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**Precise Relative Locations for Earthquakes
in the Northeast Pacific Region**

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1 **Abstract**

2 Double-difference methods applied to cross-correlation measured Rayleigh wave time
3 shifts are an effective tool to improve epicentroid locations and relative origin-time shifts
4 in remote regions. We apply these methods to seismicity offshore of southwestern
5 Canada and the U.S. Pacific Northwest, occurring along the boundaries of the Pacific and
6 Juan de Fuca (including the Explorer plate and Gorda block) plates. The Blanco,
7 Mendocino, Revere-Dellwood, Nootka, and Sovanco fracture zones host the majority of
8 this seismicity, largely consisting of strike-slip earthquakes. The Explorer, Juan de Fuca,
9 and Gorda spreading ridges join these fracture zones and host normal faulting
10 earthquakes. Our results show that at least the moderate-magnitude activity clusters along
11 fault strike, supporting suggestions of large variations in seismic coupling along oceanic
12 transform faults. Our improved relative locations corroborate earlier interpretations of the
13 internal deformation in the Explorer and Gorda plates. North of the Explorer Plate,
14 improved locations support models that propose northern extension of the Revere-
15 Dellwood fault. Relocations also support interpretations that favor multiple parallel active
16 faults along the Blanco Transform Fault Zone. Seismicity of the western half of the
17 Blanco appears more scattered and less collinear than the eastern half, possibly related to
18 fault maturity. We use azimuthal variations in the Rayleigh-wave cross-correlation
19 amplitude to detect and model rupture directivity for a moderate-size earthquake along
20 the eastern Blanco fault. The observations constrain the seismogenic zone geometry and
21 suggest a relatively narrow seismogenic zone width of 2 to 4 km.

1. Introduction

1.1 Global Oceanic Transform Faults

Although oceanic transform faults (OTF) constitute roughly 20% of the total global plate boundary length [Bird, 2003], their small maximum earthquake size, remoteness, and small tsunami hazard minimize their effect on humans. However, better constraints on OTF tectonic parameters (e.g. fault length, slip rate, and thermal structure) [Boettcher and Jordan, 2004] compared to other tectonic boundaries make OTFs a useful focus for the investigation of earthquake processes. In particular, OTF compositional structure is more homogeneous and OTF thermal structure is more accurately predicted than along other major plate boundaries [Turcotte and Schubert, 2002; Behn et al., 2007; Roland et al., 2010]. But important differences also exist between OTFs and other tectonic boundaries. Most OTFs appear to have relatively small maximum magnitude thresholds ($M < 7.0$) [e.g. Boettcher and Jordan, 2004]. Aftershocks associated with OTF earthquakes are generally fewer and smaller than those typical along other plate boundaries - the largest aftershock is usually two magnitude units smaller than the mainshock [Boettcher and Jordan, 2004]. However, transform seismicity is on average well described by a truncated Gutenberg-Richter distribution with a self-similar slope of $\beta = 2/3$ [Bird et al., 2002; Boettcher and Jordan, 2004], consistent with global averages [Stein and Wysession, 2003].

Many studies have identified strong thermal controls on the maximum seismogenic zone depth extent for OTFs [Wiens and Stein, 1983; Engeln et al., 1986; Bergman and

Solomon, 1988; Abercrombie and Ekström, 2001; Braunmiller and Nábělek, 2008; Boettcher *et al.*, 2007]. Specifically, seismicity along OTFs appears limited to temperatures less than 600°C (e.g. Abercrombie and Ekström, 2001; Braunmiller and Nábělek, 2008). Modeling by Roland *et al.* [2010] suggests that hydrothermal cooling causes thermal boundaries of the seismogenic zone to deepen more quickly near ridge-transform boundaries than half-space models suggest, and the seismogenic width remains fairly constant over the length of the fault, particularly for longer faults and slower slip rates. In addition, hydrothermal alteration of the mantle may extend the depth of brittle deformation [Roland *et al.*, 2010].

The relative deficiency in cumulative seismic moment along OTFs suggests that OTFs generally have low seismic coupling (e.g. Brune, 1968; Davies and Brune, 1971; Frohlich and Apperson, 1992; Okal and Langenhorst, 2000). Bird *et al.* [2002] noted the difficulty in identifying any fully-coupled fault segments because significant slip occurs aseismically along these faults. On average, OTF seismic coupling varies between 10% to 30%, however, coupling may vary significantly along strike [Boettcher and Jordan, 2004]. Earlier studies noted a decrease in coupling with spreading rate (e.g. Kawaski *et al.*, 1985, Sobolev and Rundquist, 1999; Bird *et al.*, 2002; Rundquist and Sobolev, 2002). Using estimated fault length and slip rate values, the observed rupture area of the largest OTF earthquakes is less than the total area above the inferred 600°C isotherm. Possible models to explain this behavior include a thin (limited depth extent) seismogenic zone with uniform coupling along strike or a wide seismogenic zone with variable coupling along strike [Boettcher and Jordan, 2004].

2. Northeast Pacific Oceanic Transform Faults

Our focus area, the northeast Pacific, west of Cascadia subduction zone, is shown in Figure 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 [Engebretson *et al.*, 1985; Lonsdale, 1991; Stock and Lee, 1994]. Around 30 Ma, the Farallon-Pacific spreading center reached the Farallon-North America subduction zone, and the Farallon split into the Juan de Fuca Plate in the north and Nazca-Cocos in the south [Stock and Molnar, 1988]. 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 to fragment [Atwater, 1989]. 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 [Riddihough, 1980; Wilson, 1986; 1989] and Explorer Plate in the north [Barr and Chase, 1974; Riddihough, 1977; 1984; Botros and Johnson, 1988].

In this study, we provide improved relative epicentroid locations and magnitudes for select OTF earthquakes in the northeast Pacific region using intermediate-period (30-80 s) relative Rayleigh wave cross-correlation measurements. While many others have used body waves to relocate events in this region using different methodologies (e.g. Sverdrup, 1987; Stoddard and Woods, 1990; Dziak *et al.*, 2000; Braunmiller and Nábělek, 2002; Cronin and Sverdrup, 2003), only limited work has been done in this region using surface waves [Boettcher and McGuire, 2009]. The relocation method we employ provides particular advantage in OTF settings. Surface waves are commonly

more clearly observed than body-waves at stations of greater distance, increasing the number of usable stations and expanding azimuthal coverage available for relocation. Additionally, surface waves are less sensitive to small changes in earth structure and measurement uncertainty.

Interpreting the observed patterns and characteristics of these improved locations provides insight into the regional tectonics of the northeast Pacific region. In the Gorda and Explorer Plate, precise locations are important for identifying and analyzing internal plate deformation. Along the BTFZ, the improved locations assist in interpreting whether slip is distributed along a single or multiple active faults. As will be discussed later for the BTFZ, the improved locations 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.

3. Methods

Multiple studies have utilized surface wave observations for earthquake location (e.g. von Seggern, 1972; Ekström, 2006; Ekström and Richards, 1994; VanDeMark, 2006; McGuire, 2008; Barmin *et al.*, 2011; Herrmann *et al.*, 2011; Wei *et al.*, 2012; Cleveland and Ammon, 2013). Because they have lower propagation speeds than body waves, surface waves can provide enhanced sensitivity to event location. Precise measurement of relative surface-wave time shifts benefits from the high waveform similarity for events with similar faulting geometry and depth. Cleveland and Ammon [2013] showed that the necessary degree of similarity in faulting geometry and depth between events for oceanic

transform fault tectonic settings is functionally broad in the intermediate-period
Rayleigh-wave band (30-80 s); small to moderate changes in depth and mechanisms do
not significantly affect the relative location accuracy. In this study, we apply methods
developed earlier (i.e. *VanDeMark*, 2006; *Cleveland and Ammon*, 2013), and combine
Rayleigh-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. As described in *Cleveland and Ammon* [2013], the term
“epicentroid” indicates the spatial average of seismic moment release sensed by signals in
the period range from about 80 to 30s. We selected 125 out of 154 earthquakes in the
region based on a search of the Global CMT (GCMT) catalog (e.g. *Dziewonski et al.*,
1981; *Ekström et al.*, 2012) 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 with acceptable signal-to-noise. We also
supplemented the strike-slip event set with 36 normal-faulting events (see electronic
supplement). Most events 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.

We used similar data processing procedures to those described by *Cleveland and Ammon*
[2013], and we focused only on short-arc Rayleigh waves (R1). We computed
displacement seismograms using a frequency-domain instrument-response
deconvolution. We excluded data from most temporary networks, but included US
Transportable Array data (Figure 2). Seismograms were graded from A to F by visual
inspection of 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 estimate the average time difference for the surface waveforms between corresponding signals from each event. Using these time differences, we invert for relative centroid latitude, longitude (an epicentroid), and time of the events using a spherical-earth version of the double-difference equations of *Waldhauser and Ellsworth* [2000]. Further details of the inversion procedure can be found in *Cleveland and Ammon* [2013].

Calculation of the locations requires specification of several parameters. Criteria must be defined to determine cross-correlation significance and event linking parameters, and average, regional 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* [2013] provide sensitivity tests of these parameters. With the exception of the slowness value, we use the same parameter values as this earlier study. While the correlation coefficient threshold can be set to values as low as 0.75 (cross-correlation normalized to 1.00) since we generally have good azimuthal coverage, we conservatively use a threshold of 0.90. Events were only linked for the inversion if they are located within 120 km of each other. We also introduce a linking criteria based on the azimuth distribution of stations (something not used in *Cleveland and Ammon* [2013]). Previously, we required at least 12 common stations with good correlation for two events to link; but if the stations are all located at a narrow azimuth region from both events, they will not provide sufficient information to characterize the variation of travel time with azimuth. In this study, we require that the largest gap in station coverage is no larger

than 310° . In other words, we require at least 50° span by the observations, which still provide enough data to reasonably constrain the azimuthal cosine pattern in relative arrival times produced by two nearby events. Application of this criterion unlinked eight events, which had very narrow azimuthal control (and thus very little control on the relative epicentroid locations of the two events). Changing the gap requirement does not significantly influence the event relocations we discuss. We compared locations calculated with a 50° limit and those without any azimuth limit and found a maximum difference in location of less than one km, and mean and median differences of 0.06 and 0.00 km. The primary effect is to remove the events that have an insufficient number of observations. Azimuthal spans significantly larger than 50° do not appreciably improve the relocation.

