead>
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<th>Epicentroid NEIC</th>
<th>Epicentroid GCMT</th>
<th>NEIC to Relocation</th>
<th>Focal Parameters GCMT</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\text{M}_w^1$</td>
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1 Calculated using the GCMT moment magnitude, $M_0$, and converted to $M_w$ using the equation:

$$M_w = \frac{2}{3} \times (\log M_0 - 16.1)$$
Table 5.2: Relocated events in common with earlier relocation studies.

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Chapter 6 | Directivity

6.1 Introduction

Important aspects of the earthquake behavior of oceanic transform faults (OTFs) remain poorly understood. Observation of cumulative seismic moment leads many to interpret OTFs as having on average low seismic coupling (e.g. [165–168]). Estimates of OTF seismic coupling vary significantly for specific fault zones, however, on average ranges between 10% to 30% [156]. However, because the width of the seismic zone of OTFs is generally poorly constrained, it is difficult to accurately assess the degree of seismic coupling [144]. Additionally, the rupture speed of earthquakes along these faults remains controversial.

Studies have identified the maximum depth of the seismogenic zone for OTFs is strongly controlled by temperature [24,144,161–164]. Specifically, seismicity along OTFs appears limited to temperatures less than 600°C (e.g. [24,144,164]). The detailed geometry of this 600°C isotherm varies with the thermal model. Using a half-space cooling model, Abercrombie and Ekström [144], predict the isotherm to have a broad arching shape, reaching maximum depth in the middle of the fault length and being pulled up to shallower depths at the ridges. Modeling by Roland et al. [159] have displayed hydrothermal circulation can heavily influence OTF thermal structure and rheology. This circulation produces hydrothermal cooling, which causes thermal boundaries of the seismogenic zone to deepen more quickly near ridge-transform boundaries than half-space models suggest. Consequently, the seismogenic width remains fairly constant over the length of the fault. Hydrothermal cooling can also cause the seismogenic zone to deepen due to isotherms shifting.
downward. This is particularly true for longer faults and slower slip rates [159], which may be consistent with studies that find a decrease in coupling with spreading rate (e.g. [160,169–171]).

Boettcher and Jordan [156] identify that, while maximum earthquake size and total moment release scale with fault length and slip rates, these measures do not directly scale with the total area above the 600°C isotherm. They propose multiple models that could account for this observation, including a thin seismogenic zone with uniform coupling along strike or a wide seismogenic zone with variable coupling along strike. Based on mineral properties, Roland et al. [159] believe the 350°C isotherm is a lower temperature limit of the seismogenic zone, which could provide support for a thin seismogenic zone model.

In addition to rupture width, the rupture speed of OTF earthquakes remains contentious. Some studies have noted that slow earthquakes are common along OTFs and that similar OTF segments experience multiple modes of rupture, including both infraseismic and ordinary (fast) rupture (e.g. [176–179,228]). Further, McGuire et al. [178] report slow rupture initiating fast. However, the idea of slow rupture along these faults remains controversial [119,144,229]. There have been multiple studies that refute reported observed slow slip [119] or have been unable to confirm the occurrence [134,230–232].

In this study, we seek to shed light on the question of OTF seismogenic zone earthquake width and rupture speed through analysis of fault length. For small faults, an earthquake source time function is well described as an impulse. However, for larger events, the entire fault does not rupture instantaneously. McGuire et al. [233,234] investigated 25 large, shallow earthquakes globally and found 80% ruptured unilaterally. Based on a simple unilateral rupture model (e.g. [235]), the observed rupture time of an earthquake will vary as a function of azimuth relative to the direction of rupture. The relationship of rupture time, $T_r$, with azimuth, $\theta$, is described by:

$$T_r = \frac{L}{V_r} - \frac{L \ast \cos\theta}{c}$$

(6.1)

where $L$ is the fault length, $V_r$ is the rupture speed, and $c$ is the wave speed [236]. In the direction of rupture, the apparent rupture time is shortest and away it is
longest. Perpendicular to the fault, $T_r$ is equal to the source time function because it is unaffected by rupture directivity. Additionally, moment is preserved in all azimuths, so as the pulse narrows, the amplitude increases and vice versa, making amplitude inversely related to pulse duration.

The variation of observed rupture time and amplitude with azimuth, called directivity, can be used to estimate rupture length and rupture speed if we can estimate slip. In this study, we use empirical Green’s Functions (eGF), small events located very near the large earthquake of interest, which we will call the “master” event. Using the Global Centroid Moment Tensor catalog (GCMT), the eGFs are selected to have similar focal mechanisms as the master. We compare the eGF and master event over a range of azimuths. The eGFs are small enough we can assume them to be point sources (no directivity), however, the larger, master event has an appreciable length and consequently observable difference in pulse duration and amplitude as a function of azimuth resulting from this length. Modeling these variations, we are able to estimate the rupture length and speed of the master event. Combined with information on slip, rupture length and speed help constrain the width of the seismogenic zone.

### 6.2 Surface-Wave Based Earthquake Directivity Measurements

#### 6.2.1 Seismogram Processing

We focused only on short-arc Rayleigh and Love waves. After acquiring long-period vertical and transverse seismograms from operational stations at the time of each event from the IRIS data archive, we computed displacement seismograms using frequency-domain instrument-response deconvolution. We excluded most temporary networks but included US Transportable Array data. A group-velocity window (5.0 to 2.75 km/s) was used to isolate Rayleigh and Love waves and we filtered signals to enhance the period range 80 to 30 seconds. We cross-correlated waveforms observed at common stations between the large event and selected surrounding events to
calculate the unnormalized correlation coefficient value, $CC$:

$$CC = \frac{\sum_{\tau=-\infty}^{\infty} b[\tau]a[\tau + t]}{\sum_{t=-\infty}^{\infty} a[t]^2}$$  \hspace{1cm} (6.2)$$

where $b(t)$ is the time series of the master earthquake and $a(t)$ is that of the eGF. In our analysis, we used the normalized coefficient value, which is similar to Equation (6.2), but instead of being divided by $\sum_{t=-\infty}^{\infty} a[t]^2$, it is represented by:

$$CC_{\text{norm}} = \frac{\sum_{\tau=-\infty}^{\infty} b[\tau]a[\tau + t]}{\sqrt{\sum_{t=-\infty}^{\infty} a[t]^2} \sqrt{\sum_{t=-\infty}^{\infty} b[t]^2}}$$  \hspace{1cm} (6.3)$$

