

# Very low frequency earthquakes in between the seismogenic and tremor zones in Cascadia?

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## Key Points:

- Three large VLFs in Cascadia were dynamically triggered by teleseismic waves.
- These VLFs likely occurred in between the seismogenic and tremor zones.
- One VLFE has a moment magnitude of 5.7, the largest VLFE detected to date and the first detected geodetically.

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## Abstract

Megathrust earthquakes and their associated tsunamis cause some of the worst natural disasters. In addition to earthquakes, a wide range of slip behaviors are present at subduction zones, including slow earthquakes that span multiple orders of spatial and temporal scales. Understanding these events may shed light on the stress or strength conditions of the megathrust fault. Out of all types of slow earthquakes, very low frequency earthquakes (VLFs) are most enigmatic because they are difficult to detect reliably, and the physical nature of VLFs are poorly understood. Here we show three VLFs in Cascadia that were dynamically triggered by a 2009 Mw 6.9 Canal de Ballenas earthquake in the Gulf of California. The VLFs likely locate in between the seismogenic zone and the Cascadia episodic tremor and slip zone, including one event with a moment magnitude of 5.7. This is the largest VLF reported to date, causing clear geodetic signals. Our results show that the Cascadia megathrust fault might slip rapidly at some spots in this gap zone, and such a permissible slip behavior has direct seismic hazard implications for coastal communities and perhaps further inland. Further, the observed seismic sources may represent a new class of slip events, whose characteristics do not fit current understandings of slow or regular earthquakes.

## Plain Language Summary

Megathrust earthquakes and their associated tsunamis pose significant hazards in Cascadia. In addition to earthquakes, episodic tremor and slip events have been discovered in the region at depths of 30 to 50 km, 10–15 km below the seismogenic zone. The expected slip behavior between the seismogenic and tremor zones remains unknown, leaving the rupture extents of future megathrust earthquakes unclear. We detect and locate three very low frequency earthquakes near this gap region, including one with a moment magnitude of 5.7. This is the largest very low frequency earthquake reported to date, and its detection not only shows that the fault slip behaviors in this gap zone can be highly heterogeneous but also challenges current understanding of slow earthquake physics.

## 1 Introduction

The Cascadia subduction zone poses serious earthquake and tsunami hazards to some of the most populous regions of the United States and Canada. Geological records reveal that at least 19 great megathrust earthquakes occurred in the region over the past ten thousand years [Walton *et al.*, 2021]. However, as an exceptionally seismically quiet subduction zone [Wang and Tréhu, 2016], large megathrust earthquakes in Cascadia have never been recorded by modern instrumentation. In contrast, slow earthquakes, which differ from regular earthquakes in their seismic radiations and rupture speeds, occur frequently across the whole subduction zone [Brudzinski and Allen, 2007; Gomberg *et al.*, 2010]. These slow earthquakes encompass a wide spectrum of slip behaviors [Peng and Gomberg, 2010], including slow slip events (SSEs) [Dragert *et al.*, 2001], very low frequency earthquakes (VLFs) [Ghosh *et al.*, 2015; Hutchison and Ghosh, 2016], low-frequency earthquakes [Brown *et al.*, 2009; Bostock *et al.*, 2012; Sweet *et al.*, 2019], and non-volcanic tremor [Wech and Creager, 2008]. In Cascadia, slow slip and non-volcanic tremor often occur concurrently with each other as episodic tremor and slip (ETS) events [Rogers and Dragert, 2003; Bartlow *et al.*, 2011].

These ETS events recur semi-regularly every 11 to 15 months and can propagate unilaterally or bilaterally [Dragert and Wang, 2011; Wech and Bartlow, 2014]. They can have moment magnitudes equivalent to Mw 6.7 earthquakes with the SSEs releasing most of their moments [Dragert *et al.*, 2001; Kao *et al.*, 2010]. These events accommodate a portion of the slip deficit at the subduction zone and concentrate along a band at depths of 30–50 km, about 10 to 15 km deeper than the downdip edge of the seismogenic zone [Brudzinski and Allen, 2007; Gomberg *et al.*, 2010; Wang and Tréhu, 2016; Walton *et al.*, 2021]. Additionally, typical VLFs in the region can have equivalent moment magnitudes ranging from 2.1 to 4.1 [Ghosh *et al.*, 2015; Hutchison and Ghosh, 2016; Ide, 2016]. In between the tremor

66 zone and the seismogenic zone on the fault, there is a gap that is not fully locked, yet devoid  
67 of slow earthquakes [Hyndman and Wang, 1995; Wang et al., 2003; Brudzinski and Allen,  
68 2007; Gombert et al., 2010; Priest et al., 2010; Schmalzle et al., 2014]. Understanding the  
69 slip behaviors in this gap zone gives insight into the stress and strength conditions of the  
70 megathrust fault, and can lead to improved forecasting of future earthquake rupture scenar-  
71 ios [Bruhat and Segall, 2016; Ramos and Huang, 2019]. Studies of this gap zone are largely  
72 hindered by a lack of robust observations, or the loss of resolution of onshore instruments;  
73 hence, little is known about the nature of the gap zone or its relation to the locked zone and  
74 the tremor zone.

75 VLFs differ from tremor and regular earthquakes with similar magnitudes as they are  
76 rich in low-frequency radiation content in a band of  $\sim 20$ – $50$  seconds and depleted in high-  
77 frequency radiation ( $\geq 1$  Hz) [Obara and Ito, 2005; Ito and Obara, 2006; Ito et al., 2007].  
78 Their apparent correlations with ETS events in Cascadia seem to indicate that VLFs are an-  
79 other seismic manifestation of slow slip, similar to tremor and low frequency earthquakes but  
80 responsible for a larger portion of the moment release [e.g., Ghosh et al., 2015]. However,  
81 the relationship between VLFs and SSEs are more complicated than those between tremor  
82 and SSEs [e.g., Bartlow et al., 2011]. For example, VLFs in Cascadia can correlate with  
83 ETS events either spatiotemporally [Ghosh et al., 2015] or only temporally [Hutchison and  
84 Ghosh, 2016]. In either case, previously identified VLFs in Cascadia were all located in  
85 the tremor zone at depths of 30–50 km, and they are almost always accompanied by tremor  
86 [Ghosh et al., 2015; Hutchison and Ghosh, 2016; Ide, 2016].

87 In a search of the USArray seismic data from August to October 2009 (see Open Re-  
88 search), we find that three VLFs occurred over a 5-minute period. Two of the VLFs likely  
89 occurred in the gap zone between the seismogenic locked zone and the tremor zone, and one  
90 is likely adjacent to the gap zone (Figure 1). Geodetic data confirm the best resolved event,  
91 with clear observations of static strains that are consistent with the seismically derived fo-  
92 cal mechanism. The VLFs coincide with surface wave arrivals of the August 2009 Mw 6.9  
93 Canal de Ballenas earthquake in the Gulf of California, Mexico, and we examine whether  
94 the VLFs may have been triggered by the dynamic stresses from the passing waves. These  
95 events were also close to the onset area of the 2009 Cascadia ETS event and occur three days  
96 before its reported initiation [Bartlow et al., 2011]. This spatiotemporal correlation between  
97 the VLFs and ETS event suggests that these previously unknown VLFs were either diag-  
98 nostic of, or played a role in, the nucleation process of the 2009 Cascadia ETS event. Most  
99 importantly, identifying these VLFs offers new insight into the physical nature of the gap  
100 zone.

## 101 2 Datasets and Methods

### 102 2.1 Detecting and Locating VLFs in Cascadia

103 We use an array-based surface wave detector that is developed from the AELUMA  
104 method (Automated Event Location Using a Mesh of Arrays) [de Groot-Hedlin and Hedlin,  
105 2015; Fan et al., 2018]. The method takes advantage of local coherence of the recorded sig-  
106 nals, and then forms an inverse problem to locate the signal sources assuming that the waves  
107 propagate along great circle paths [de Groot-Hedlin and Hedlin, 2015; Fan et al., 2018].  
108 Here we follow the same data processing protocol outlined in Fan et al. [2018] and use the  
109 same empirical parameters that have been implemented to investigate stormquakes and sub-  
110 marine landslides in the Gulf of Mexico [Fan et al., 2019, 2020]. The only difference is that  
111 we use 360 s time-window and 180 s time-step for the beamforming procedure instead of  
112 using the 600 s time-window and 300 s time-step as used in previous studies.

113 We first divide the large arrays into small subarrays, each comprising three stations.  
114 Second, tau-p beamforming analysis is applied to continuous data that are filtered in the 20 to  
115 50 s period band to detect signals, and the detections are screened through a quality control

116 procedure, such as examining cross-correlation coefficient, phase velocity, and beam-power  
117 value of each detection [see details in *Fan et al.*, 2018]. The records (LHZ component) are  
118 downloaded from available stations in the contiguous United States during the study period  
119 (see Open Research). We use vertical records because Rayleigh waves are used to detect the  
120 seismic sources, and vertical components generally have lower noise levels compared to hor-  
121 izontal components. Due to the signal to noise ratios and the quality control steps, not all  
122 records are used for the final location. Third, the remaining detections are grouped into non-  
123 overlapping clusters. Fourth, detections of each cluster are used to locate one seismic source  
124 and its location uncertainty is empirically estimated [*Fan et al.*, 2019, 2020]. During the lo-  
125 cation step, possible arrival angle anomalies are empirically corrected using earthquakes  
126 reported in the Global Centroid Moment Tensor Project [*Ekström et al.*, 2012]. Finally, the  
127 quality of each located seismic event is assessed to avoid duplicates and a catalog is popu-  
128 lated with the located events. Our approach is data-driven with few assumptions about the  
129 nature of the seismic sources. The AELUMA method is particularly well-suited for detecting  
130 unconventional seismic sources that are commonly missed in standard catalogs [*Fan et al.*,  
131 2019, 2020]. This is because the method applies to continuous waveforms and can detect and  
132 locate any source of seismic radiation without phase picks or knowing the source types [*Fan*  
133 *et al.*, 2018].