The misfit to the double-difference times is only slightly sensitive to average slowness values (within a range of reasonable values). We determined optimal surface-wave slowness by using the value that minimized the average misfit between the revised and original US Geological Survey 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 BTFZ, the optimal slowness was slightly higher, 0.28 s/km (3.57 km/s). The difference between slownesses of 0.26 and 0.28 s/km expanded/contracted the event locations along strike from the center of the epicentroid cluster. The mean and maximum relocation differences between the two slownesses were about one and six km, respectively; this difference did not change the event distances normal to the fault trace. Compared with values corresponding to PREM and model GDM52 [Ekström, 2011], a

slowness of 0.26 s/km is reasonable for all three regions. For these reasons, we use a slowness of 0.26 s/km for the entire dataset. Previous work by *Cleveland and Ammon* [2013] 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.

4. Relative Relocation Results for the Northeast Pacific

4.1 Earthquake Relocations

Our improved relative locations are displayed in Figure 3 and listed in Table S1. Our relative locations within each earthquake cluster are precise, but the entire cluster can be shifted. If there is a regional bias in the original event locations, as has been noted for events in this study region (e.g. *Dziak et al.*, 1991; *Cronin and Sverdrup*, 2003; *Braunmiller and Nábělek*, 2008), the relocated events inherit this bias. Ground truth information on one or more earthquakes could be used to compute absolute locations. While there have been ocean bottom seismometer studies in this region (i.e. *Sverdrup et al.*, 1985), unfortunately none include events common with this study. The event linking criteria separates the events into two groups, those in the Explorer Plate and those along and in the Blanco and Gorda Block.

We link both strike-slip and normal faulting events together, linking near vertical and 45°-dipping normal faults does not introduce a source-related phase shift when the strikes are parallel or orthogonal, allowing their cross-correlation to measure the locations-based

phase shift (see electronic supplement). Considering the acceptable range of focal parameters, we use the GCMT catalog to limit the included strike-slip events to those with null axis plunge of 70° to 90° and the normal faulting events with null and tension axis plunge of 0° to 20° . One must also be concerned about differences in strike, particularly when linking strike-slip and normal faulting events. Inspection of the included events showed that the normal faulting events have strikes nearly parallel or normal to the nearby strike-slip events. To insure that the mechanism differences do not affect the results, we estimated locations using only strike-slip events and using a combined data set including strike-slip and normal faulting events. Locations differences were minor (~ 5 km on average) between the inversions, providing confidence that linking with normal faulting events does not adversely effect event locations. Additionally, we visually analyzed all the azimuthal travel-time patterns corresponding to linked normal and strike-slip events, guided by the theoretical discussion and synthetic seismogram calculations. We did not observe appreciable azimuthal artifacts that would produce erroneous effect on the relocations. For all events, examination of the azimuth travel-time patterns did not indicate any nodal effects large enough to influence the relocations.

The distance between the initial epicenter and final epicentroid locations for all 161 events range from 0 to 69 km, with a median shift of about 12.9 km (Figure 4). The absolute value of the centroid time shifts for all events range from 0.01 to 9.37 s with a mean absolute shift of 2.08 s; the overall mean and median shifts are about 0.00 s and 0.05 s, respectively. Part of the centroid time shifts reflects the difference between time of initiation of rupture and the centroid time, which can (and should) 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. The Explorer events (those north of 46°N) have a median distance shift of 19.6 km and mean and median time shifts of 0.00 s and 0.36 s. The southern events have a median distance shift of 11.2 km and mean and median time shifts of 0.00 s and 0.03 s.

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 region, six events moved more than 25 km; twenty-eight moved more than this distance in the north. The three events in the south that moved the farthest were all located along the BTFZ (08 Aug 2012 17:45:33, M_w 4.9, moved 69.2 km; 10 Jan 2008 01:37:19, M_w 6.3, moved 54.6 km; and 27 Oct 1994 17:45:58, M_w 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 are more collinear with nearby events than their original locations. While the northern events had a higher median shift, only two events moved more than 40 km (11 Sept 2008 23:26:19, M_w 5.2, moved 56.3 km; and 14 Sept 2001 04:45:08, M_w 6.0, moved 40.4 km). The 2008 event moved to where a cluster of seismicity was consolidated by the relocation. The 2001 event is located in the region of diffuse seismicity of the Sovanco Fracture Zone. As with the events in the south, visual inspection of the observed azimuth travel-time patterns support the relocations for these events.

For those events in common, we compare our locations to those of other studies (Figures 5-7). To facilitate comparison, reduce possible regional bias in our work, and achieve values closer to absolute locations, we applied static shifts to our relative locations. Given the paucity of seismicity along the Juan de Fuca Ridge, seismicity around the Explorer Plate does not link with any BTFZ 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 *Braunmiller and Nábělek* [2002] and fix the 06 April 1992 M_w 6.7 earthquake to the location carefully calculated by *Cassidy and Rogers* [1995] (50.55°N , -130.46°E); *Cassidy and Rogers* [1995] located the 1992 event using P - and S -wave travel-time residuals and corrected for regional bias using *Wahlström and Rogers* [1992]. Comparing our locations to those of *Braunmiller and Nábělek* [2002], we relocated eight events contained in this earlier study (Figure 5, Table S2). Excluding the 06 April 1992 event, the seven remaining events differ in location between the two catalogs by ~ 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 of the NEIC location and over 30 km southwest of the closest relocated events. Our location aligns the 1992 event with the pattern of the surrounding events, providing confidence in our location. Other than the 07 April 1992 00:42:21 event, which *Braunmiller and Nábělek* [2002] located well west of the other events, *Braunmiller and Nábělek's* locations are generally in agreement with the trends that ours display.

262 In the south, we do not have a viable single event to use as an absolute reference location.
263 Instead, we shift our BTFZ events 30 km at 200° azimuth, a minimal amount that appears
264 to best correlate the seismic trend with the bathymetry (Figure 6). Similar shifts were
265 applied by *Cronin and Sverdrup* [2003] (28 km at 186°) and *Braunmiller and*
266 *Nábělek* [2008] (25 km at 212°). Other studies have used locations calculated using T-
267 phase data to provide reference locations. Considering the influence bathymetry local to
268 the source can have on these locations, we do not feel this method is necessarily any
269 more accurate. For this region, we found four other catalogs with relocations of common
270 events as ours. With the static shift applied, we find our locations provide the best
271 agreement with the local variations in bathymetry. The general patterns are similar as
272 those represented in *Dziak et al.* [2000] and *Braunmiller and Nábělek* [2008]; however,
273 our observations locate the events along about half as narrow a band of seismicity than
274 *Braunmiller and Nábělek* [2008]. The locations of *Cronin and Sverdrup* [2003] again
275 display broader scatter in locations and only a general agreement with local bathymetry.
276 We did not apply a static shift to the events in and around the Gorda block because the
277 locations appear to correlate well with the Mendocino Transform Fault in the north-south
278 direction and are about 20 km shifted east relative to the Gorda Rise (Figure 7). The
279 azimuthal orientation of our normal faulting earthquake locations matches the
280 bathymetric trend defining the Gorda Rise. The most comprehensive catalogs of
281 earthquake relocations in this region predate the period of our catalog [*Stoddard and*
282 *Woods*, 1990; *Sverdrup*, 1987], so we cannot compare similar events, however, the
283 general seismicity patterns in these earlier studies appear to agree with our results. The

following sections provide a more detailed analysis of seismicity of the Gorda Block, Blanco Transform Fault, and Explorer Plate.

4.2 The Gorda Block

The Gorda block is young oceanic lithosphere (≤ 5 Ma) [Stoddard and Woods, 1990] that subducts beneath Oregon and California. The Mendocino Transform Fault (MTF) provides the southern border and the Gorda Ridge (GR) the western. To the north is the BTFZ. The GR and MTF separate the Pacific Plate to the west and south from the Gorda block. In one model for the region, Riddihough [1980] identified three GR sections, each divided by small transforms [Chaytor *et al.*, 2004]. Spreading rates of these sections range from 55 to 25 mm/yr and decrease toward the south [Riddihough, 1980]. Magnetic anomaly patterns provide evidence for recent and ongoing deformation within the Gorda Block (e.g. Menard and Atwater 1968; Atwater and Mudie, 1973; Riddihough, 1980, Carlson and Stoddard, 1981).

Our catalog of the Gorda intraplate seismicity is relatively sparse, and the relocations do not significantly change existing seismicity patterns, other than collapsing several of the events into a tighter cluster along the northern Gorda Ridge (assuming the event location offset from the ridge is a result of the initial location bias). While some locations move over 20 km, these do not significantly affect the tectonic interpretation that the Gorda block is long known to deform internally.

303 4.3 Blanco Transform Fault Zone

304 The Blanco Transform Fault Zone (BTFZ) is a strike-slip fault that extends ~360 km
305 between the Gorda Ridge to the east and Juan de Fuca Ridge to the west (Figure 1). The
306 BTFZ is commonly described as composed of five transform segments, separated by
307 extension basins. These basins include the Gorda Depression, Cascadia Depression,
308 Surveyor Depression, and East and West Blanco Depressions, [McManus, 1965; Embley
309 *et al.*, 1987]. The easternmost portion of the BTFZ is the Gorda Depression Segment.
310 Earthquakes in this region are generally shallow (4-6 km) and small ($M_w \leq 5.3$). Normal
311 faulting is observed at the northern Gorda Ridge and within the Gorda Depression.
312 Braunmiller and Nábělek [2008] calculated 15% seismic coupling for this segment.