We select the eGFs based on their size, proximity, and similarity in rupture parameters to the master event; the moment of the eGFs are at least one order of magnitude smaller (most $M_w \leq 6.0$) and no more than 100 km from the epicenter of the master. Because the eGF is small enough to be interpreted as an impulse source in the selected frequency band and assuming the rupture parameters and location (including depth) are similar between the eGF and master, any differences in correlation coefficient as a function of azimuth results from changes in the observed rupture time and amplitude of the master event. While the rupture time changes with azimuth as a cosine function (Equation (6.1)), because the correlation coefficient is a function of both rupture time and amplitude, the amplitudes of the observed pattern is not exactly the same as that represented in Equation (6.1). In order to compare multiple events, we scale the correlation coefficient by the ratio of the moments of the eGF to master and subtract the mean. For this step, we use the National Earthquake Information Center (NEIC) $M_w$ value and convert it to moment using:

$$M_w = \frac{2}{3}(\log M_0 - 16.1).$$  \hspace{1cm} (6.4)$$

We will refer to this value as $\Omega_{cc}$. Uncertainty in the ratio of moments will be mapped into the amplitudes we observe, making $\Omega_{cc}$ a function of directivity and moment ratio. We use synthetics waveforms to investigate the expected patterns.
and possible influences of depth and faulting-geometry. For analysis, we group the measurements by azimuth into 5° half-overlapping bins and calculate the median of each bin. We then fit an L1 least squares optimized cosine curve to the binned medians. As will be described in the next section, a cosine curve is not always an accurate representation of the data, however, it does accurately represent the azimuth with the largest amplitudes and generally produces a reasonable estimate of the amplitude.

6.2.2 Synthetic Waveform Modeling

In order to model the influence of directivity, we use fundamental-mode synthetic seismograms [139] of vertical strike-slip sources computed with a flat-layered, earth model adapted from the 0-20 Myr oceanic lithosphere model of Anderson and Regan [27]. While the synthetics are generated for a point source, we model rupture length by adding synthetic waveforms from a series of subfault sources spaced 0.5 km apart and shifted in time depending upon the rupture speed. We investigated fault lengths varying from 20 to 120 km and rupture speeds of 1.0, 1.33, 2.0, and 4.0 km/s. Due to the sampling rate, the highest rupture speed, 4.0 km/s, uses a subfault spacing of 1.0 km. For this study, we focus on a period range of 80 to 30 seconds. Shorter periods appear more sensitive to differences in sources, particularly depth, and generally display more complex azimuthal patterns and lower normalized correlation coefficient values.

When the master event and eGF have similar focal parameters, the cross-correlation coefficient varies smoothly with azimuth in a “cosine-like” pattern. This azimuthal pattern reflects several aspects of the rupture behavior. The maximum value identifies the direction of rupture because the amplitude of the master event relative to the smaller event is at a maximum in that direction. Either increasing the length of the fault or decreasing the rupture speed increases the amplitude of the cosine pattern. A bilateral rupture produces a very different pattern that is readily recognized – two downward dipping curves symmetric about the azimuth of the fault strike.

For short rupture lengths (> 30 km), the azimuthal pattern is modeled well as a cosine at all rupture speeds. At greater fault lengths, we begin to observe a flattening of the peak or trough and extension of the opposing trough/peak
depending upon the rupture speed (Figure 6.1). Below rupture speeds of 4 km/s, the trough of the pattern (away from rupture direction) becomes increasingly flattened at long rupture lengths (Figure 6.2). Additionally, at these speeds the peak becomes slightly increased. For speeds near 4 km/s, the values away from the direction of rupture display a systematic decrease in normalized coefficient value. The normalized coefficient is a measure of waveform likeness; this value indicates the master and eGF waveforms generally become less similar in the direction away from rupture. We also observe that the normalized coefficients generally become lower as the fault length increases and at lower rupture speeds (Figure 6.2). Ben-Menahem and Singh [237] display how directivity produces asymmetric P- and S-wave radiation patterns. In sub-shear conditions ($V_r < c$), this produces an elongation of the radiation pattern lobe in the direction of rupture. This could explain what is observed at rupture speeds less than 4 km/s; the master event amplitudes become asymmetrically larger in the direction of rupture. Comparatively, away from rupture, the amplitudes become more similar, causing a flattening of the correlation value. Based on Ben-Menahem and Singh’s [237] calculations, this effect is intensified as rupture speed nears the wave speed. For the earth model and period band (80 to 30 s) we use, the wave speed is just below 4 km/s. A rupture speed of 4 km/s is super-shear, which may cause a reversal of the effect just described; consequently this could cause amplitudes to become more similar in the direction of rupture, producing an observed flattening. However, this theory is not universally complete because at very large rupture lengths (e.g. 120 km), we observe flattening of both the peak and trough at 4 km/s rupture speed. While not investigated in detail here, these asymmetric patterns appear to be more pronounced at shorter periods. These observations suggest that given strong azimuthal coverage, it is possible to model rupture speed based on the observed flattening and asymmetry of the pattern.

We used synthetic seismograms to investigate the influence that differences in faulting parameters have on our measurements. Given two collocated events, the correlation coefficient is sensitive to differences in faulting geometry and produces a distinctive effect, which is very different from the cosine-like pattern produced by directivity. The effect of these differences is observed strongly at the radiation pattern nodes, where the correlation value suddenly reverses polarity. Because this pattern is very different from the cosine-like pattern resulting from directivity, its
influence would be easily identified. Additionally, there is a 45° shift between Love and Rayleigh wave radiation pattern nodes (both of which we use), which further enables identification of this effect.

A final consideration when cross-correlating the master and eGFs is the influence differences in depth may have on the correlation value. Keeping the master event at 12 km depth, we tested eGF depths of 5, 8, 12, 20, and 25 km. Increasing the depth of the eGF relative to the master causes the amplitude of the correlation coefficient versus azimuth pattern to increase. The observed displacement of surface waves decreases with depth [151], which causes the amplitude of the master relative to the eGF to increase. Only the scaling of the pattern amplitude is affected; phase remains unchanged. The change in amplitude between depths of 5 to 12 km is generally around 60%. This is enough to strongly affect the interpretation of rupture length and speed. However, Love and Rayleigh wave displacements decrease at different rates as a function of depth [238] making it possible to identify differences in depth through comparison of measurements of these two waves. Additionally, using multiple eGFs with the same master event would also provide a method of identifying depth differences. Again, this analysis focusses only on a period band of 80 to 30 s, shorter periods display increased sensitivity to differences in depth.

6.3 Global Strike-Slip Directivity Results

We analyzed 29 large, vertical strike-slip earthquakes occurring since 1980. Of those that provided enough station coverage to interpret patterns, 21 display noticeable directivity. In this study, we focus on 5 of these 21 because they provide multiple viable eGFs with strong station coverage (Tables 6.1 and 6.2). These five events are located in four regions: the Gulf of California (Figure 6.3), northern Mid-Atlantic Ridge (Figure 6.6), off the coast of the Pacific Northwest (Figures 6.8 and 6.10), and the Carlsberg Ridge (Figure 6.12).