134 We detected three seismic sources (E1, E2, and E3) in Cascadia within 15 minutes af-  
135 ter the 2009 Canal de Ballenas earthquake in the Gulf of California, Mexico that are likely  
136 VLFs (Figures 1,2). Due to the temporal correlation between the earthquake surface wave  
137 arrivals and the detected sources, we hypothesize that the detected sources were triggered by  
138 the 2009 earthquake. We will discuss this hypothesis in later sections. These three VLFs  
139 were detected by 84, 57, and 187 subarrays, respectively. The larger temporal separation be-  
140 tween E3 and the 2009 earthquake likely allows for more detections of E3. Consequentially,  
141 E3 can be clearly seen in the record section when the traces are aligned with respect to its  
142 location (Figure 3). The location uncertainties for all three VLFs are shown as the dashed  
143 lines in Figure 1. The location uncertainty of the detected seismic sources are computed by  
144 examining the spatial structure of a suite of grids within a misfit threshold [*Fan et al.*, 2018].  
145 Based on the obtained location, grids that can minimize the misfit values within 25% of the  
146 minimum value are taken as possible source locations [*Fan et al.*, 2018]. From the set of pos-  
147 sible sources, we compute a distance covariance matrix and use its eigenvectors and eigen-  
148 values to define an uncertainty ellipse around the source location solution [*Fan et al.*, 2018]  
149 (Figure 2). This approach can provide a formal way to address statistical location uncertainty  
150 due to data availability. However, the misfit threshold is chosen subjectively. In our case, the  
151 25% of the optimal value is a conservative choice, and the results represent the lower-bound  
152 of the resolution. In later parts, we will evaluate the event locations with local strainmeter  
153 records to provide independent constraints on the results.

## 154 2.2 Determining Focal Mechanisms of the VLFs

155 We use a cross-correlation method to estimate the focal mechanisms (Figure 4). The  
156 approach shares similarities with the grid-search centroid moment tensor inversion method,  
157 which has been applied to search VLFs in Cascadia and offshore Japan [e.g., *Ito and Obara*,  
158 2006; *Ghosh et al.*, 2015]. The near-field stations in the Pacific Northwest (inset, Figure 1)  
159 are not used to analyze E3 due to the interference between its surface waves and those of the  
160 Canal de Ballenas earthquake (Figure S1).

161 Our method resolves the event focal mechanism, duration, and the event depth based on  
162 a VLF catalog (Figure 4). Instead of searching the possible combinations of the fault geom-  
163 etry (strike, rake, dip) and event depth, we fix the E3 epicenter as the resolved location from  
164 our surface wave detector and use a VLF catalog of events beneath southern Vancouver Is-  
165 land and northern Washington State [*Ide*, 2016] to forward calculate synthetic seismograms.  
166 Based on the amplitudes of the VLF-related waves, we initially assume the event has a  
167 seismic moment of  $2 \times 10^{18} \text{ N} \cdot \text{m}$ . The catalog has 112 events, and for each focal mecha-

168 nism (Figure 4), we compute three-component synthetic waveforms for sources at depth from  
169 5 km to 50 km with a 5 km increment. We also investigate a set of source durations assum-  
170 ing a Gaussian function shape with the duration as 6 times the standard deviation; we test  
171 durations from 0.9 s to 257.1 s.

172 The synthetic waveforms are computed for each station in the vertical, north-south, and  
173 east-west directions up to 3600 s with the Instaseis method [Driel *et al.*, 2015]. The Instaseis  
174 method pre-computes a Green's function database with the axisymmetric spectral-element  
175 method AxiSEM [Nissen-Meyer *et al.*, 2014]. Here, we use the Green's functions calcu-  
176 lated with the anisotropic version of the PREM model up to 5 s [Dziewonski and Anderson,  
177 1981]. These synthetic seismograms are then filtered at 25 to 50 s period band and are cross-  
178 correlated with the observed three-component waveforms (e.g., 300 s waveform outlined  
179 by the yellow band in Figure 3) of the best-resolved event, E3, in the same frequency band.  
180 We focus on stations east of the source with epicentral distances from 500 km to 3300 km  
181 (up to 30° epicentral distance, Figure 5). For each station, a representative cross-correlation  
182 coefficient is taken as the geometric mean of the cross-correlation coefficients of the three  
183 components (e.g., Figure 5), the preferred depth for the focal mechanism maximizes the total  
184 summation of the representative cross-correlation coefficients from all stations. The optimal  
185 solution, including both the focal mechanism and the event depth, has the maximum total  
186 summation of cross-correlation coefficients. After obtaining an optimal solution, we calcu-  
187 late the amplitude ratios between the synthetic waveforms and the observations for all the  
188 stations and components (Figure 4d), and the median value of the ratio distribution (0.25 for  
189 E3) is used to scale the initial seismic moment to compute the VLFE moment.

190 Given the noise level of the records, we can only estimate the focal mechanism for one  
191 of the detected seismic sources in Cascadia (E3), which has waveforms that are separated  
192 from the surface waves of the Canal de Ballenas earthquake (Figure 3). It is challenging to  
193 analyze events E1 and E2 in more detail because the high amplitude coda waves from the  
194 Canal de Ballenas earthquake mask the VLFE signal (Figure 3).

## 195 **2.3 Dynamic and Static Deformation**

196 With our starting estimates that seismic moments of the VLFEs are on the order of  
197  $10^{18}$  N · m, geodetic methods may detect the associated static deformation and verify our  
198 results. We use strainmeters because they are generally sensitive to static strains from small-  
199 to-moderate crustal earthquakes, and can give precise onset timing of the static deformation,  
200 unlike with more commonly used space geodetic techniques (i.e., GNSS, InSAR). They can  
201 also measure broadband dynamic strains from the Canal de Ballenas event, which allows us  
202 to robustly estimate dynamic stresses at the times of the detected seismic sources. We will  
203 later examine the relations between the dynamic stress and the observed VLFEs.

204 In particular, we use strain data from borehole strainmeters (BSMs) in the Network of  
205 the Americas (NOTA) (Figure 6a). These BSMs are four-component Gladwin-type differen-  
206 tial capacitance strainmeters [Gladwin, 1984]. Unprocessed data given in capacitance bridge  
207 counts are converted to linear strains using standard linearization procedure [Barbour and  
208 Crowell, 2017]. We outline the steps taken to analyze both dynamic strains from the source,  
209 and static strains from the VLFEs.

### 210 **2.3.1 Dynamic Strains from the Canal de Ballenas Earthquake**

211 The 2009 Mw 6.9 Canal de Ballenas earthquake was a strike-slip event in the north-  
212 central region of the Gulf of California, Mexico [Castro *et al.*, 2011]. The earthquake rup-  
213 tured a segment of an *en echelon* transform fault system with a shallow hypocenter close to  
214 the seafloor [Castro *et al.*, 2011; Plattner *et al.*, 2015]. The Canal de Ballenas earthquake  
215 generated strong Rayleigh waves, and the observed dynamic strains at NOTA stations were

216 between 2.1 and 15.3 (mean 7.3) times larger than those of most Mw 6.9 teleseisms, accord-  
217 ing to the relations of *Agnew and Wyatt* [2014].

218 For analyses of the teleseismic waves, we use the root-mean-square strain timeseries  $\epsilon$   
219 for the given time window, given by  $\epsilon = \sqrt{(\sum_{i=1}^n g_i^2) / n}$ , where  $g_i$  are the linear strain time-  
220 series for the  $n$  strain channels; for these BSMs,  $n = 4$  under normal operating conditions.  
221 We then calculate the peak value of the RMS strain timeseries,  $\hat{\epsilon}$ , after applying a two-pass  
222 Butterworth highpass filter with a corner frequency of 0.004 Hz (250 s period) to mask out  
223 all non-seismic signals that strainmeters have well-known sensitivities to (e.g., tides, atmo-  
224 spheric pressure, etc.); this is the peak dynamic strain (PDS).

225 Following *Hill* [2008], we estimate peak dynamic stress ( $\hat{\delta}$ ) as the observed PDS scaled  
226 by twice the crustal shear modulus  $\mu$  ( $\hat{\delta} = 2\mu\hat{\epsilon}$ ). We use  $\mu = 30 \times 10^9$  Pa to be consistent with  
227 the crustal velocity and density model used to locate the VLFES. This is a simplistic estimate  
228 of the true stress perturbation, which might be larger if the event occurred where material  
229 properties contrast strongly; however, at Cascadia, contrasts in S-wave velocity ( $V_S$ ) at the  
230 slab interface are generally within a few percent [*Porritt et al.*, 2011], which translates to a  
231 smaller perturbation in  $\mu$ , given that  $\mu = \rho V_S^2$ , where  $\rho$  is density.

### 232 2.3.2 Static Strains from the Local VLFES

233 Theoretically, the lowest detectable static strain is about  $0.1\text{--}0.2 \times 10^{-9}$  [parts-per-  
234 billion (ppb), or nanostrain]. Following *Wyatt* [1988] this implies that strain from an event  
235 with  $10^{18}$  N · m seismic moment will be undetectable beyond a few hundred kilometers.  
236 However, because of noise and other unrelated signals, the practical limit of detection of  
237 an event of this size is  $\sim 100\text{--}130$  km. Relative to the location of VLFE event E3, this lim-  
238 itation leaves 14 possible NOTA stations. However, data from four of these stations are ei-  
239 ther unavailable or too contaminated with non-seismic signals such that only stations B003,  
240 B004, B014, B007, B001, B013, B009, B010, B011, and B926 are useful for analyzing static  
241 strains (Figure 7).