313 West of the Gorda Depression is the Cascadia Segment, which includes the Blanco Ridge
314 and Cascadia Depression. In its eastern limit, the Blanco Ridge is the narrow, steep
315 northern border of the Gorda Depression. With a ~150 km length, the 5 Ma Blanco Ridge
316 [Embley and Wilson, 1992] is the longest uninterrupted ridge segment of the BTFZ. The
317 peak seismic slip rates [Braunmiller and Nábělek, 2008] and largest recorded earthquake
318 [Dziak *et al.*, 1991] of the BTFZ occurred along the Blanco Ridge. Estimated average
319 seismic coupling values along the entire BTFZ vary considerably (e.g. Hyndman and
320 Weichert, 1983; Boettcher and Jordan, 2004; Willoughby and Hyndman, 2005;
321 Braunmiller and Nábělek, 2008), including 10% [Boettcher and Jordan, 2004] to 36%
322 [Braunmiller and Nábělek, 2008] seismic coupling. These differences reflect the
323 calculation's dependence upon assumed vertical extent of seismogenic region and the
324 assumed value of the shear rigidity. East of 127.9°W, the ridge is more seismically

325 active, including six of the ten $M_w \geq 6$ earthquakes recorded along the BTFZ [NEIC,
326 GCMT], leading *Braunmiller and Nábělek* [2008] to describe this region as having full
327 seismic coupling. *Braunmiller and Nábělek* [2008] describe lower seismicity rate and
328 earthquake sizes west of 127.9°W and estimate the seismic coupling to be less than 50%.
329 The Blanco Ridge segment terminates in the west along the southern border of the
330 Cascadia Depression.

331 Normal faulting seismicity is observed between 128.5° to 129.2°W in the Cascadia and
332 Surveyor Depressions [*Braunmiller and Nábělek*, 2008]. The Surveyor Depression is
333 located ~ 10 km northwest of the northwest corner of the Cascadia Depression. The
334 seismicity along the Cascadia and Surveyor Segments is generally deep (6-9 km) and
335 small ($M_w \leq 5.6$). *Braunmiller and Nábělek* [2008] calculated 15% seismic coupling in
336 these regions. The T-axis of the faults and geologic features suggest the Surveyor
337 Depression is a pull-apart deformation and the Cascadia Depression is a spreading center
338 [*Braunmiller and Nábělek*, 2008; *Embley and Wilson*, 1992]. The Cascadia Depression is
339 believed to have formed 5 Ma when PAC-JDF plate motion changed from 100° to 110°E .
340 The Surveyor Depression is significantly younger, 0.35 to 0.4 Ma [*Embley and Wilson*,
341 1992].

342 The westernmost portion of the BTFZ, the Western Blanco Segment, is structurally more
343 complex than the eastern segments. Like the Surveyor Depression, this segment is
344 believed to be younger, with reorganization 0.4 Ma [*Embley and Wilson*, 1992].
345 Seismicity is largely strike-slip and shallow (4-6 km) [*Braunmiller and Nábělek*, 2008].
346 *Braunmiller and Nábělek* [2008] calculated 33% coupling in the Western Blanco

Segment. In addition to the main fault along the north wall of the Western Blanco Depression, additional transform fault strands are proposed for this region (e.g. *Embley and Wilson*, 1992; *Dziak et al.*, 2003; *Dauteuil*, 1995; *Delaney et al.*, 1981). *Braunmiller and Nábělek* [2008] could not delineate these strands based on earthquake locations, but noted that slip vectors support the presence of three strands.

Adding a collective shift to our BTFZ earthquake relocations results in a strong correlation between seismicity and bathymetry. The relocations for events along the BTFZ generally concentrate the set of events into a more compact linear feature, however, not all of the events arrange into a single line. The results support interpretations that multiple active fault segments constitute the plate boundary (e.g. *Embley and Wilson*, 1992; *Dziak et al.*, 2003; *Dauteuil*, 1995; *Braunmiller and Nábělek*, 2008). We lack the number of events to delineate specific fault traces, but visual analysis of our measurements for particular event pairs confirms that the events are not collinear. We observe a greater variance in the location of seismicity normal to the fault zone along the west BTFZ than the east [*Braunmiller and Nábělek*, 2008]. The difference between the east and west BTFZ system might reflect an increased fault-system complexity of the west relative to the east. Considering that the western BTFZ could be up to 5 Ma younger than the eastern (i.e. *Embley and Wilson*, 1992), the fault-system complexity supports the idea that as a transform fault matures, slip localizes to fewer active parallel faults.

Observing the recent seismicity through time (Figure 8), a noticeable difference in recent event magnitude and rupture behavior is apparent between western and eastern BTFZ. As mentioned earlier, the western half has ruptured in three $M \geq 6.0$ between 1965 and 2015,

while the eastern half has experienced seven (NEIC); the eastern half has also experienced larger events than the western. The east tends to rupture in single events, with little or no aftershock sequences. In the west, there have been possibly repeating swarm-type rupture events between -130°E to -129°E . These differences in rupture behavior might also reflect the maturing of a single dominant fault, enabling larger single ruptures. Normal faulting event locations in the middle of the fault zone coalesce into a tighter cluster, which is consistent with a visual inspection of the cross-correlations between these events. Our catalog does not include any normal faulting events in the eastern half of the BTFZ.

In the original NEIC hypocenters, there is a noticeable seismicity gap that includes regions both east and west of the Cascadia Depression (Figure 6 and 8). The relocation of the 27 October 1994 (M_w 6.29; 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 at least a partial reactivation of the structures involved in the 1994 event. Prior to 1994, there were two M_w 6.4 (GCMT) events in 1981 and 1985 that could be related to the same features. Two other strong events occurred in 2000 (M_w 6.0) and 2012 (GCMT: M_w 6.0). Our observations relocated the 2000 event ~ 30 km to the east of the NEIC location (43.65°N , -127.61°E). We did not include the 2012 event in the final inversion because the GCMT dip, 62° , of one of the nodal planes was outside our acceptable range, however, preliminary relocations placed it near the 2000 event, suggesting its rupture could overlap that of the

2000 event. *Boettcher and McGuire* [2009] relocated the 1981, 1985, 1994, 2000 and 2008 events. This study concluded the 1981, 1994, and 2008 earthquakes were co-located, while the 1985 and 2000 events ruptured a similar region ~25 km east of the 1981/1994/2008 events. The original NEIC hypocenters suggest that the fault segment between -128.5°E to -127.75°E did not host any large events since 1981. Our relocations suggest that is not the case. If the 1994/2008 events ruptured the entire region between the Cascadia Depression and the 2000/2012 epicentroids, the 1994/2008 events could account for slip along ~60 km of the boundary. Below, we describe how we used modeling to estimate the rupture length and seismogenic zone width of the 1994 event.

Estimating the length of moderate-size ruptures is difficult. We performed a limited test for a long, unilateral rupture along a transform fault by looking for directivity effects. We computed the relative amplitude of intermediate-period R1 waves for signals from the 1994 earthquake relative to signals from all nearby events. The results are displayed as a function of azimuth in Figure 9. The results are noisy, but show a systematic variation that resembles a sinusoid, which is not representative of differences in faulting geometry. Some of the scatter arises because the amplitude is a measure of the relative directivity between 1994 and each of the other events. We interpret the amplitude modulation to be the result of rupture directivity in the 1994 event. We can reproduce the pattern using fundamental-mode synthetic seismograms [*Herrmann, 2002*] corresponding to vertical strike-slip sources computed with a flat-layered, earth model adapted from the 0-20 Ma oceanic lithosphere model of *Anderson and Regan* [1983]. We model rupture directivity using superposition of a series of point sources spaced one-km apart and shifted in time

413 based upon the rupture speed (see Figures S1 and S2 in the electronic supplement for
414 details on the rupture directivity modeling using synthetic waveforms).

415 The pattern reveals several aspects of the rupture behavior. Since the ratio is 1994 scaled
416 by other events, the maximum amplitude value is in the 1994 rupture direction. In our
417 bandwidth, increasing the rupture length or decreasing the rupture speed enhances the
418 amplitude of the pattern. The observations suggest that the 1994 event ruptured to the
419 northwest (Figure 9). *Braunmiller and Nábělek* [2008] also reported northwestward
420 rupture directivity for this event. The best fit to the data is an azimuth of $\sim 282^\circ$; the 1994
421 event GCMT strike is 296° . The amplitudes vary by roughly $\pm 30\text{-}40\%$. Synthetic
422 modeling of the pattern suggests that the rupture was predominantly unilateral; bilateral
423 rupture results in a distinct pattern inconsistent with the observations. We are unable to
424 produce amplitudes close to those observed using unilateral rupture lengths shorter than
425 30 km. A 30 km northwestward rupture is consistent with aftershock locations for the
426 1994 event reported in *Dziak et al.* [2000]. To fit the observations with a length of 30 km
427 requires a rupture speed near or less than 1 km/s, while a rupture length of 60 km, the
428 entire width of the “gap,” requires a rupture speed over 4 km/s.

429 Using relative magnitude calculations documented in [*Cleveland and Ammon*, 2015] and
430 listed in Table S1, we estimate that the relative moments of the 1994 and 2008 events
431 differ by about 10%, which is consistent with *Boettcher and McGuire* [2009]. If we
432 assume the 2008 event is a repeat of the 1994 and there was no slip between events and
433 each event released all the plate motion strain accumulated between the events, based on
434 MORVEL [*DeMets et al.*, 2010], the approximate seismic slip is ~ 64 cm (using

~ 49 mm/yr). Assuming a mean shear modulus of 44.1 GPa [Boettcher and Jordan, 2004], the approximate rupture area would be ~115 km². Assuming a unilateral rupture of 60 km would require the width of the fault to be ~2 km; a 30 km unilateral rupture would require a width ~4 km. Studies suggest 600°C defines the lower bound of the seismogenic zone in oceanic lithosphere [e.g. Abercrombie and Ekström, 2001; Roland *et al.*, 2010]. Based on mineral properties, Roland *et al.* [2010] suggest that the upper bound corresponds to a temperature of about 350°C. With these thermal bounds, thermal modeling by Roland *et al.* [2010], predicts a near constant seismogenic zone width of ~4 to 5 km along the portion of the plate boundary ruptured in the 1994 earthquake. 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 suggest the rupture width could be even narrower than these models predict.