6.3.1 Gulf of California

Two of the largest earthquakes to occur in the Gulf of California occurred on 03 August 2009 (NEIC: $M_w$ 6.9; 29.04°N, -112.90°E; 10.0 km) and 12 April 2012 (NEIC: $M_w$ 7.0; 28.70°N, -113.10°E; 13.0 km)(Figure 6.3). These two events appear to be
located on parallel fault segments of the plate boundary. Analysis of directivity measurements suggests the 2009 event ruptured to the northwest at about 320° azimuth (Figure 6.4). The the 2012 event appears to have ruptured in the opposite direction, about 145° (Figure 6.5). Both azimuths are in good agreement with the strikes in the GCMT (2009: strike = 131°; 2012: strike = 131°). Castro et al. [239] investigated the directivity of the 2009 earthquake using body-wave based epicenter relocations and fault mapping by Aragón-Arreola and Martín-Barajas [240]. They estimated the event ruptured a 50 km long and 12 km wide fault segment. Comparing the amplitude of the pattern we measure with model results using synthetic seismograms, we estimate a fault length of 50 km would require a rupture speed of approximately 2 km/s. We measured a slightly higher amplitude in the 2012 event, suggesting it either ruptured a greater length or at a lower rupture speed.

The measurements for the 2009 event display significant scatter between ∼325° to 360°. While this is located near a Rayleigh wave node, similar scatter is observed in the Love waves (not nodal), making us believe it is not a nodal effect. However, the Rayleigh waves appear to be consistently high, while the Love are low. Similar broad scatter is observed in the events located offshore of the Pacific Northwest in the direction of western United States. Analysis of other events not described in this study observed similar scatter in azimuths directed toward a similar region. These observations lead us to believe the scatter may result from propagation through western United States earth structure.

### 6.3.2 Northern Mid-Atlantic Ridge

We investigated a large earthquake rupturing along the northern Mid-Atlantic Ridge north of Azores Island occurring on 02 April 2003 (Mw 6.2; 35.21°N, -35.70°E; 12.4 km)(Figure 6.6). Despite being such a relatively small event, it displays clear directivity (Figure 6.7). Our measurements indicate directivity near 90° azimuth, which is compatible with the GCMT (strike = 280°) and results modeled by Pro et al. [241] (strike = 95°). Considering the numerous large prior events to the west of the 2003 event, it is reasonable for it to rupture to the east. Pro et al. [241] calculated a rupture speed ∼1.5 km/s and length of either 28 or 15 km using Rayleigh wave directivity or P-SH waveform modeling, respectively. Using our
modeling, these lengths seem too low to match the observed pattern, even for this slow rupture speed. We estimate this event would require at least 40 to 60 km length. At these lengths the speed could be up to nearly 2 km/s. Without good constraints on the amount of slip during the 2003 event, we are unable to estimate the width of the seismogenic zone.

6.3.3 Off the Coast of the Pacific Northwest

Abundant strike-slip seismicity occurs offshore of the North American Pacific Northwest. We focus on two events, one ruptured along the Blanco Fracture Zone on 27 October 1994 (Mw 6.3; 43.54°N, -127.54°E; 15.9 km) (Figure 6.8). This event was relocated in the previous chapter (Chapter 5). This study inferred the fault segment that ruptured in 1994 re-ruptured on 10 January 2008. There is a noticeable gap seismicity in this region, providing up to about 60 km of possible fault length. Similar to this earlier study, our measurements indicate the event ruptured at an azimuth ∼290° (Figure 6.9). This is consistent with the GCMT strike of 296°. The observed amplitude of the Ωcc pattern appears best fit with fault lengths of at least 60 km and rupture speed of around 2 km/s. While each azimuth displays a range of Ωcc values, generally the pattern is similar for all eGFs, but simply shifted vertically. We required the observation stations to be located at least ten degrees from the master event; closer stations display great variance between azimuths of ∼125° to 150°. As described for the Gulf of California events, we infer this scatter to be related to propagation through western United States.

As described in the former chapter (Chapter 5), we can use a simple approximation to estimate the dimensions of the 1994 earthquake rupture plane. In chapter 5, we calculated a relative moment of ∼3.2e25 dyne-cm for the 1994 event. Assuming the 2008 event is a repeat of the 1994 earthquake and there was no slip between events, based on MORVEL [13], a complete stress drop would include ∼64 cm (∼49 mm/yr) of the seismic slip. Using a shear modulus of 44.1 GPa [156] and Equation (6.4), we calculate the rupture area to be ∼115 km². Consequently, a unilateral rupture of 60 km causes the width of the fault to be ∼2 km. This estimate is compatible with thermal modeling by Roland et al. [159], who predicts a near constant seismogenic width of ∼4 to 5 km along the portion of the fault ruptured by the 1994 event.
Another large strike-slip event ruptured within the Gorda Plate on 15 June 2005 (M\text{w} 7.2; 41.20°N, -125.98°E; 9.9 km) (Figure 6.10). As described in Chapter 5, this region is unusual given the abundance of intraplate seismicity. The surrounding seismicity does not provide a clear indication of the strike of this rupture. Our measurements suggest the rupture propagated to the west-southwest at an azimuth of ∼250° (Figure 6.11). While different from the strike of 47° recorded by the GCMT, with more observations it is possible for the peak amplitude to shift closer to 225°. Interestingly, despite being nearly one magnitude greater than the 1994 Blanco event, the 2005 event displays only about 65% of the amplitude of directivity as the earlier event. This could be partially accounted in a difference of rupture speed, but given the significantly higher moment, this suggests a significantly wider fault zone and or seismic slip. A difference in down-dip dimension of the fault zone may be related to the 2005 event being intraplate while the 1994 was interplate.

6.3.4 Carlsberg Ridge

The 15 July 2003 (M\text{w} 7.5; -2.66°N, 68.33°E; 10.0 km) (Figure 6.12) Carlsberg Ridge earthquake is unique considering its size and location. Besides the 2003 event, the largest local seismicity were low M\text{s} 6s in the early 1980s hundreds of kilometers south. Additionally, looking at Figure 6.12, the 2003 event appears to have occurred at a location where a northeast striking fault strand intersects the north-northwest striking Carlsberg ridge. Seismicity on this northeast striking strand appears to only extend for a couple hundred kilometers. The GCMT catalog records the 2003 event as having strikes of 216° or 307°. Our measurements suggest a rupture direction of about 35° azimuth, suggesting it ruptured the strand off of the main Carlsberg ridge. Our measurements also display a high amplitude of directivity.