242 The distances from these stations to VLFE event E3 range from 34 to 116 km, which  
243 implies that static strains will be much less than 100 ppb [*Wyatt*, 1988]; at these levels, the  
244 observed PDS from the 2009 Canal de Ballenas earthquake is at least 3–4 times but possibly  
245 10–100 times larger than the static signal from the VLFE. For this reason, we first detrended  
246 the records based on the data seen between the origin time and the first surface-wave arrivals;  
247 we then applied a causal, lowpass filter [*Agnew and Hodgkinson*, 2007] to the detrended  
248 records to preserve the time-independence of these signals, for comparison with the timing  
249 of the VLFES. Static offsets are computed from these filtered, detrended strain records ( $g$ ),  
250 and are then transformed to tensor strain values ( $E$ ) using the coupling equation:

$$251 \quad E = Cg \quad (1)$$

As described above  $g$  is a matrix of strain timeseries from the instrument's 4 strain gauges:

$$252 \quad g = [g_1, g_2, g_3, g_4]' \quad (2)$$

253 The matrix  $C$  is a  $3 \times 4$  matrix of calibration coefficients determined by tidal analyses [e.g.,  
254 *Hodgkinson et al.*, 2013]; it transforms  $g$  into tensor strain components, with coefficients  
255 that vary by station. The resulting matrix  $E$  contains the areal strain and two engineering  
256 shear strains in an east-north (e-n) coordinate reference system, where extensional strains are  
positive:

$$257 \quad E = [E_{ee} + E_{nn}, E_{ee} - E_{nn}, 2E_{en}]' \quad (3)$$

258 For instance, the value  $E_{ee}$  represents uniaxial, extension in the east direction. Thus, the  
259 rms extension is found through the quadrature sum of the components of  $0.5E$ , or  $E_{\text{RMS}} =$   
 $\sqrt{(E_{ee}^2 + E_{nn}^2 + E_{en}^2)/3}$ , and similarly the shear components of  $0.5E$  give the maximum shear  
260 strain:  $\tau_{\text{max}} = \sqrt{(E_{ee} - E_{nn})^2/4 + E_{en}^2}$ . The calibration matrices ( $C$ ) used for these strain-  
261 meters are from *Roeloffs* [2010] and *Hodgkinson et al.* [2013] as detailed in Table S3.

### 3 Results

In Cascadia, we detect three new seismic sources that are likely VLFs (Figure 2). These sources generated coherent, transcontinental wavefields, and were detected by our surface wave detector [*de Groot-Hedlin and Hedlin, 2015; Fan et al., 2018*]. With the measured centroid times and surface-wave propagation directions, we identify three seismic sources offshore Cascadia, E1–E3 (Figure 2). The VLFs coincide with surface wave arrivals of the 3 August 2009 Mw 6.9 Canal de Ballenas earthquake.

The best resolved event (E3) occurred at 18:13:10 UTC, ~764 seconds after the Canal de Ballenas earthquake origin time; its epicenter is near the entrance of the Strait of Juan de Fuca, ~2360 km away from earthquake epicenter (Figure 1). Therefore, E3 occurred coinciding with the earthquake Rayleigh waves, assuming a group velocity of 3–3.5 km/s. Its coherent waveforms can be clearly identified from the aligned traces (Figure 3). All of the VLF events (E1–E3) occurred immediately after the passing seismic waves from the Canal de Ballenas earthquake, and were most likely dynamically triggered by the earthquake. We could not analyze seismic data from stations in the near-field confidently because of the near-instantaneous triggering responses: the long lasting coda waves of the Canal de Ballenas earthquake masked signals of E1–E3 at stations in the Pacific Northwest (Figures 3, S1). However, none of the events (E1–E3) produced visible, high-frequency body-wave phases (Figure S1), nor are they listed in standard catalogs, refuting the possibility that instead they are regular earthquakes. Such a lack of high-frequency seismic radiation suggest that these sources are likely VLFs.

We verify this hypothesis by modeling the E3 focal mechanism (Figure 4). The waveforms associated with E3 are well-separated from the surface waves of the Canal de Ballenas earthquake in the far-field, permitting such an analysis; the other two VLFs are too difficult to model due to the poor signal-to-noise ratios of the records (Figures 3, S2). Therefore, we focus our discussions on event E3 in this study and only report the detections of E1 and E2 (Figure 2). The preferred solution suggests that E3 lasted less than 20 s (a point source) and has a mechanism with a strike of 125°, dip of 1°, and rake of -117° at a depth of 15 km (Figure 5). The subhorizontal dip of E3 deviates away from the local slab geometry and likely suffers uncertainties.

The focal mechanism and depth solution suffers from uncertainties because it is based on a catalog, and the teleseismic surface waves used for the analysis were filtered in a narrow period-band (Figure 4). Since the likely depth range for E3 is 15–25 km (Figures 6 and 8), it is difficult to determine how the source depth deviates from the plate interface geometry [*Hayes et al., 2018*]. However, it is worth noting that the E3 depth range is shallower than the tremor and slow slip zone depths of 30–50 km [*Bartlow et al., 2011; Bartlow, 2020*]. After resolving the focal mechanism and the event depth, we use the amplitude ratios between the synthetic waveforms and the observations to estimate the VLF moment magnitude. The E3 event has an estimated moment magnitude ( $M_w$ ) of 5.7 ( $0.5 \times 10^{18}$  N · m, Figure 4), which is much larger than those of other VLFs ( $M_{2.1-4.1}$ ) in the region [*Hutchison and Ghosh, 2016; Ide, 2016*].

At multiple stations near E3, we observe static strain offsets after the E3 occurrence (Figure 6). A table of observed offsets can be found in the Supplement. We ruled out the possibility that these are spurious strains [e.g., *Barbour et al., 2015*] by confirming the absence of static offsets at distant stations in the region with similar dynamic strain amplitudes (Figure 9). We also note that the observed static strains are not apparent until soon after the seismically-determined origin time of E3 (Figure 6). With the source parameters, we model the static strains due to E3 with an edge dislocation in an elastic halfspace [*Okada, 1985*], and compare the model-predictions with observations at nearby strainmeter stations of the NOTA network (Figures 6 and 7). With the exception of station B003, the overall spatial pattern of the observed static strains from the other nine stations is consistent with the synthetic strains. This confirms the event E3 and its source model, suggesting that these strain data

314 represent the first set of direct observations of static crustal deformation associated with a  
315 VLFE at any subduction zone.

## 316 4 Discussion

### 317 4.1 Resolutions and Uncertainties

318 The detected VLFEs in Cascadia are unlikely to be data artifacts: their radiated surface  
319 waves, particularly from E3, span most of the United States (Figure 2) and the direct geode-  
320 tic observations conclusively confirm E3 and that the event occurred near the seismically  
321 determined location (Figures 6 and 8). Further, E3 can be directly identified from aligned  
322 waveforms, furthering confirming its location (Figure 3).

323 In addition to triggered seismic events, heterogeneous subsurface structure can cause  
324 a secondary coherence surface wavefield by reflecting or converting the incoming waves  
325 [Obara and Matsumura, 2010; Maeda et al., 2014; Buehler et al., 2018; Yu et al., 2017,  
326 2021]. Here we considered the possibility that the detected VLFEs are actually scattered  
327 energy from the Canal de Ballenas event seismic waves rather than a unique local source.  
328 For example, the observations could have been from S-wave to surface-wave conversions  
329 that have been observed from the US west coast [Buehler et al., 2018; Yu et al., 2021]. How-  
330 ever, we found that this hypothesis violates the observations in a number of ways. First, if  
331 the VLFEs are S-to-Rayleigh wave or P-to-Rayleigh wave scatterers, these seismic sources  
332 would occur upon the body wave arrivals. However, the observations show that the detected  
333 seismic sources occurred after the surface waves (Figure 3). Second, previous surface-wave  
334 reflections from a single scatter would last longer than 200 s [Obara and Matsumura, 2010],  
335 which contradicts to what we observe for the surface waves of E3 in Figure 3 (duration of  
336 E3 is less than 20 s). Lastly, if the detected triggered sources are structural heterogeneities  
337 (scatterers), they would cause the same scattering for earthquakes from the same region. In  
338 that case, the seismic sources detected by ALEUMA would be located at the same location  
339 and the measured propagation directions would be identical after large triggering earthquakes  
340 from the nearby region [Obara and Matsumura, 2010]. However, this is inconsistent with  
341 our observations (Figure S3). We observe no triggered seismic sources in Cascadia after the  
342 2010 Mw 7.2 El Mayor-Cucapah earthquake, 2012 Mw 7.0 Baja California earthquake, or  
343 the 2019 Mw 7.0 Ridgecrest earthquake (Figure S3). These observations falsify the scattering  
344 hypothesis and confirm the observed VLFEs, particularly E3.

345 Overall there is good spatial agreement between the observed static strains and mod-  
346 eled strains (Figure 7a), and there is no apparent influence from peak dynamic strain levels  
347 (Figure 7b). However, observations from the closest station, B003, are notably smaller than  
348 suggested by the dislocation model. We believe this misfit arises for a few important reasons,  
349 namely the effects of ocean loading on the earth tides, and errors in source mechanism pa-  
350 rameters. The loading of the crust by the ocean causes highly nonlinear distortions in the  
351 phases and amplitudes of tidal constituents near the land-sea interface [e.g., Lambert, 1970;  
352 Farrell et al., 1973]. With strainmeter data located near the coast, it is notoriously difficult to  
353 model tidal strains needed for an accurate calibration procedure [see Hodgkinson et al., 2013;  
354 Kamigaichi et al., 2021]. Consequently, there are significant uncertainties in the strainmeter  
355 calibration coefficients in this region owing to their proximity to the coast. At present, the  
356 tidal models for all the stations considered here are adequate, with the exception of B003 [see  
357 Roeloffs, 2010; Hodgkinson et al., 2013; Reuveni et al., 2014]. The second major influence in  
358 near-source results comes from variations in focal mechanism parameters; these have a sim-  
359 ilarly strong influence on the spatial pattern of coseismic strain in the near-to-intermediate  
360 field, layered structure notwithstanding. For instance, the locations of nodes of null deforma-  
361 tion are strongly influenced by strike and dip, and some stations are located close to these  
362 nodes, between lobes of significant deformation (e.g., Figure 6a) where small changes in  
363 strike or dip would have the strongest effect. Unfortunately, there are too few strainmeters  
364 near E3 to perform an independent source inversion; but, the current set of static strain obser-



365 vations can conclusively confirm the E3 occurrence and its relatively large moment magni-  
366 tude.