4.4 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 [Bebel *et al.*, 1992]. Like the Gorda block, the Explorer Plate experiences significant intraplate seismicity. However, unlike the Gorda, magnetic lineations do not indicate substantial internal deformation in the Explorer [Bebel *et al.*, 1992]. Several studies (i.e. Hyndman *et al.*, 1979; Riddihough, 1984; Botros and Johnson, 1988) propose that the Explorer separated from the Juan de Fuca at about 4 Ma with 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 [Riddihough, 1984; Botros and Johnson, 1988]. 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 $\sim 105^\circ$ 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 clockwise to 120° azimuth [Riddihough, 1984]. 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 [Braunmiller and Nábělek, 2002]. Multiple hypotheses, including heightened buoyancy [Riddihough, 1984] and reduced slab pull forces [Braunmiller and Nábělek, 2002], have been proposed to explain the slowing and rotation of convergence.

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 [Riddihough, 1977; 1984]. Alternatively, based on analysis of seismicity and detailed bathymetry, several studies [Barr and Chase, 1974; Rohr and Furlong, 1995] proposed the eastern portion of the Explorer Plate is becoming coupled with North America, whereas the western is accreting to the Pacific. This model suggests the formation of a new Pacific-North American Plate transform boundary, splitting the Explorer Plate. GPS measurements in northern Vancouver Island [Dragert and Hyndman, 1995; Henton *et al.*, 1996] support this interpretation. Initially, slip vector orientations of intraplate seismicity [Wahlström *et al.*, 1990; Kreemer *et al.*, 1998] and inversion of seismic strain distribution within the Explorer Plate [Kreemer *et al.*, 1998] also supported

478 this interpretation, however, measurements leading to this conclusion were later shown to
479 be erroneous by *Braunmiller and Nábělek* [2002]. *Rohr and Furlong* [1995] inferred that
480 the new fault is forming as a result of a southern lengthening of Queen Charlotte Fault.
481 This would result in the extinction of the Explorer Ridge as the Pacific Plate captures it.
482 *Braunmiller and Nábělek* [2002] proposed a different interpretation of the seismicity.
483 They believe the Explorer continues to move independently of North America and the
484 Pacific. Ongoing adjustments along the southeastern Explorer-Pacific boundary to bring
485 it in agreement with the Revere-Dellwood fault is resulting in the formation of a new
486 transform fault, called the Southwest Explorer transform boundary, separating the
487 southwest corner of the Explorer Plate from the rest of the plate, causing just this corner
488 to become a part of the Pacific Plate. North of the Southwest Explorer transform
489 boundary, however, *Braunmiller and Nábělek* [2002] do not recognize the formation of a
490 new fault boundary and believe the Explorer continues to move independently. A key
491 reason for this conclusion was that *Braunmiller and Nábělek* [2002] found the azimuth of
492 slip vectors of earthquakes in this region are inconsistent with Pacific-North America
493 motion; azimuths largely ranged between *Riddihough's* [1984] Pacific-Explorer motion
494 (310°) and Pacific-North America motion (340°), with an average value of 323° .
495 Additionally, *Braunmiller and Nábělek* [2002] point out that much of the seismicity
496 considered intraplate by *Rohr and Furlong* [1995] is actually an artifact of systematic
497 mislocation to the northeast of regional seismicity. Earthquake relocation by *Braunmiller*
498 *and Nábělek* [2002] shifted the location of much of the seismicity to the expected plate
499 boundaries.

Even with our improved relative locations, the Explorer region seismicity remains spatially complex – a mix of small clusters, linear trends, and regions of diffuse seismicity. The seismicity patterns displayed by our relocations are consistent with those of *Braunmiller and Nábělek* [2002], including the observation that initial locations are biased toward the northeast. Although not a new result, our work confirms earlier conjectures that the deformation of the Sovanco Fracture Zone (FZ) is diffuse and not simply a result of low-quality locations. Much of the seismicity in the Sovanco Fracture Zone is smaller than M 4.5, which is below the magnitude threshold used in this study. South of the Sovanco, two tight clusters of seismicity are observed near 47.75°N, -128.75°E and 48.25°N, -129°E, falling along short transform step-overs resulting from overlapping spreading segments of the northern Juan de Fuca ridge [*Weekly et al.*, 2014].

Through the middle of the Explorer, from around 49°N, -129°E to 49.5°N, -128.75°E, eight events align linearly NE-SW with the relocations, a pattern not observed with the initial scattered locations. The earthquakes appear to align in two or more ~220° trending en-echelon, linear patterns, consistent with the strike of the events. This linear pattern of intraplate seismicity, not included in any of the current tectonic models, suggests northeast-southwest oriented internal deformation of the Explorer Plate, possibly delineating faults that accommodate some slip previously attributed solely to the Nootka Fault. Southeast of this 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 extend to the west before bending back east and following the coast of Moresby Island, British

Columbia, Canada. This linear pattern was also observed in *Braunmiller and Nábělek* [2002], however, our locations generally display ~ 10 km less scatter. *Braunmiller and Nábělek* [2002] noted that while seismicity through the Explorer and up to Queen Charlotte Island roughly aligns with the direction of Pacific-North American plate motion, based on slip vectors they interpreted the seismicity as a series of right-stepping strike-slip faults as opposed to the formation of a new transform fault proposed by *Rohr and Furlong* [1995]. Our results do not support a simple through-going transform, and support the complex structure described by *Braunmiller and Nábělek* [2002].

North of $\sim 51.25^\circ\text{N}$, our relocations do not follow the plate boundary inferred by the *Bird* [2003] model. Instead of stepping to the east, rupturing along the Queen Charlotte fault, seismicity appears to continue along a northern extension of the Revere-Dellwood fault (RDF). This is particularly true for the 17 November 2009 M_w 6.6 strike-slip earthquake, which displays a clear southwestern separation from the Queen Charlotte fault by ~ 17 km. We also relocated two strike-slip events farther west of this RDF extension and the 2009 M_w 6.6 event, including an aftershock of this 2009 event (17 November 2009 M_w 5.8) and 17 June 2011 M_w 4.9. These locations are consistent with the tectonic model proposed by earlier studies (i.e. *Braunmiller and Nábělek*, 2002; *Allan et al.*, 1993; *Rohr*, 2015; *Rohr and Tryon*, 2010), in which the RDF extends north to $\sim 52^\circ\text{N}$, overlapping the Queen Charlotte fault by ~ 80 km. Seismic reflections presented in *Rohr* [2015] identify a second fault structure ~ 15 km southwest of the RDF extension. Our relocations display a southwest offset for these two smaller strike-slip events consistent with these structures. The M_w 6.6 mainshock triggered aftershock activity on the parallel fault to the southwest. The six northern-most normal faulting events we relocate are all aftershocks of the 2012

M_w 7.8 Haida Gwaii underthrusting earthquake. Based on slip models by *Lay et al.* [2013], these events are oceanward of peak slip regions of the 2012 rupture, likely indicating what can be viewed as typical outer rise responses to a large underthrust event.

5. Conclusions

We utilize cross-correlation of Rayleigh waves to calculate precise relative earthquake epicentroid locations in the northeast Pacific region. Our BTFZ relocations demonstrate the breadth (spread perpendicular to average strike) of the fault zone seismicity is more compact than initial locations. However, the seismicity does not collapse to a single trend, suggesting the presence of multiple active parallel faults along the fracture zone. The western and eastern BTFZ seismicity characteristics differ; seismicity is more scattered in the west, possibly relating to this portion of the fault being younger. We used directivity of the 27 October 1994 (M_w 6.29; 43.75°N, -127.92°E) earthquake along the BTFZ to constrain the rupture length to a range between 30-60 km, which when combined with a slip estimate based on plate rates suggests a narrow seismogenic zone width in the range of 2-4 km. The results are consistent with thermal bounds on OTF faulting. Relocations in and along the Explorer Plate identify a linear trend in seismicity north of ~49.5°N extending about 180-200 km to the north-northwest. North of ~51.25°N, seismicity pattern support models proposing northern extension of the Revere-Dellwood fault [*Braunmiller and Nábělek*, 2002; *Allan et al.*, 1993; *Rohr and Tryon*, 2010], offset and overlapping the Queen Charlotte fault by ~80 km [*Rohr*, 2015]. South of ~49.5°N, seismicity within the Explorer is complex, reflecting evolving plate interactions. Our locations contain a SSW-NNE linear pattern of seismicity perpendicular

to the Sovanco Transform and extending 50 km into the Explorer Plate, possibly suggesting the development of a fault that accommodates some of the SSW-NNE slip previously attributed solely to the Nootka Fault.