Using inversion of teleseismic P and SH waves to model the slip distribution, Antolik et al. [229] inferred the 2003 event ruptured the entire length of the strand toward the northeast, calculating a rupture length of about 200 km and width of over 30 km, however, noted the complexity of the rupture. The complexity of this rupture is also represented in the large non-double couple component of the focal mechanism (Figure 6.12). Bohnenstiehl et al. [242] estimated a similar fault length using aftershock data. While their fault dimensions estimates low average seismic slip (< 1 m), modeling by Antolik et al. [229] led them to conclude that the event
displayed minor slip in the first 100 km and most of the moment release occurred at around 175 km from the hypocenter. The large rupture width of Antolik et al. [229] is controversial considering it exceeds the 600°C isotherm popularly viewed as an upper limit to the seismogenic zone (e.g. [24,144,164]). Modeling by Antolik et al. [229] also estimated a rupture speed near 3.5 km/s. At a rupture speed near 4 km/s, our synthetic waveform modeling suggests the observed directivity amplitudes could be achieved in only around 100 km of rupture length. For our investigation, at these great fault lengths of the main event (> 120 km), waveform correlation with the eGF begins to suffer in our modeling, making it difficult to confidently estimate rupture lengths and speeds.

6.4 Conclusions

Multiple aspects of the rupture mechanics of oceanic transform faults are poorly understood. Specifically, it is unclear at what speed rupture propagates and how wide the seismogenic zone is along the fault surface. In this study, we present methods for analyzing rupture directivity using cross-correlation of surface waves. We apply the cross-correlation method to investigate large strike-slip events occurring in the Gulf of California, northern Mid-Atlantic, offshore of the Pacific Northwest, and along the Carlsberg Ridge. Through modeling fault ruptures with synthetic waveforms, it is possible to estimate rupture length and speed. We generally find appropriate rupture speeds to be near 2 km/s. The 15 July 2003 Carlsberg Ridge (M_w 7.5) event possibly had a higher rupture speed near 4 km/s, however, multiple complexities of this rupture make it difficult to confidently estimate the rupture speed and length using the cross-correlation method.

Rupture length, combined with the amount of seismogenic slip and the moment, can be used to estimate the width of the seismogenic zone. Using an estimated seismic slip value and the rupture length modeled in this study for the 27 October 1994 (M_w 6.3) Blanco Fracture Zone (BFZ) earthquake, we approximate the seismogenic zone along this portion of the BFZ to be about 2 km wide. Unfortunately, we lack good constraints on seismic slip for the other events investigated in this study, prohibiting estimation of seismogenic width in other fault systems. Rupture length estimates and magnitude of the 02 April 2003 (M_w 6.2) northern Mid-Atlantic Ridge earthquake are similar as the 1994 BFZ event. This may indicate the Mid-Atlantic
Ridge event also had a narrow rupture width, but we cannot be certain without a good estimate of seismic slip for the 2003 event. Comparing the 1994 BFZ event with the nearby 15 June 2005 (Mw 7.2) earthquake rupturing within the Gorda Plate, we observe the larger magnitude, intraplate Gorda Plate event displays smaller directivity effects than the interplate BFZ event. While more comparative examples need to be studied, this could reflect differences in the rupture behavior (e.g. seismogenic width, seismic slip) between inter- and intraplate seismicity.

In addition to providing a method of modeling rupture length and speed, this study has also highlighted multiple new applications of this cross-correlation method. Our sensitivity tests with synthetic waveforms suggests that the method we use is sensitive to differences in depth and focal parameters between the master and sub-fault. While consequently requiring careful application of this method, these sensitivities also present the possibility of exploiting additional information from the measured correlation values. Observation of correlation values also suggests possible surface wave scattering occurring to waveforms as they propagate through western United States.
Figure 6.1. The cross-correlation coefficient normalized by moment ratio, $\Omega_{cc}$, at varying azimuths using synthetic waveforms (band 30 to 80 s). The master events are 60 km long in both examples, but the rupture speed varies (top: 2.0 km/s, bottom: 4.0 km/s). At lower speeds, the direction away from rupture (the tough) becomes flattened; the opposite pattern is observed at higher speeds.
Figure 6.2. The cross-correlation coefficient normalized by moment ratio, $\Omega_{cc}$, at varying azimuths using synthetic waveforms (band 30 to 80 s) with master events of varying length. All master events have a rupture speed of 2.0 km/s. At shorter lengths, the pattern is well described by a cosine function (gray, dashed line), but as the length increases, the peak (direction of rupture) becomes increasingly amplified while the opposite direction becomes flattened. Additionally, as the length increases the waveforms become increasingly dissimilar, particularly in the direction away from rupture; this is indicated by the systematic decrease in normalized correlation coefficient.
Figure 6.3. Map of seismicity in the Gulf of California. All events in the NEIC catalog since 1973 are mapped; the focal mechanisms of events included in the GCMT catalog are plotted at the NEIC epicenters. All events are scaled to magnitude. The two events in blue are those investigated in this study; the northern is the 03 August 2009 $M_w$ 6.9 and the southern the 12 April 2012 $M_w$ 7.0.
Figure 6.4. Measured cross-correlation coefficients normalized by moment ratio, $\Omega_{cc}$, as a function of azimuth of the $M_w$ 6.9 03 August 2009 earthquake located in the Gulf of California compared to multiple local eGFs (color-coded). The median binned values are fit with an cosine function optimized using L1 least squares. These measurements indicate the rupture of the 2009 event propagated to about a $320^\circ$ azimuth.
Figure 6.5. Measured cross-correlation coefficients normalized by moment ratio, $\Omega_{cc}$, as a function of azimuth of the $M_w$ 7.0 12 April 2012 earthquake located in the Gulf of California compared to multiple local eGFs (color-coded). The median binned values are fit with an cosine function optimized using L1 least squares. These measurements indicate the rupture of the 2012 event propagated to about a $145^\circ$ azimuth.
Figure 6.6. Map of seismicity of a portion of the northern Mid-Atlantic Ridge. All events in the NEIC catalog since 1973 are mapped; the focal mechanisms of events included in the GCMT catalog are plotted at the NEIC epicenters. All events are scaled to magnitude. The event in blue is the 02 April 2003 $M_w$ 6.2 investigated in this study.
Figure 6.7. Measured cross-correlation coefficients normalized by moment ratio, $\Omega_{cc}$, as a function of azimuth of the $M_w$ 6.2 02 April 2003 earthquake located along the northern Mid-Atlantic Ridge compared to multiple local eGFs (color-coded). The median binned values are fit with a cosine function optimized using L1 least squares. These measurements indicate the rupture of the 2003 event propagated to about a 90° azimuth.
Figure 6.8. Map of seismicity Along the Blanco Fracture Zone. We display the GCMT focal mechanisms at the improved epicentroids described in Cleveland and Ammon [18] in dark red and blue. The event in blue is the 27 October 1994 M\textsubscript{w} 6.3 investigated in this study. The transparent, gray events include all events in the NEIC catalog since 1973 and the focal mechanisms of events included in the GCMT catalog plotted at the NEIC epicenters. All events are scaled to magnitude.
Figure 6.9. Measured cross-correlation coefficients normalized by moment ratio, $\Omega_{cc}$, as a function of azimuth of the $M_w 6.3$ 27 October 1994 earthquake located along the Blanco Fracture Zone compared to multiple local eGFs (color-coded). The median binned values are fit with an cosine function optimized using L1 least squares. These measurements indicate the rupture of the 1994 event propagated to about a $290^\circ$ azimuth.
Figure 6.10. Map of seismicity of the Gorda Plate. All events in the NEIC catalog since 1973 are mapped; the focal mechanisms of events included in the GCMT catalog are plotted at the NEIC epicenters. All events are scaled to magnitude. The events along the south delineate the Mendocino Fracture Zone; those along the west lie along the Gorda Ridge. The event in blue is the 15 June 2005 $M_w$ 7.2 investigated in this study and is located within the Gorda Plate.
Figure 6.11. Measured cross-correlation coefficients normalized by moment ratio, $\Omega_{cc}$, as a function of azimuth of the $M_w$ 7.2 15 June 2005 earthquake located in the Gorda Plate compared to multiple local eGFs (color-coded). The median binned values are fit with a cosine function optimized using L1 least squares. These measurements indicate the rupture of the 2005 event propagated to about a 250° azimuth.
Figure 6.12. Map of seismicity of a portion of the Carlsberg Ridge. All events in the NEIC catalog since 1973 are mapped; the focal mechanisms of events included in the GCMT catalog are plotted at the NEIC epicenters. All events are scaled to magnitude. The event in blue is the 15 July 2003 $M_w$ 7.5 investigated in this study.
Figure 6.13. Measured cross-correlation coefficients normalized by moment ratio, $\Omega_{cc}$, as a function of azimuth of the $M_w$ 7.5 15 July 2003 earthquake located along the Carlsberg Ridge compared to multiple local eGFs (color-coded). The median binned values are fit with an cosine function optimized using L1 least squares. These measurements indicate the rupture of the 2003 event propagated to about a 35° azimuth.
Table 6.1: Master (italics) and empirical Green's Function earthquakes used for measurements.