367 Instead, to independently test the seismic-based location of E3, we forward modeled  
368 the same focal mechanism at every point on the Slab2.0 subduction zone interface [*Hayes*  
369 *et al.*, 2018] and calculated the source likelihood from the strain data misfit. Owing to station  
370 coverage there is a relatively broad zone of plausible source locations (Figure 7c). The uncer-  
371 tainty in the geodetic location is unavoidably large because additional stations are needed to  
372 rigorously constrain the source based on the static deformation pattern. There are two sepa-  
373 rate areas with a reduced Chi-squared ( $\chi^2$ ) misfit equal to 1, meaning that a source like E3  
374 located in either region would fit the data equally well. One area sits within the seismic-  
375 derived uncertainty whereas the other falls far outside of it (Figure 7c). The seismic loca-  
376 tion uncertainty is estimated based on a very conservative criteria (within 25% of the misfit  
377 minimum instead of the commonly used 5% threshold), which likely overestimates the un-  
378 certainty ellipse; thus, the most likely location based on strain observations alone is within  
379 tens of kilometer of the seismic location (Figures 7c and 8). Further, teleseismic surface  
380 waves are sensitive to source depth, and a change of E3 depth from 15–20 km to 30–50 km  
381 or greater is also highly unlikely (Figure 4b).

382 We also compared the timing of the surface waves and VLFE detections with long-  
383 term strain records in Cascadia and the detected tremor events from the World Tremor Database  
384 (WTD)[*Idehara et al.*, 2014] in Figure 9. These data cannot definitively rule out deformation  
385 signals related to slow slip occurring prior to 2009/8/3, but they do show that if slow slip re-  
386 lated to the 2009 ETS event initiated before these arrivals, the strain signals are undetectable  
387 relative to the non-tectonic noise seen at these stations. Further, the tremor rate increases  
388 roughly 10 hours after the triggered VLFE events, as the slow slip event is apparently de-  
389 veloping; this is juxtaposed by a multi-day quiescence and a lack of slip-related signals in  
390 GNSS data prior to the passing seismic waves (Figure 9c).

## 391 4.2 Triggering and Interaction

392 Slow earthquakes interact and trigger each other frequently [*Obara and Kato*, 2016].  
393 For example, slow slip events can drive tremor, causing ETS events in Cascadia [*Rogers and*  
394 *Dragert*, 2003; *Bartlow et al.*, 2011], and VLFEs have been triggered by long-term SSEs  
395 offshore Japan [*Hirose et al.*, 2010; *Araki et al.*, 2017; *Katakami et al.*, 2020]. The close spa-  
396 tiotemporal correlation between the observed VLFEs and the 2009 slow slip event in Casca-  
397 dia suggests that they are likely physically related [*Rubinstein et al.*, 2009]. One possibility is  
398 that these large-magnitude VLFEs, caused by the passing seismic waves, may have initiated  
399 a slow slip event which eventually developed into the 2009 Cascadia ETS event. Presently,  
400 we cannot confirm this cascading process, as neither GNSS stations or borehole strainmeters  
401 in Cascadia detected slip-related deformation before the 2009 ETS event above background  
402 noise levels (Figure 9). Such ambiguity cannot exclude that there were slow slips triggered  
403 by the passing seismic wave, which would suggest that the 2009 ETS event was triggered by  
404 the Canal de Ballenas earthquake. A long-term, systematic investigation of the onset timing  
405 of Cascadia ETS is necessary to illuminate the susceptibility of slow slip events in the region  
406 due to remote earthquakes. Despite the ambiguity in the timing of the 2009 ETS event rela-  
407 tive to the VLFEs, our observations suggest that complex slip interactions may occur more  
408 frequently at Cascadia than previously documented.

409 Slow earthquakes can be susceptible to triggering due to small external stress perturba-  
410 tions [*Obara and Kato*, 2016; *Katakami et al.*, 2020; *Araki et al.*, 2017], which is best illus-  
411 trated by the sensitivity of tremor occurrence to Earth tides and passing seismic waves [*Ru-*  
412 *binstein et al.*, 2008, 2009; *Hawthorne and Rubin*, 2010; *Chao et al.*, 2013; *Houston*, 2015;  
413 *Miyazawa*, 2019]. For example, remote triggering of VLFEs by surface waves from a moder-  
414 ate to large, distant earthquake has been reported in the Nankai subduction zone [*Miyazawa*,  
415 2019]. Passing seismic waves also triggered aseismic slip events on the San Andreas fault

416 that led to migrating tremor [*Shelly et al.*, 2011]. Such dynamically triggered cascading slip  
417 events may be similar to what we observe in this study. With direct measurements of dy-  
418 namic strain, we estimate the dynamic stresses associated with the passing seismic waves.  
419 Assuming a shear modulus of 30 GPa, the dynamic, elastic stress perturbations from the  
420 Canal de Ballenas earthquake were likely ~20–30 kPa at E3. The true triggering stresses  
421 at E3 could vary within a few percent depending on the depth dependence of surface waves,  
422 fault geometry, fault frictional properties and pre-stress, and dynamic pore pressure effects.  
423 Other earthquakes with similar magnitudes near the Canal de Ballenas earthquake did not  
424 trigger events in Cascadia (Figure S3), suggesting that earthquake rupture process might play  
425 an important role in generating strong teleseismic ground motions that can cause dynamic  
426 triggering cases.

427 The observed triggering process suggests that the E3 patch in the fault gap was at a  
428 critical state prior to the surface wave arrivals. Alternatively, the fault could have been very  
429 weak, such that the dynamic stress changes from the Canal de Ballenas earthquake were  
430 sufficient to trigger an unstable dynamic rupture; in that case, triggered VLFs would be  
431 a commonly-observed phenomenon rather than the rarely reported phenomenon they are  
432 presently [*Miyazawa*, 2019]. Nonetheless, in this study, the observed VLFs show that in  
433 between the seismogenic zone and the ETS zone, some patches of the megathrust fault gap  
434 are capable of hosting M5.7 seismic events that are sensitive to transient stress perturbations.

#### 435 4.3 Physical Conditions in Between the Seismogenic and Tremor Zones

436 Event E3 occurred at depths shallower than other slow earthquakes in the Cascadia  
437 subduction zone [*Gomberg et al.*, 2010; *Brudzinski and Allen*, 2007]. Interestingly, neither  
438 tremor nor slow slip signals were detected in the region during these triggered VLFs [*Wech*  
439 *and Creager*, 2008; *Bartlow et al.*, 2011]; this behavior differs from typical VLFs in this  
440 region that are often coincident with tremor and slow slip [e.g., *Hutchison and Ghosh*, 2016;  
441 *Ide*, 2016]. The relatively shallow depth of E3 corresponds to the deepest part of the locked  
442 zone – a gap in between the seismogenic zone and the tremor zone [*Hyndman and Wang*,  
443 1995; *Wang et al.*, 2003; *Priest et al.*, 2010; *Schmalzle et al.*, 2014; *Bruhat and Segall*, 2016].  
444 In northern Cascadia, slow slip events have penetrated upward into this gap zone during previ-  
445 ous ETS events, but tremor has been scarce there [*Wang et al.*, 2008; *Wech et al.*, 2009;  
446 *Hall et al.*, 2018]. Further, sporadic weak slips are observed in this gap zone across all of  
447 Cascadia [*Bartlow*, 2020; *Nuyen and Schmidt*, 2021].

448 Given the magnitude and location of E3, our observed VLFs may relate to these  
449 aseismic slips, and in combination, their slip contributions may be analogous to the longer-  
450 term SSEs in Nankai, where tremors do not overlap with SSEs in such a gap zone [*Hirose*  
451 *et al.*, 2010; *Kobayashi*, 2014]. However, the occurrence of E3 also suggests that this gap  
452 region is not creeping steadily or slipping aseismically [*Holtkamp and Brudzinski*, 2010;  
453 *Hyndman*, 2013; *Schmalzle et al.*, 2014]. Instead, the gap zone is likely partially locked and  
454 rupture might be able to propagate sufficiently fast at some patches. Such a locking transition  
455 zone is physically intuitive, but the type of seismic slip events needed to confirm its proper-  
456 ties are rarely observed in the region. The magnitude of E3 indicates that the fault's strength  
457 may increase in the gap zone, compared to the ETS zone, suggesting that loading stresses  
458 from the slow slip events fail to generate seismic failures during conventional Cascadia ETS  
459 events, possibly due to the size of the locked fault patch and frictional properties [*Hall et al.*,  
460 2018].

461 Conceptually, the gap zone may represent a semi-frictional phase connecting the seis-  
462 mogenic and the tremor zone [*Gao and Wang*, 2017], which may be comprised of large  
463 strong fault patches that can generate VLFs, but with a lower level of asperity heterogeneity  
464 on fault than the tremor zone [e.g., *Wu et al.*, 2019], surrounded by a ductile matrix [*Chestler*  
465 *and Creager*, 2017] that could slip aseismically [*Wang et al.*, 2008; *Hall et al.*, 2018]. There-  
466 fore, our observed VLFs could be a manifestation of a transitioning semi-frictional be-

467 haviour as an intermediate stage between seismic rupture and viscous creep [*Gao and Wang,*  
 468 2017]. Such cases may explain our observed VLFs with a lack of tremor activity in the  
 469 gap zone, presumably related to the nonstationary shear stress rates inferred from decadal-  
 470 averaged crustal deformation rates [*Bruhat and Segall,* 2016].