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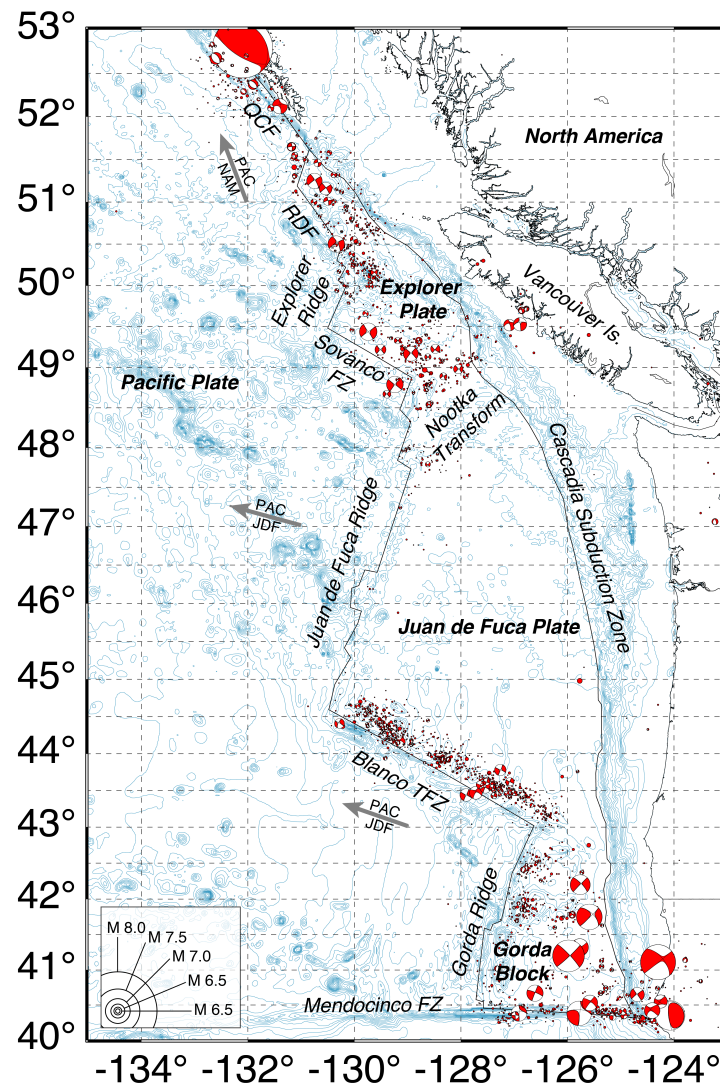
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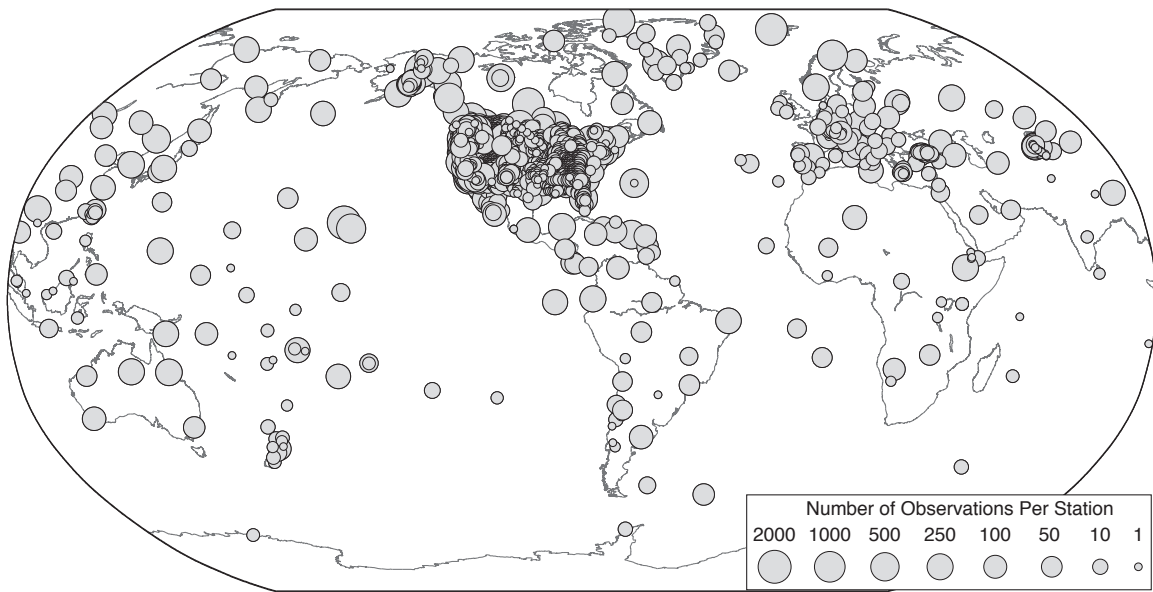
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- 820

821 **Figures**

822

823 Figure 1: Regional seismicity of the northeast Pacific. Shown are locations and
 824 magnitudes of all events (including mechanisms) $M \geq 4.0$ in the NEIC catalog (since
 825 1973). Focal mechanisms from the GCMT catalog are also shown for all events available.
 826 These mechanisms are located at the NEIC hypocenter and scaled in size according to
 827 NEIC magnitude. Plate motion directions are calculated using MORVEL [DeMets *et al.*,
 828 2010]. Plate boundaries (solid black line) are based on Bird [2003].