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G.Cali: Gulf of California
Atlantic: Mid-Atlantic Ridge
PNW: Pacific Northwest
Carlsberg: Carlsberg Ridge
Table 6.2: Directivity measurements based on an L1 least squares fit to the binned (5° overlapping bins) median $\Omega_{cc}$ value.

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Appendix A
Uniqueness and Commonality of Outer Rise Earthquakes: Southern Vanuatu, 16 May 1995 ($M_w$ 7.7)

A.1 Seismic Moment Distribution Estimation Methods

A.1.1 The Ammon Approach: The Search-Based Approach

The search-based model uses a search driven inversion of body waves and surface-wave derived source time functions to analyze the events. The inversion method utilizes teleseismic P and SH waves from stations composing a robust suite of azimuthal coverage. Station distances range from 30° to 90° to avoid contamination from PP and Pdif phases. Waveforms are bandpass filtered at 100 s to 5 s period and windowed to include the first 100 s after the first P wave arrival, cutting off before the PP and SS phase arrivals. We select only waveforms that enable accurate P wave first arrival identification. A similar approach is applied for SH waveforms. As noted by Hartzell and Heaton [243], inaccuracy as small as 1 s in picking of the first arrival can produce significant error in location of the rupture front. Given the inherent difficulty in accurately picking SH wave first arrival, these waveforms
are included for comparison to the predicted, however, their contribution to the model has been down-weighted as to not erroneously skew the model results.

The unknown model source parameters include the strike, dip, rake, rupture speed, and fault plane dimensions. A grid search based on minimization of the objective function is used to find strike and dip with rake either fixed or allowed to vary. We fix the target hypocenter based on the NEIC catalog location. The location of the fault plane relative to the hypocenter is variable, but is resolved from the distribution of slip calculated by the model. The fault dimensions are also found with similar methods. Iterations are run for rupture speeds between $V_r = 1.0$ to $3.5$ km/s. The model computes a body wave Green’s function for a simple 4 km thick ocean layer over a half space.

We divide the fault plane into a grid with a spacing of 3 km. The model is inherently a slip-pulse model. The slip at each grid cell (subfault) is a variable width half cosine wavelet. Starting with smooth perturbations, an initially blank model is perturbed randomly and the objective function is compared with the current best estimate of the model. If the perturbed model has a lower objective function than the current best estimate, it replaces the current best. The search continues for a fixed number of trials (but can be continued if the fit at that point is insufficient). The objective function is a combination of the waveform misfit, the model roughness, the model length, and a measure of the deviation of subfault rakes from the a priori rake. The rake, roughness, and length norms are weighted relative to the signal misfit. We begin with reconstruction of the smooth aspects of the moment (or slip) model first, and then increase the complexity of the model to improve the fit to the observations. Allowing the search to continue while slowly increasing the wavenumber of perturbations fine-tunes the smooth model to fit the more subtle features in the observed waveforms.

A.1.2 The Kikuchi-Kanamori Approach

Like the search-based model, the Kikuchi-Kanamori (KK) finite fault model utilizes a suite of broadband P and SH waveforms bandpass filtered at 50 to 10 s period. The waveforms are windowed to include 10 seconds before the P wave first arrival and 65 seconds after. This set of waveforms constitutes a full azimuthal distribution of stations between $30^\circ$ and $90^\circ$. As described with the search-based model, due
the limitations in accurate identification of SH first arrival, these waveforms are
down-weighted to have minimal influence on the inversion calculations. Also like
the search-based model, the defined fault plane is divided into a grid of subfaults.
For this model, we use a grid spacing of 5 km. Unlike the search-based algorithm
of the search-based model, the KK model applies a similar inversion method as
described in *Hartzell and Heaton* [243], solving an overdetermined system of linear
equations using a stabilized nonnegative linear-least squares method to invert for
the amount of slip along each subfault. A smoothing constraint is applied to
stabilize the solution. A grid search is used to find the optimal strike, dip, rake,
and rupture speed values. This optimization is based upon the variance between
the observed and predicted waveforms and the quality of the source time function
and slip distribution. The Green’s function is calculated using a simple three-layers
(including a 4 km thick water layer) over a half-space earth structure.