471 From the seismogenic zone to the tremor zone in northern Cascadia, there is a change  
 472 in the reflection character on seismic images from a thin sharp reflection band to a broad  
 473 reflection band [*Nedimović et al.,* 2003]. The seismic images were obtained from active  
 474 surveys that were ~50 km north of E3, and the images show prominent broad reflectors at  
 475 a depth range of 15–25 km, coinciding with the gap zone [*Nedimović et al.,* 2003]. Such  
 476 reflectors are regionally extensive and may represent interlayered mafic and/or sedimentary  
 477 rocks or intensely sheared sediments with trapped fluids [*Yorath et al.,* 1985; *Clowes*  
 478 *et al.,* 1987; *Calvert and Clowes,* 1990], and E3 seems to relate to these reflectors. Shear-  
 479 ing and fluid can lead to prevalent ductile banding there, which would produce a thick semi-  
 480 frictional and semi-ductile zone where either frictional or ductile slip can occur depending  
 481 on the local conditions [*Nedimović et al.,* 2003; *Gao and Wang,* 2017]. In such a scenario,  
 482 some kilometer-scale sub-horizontal fault patches that are confined with high fluid pressure  
 483 would preferably slip seismically, which might have been the structural cause of the observed  
 484 VLFs.

485 Seismic sources in the region such as E3 might be used to infer dimensions and stress  
 486 states of the locked fault patches. In conjunction with the sporadic weak slip in the zone [*Bart-*  
 487 *low,* 2020; *Nuyen and Schmidt,* 2021], the few VLFs (E1–E3) seem to suggest that fault  
 488 properties of the gap zone are highly heterogeneous. In that case certain patches could slip  
 489 fast, possibly allowing a deep rupture extent at these spots, which could change the distribu-  
 490 tion of strong shaking and cause more intense ground motion along the densely populated  
 491 margin [*Melgar et al.,* 2016; *Frankel et al.,* 2018; *Wirth et al.,* 2018; *Ramos and Huang,*  
 492 2019]. However, given the relatively large location uncertainties for these VLFs, future sys-  
 493 tematic investigations are warranted to delineate the downdip rupture limit of large megath-  
 494 rust earthquakes in Cascadia.

#### 495 **4.4 Breakdown of the Slow Earthquake Scaling Relationship**

496 It has been suggested that slow earthquakes, including VLFs, follow an apparent  
 497 moment-duration scaling relationship where the moment rate of these events is likely con-  
 498 stant and the final seismic moment is proportional to the characteristic duration [*Ide et al.,*  
 499 2007]. This would be different from the scaling of regular earthquakes, for which moment  
 500 scales linearly with the cube of the characteristic duration [*Abercrombie,* 1995; *Houston,*  
 501 2001]. The difference was hypothesized to arise from different rupture dynamics between  
 502 slow and regular earthquakes [*Ide et al.,* 2007]. Further, slow earthquakes are thought to rup-  
 503 ture faster than the plate movement rate, but their rupture speeds cannot accelerate to those  
 504 of typical earthquakes [*Gao et al.,* 2012; *Bletery et al.,* 2017]. If those empirical scaling rela-  
 505 tionships hold true, we would expect a M5.7 VLFE or slow earthquake to last from 6 days to  
 506 a month [*Ide et al.,* 2007]. Consequentially, the rupture speed of such an event would be too  
 507 slow to generate seismic signals that can be observed in the far field [*Gao et al.,* 2012].

508 Here, seismic and geodetic observations directly refute such slow earthquake scaling  
 509 relationships: waveform modeling shows that E3 was likely a transient event, whose duration  
 510 is much less than the predicted duration from the slow earthquake scaling relationship (Fig-  
 511 ure 4) but is in closer agreement with the scaling of regular earthquakes [*Houston,* 2001],  
 512 and static strains developed within the duration timescale (Figure 6). Our observations agree  
 513 well with recent findings showing that slow slip events and regular earthquakes follow simi-  
 514 lar scaling laws that the slow slip moment scales with the slip duration cubed [*Michel et al.,*  
 515 2019; *Frank and Brodsky,* 2019]. Numerical simulations also show that slow self-arresting  
 516 rupture within the rupture patch would generate VLFs with source properties following the  
 517 cubed duration scaling [*Wei et al.,* 2021]. Collectively, these results show that slow and regu-

lar earthquakes are likely governed by similar dynamic properties [Michel *et al.*, 2019; Frank and Brodsky, 2019]. Alternatively, if the events detected here are not actually VLFs, event E3 is still not a typical earthquake as no high-frequency seismic radiation was observed at seismic stations in the Pacific Northwest (Figures S1 and S4). Our reported VLFs seem to be distinct from other Cascadia VLFs or slow earthquakes in Nankai [Ghosh *et al.*, 2015; Hutchison and Ghosh, 2016; Obara and Kato, 2016; Ide *et al.*, 2007]: they do not seem to be accompanied by tremor or SSEs but are independent slip events  $\sim 10$ – $15$  km shallower than typical VLFs in the region. Further, E3 is the largest magnitude that has been reported at all subduction zones, producing measurable static geodetic signals. Our findings raise new questions about the physical nature of the gap zone: is there a new class of slip events that represents a bridge between future megathrust earthquakes and tremor in Cascadia?

## 5 Conclusions

By analyzing continuous data from seismic stations across the United States, we identify and locate 3 previously unknown VLFs that are close to the 2009 Cascadia slow slip event and that occurred roughly three days before the initiation of the associated tremor activity. One VLF is located in the critical gap zone with a moment magnitude of 5.7, which is the largest VLF that has been identified across all subduction zones. Further, this is the first time that a VLF is recorded geodetically, with an array of strainmeters showing clear deformation signals associated with the event. Our findings suggest that some patches of the gap zone are capable of hosting large, fast slip events, indicating possible deep ruptures at sporadic spots in Cascadia. Our observed VLFs also show that the Cascadia megathrust is weak and is sensitive to transient stress perturbations. Lastly, the identified VLFs challenge the current understanding of slow earthquake physics, with characteristics that deviate away from the empirical scaling relations of slow earthquakes.

## Open Research

The seismic data were provided by Data Management Center (DMC) of the Incorporated Research Institutions for Seismology (IRIS). The facilities of IRIS Data Services, and specifically the IRIS Data Management Center, were used for access to waveforms, related metadata, and/or derived products used in this study. IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience and EarthScope (SAGE) Proposal of the National Science Foundation (NSF) under Cooperative Agreement EAR-1261681. High-frequency strain data from the Network of the Americas (NOTA) network were also obtained from the IRIS DMC; this material is based on services provided by the Geodesy Advancing Geosciences and EarthScope (GAGE) facility, operated by UNAVCO, Inc., with support from the NSF and the National Aeronautics and Space Administration (NASA) under NSF Cooperative Agreement EAR-1724794. NOTA Level 2 strain and GNSS data were obtained from UNAVCO web-services (<https://www.unavco.org/data/web-services/web-services.html>). The tremor catalog is obtained from the Pacific Northwest Seismic Network and the World Tremor Database (<http://www-solid.eps.s.u-tokyo.ac.jp/~idehara/wtd0/welcome.html>). The earthquake catalogs used in this study are from the Global Centroid Moment Tensor project (GCMT) [Ekström *et al.*, 2012].