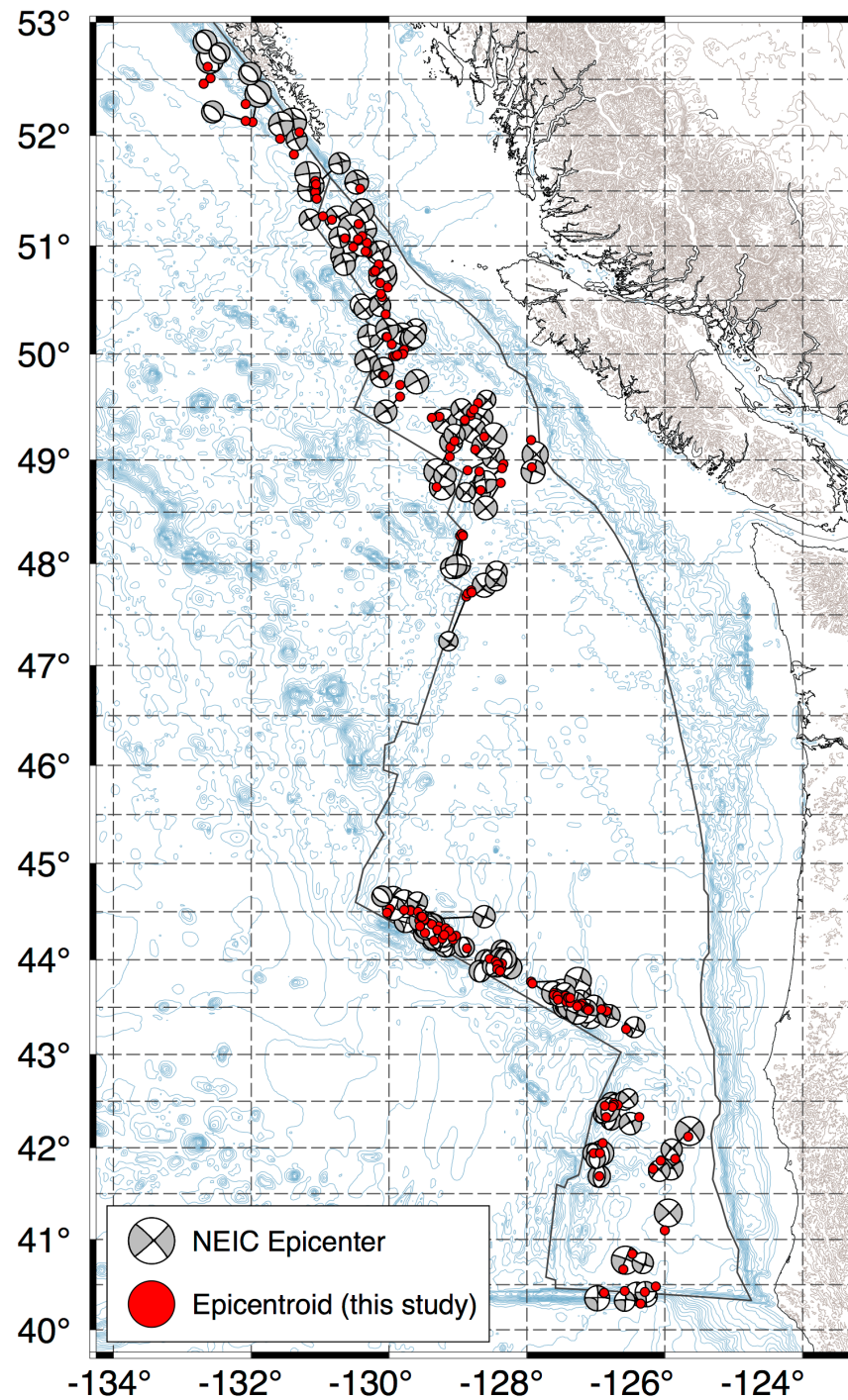


829

830 Figure 2: Locations of seismic stations used for relocation of earthquakes in this study.

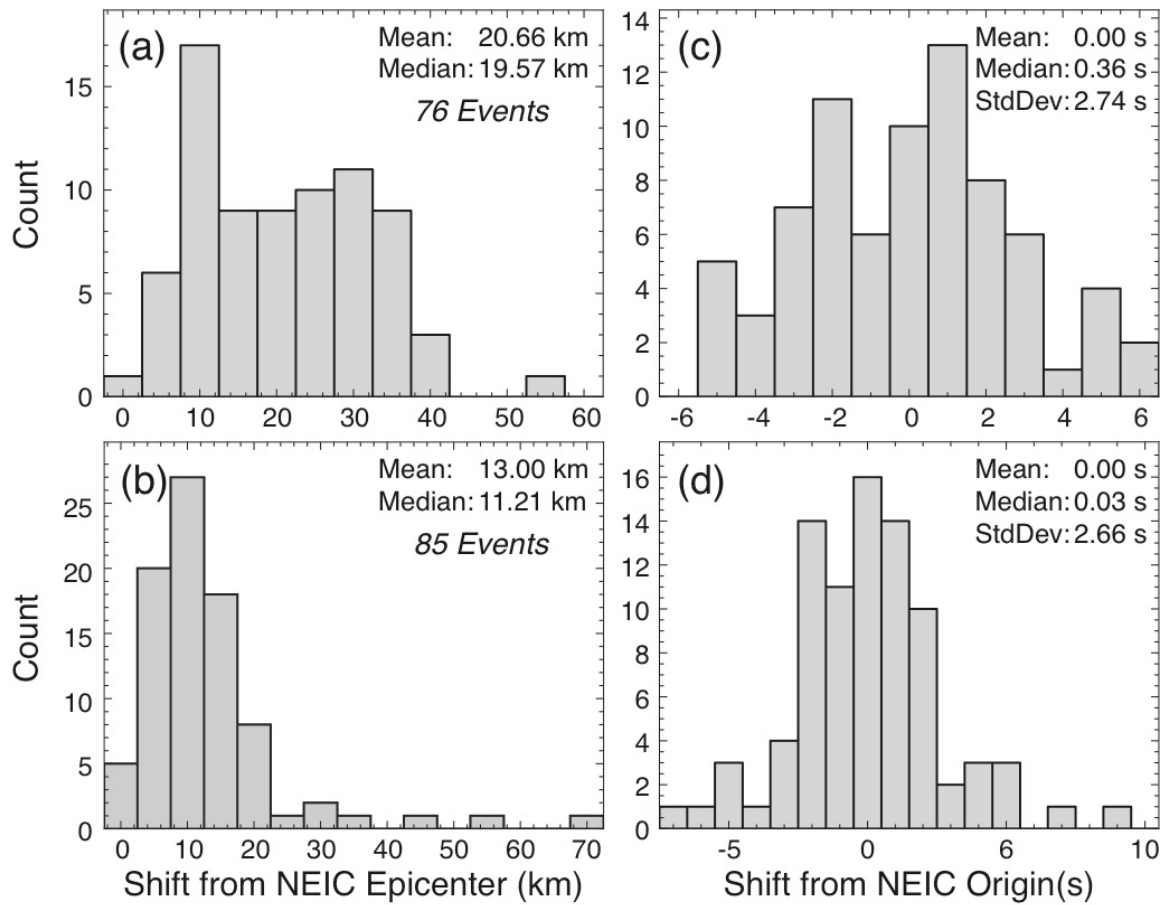
831 The markers are scaled in size to the number of times waveforms from each station were

832 used to calculate differential travel times between events.



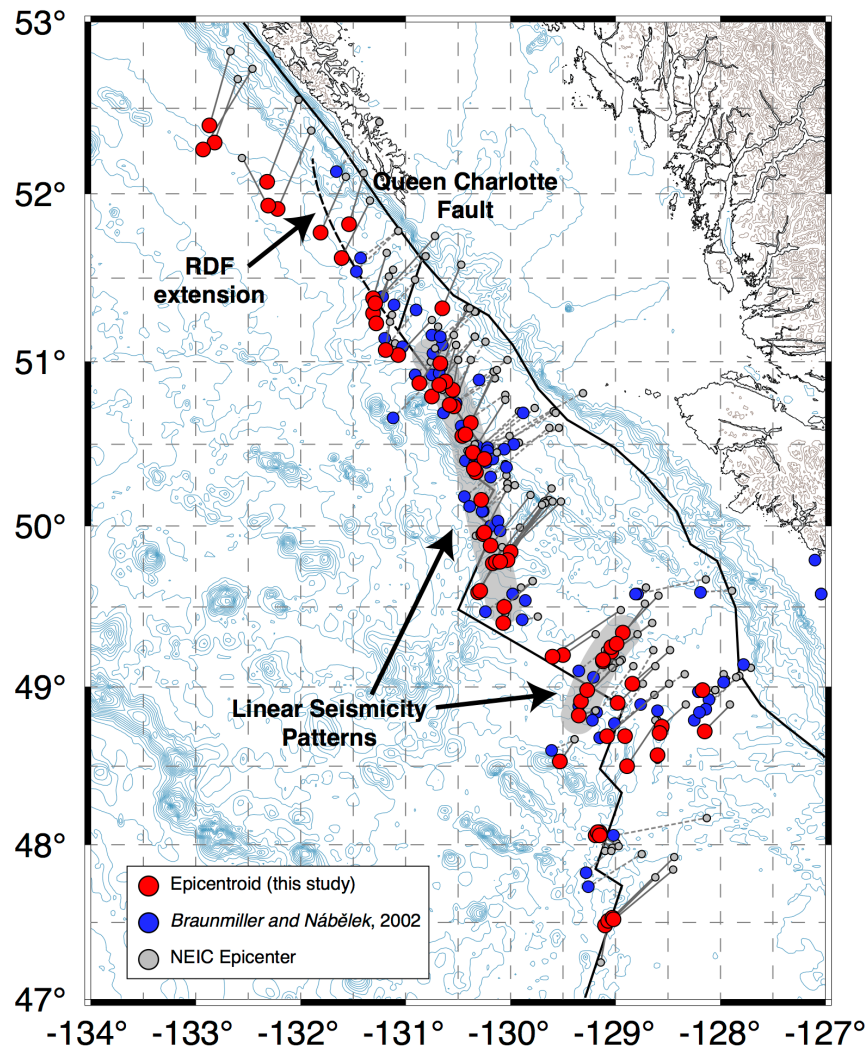
833

834 Figure 3: Relocated epicentroids from this study (red circle) compared to original NEIC
 835 locations (gray) labeled by the respective GCMT focal mechanism. Plate boundaries
 836 (solid black line) are based on *Bird* [2003].



837

838 Figure 4: Comparison of the epicentroid locations with the original NEIC locations
 839 (subplots a and b) and origin time (subplots c and d) for events north (76 events, subplots
 840 a and c) and south (85 events, subplots b and d) of 46°N.



841

842 Figure 5: Explorer Plate relocated epicentroids from this study (red circle) compared to
 843 original NEIC locations (gray) and relocations from *Braunmiller and Nábělek* [2002]
 844 (smaller, blue). Following the method of *Braunmiller and Nábělek* [2002], we shift our
 845 relative relocations to fix the 06 April 1992 M_w 6.7 earthquake to the location carefully
 846 calculated by *Cassidy and Rogers* [1995] (50.55°N , -130.46°E). Plate boundaries (solid
 847 black line) are based on *Bird* [2003]. The dashed black line shows the proposed extension
 848 of the Revere-Dellwood fault (RDW) by *Rohr* [2015; *Rohr and Tryon*, 2010].
 849 Highlighted in gray are two linear patterns observed in the improved seismic locations.