A fundamental difference between the Ammon and KK finite fault models is
the manner in which the subfault source time function is defined. While the search-
based model defines the subfault slip as a variable width single half cosine, the KK
model applies the time function as a set of overlapping triangles or set of adjacent
boxcar functions. Using triangles, the width, overlap, and number of triangles are
user defined. Our preferred method is the adjacent boxcars, in which the width
and number are boxcars are user defined. A smoothness constraint is also applied.
The optimal values for the subfault source time function variables is determined in
a similar grid search in which the fault parameters are determined. Each triangle
allows the rake to vary from the user specified value $\pm 45^\circ$. The subfault rake is
subsequently the average of the set of boxcars, and the total subfault slip is simply
the net slip of the boxcar set.
Appendix B
Large Earthquake Processes in the Northern Vanuatu Subduction Zone

B.1 Relocation

We used two methods to relocate the 1980 and 2009 events, including Rayleigh wave cross-correlation and body-wave arrival time based double difference. Body waves suffer somewhat from the short time interval between the 2009 earthquakes, which produces interference, particularly for the first two events. Using Rayleigh wave cross-correlation, we were able to measure the timing of the two later 2009 events relative to the first. The change in delay between the events as a function of station azimuth allows estimation of the relative centroid locations. The mean value of the cosine distribution indicates the second and third events occurred ∼15.75 and 70.50 minutes after the first, respectively. The centroid of the second event was approximately due north of the first, and the third was due south (Figure 3.7). Repeating this process for the 1980 events, which suffers from a severely limited azimuthal coverage of seismic stations, we estimate that the July 17 event centroid was located at an azimuth of ∼346° relative to the July 8 event centroid. The limited common station coverage between the 1980 and 2009 events causes the relative centroid location between these two sequences to be inconclusive. This method also proved inconclusive for relocation of the 1997 mainshock relative to the first 2009 event. We also relocated the 1997 and 2009 hypocenters using the
double difference method [15] (Figure 3.2). Events within 40 km of each other were linked but the relative travel time differences were also weighted as a function of distance. We used the USGS EDR P, Pn, and Pg phase picks. The relocated events from this procedure are not significantly different than the original. The primary effect was minor re-alignment of several scattered events along the dip into linear arrangements.

We also present relative relocations of 19 aftershocks following the 2013 event (Table B.1) using double difference. Time lags were measured using cross correlation of the R1 waves following methods described by Cleveland and Ammon [18]. Because data were filtered to emphasize the band between 80-30 seconds, we describe these locations as epicentroids, indicating the seismic moment spatial average. The difference between epicenter and epicentroid is minor for moderately sized events, but is noticeable for several of the largest events. This is particularly true in consideration of the time shift, reflecting of the difference between the origin and centroid time. The relocation events did not move significantly from their original locations. Several of the events moved to a slightly tighter clustering.

B.2 Finite Fault Modeling

As described in Section 3.6, using finite fault modeling, we produce the best fit to the P- and SH-waves with similar faulting parameters as the USGS W-Phase moment tensor solution \((\phi = 309^\circ, \delta = 17^\circ, \text{ and } \lambda = 58^\circ)\). This model uses the NEIC hypocenter \((10.738^\circ \text{S}, 165.138^\circ \text{E})\) with a depth of 18 km. We also tested the deeper source of 28.7 km as listed by the NEIC. These tests included both a dip similar to the GCMT and NEIC W-Phase moment tensor solutions \((\phi = 309^\circ, \delta = 17^\circ, \text{ and } \lambda = 58^\circ)\) (Figures B.3, B.4) and a dip like the NEIC centroid moment tensor solution \((\phi = 309^\circ, \delta = 30^\circ, \text{ and } \lambda = 57^\circ)\) (Figures B.5, B.5). Both of these models produce similar slip patterns as the shallower source, but with slightly poorer fit to the waveforms.

B.3 Kikuchi – Kanamori Inversion Modeling

Like the finite fault modeling for the 2013 event, we performed the Kikuchi – Kanamori inversion modeling with both a shallow and deep source for this earth-
quake. In section 3.6, we present the shallow source model because it provides the best fit to the waveforms (Figure 3.15). The deeper source produces generally similar results, however, the variable mechanism inversion displays more variability in the focal parameters of the subfaults (Figures B.10, B.11, B.12). Additionally, the total faulting mechanism of these subfaults has a very large non-double couple component, which is not observed in the actual event.
**Figure B.1.** Left panel: Seismicity of the northern Vanuatu subduction zone, displaying all NEIC earthquake hypocenters since 1973. The Australian plate subducts beneath the Pacific in nearly trench-orthogonal convergence along the Vanuatu subduction zone. Right panel: All GCMT moment tensor solutions for $M_w \geq 5$ since 1976, scaled with moment. This region experiences abundant moderate and large seismicity events, but lacks any great event ($> M_w 8$) since 1900. The largest events are displayed with dotted outlines of the magnitude-scaled circle. Convergence rates are calculated using the MORVEL model [13].

**Figure B.2.** All P and SH waveform fits for the finite fault modeling of the 2013 mainshock shown in Figure 3.14 (source depth = 18 km). The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 10.5 s prior to the P- and SH-wave arrivals.
Figure B.3. Finite fault model based on teleseismic P- and SH-wave inversion for the 2013 event using a kinematically constrained linear least-squares procedure. This shows the model for a deep source (28.7 km) with shallowly dipping plane (dip = 17°) (Figure B.4).
Figure B.4. All P and SH waveform fits for the finite fault modeling of the 2013 mainshock shown in Figure B.3 (source depth = 28.7 km, dip = 17°). The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 10.5 s prior to the P- and SH-wave arrivals.
Figure B.5. Finite fault model based on teleseismic P- and SH-wave inversion for the 2013 event using a kinematically constrained linear least-squares procedure. This model shows a deep source (28.7 km) with steeply dipping plane (dip = 30°) (Figure B.6).
Figure B.6. All P and SH waveform fits for the finite fault modeling of the 2013 mainshock shown in Figure B.5 (source depth = 28.7 km, dip = 30°). The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 10.5 s prior to the P- and SH-wave arrivals.