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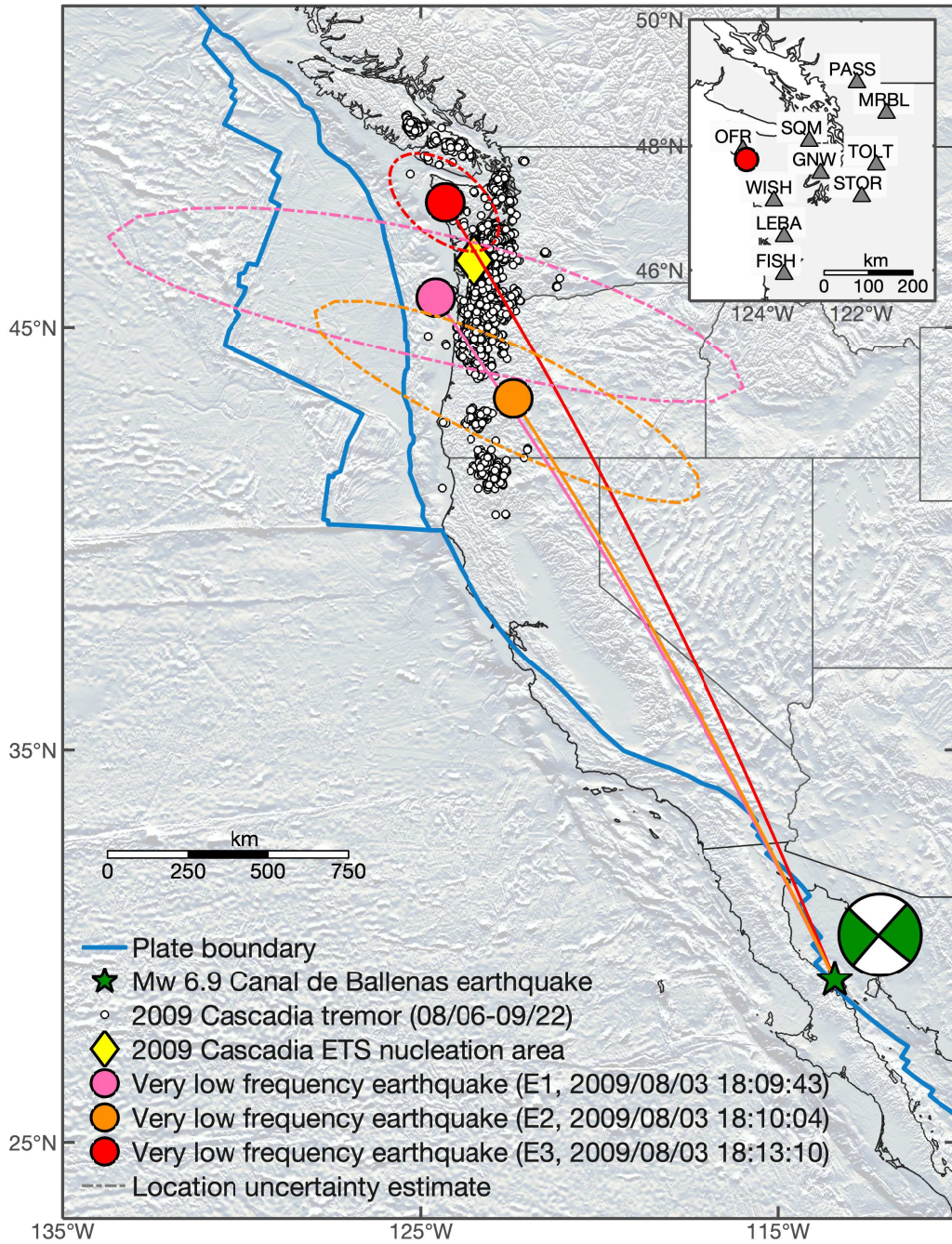
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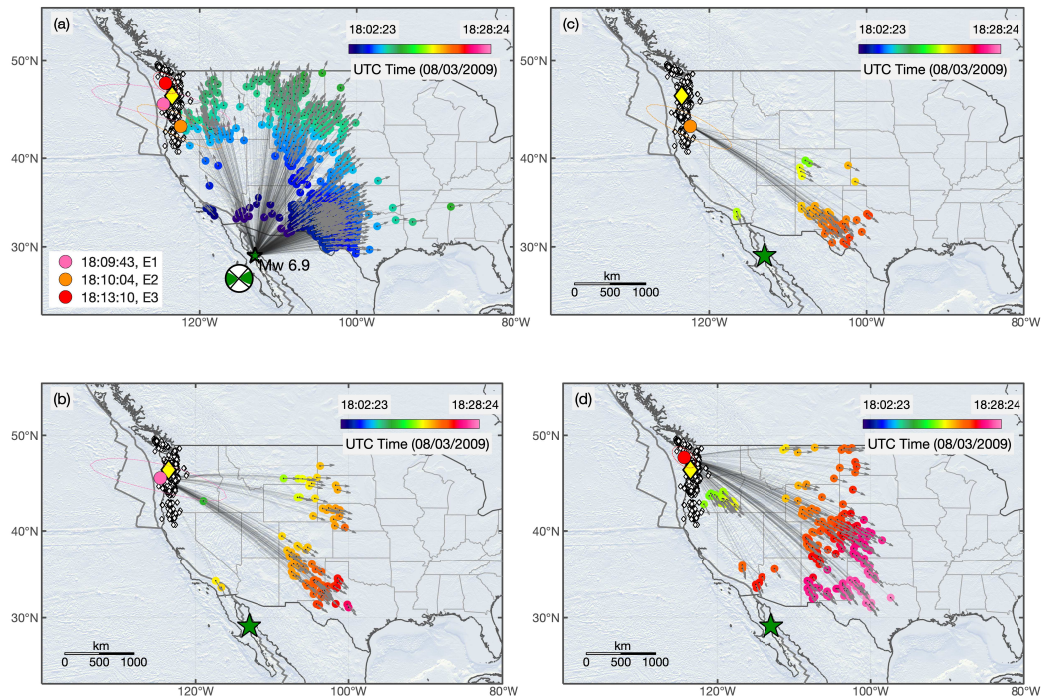
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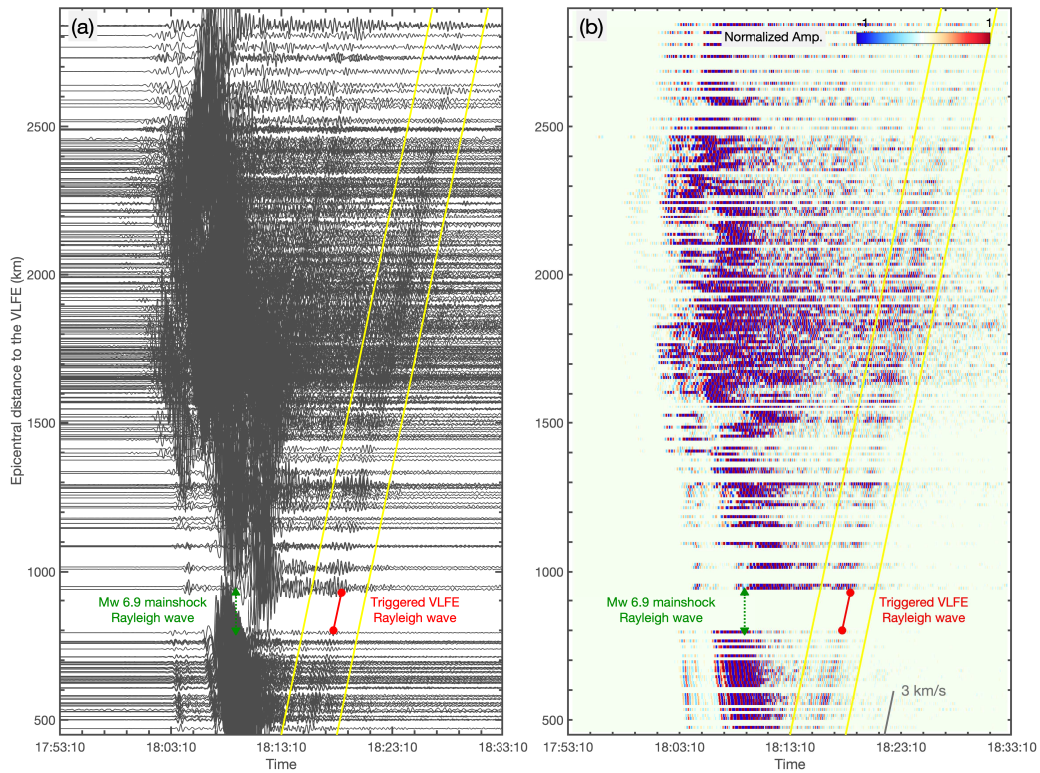
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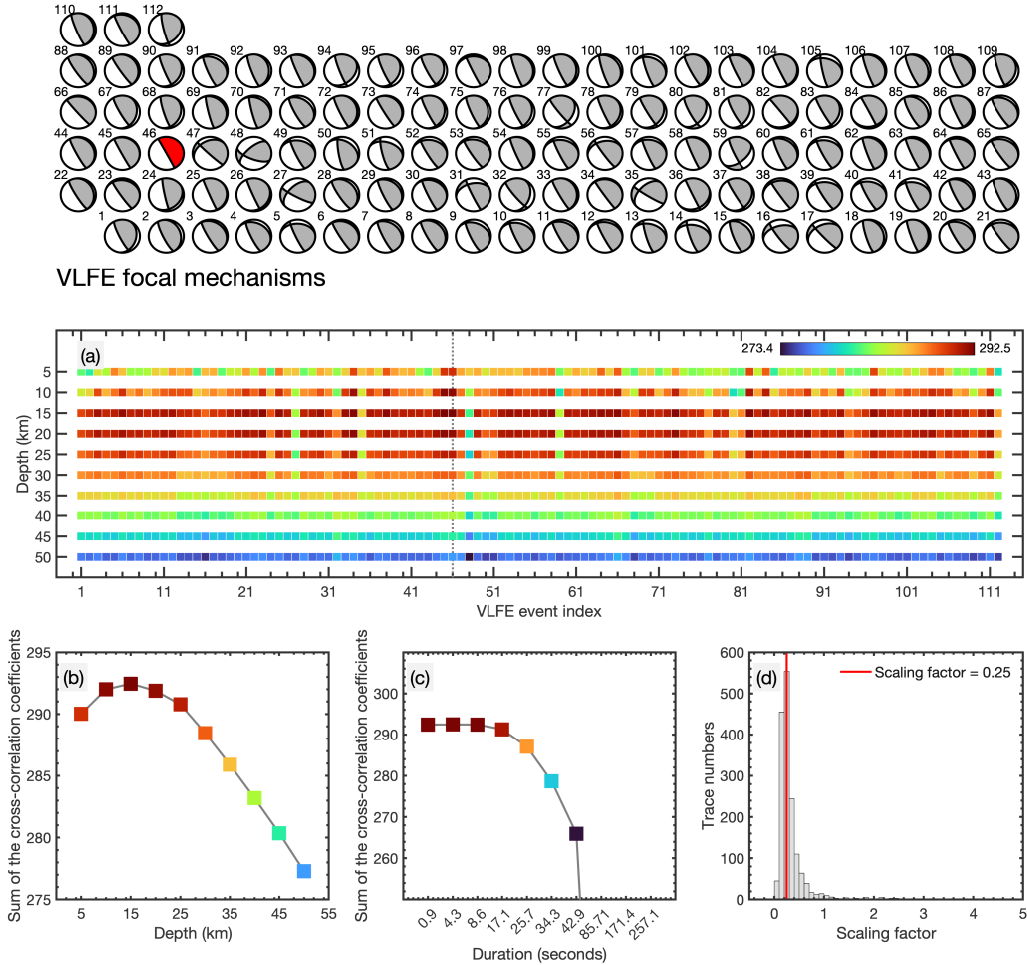
**Figure 1.** The 2009 M6.9 Canal de Ballenas earthquake, the 2009 Cascadia episodic tremor and slip (ETS) event, and three dynamically triggered very low frequency earthquakes (VLFEs, E1–E3). The tremor catalog is from the Pacific Northwest Seismic Network (PNSN). Inset: broadband near-field stations in the Pacific Northwest.