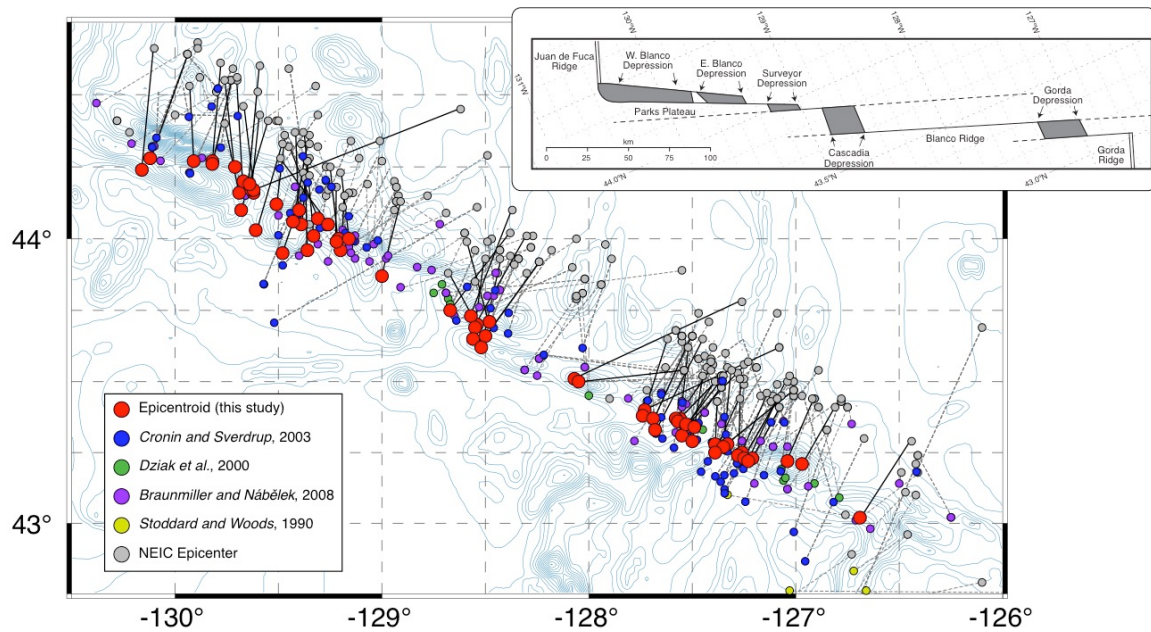
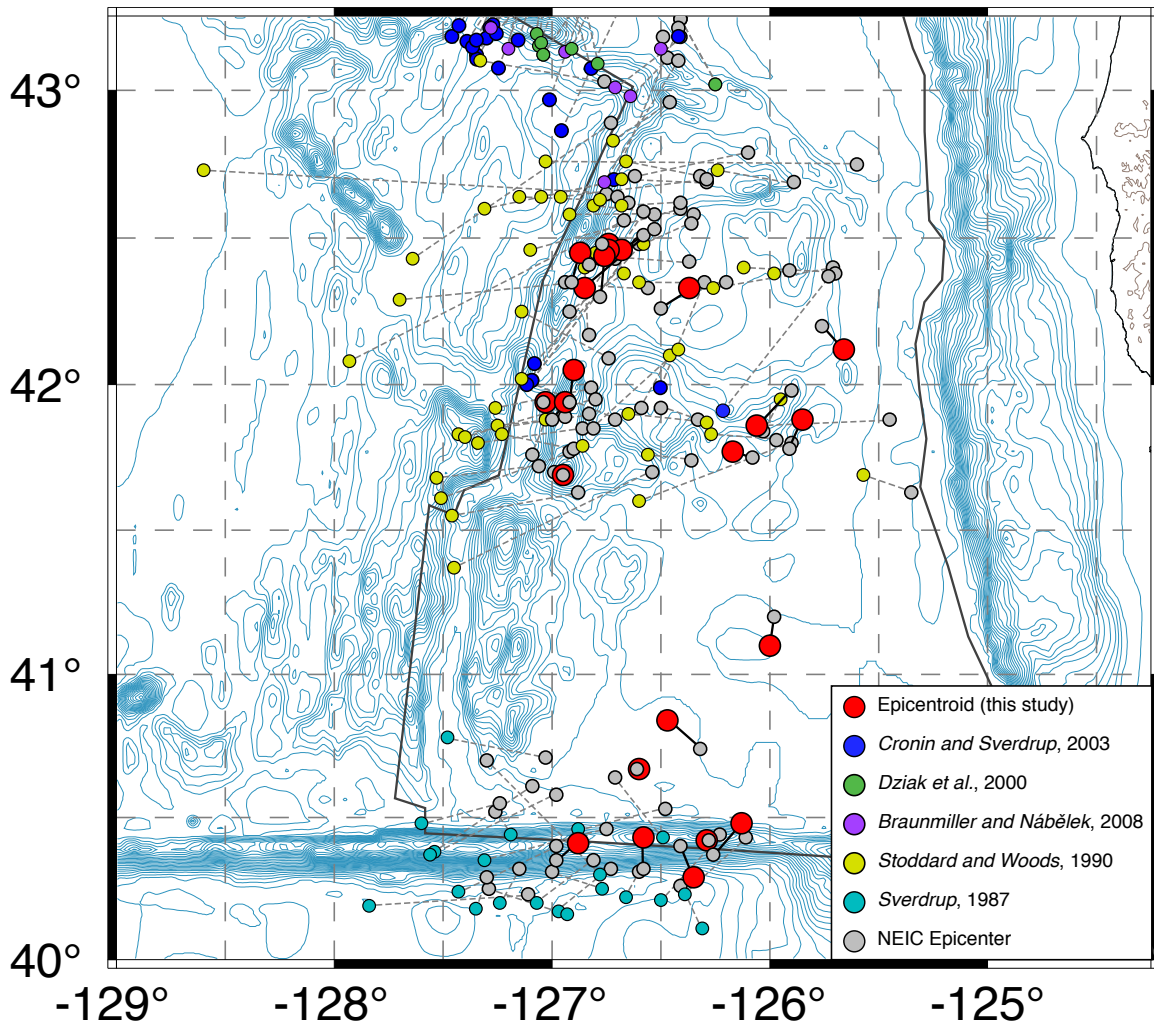
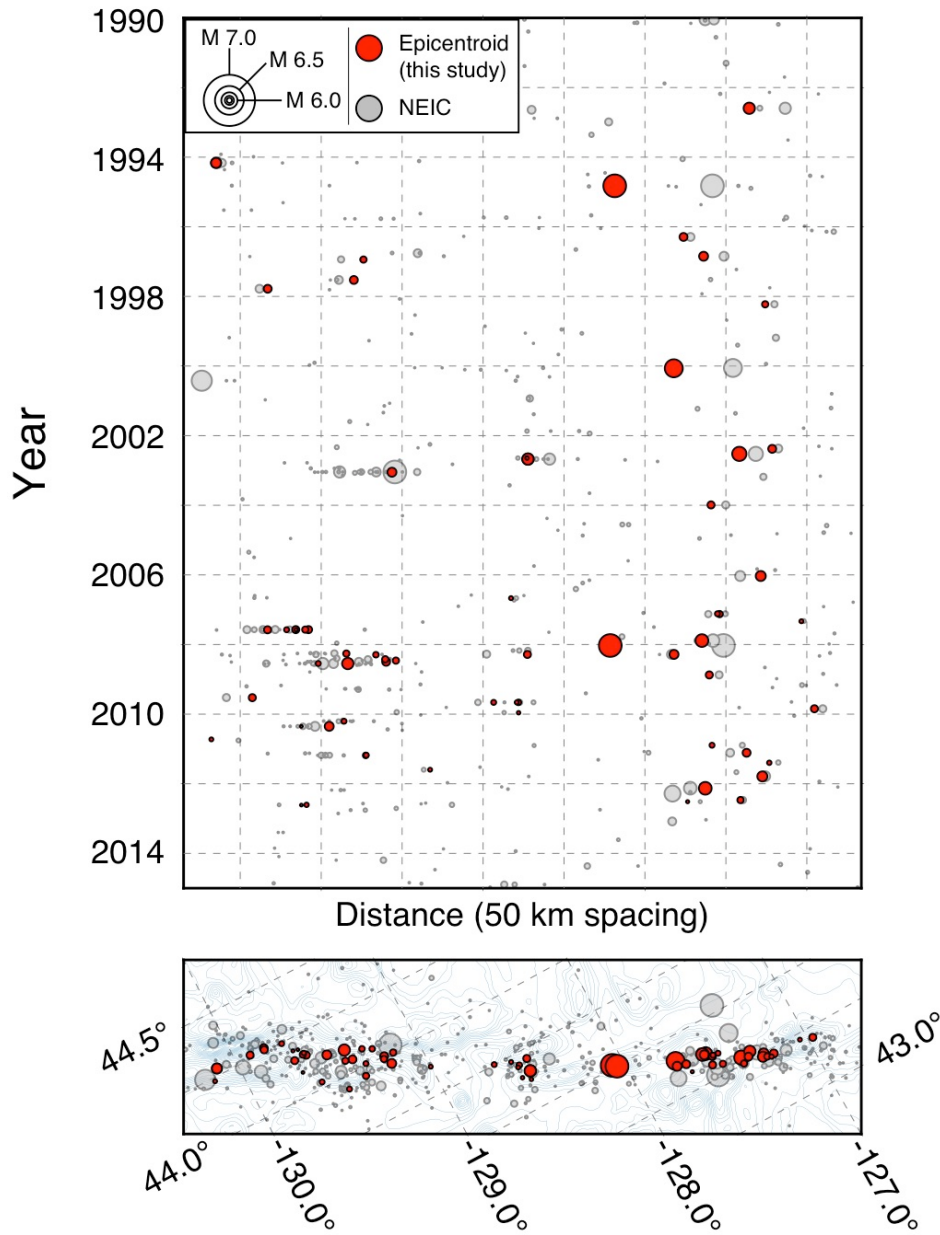


Figure 6: Blanco Transform Fault Zone relocated epicentroids from this study (red circle) compared to original NEIC locations (gray) and relocations from *Cronin and Sverdrup* [2003] (blue), *Dziak et al.* [2000] (green), *Braunmiller and Nábělek* [2008] (purple), and *Stoddard and Woods* [1990] (yellow). We shift our relative relocations a minimal amount that appears to best correlate with the bathymetry (30 km at 200° azimuth). The inset diagram is based on the bathymetry map in *Embley and Wilson* [1992].



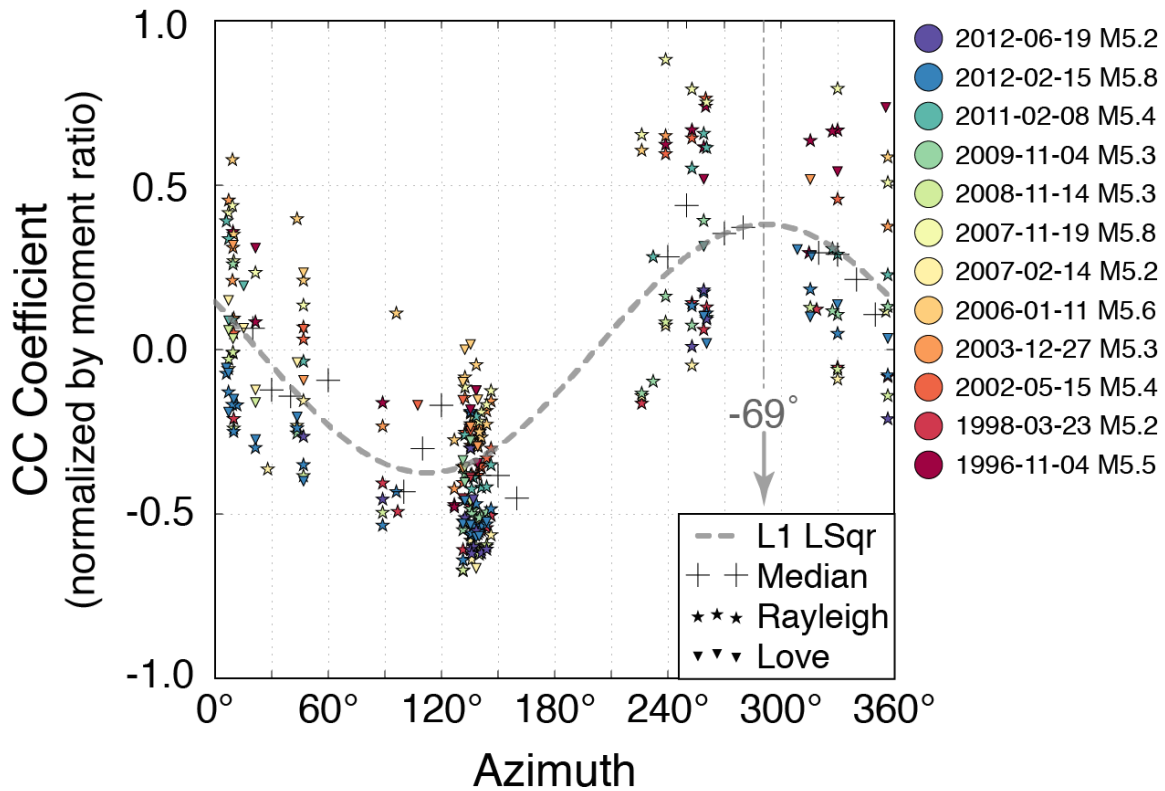
857

858 Figure 7: Gorda block relocated epicentroids from this study (red circle) compared to
 859 original NEIC locations (gray) and relocations from *Cronin and Sverdrup* [2003] (blue),
 860 *Dziak et al.* [2000] (green), *Braunmiller and Nábělek* [2008] (purple), *Stoddard and*
 861 *Woods* [1990] (yellow), and *Sverdrup* [1987] (light blue). There is no shift in our relative
 862 relocations. Plate boundaries (solid black line) are based on *Bird* [2003].