Figure B.7. Moment distribution for the finite fault for the 2013 earthquake for a shallow source (18 km) and shallowly dipping plane (dip = 17°) (Figure 3.14).
Figure B.8. All P and SH waveform fits for the Kikuchi-Kanamori fixed mechanism inversion modeling of the 2013 mainshock shown in Figure 3.15 (source depth = 18 km). The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 11 s prior to the P- and SH-wave arrivals.
**Figure B.9.** All P and SH waveform fits for the Kikuchi-Kanamori variable mechanism inversion modeling of the 2013 mainshock shown in Figure 3.15 (source depth = 18 km). The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 11 s prior to the P- and SH-wave arrivals.
Figure B.10. Kikuchi – Kanamori inversion models [17] for a fixed (left) (S11) and variable mechanism (right) (S12) of the 2013 mainshock. Both inversion models use a deep source (30km), similar to that used in the finite fault model (S5) and four subevents. Results for the fixed mechanism show the location of the subevents on the rupture plane. Results for the variable mechanism inversion shows the location of each subevent along strike. Below the subevents is the total mechanism for the entire event as a whole. The contribution of each subevent to the moment rate function is labeled.
Figure B.11. All P and SH waveform fits for the Kikuchi-Kanamori fixed mechanism inversion modeling of the 2013 mainshock shown in Figure B.10 (source depth = 30 km). The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 11 s prior to the P- and SH-wave arrivals.
Figure B.12. All P and SH waveform fits for the Kikuchi-Kanamori variable mechanism inversion modeling of the 2013 mainshock shown in Figure B.10 (source depth = 30 km). The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 11 s prior to the P- and SH-wave arrivals.
Figure B.13. All P waveform fits for the finite fault modeling of the 1997 mainshock shown in Figure 3.5. The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 11 s prior to the P-wave arrivals.
Figure B.14. All P waveform fits for the Kikuchi-Kanamori fixed mechanism inversion modeling of the 1997 mainshock shown in Figure 3.17. The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 11 s prior to the P-wave arrivals.
Figure B.15. All P waveform fits for the Kikuchi-Kanamori variable mechanism inversion modeling of the 1997 mainshock shown in Figure 3.17. The thick black lines show the observed waveforms, while the thin are the predicted for each station. Each set of waveforms is accompanied by the peak-to-peak amplitude in microns, station code, and azimuth (degrees). The timescale is in seconds, beginning 11 s prior to the P-wave arrivals.
Figure B.16. Mechanisms of all events in northern Vanuatu in the GCMT catalog. Events are located and scaled according to the GCMT catalog. The color indicates the centroid depth and the diameter is scaled to the moment. Convergence rates are calculated using the MORVEL model [13].
Figure B.17. Space-time plot (right) of all $M > 5.0$ seismicity in Vanuatu recorded in the NEIC catalog as a function of distance south of -9.5°N, 165.5°E. The figure on the left shows the location of the seismicity on a map rotated to make the trench oriented vertically.
Figure B.18. Seismicity of the Vanuatu subduction zone, displaying all NEIC earthquake hypocenters since 1973. The Australian plate subducts beneath the Pacific in nearly trench-orthogonal convergence along the Vanuatu subduction zone. This region experiences abundant moderate and large seismicity events, but lacks any great event (Mw > 8) since 1900. Convergence rates are calculated using the MORVEL model [13].
Table B.1: Relative relocations of a set of 2013 aftershocks using R1 waves

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1 National Earthquake Information Center catalog (NEIC)
2 Measured from NEIC hypocenter to this study’s relocated epicentroid
Appendix C
Precise Relative Earthquake Location Using Surface Waves

C.1 Introduction

In constructing the surface wave relocations, there are several choices that must be made, notably the criteria for considering a cross-correlation as significant, the distance across which to link two events, the minimum number of double-difference observations required to link two events, the double-difference weight criteria, and the average surface-wave slowness used to map time shifts to distances. The following sections describe our numerical sensitivity tests of these parameters. Our preferred values link events only when there are at least 12 double-differences that correlate at least as well as 0.90, a maximum event-to-event link distance of 120 km, and an average Rayleigh-wave slowness of 0.245 s/km. In general, we used three iterations in the inversion, since there was little change after that (the iterative inversion had converged). The tests were performed on a preliminary data set with most of the final 81 events that are used in the final inversion and are part of the interpretation. The extra five events do not affect the test conclusions significantly, with perhaps the exception of northernmost part of the study area, which is a region where the event correlations were weakest.

C.1.1 Event Linking Criteria – Cross Correlation Threshold

In the inversion, event locations are related to one another through a set of inter-event “links.” Events are linked based on the number of common, well-
correlated observations available to compute double-difference observations, and on the distance between the events. We explored the effects of the correlation threshold on the results using numerical tests before deciding on a threshold of 0.90. To illustrate the sensitivity of the cross-correlation threshold on an inversion of many events, we assumed a range of threshold values and performed several relocations. Our tests show that for this data set, the first-order results are relatively insensitive to correlation coefficient thresholds between 0.75 and 0.99. Example relocation results for three different cross correlation thresholds (0.75, 0.90, 0.99) are shown in Figure C.1. The results are robust for the threshold range from 0.75 to 0.90. In fact, the 0.75 and 0.90 results differ significantly for only a few events. The reason for the insensitivity in this case is the high-quality of the observations, and the use of an iteratively reweighted inversion procedure, which down-weights the outliers. An exceptionally high cross-correlation threshold (0.99, right panel) affects about one dozen events, particularly in the north, because at the greater threshold these events no longer have enough double-differences to be relocated.

C.1.2 Event Linking Criteria – Minimum Number of Links

The minimum number of double-differences required for event pairing is another variable to be defined. If the minimum is set too low then relative time patterns become highly sensitive to each link and possibly poorly constrained. Defining this variable is a balance between providing enough constraint on the relative time sinusoidal pattern without requiring too many, preventing events from being relocated. Location estimates assuming three different criteria are show in Figure C.2. In this study we required at least 12 links. Sparing a few outliers that suffer from a reduced number of links, the PFZ relocations are relatively insensitive to our choice. The robustness of the results is again a reflection of the high-quality observations available from the PFZ region. Most events simply are not near the minimum number required, so with the exception of a couple of events, our results are not affected the minimum number specified – the insensitivity in this case is clear, but not generally applicable to all data sets. This is particularly the case for studies of smaller events that will have fewer observations.
C.1.3 Event Linking Criteria – Maximum Distance Between Linked Events

The maximum distance over which to allow two events to be linked is yet another choice that we must make to set up the relocation problem. If two events are too far apart, then waveform travel paths from each event to a station may be too different for our simple model. Consequently, our procedure requires two events to be within a specified distance threshold to be linked. The shortest horizontal wavelength of the Rayleigh waves in this analysis is on the order of one hundred kilometers. Our tests show that any distance around one wavelength does not make any appreciable difference in the locations (Figure C.3). Still, the variation in slowness that may exist in the actual Earth is certainly going to increase the farther apart two sources are. One way to account for this is to weight the double differences in the inversion as a function of the distance between the events. This way, observations connecting closer events are more important. We use a weighting scheme that weights close events by unity and then decreases in equidistant steps from 0.75, 0.50, and 0.00 as the distance threshold is approached. We tested the influence of this weighting scheme and the results are shown in Figure C.4. The northern events are again most affected by introducing distance-dependent double-difference weighting. This could be due to poorer station azimuthal coverage for these events or greater dependence of their location on events at larger distances. Generally, variability in these inversion parameters produced only minor changes in the relocations and this effect was isolated to only several specific events.