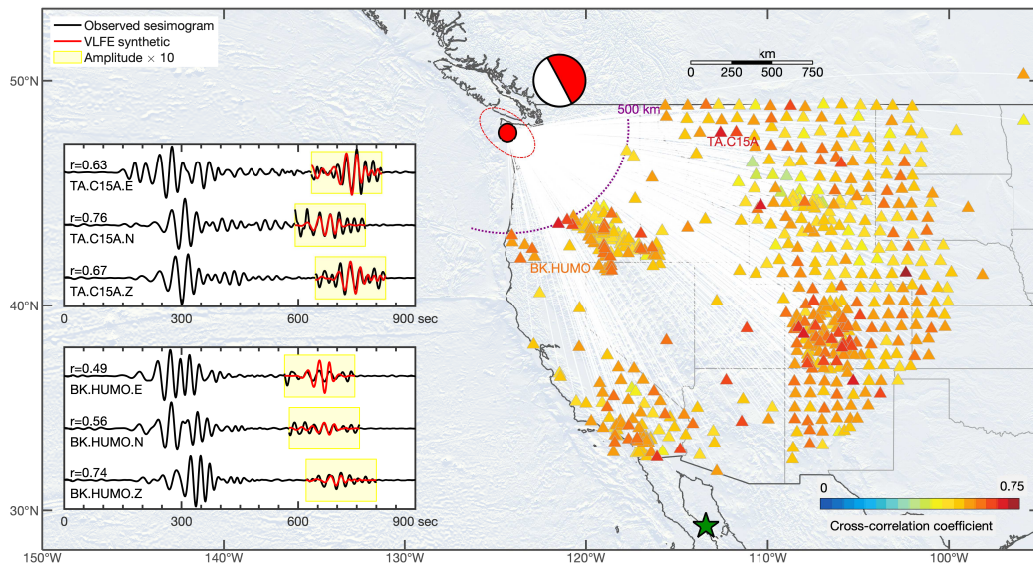
837 **Figure 2.** The very low frequency earthquakes and the triggering 2009 M6.9 earthquake (a). The legends  
 838 are similar to those of Figure 1. The Rayleigh wave arrival times and propagation directions are shown as  
 839 the colored dots and arrows. The thin gray lines show the great circle paths from the source to the subarrays.  
 840 The four events share the same colorbar. These three VLFs were detected and located by 84 subarrays at  
 841  $45.67^{\circ}/-124.58^{\circ}$  (b), by 57 subarrays at  $43.37^{\circ}/-122.41^{\circ}$  (c), and by 187 subarrays at  $47.78^{\circ}/-124.32^{\circ}$  (d).



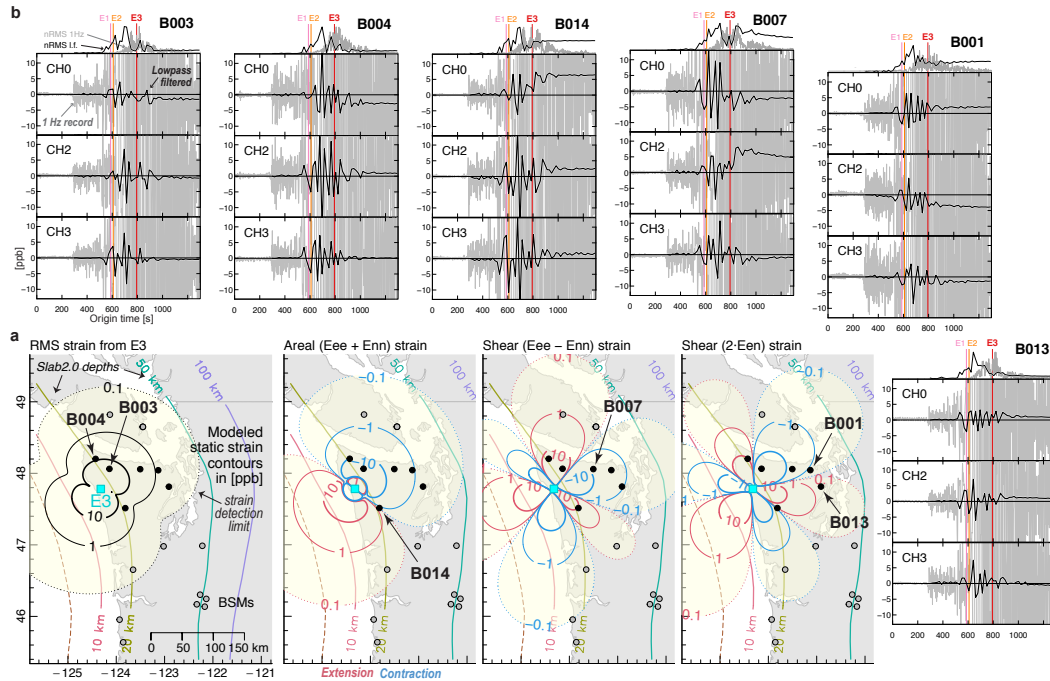
842 **Figure 3.** Record sections that are aligned with the epicenter of the VLFE E3 in Figure 2d. The records  
 843 are self-normalized and bandpass-filtered to show signals in the 20–50 s period band. The yellow lines show  
 844 a 3 km/s reference move-out velocity, windowing the VLFE waveforms. (a), waveform records. (b), polarity  
 845 plot of the records that red color indicates positive polarity while blue color indicates negative polarity.  
 846 Coherent phases can be easily identified by tracing coherent polarities.



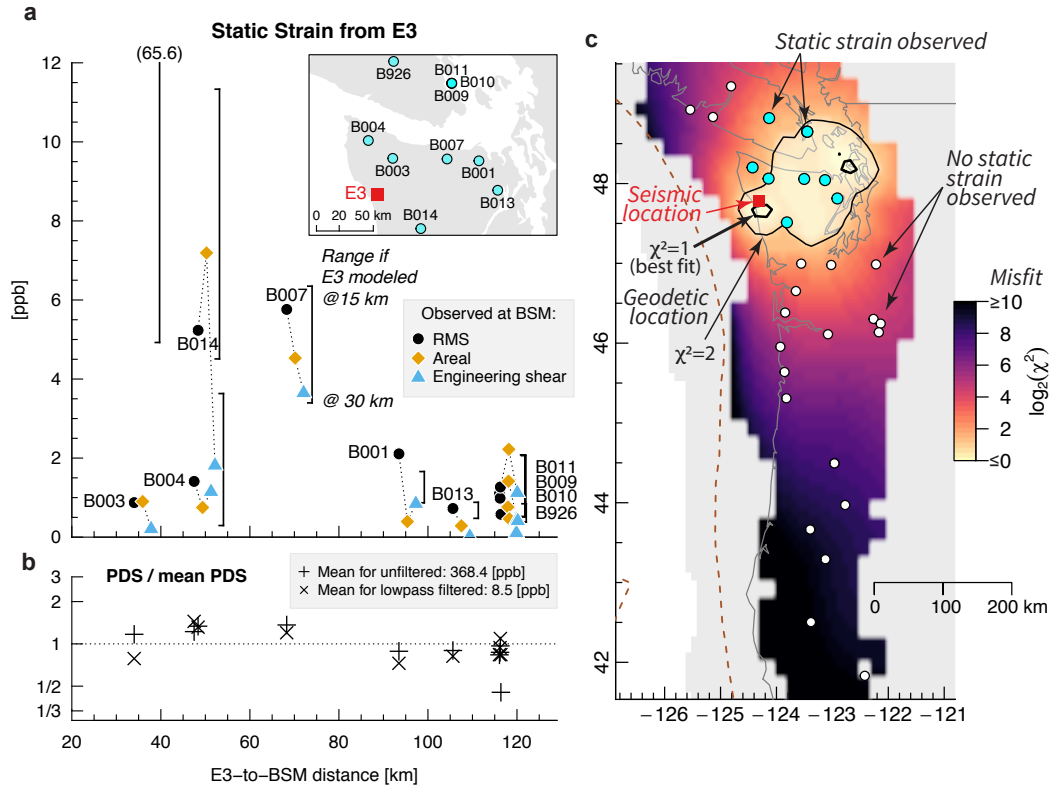
**Figure 4.** Determining the focal mechanism of VLFE E3. The VLFE focal mechanisms are shown in the top panel with their event index listed at their upper left corner [Ide, 2016]. The optimal focal mechanism is denoted as the red beachball. (a), total cross-correlation coefficients of the 112 candidate focal mechanisms. The total cross-correlation coefficient for a focal mechanism is the sum of the average cross-correlation coefficients of all the analyzed stations. (b), VLFE depth of E3 event showing total cross-correlation coefficients for the optimal focal mechanism at depth from 5 to 50 km (c), VLFE duration of E3 showing total cross-correlation coefficients for the optimal focal mechanism with duration from 0.9 to 257.1 seconds. We assume a Gaussian function shape with the duration as 6 times the standard deviation. (d) scaling factor of the VLFE moment. The testing moment is  $2 \times 10^{18} \text{ N} \cdot \text{m}$ . With the scaling factor, the VLFE moment is  $0.5 \times 10^{18} \text{ N} \cdot \text{m}$ , equivalent to a moment magnitude of 5.7.



857 **Figure 5.** Seismic observations of E3 showing the average cross-correlation coefficients of the synthetic  
 858 and observed waveforms of the VLFE E3. The average cross-correlation coefficient of a station is obtained  
 859 by geometrically averaging coefficients of the three-component records. The legends are similar to those in  
 860 Figure 1. The green star shows the 2009 M6.9 Canal de Ballenas earthquake. The red circle is VLFE E3. The  
 861 beachball focal mechanism shows the preferred solution for VLFE E3. Triangles show seismic stations with  
 862 their colors corresponding to the average cross-correlation coefficients. Waveforms of two annotated stations,  
 863 TA.C15A and BK.HUMO, are shown in the insets. Insets: example three-component waveforms of the main-  
 864 shock and the VLFE, overlain with synthetic waveforms of the VLFE. The two stations are at the eastward and  
 865 the southward directions of the VLFE, respectively. The yellow shaded regions show records amplified by ten  
 866 times.