863

864 Figure 8: Timeline plot of Blanco Transform Fault Zone seismicity since 1992. Shown
 865 are both relocated epicentroids from this study (red) and all NEIC epicenters $M \geq 4.0$
 866 (gray). A similar static shift as the relocated epicentroids (20 km at 200°) has been
 867 applied to the NEIC locations. Note differences in event locations are apparent,
 868 particularly for the large 1994, 2000, and 2008 events.



869

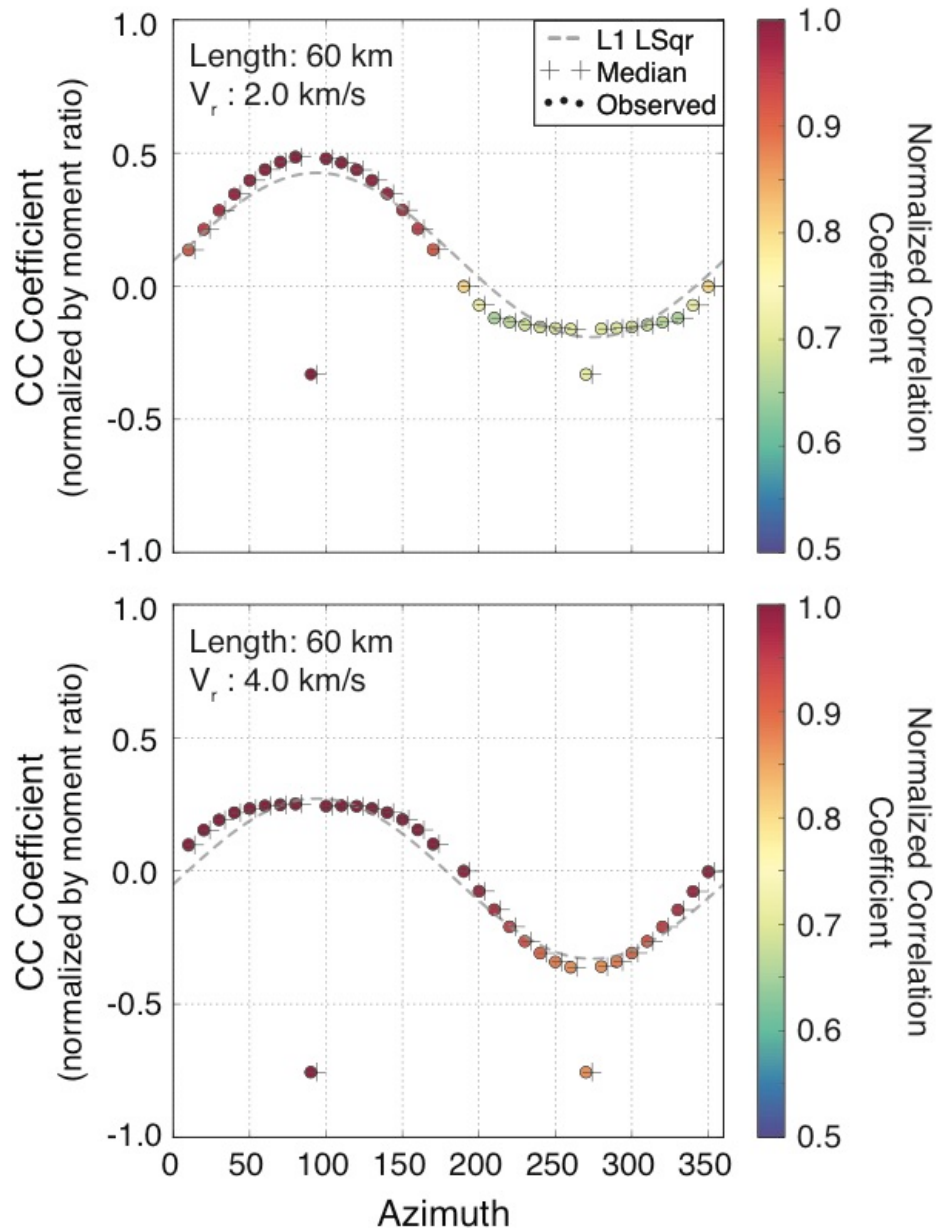
870 Figure 9: Unnormalized cross-correlation coefficient values (divided by the mean value
 871 and shifted to be centered at zero) between the 27 October 1994 M_w 6.3 Blanco
 872 Transform Fault Zone event and other surrounding events (labeled by color) at varying
 873 event to station azimuths. The observed cosine pattern (L1 optimized fit displayed with
 874 the gray, dashed line) results from rupture directivity of the 1994 event. The peak
 875 amplitude indicates the direction of rupture. Labeled is the observed strike of bathymetric
 876 ridges (-69°) [Embley and Wilson, 1992].

877 **Tables (Electronic Supplement)**

878 Table S1: Event relative relocations, shifted relocations and relative magnitudes.

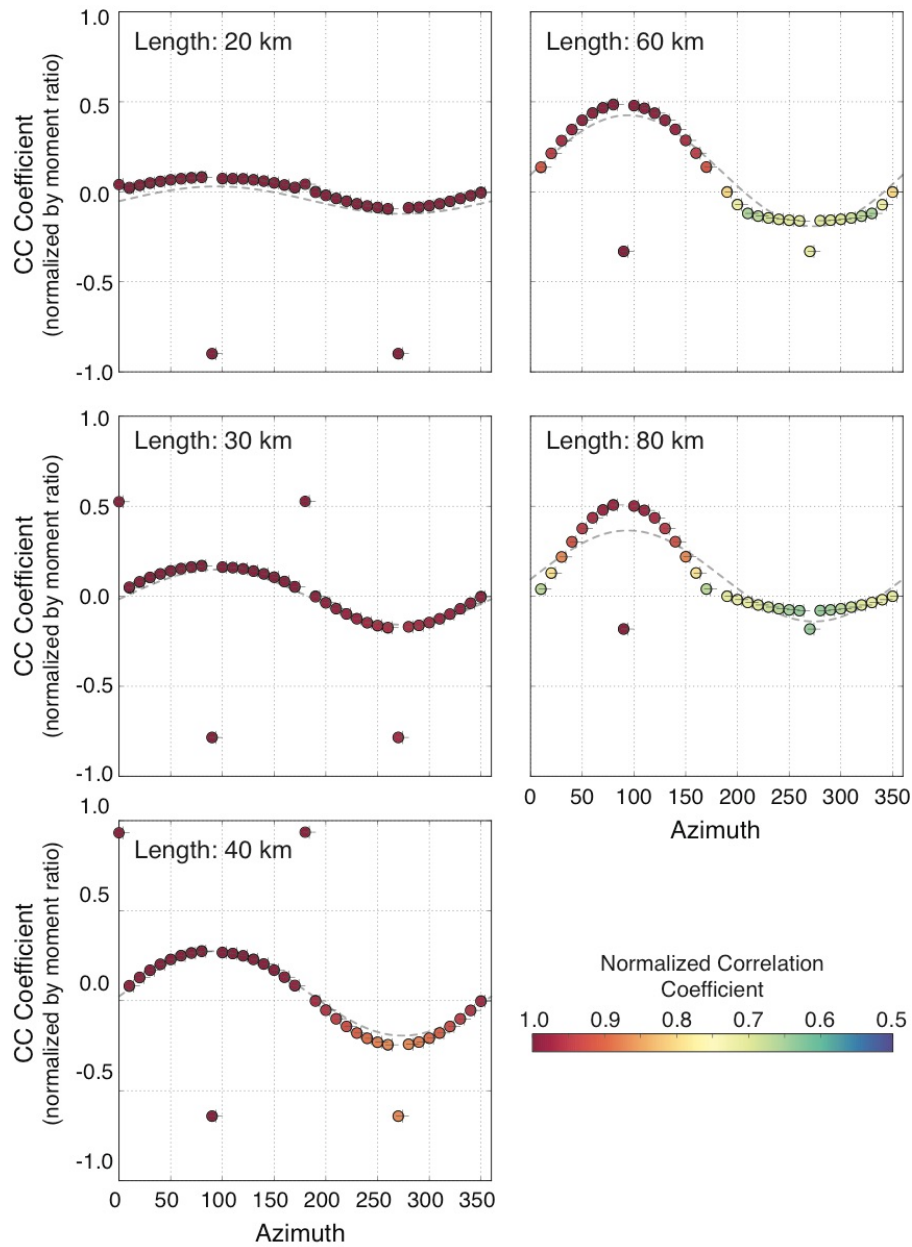
879 Table S2: Relocated events in common with earlier relocation studies.

880

881 **Figures (Electronic Supplement)**

882

883 Figure S1: The cross-correlation coefficient normalized by moment ratio, at varying
 884 azimuths using synthetic waveforms (band 30 to 80 s). The master events are 60 km long
 885 in both examples, but the rupture speed varies (top: 2.0 km/s, bottom: 4.0 km/s). At lower
 886 speeds, the direction away from rupture (the trough) becomes flattened; the opposite
 887 pattern is observed at higher speeds.



888

889 Figure S2: The cross-correlation coefficient normalized by moment ratio at varying
 890 azimuths using synthetic waveforms (band 30 to 80 s) with master events of varying
 891 length. All master events have a rupture speed of 2.0 km/s. At shorter lengths, the pattern
 892 is well described by a cosine function (gray, dashed line), but as the length increases, the
 893 peak (direction of rupture) becomes increasingly amplified while the opposite direction
 894 becomes flattened. Additionally, as the length increases the waveforms become
 895 increasingly dissimilar, particularly in the direction away from rupture; this is indicated
 896 by the systematic decrease in normalized correlation coefficient.