C.1.4 Mean Surface-Wave Slowness Tests

Perhaps the most important assumption on the resulting locations is that of uniform Rayleigh-wave slowness across the region. In reality, the slowness varies laterally within Earth and with period, so the average is an approximation. We view the assumption of a uniform slowness as equivalent to assuming a spherically symmetric reference model typically used in body-wave location. The value of slowness that is appropriate depends on the local near-source structure and the bandwidth of the surface waves used in the measurements. We explored a range of values from 0.225 to 0.350 s/km during our tests. The tests show the intuitive result that the main sensitivity of the epicentroid locations to slowness is the expected dilation.
or contraction of the overall epicentroid location pattern about the centroid of a cluster of events. Higher slowness values cause the events to contract towards the event-cluster centroid, while lower slowness has the opposite effect. Results from three different slowness values are shown in Figure C.5. The red circles show the epicentroids for an assumed group slowness of 0.250 s/km (4.00 km/s), the gray circles show the locations for a lower slowness value of 0.225 s/km (4.44 km/s) and higher slowness value 0.300 s/km (3.33 km/s). This is an unusually large range of values. Examination of Figure C.5 shows that events further from the geometric centroid of the cluster of events are most affected by changes in slowness. The events most distant from the center of the overall distribution are moved by as much as 20 km towards or away depending on whether the slowness is decreased or increased. To choose an appropriate value for the Rayleigh-wave average slowness we computed synthetic Rayleigh waveforms for an ocean-lithosphere earth model (described in section 4.3.2) to a single station from a linear array of sources spaced 25 km apart and spanning 500 to 3,000 km from the station. We assumed a vertical strike-slip faulting geometry compatible with the observations, applied the same filters, and then cross-correlated the synthetic Rayleigh waveforms. Comparing the travel time as a function of distance shows a near constant slowness of ∼0.253 s/km (3.95 km/s), which roughly corresponds to the group velocity for oceanic lithosphere in the 80 to 30 s period band we employ. The group and phase velocity are roughly constant for oceanic structures in this band, hence the two dispersion values are roughly equal (the phase velocity was slightly lower than the group in this model) (Figure C.6). As mentioned in section 4.3.2, a slowness of 0.25 s/km is ∼5-7% lower than tomography studies of this region [153,244].

To determine an optimal choice for the mean group slowness, we tracked how far the epicentroids moved from the original NEIC epicenter locations for each assumed value of slowness that we tried in multiple inversions of the PFZ observations. A slowness value near 0.245 s/km was optimal in this regard, minimizing the differences in our locations and those of the USGS.
Figure C.1. Effect of varying the minimum acceptable correlation coefficient (normalized to 1.0) in location process. Red circles show the locations with the preferred value of 0.9. Gray circles show the locations with values of 0.75 (left) and 0.99 (right). Open circles show the original NEIC locations for events lacking the sufficient number of links to be located. The minimum number of observations required for a link was 12 and the maximum link distance was 120 km.
Figure C.2. Effect of varying the minimum number of links used in the location process. This variable defines how many surrounding events a specific event must be linked to in order to be located. Red circles show the locations with the preferred number of links of 12. Gray circles show the locations with links of 8 (left) and 24 (right). Open circles show the original NEIC locations for events lacking the sufficient number of links to be located. The largest effects result from a very high minimum link distance, causing more events to lack the sufficient number of links.
Figure C.3. Effect of varying maximum link distance in location process. Maximum link distance requires events to be no further than the designated value in order to be linked in the location process. Red circles show the locations with the preferred distance of 120 km. Gray circles show the locations with distances 100 km (left) and 150 km (right). Open circles show the original NEIC locations for events lacking the sufficient number of links to be located. Variation has only minor effects, largely affecting events located further from most other events.
Figure C.4. Effect of weighting by distance in the location process. If weighting is used, the influences of linked events at greater distances are down-weighted compared to closer links. Red circles show the locations using weighting. Gray circles show the locations without using distance weighting. Open circles show the original NEIC locations for events lacking the sufficient number of links to be located. Weighting has minimal effects, having the greatest effect on events in the north, showing a dependence on links of greater distances.
Figure C.5. Effect of varying slowness in location process. Red circles show the locations with slowness of 0.25 s/km. Gray circles show the locations with slowness of 0.22 s/km (left) and 0.29 s/km (right). Open circles show the original NEIC locations for events lacking the sufficient number of links to be located. Lines show the effect of changing the slowness between the preferred and variations. Decreasing slowness spreads the events away from the center of the cluster and higher slowness has the opposite effect.
Figure C.6. Phase and group velocities of the earth model used for the synthetic waveform tests. This model is an isotropic version of the 0-20 Myr oceanic lithosphere model of [27]. This model includes a 6.2 km thick oceanic crust and a 20 km thick seismic lid (LID). The shaded region highlights the period band used in this study.
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### Figures

**Figure C.7.** Different velocity models tested.

**Figure C.8.** Group and phase velocities of the different velocity models tested.
Figure C.9. Stack of synthetic waveforms for velocity models of varying crustal thickness. The highlighted region shows velocities ranging from 3.5 - 4.0 km/s.

Figure C.10. Dispersion of synthetic waveforms for velocity models with 6.15 km thick crust.
Table C.1: List of the 81 relocated events from this study.
At the bottom of the table are the 5 events analyzed in this study but not used in final locations. These events were removed from relocation due to poor correlation with surrounding events.

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*Fault rupture segmentation*  
PhD Minor: Computational Science, Penn State University  
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**HONORS AND AWARDS**

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Graduate Colloquium Award – 2nd in Poster Presentation, 2010; 2nd in Oral Presentation, 2013  
Anne C. Wilson Award, 2007  
**Los Alamos National Lab**  
Student Symposium – 1st in Geophysics Poster Presentation, Summer 2011  
**Skidmore College**  
Phi Beta Kappa Society – National Honors Society Member, 2006  
Mente et Malleo Research Grant, 2005  
**University of Cincinnati**  
Alpha Lambda Delta – National Honors Society Member, 2001  
Cincinnatus Scholarship – 2001

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**Los Alamos National Laboratory**, June-August 2011 – Seismology Internship  
**Skidmore College**, September 2005-May 2006 – Cosmogenic Isotope Laboratory, Lab Assistant

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**ABSTRACTS**

*Published first author abstracts from major conferences:*