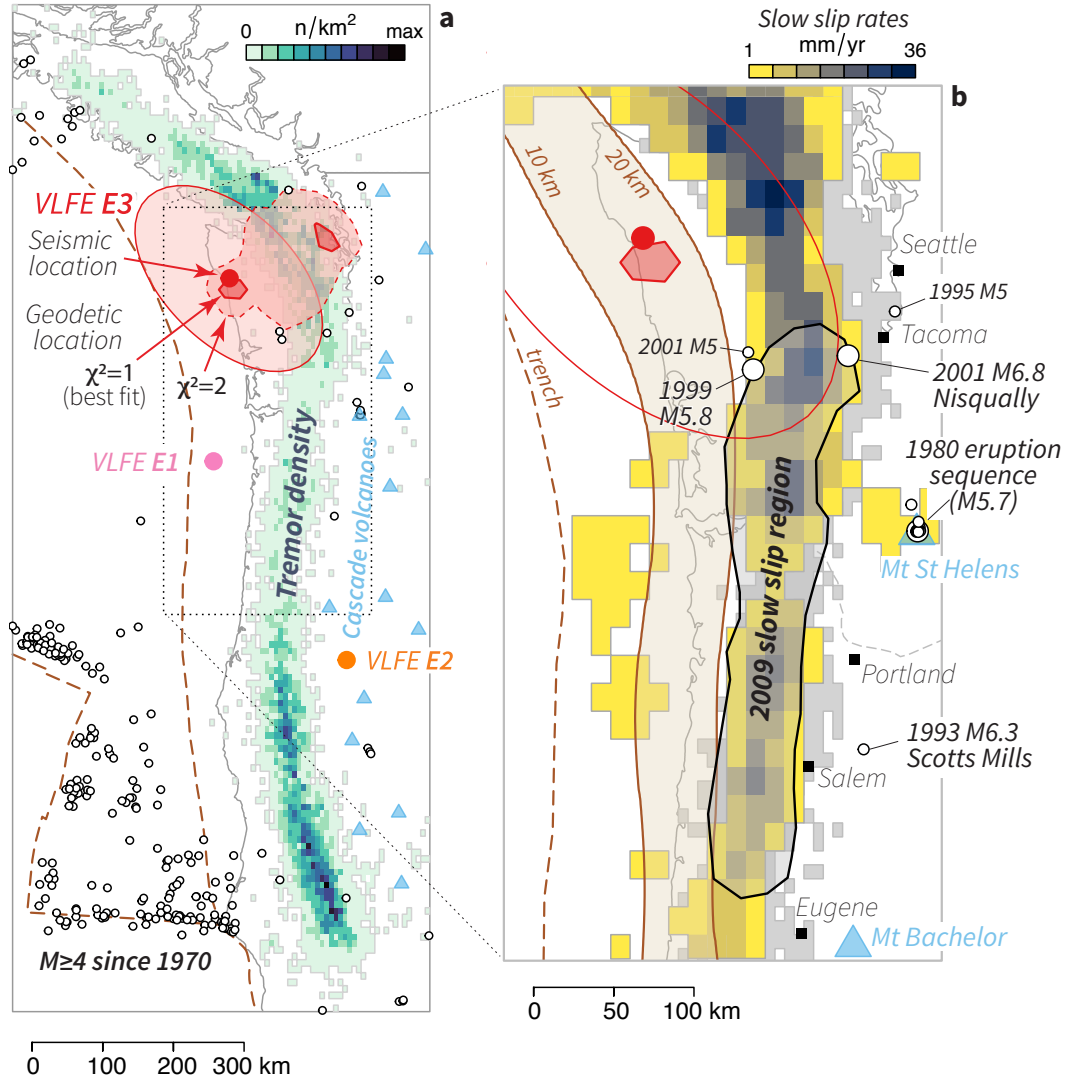


**Figure 6.** Static strains associated with triggered VLFE detection E3. (a) Contours of modeled static deformation from the best fitting focal mechanism for E3 (5), including root-mean-square (RMS) strain, areal strain ( $E_{ee} + E_{nn}$ ), and the two engineering shear strains  $E_{ee} - E_{nn}$  and  $2E_{en}$ . Contours are limited to the theoretical strain detection limit of  $\sim 0.1$  ppb. Stations in (b) are shown as filled circles. (b) Observations of static strains in high-frequency (1 Hz) strain records from B003, B004, B014, B007, B001, and B013. For each strain channel, we show the lowpass filtered record, obtained with a causal filter with a 18 s corner period (56 mHz), overlain on the original record. (Note that the 1 Hz records are not clipped; rather, the vertical scales are set to highlight signals in the low pass filtered record.) Vertical lines show the origin times of the VLFE detections E1–E3: static strains are not apparent until after E3. Self-normalized RMS strain records are shown at the top: E3 occurs around the time of maximum 1 Hz RMS strain (grey), after the peak in low-frequency RMS strain (black).

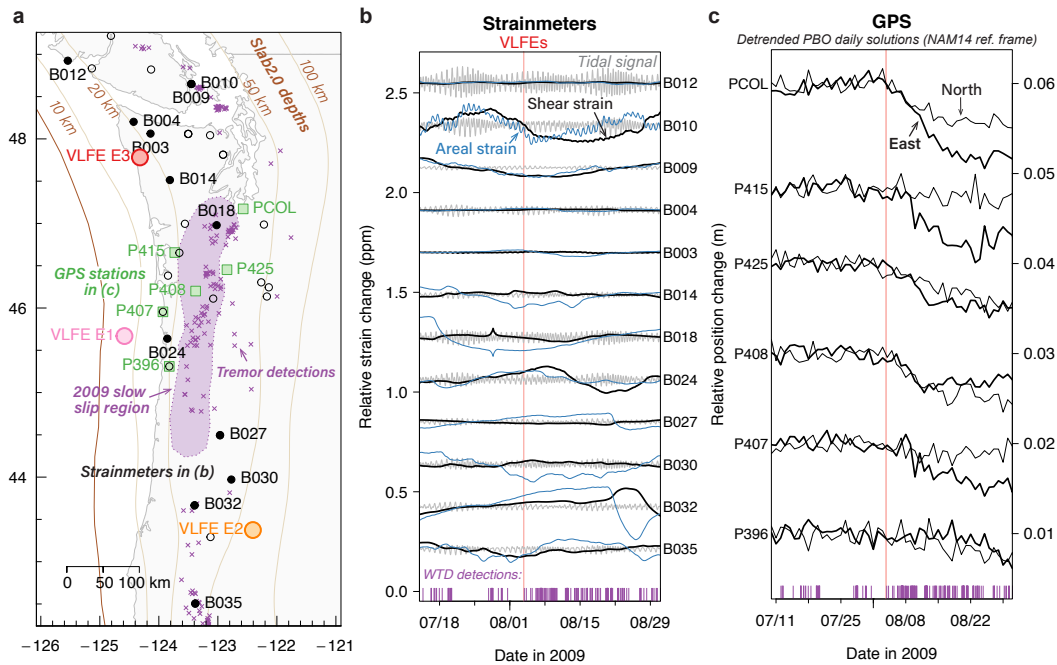


878 **Figure 7.** Observed static strains compared to variability in modeled strain due to depth and location un-  
 879 certainty. (a) Observed tensor strains and the range of model results found by fixing the source epicenter  
 880 but varying the source depth from 15 km to 30 km. (b) Peak dynamic strains (PDS) from the teleseismic waves.  
 881 Points show the ratio of the PDS at each station to the mean value for all stations, for both unfiltered (1 Hz)  
 882 and lowpassed (56 mHz) records; values are shown on a log scale. As expected, the PDS is relatively con-  
 883 sistent across the study region. (c) Map of source likelihood found by moving the E3 source to each point on  
 884 the Slab2.0 depth surface [i.e., *Hayes et al.*, 2018], forward modeling the static strains, and computing the  
 885 misfit. Colors show the base-2 logarithm of the reduced Chi-squared misfit ( $\chi^2$ ); the thick black line shows  
 886 the region where residual variance is equal to observational variance ( $\chi^2 = 1$ ). There are two small regions of  
 887 lowest-misfit; one is near the location of E3 determined independently with seismic data.





**Figure 8.** VLFE E3 in the seismogenic transition zone. (a) Locations of VLFEs E1–E3, seismic and geodetic location uncertainties for E3, tectonic tremor densities [e.g., *Wech and Creager, 2008*], and crustal seismicity with  $M \geq 4$ , since 1970 [e.g., *U.S. Geological Survey, 2017*]. (b) Zoom in map of the location of E3, which both seismic and geodetic data indicate is between the 10 and 20 km depth contours according to the Slab2.0 model [i.e., *Hayes et al., 2018*]. This is outside of the tectonic tremor region [grey area, from (a)], the 2009 slow slip region [*Bartlow et al., 2011*], and the region where average slow slip rates are resolvable at  $> 1$  mm/yr levels [*Bartlow, 2020*].



895 **Figure 9.** Timing of VLFs compared to the 2009 slow slip event. (a) Map of the VLFs, the slow slip  
 896 region [Bartlow *et al.*, 2011] and the primary GPS stations used to constrain the slip patch, tremor detections  
 897 from the World Tremor Database (WTD) [Idehara *et al.*, 2014], and the NOTA BSM network (circles). (b)  
 898 Shear and areal strains from the NOTA strainmeters highlighted in (a), from 2009/7/12 to 2009/9/1. Strains  
 899 have been corrected for atmospheric pressure and tides, detrended, and lowpass filtered with a causal filter  
 900 with a corner period of 2.5 days. Shown below these timeseries is the WTD catalog (see a); the first event  
 901 on 2009/8/3 occurred at 21:24:14 UTC, approximately 10 hours after the VLFs. (c) Detrended timeseries  
 902 of daily GPS positions at the stations shown in (a) during the same period. Noise levels notwithstanding, the  
 903 initiation of the 2009 ETS slow slip event appears to coincide with the VLFs; slip is modeled to begin after  
 904 8/3 and is clearly developed by the 7th [Bartlow *et al.*, 2011].