1 Measuring earthquake source parameters in
2 the Mendocino Triple Junction region using a
3 dense OBS array: implications for fault
4 strength variations

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7 **Abstract**

8 Subduction zones produce earthquakes on a set of faults that op-
9 erate under a wide variety of conditions resulting from considerable
10 variations in depth, temperature, rock type, and fluid pressure. These
11 variations likely lead to variation in the stress levels that drives par-
12 ticular earthquakes and that in turn effects the magnitude of seismic

13 shaking they produce. In the Mendocino Triple Junction (MTJ) re-
14 gion, intraplate faults within the mantle of the subducting plate fail
15 regularly in energetic earthquakes while the adjacent thrust interface
16 of the Cascadia subduction zone remains seismically quiet despite the
17 likelihood that it operates at much lower levels of stress and strength.
18 In 2012, as part of the Cascadia Initiative community experiment, an
19 ocean bottom seismometer (OBS) array was deployed in the MTJ area,
20 providing unusually dense data covering both the inter- and intra-
21 plate earthquakes. Combining these data with onshore networks, we
22 detect and relocate 1137 earthquakes with a three dimensional veloc-
23 ity model. We perform detailed spectral and time domain analysis
24 to study variations in earthquake source properties between the dif-
25 ferent types of faults. We observe a wide variability of stress drops
26 and systematic lateral and depth variations in the earthquake source
27 spectra resulting from the different types of tectonic fault systems in
28 this region: intraplate faults within the subducted oceanic mantle, the
29 Mendocino transform plate boundary fault, and the thrust interface of
30 the Cascadia subduction zone. Some of the depth variability of source
31 spectra can be explained by the expected increase in rupture velocity
32 with depth. However, the overall variation in stress drop estimates is
33 consistent with the highest stress drop earthquakes occurring in the
34 depth range predicted by strength envelopes. Moreover, the earth-
35 quakes in the vicinity of the thrust interface, likely including some
36 within the subducted oceanic crust, show clearly lower stress drops

37 and depletion of high-frequency radiation for their size compared to
38 events in the oceanic mantle and along the transform fault at a given
39 depth. Our results suggest that the variability in earthquake source
40 spectra that can be expected in the Cascadia subduction zone is sys-
41 tematic and that the thrust interface operates at a lower stress level
42 than the core of the subducted plate a few kilometers away.

43 1 Introduction

44 Subduction zones produce the complete spectrum of earthquake rupture dy-
45 namics in terms of the partitioning of energy into seismic radiation for a
46 given amount of fault slip. Along the thrust interface, observations from the
47 shallowest ~ 15 km indicate that seismic moment in ordinary earthquakes is
48 released relatively slowly [Bilek and Lay, 1999] and even more-so in the so
49 called Tsunami earthquakes that slip tens of meters while radiating mini-
50 mal high-frequency energy [Kanamori and Kikuchi, 1993]. Moving deeper,
51 the high-frequency radiation from M 6-7 earthquakes increases steadily with
52 depth. Moreover, recent observations have shown that even within one great
53 thrust earthquake the shallower and deeper portions of the thrust interface
54 produce significantly different seismic radiation in terms of frequency con-
55 tent with high-frequency energy preferentially coming from the base of the
56 seismogenic zone [Lay et al., 2012]. Just below the seismogenic zone, failure
57 occurs as slow slip events with only minimal radiation in the form of tremor

[Peng and Gomberg, 2010]. In contrast, intraslab earthquakes at this depth (25-60 km) that rupture primarily in the mantle of the subducting plate generate strong shaking due to both their short-duration (per moment) ruptures and the relatively low attenuation between their location and the surface [Youngs et al., 1997]. This great variability in the energy balance involved in fault failure must result from spatial variations in both frictional properties as well as the stress and strength levels which in turn likely reflect variations in fault-zone composition, bulk elastic properties, and fluid pressure.

A systematic variation in radiation with depth is expected owing to the increase in seismic wave speeds, and hence rupture velocity V_r , with depth. The increase in V_r leads to ruptures reaching larger areas and producing a greater amount of radiation in a given amount of time. This effect can account for much of the observed scaling of duration in the shallowest 50km [Bilek and Lay, 1999]. Increasing V_r also contributes to shorter rupture durations deeper in slabs but there are indications that either stress drop (proportional to the ratio of slip to fault dimension) must increase or other rupture parameters must change in the 100-600 km depth range, in order to explain the observed rupture duration changes [Vidale and Houston, 1993]. In the extreme, faults in the mantle of the subducting plate at depths of 500-600 km likely melt in at least some large deep earthquakes producing extremely large (100 MPa) stress drops [Kanamori, 1998].

While this overall trend is clear, it is difficult to untangle the relative contributions of increasing rupture velocity and increasing stress drop with

81 depth for most subduction zone earthquakes shallower than ~ 50 km due to
 82 the limited observational constraints available. Most thrust interface rup-
 83 tures occur offshore without seismic or geodetic data immediately above the
 84 slipping region and intraslab ruptures are relatively deep (≥ 25 km), placing
 85 strong limitations on the resolution of seismic and geodetic inversions. Stud-
 86 ies of the frequency dependence of large thrust ruptures often have resolution
 87 only at length scales of 50 km or more [Lay et al., 2012] while studies of M7
 88 intraslab events often cannot distinguish between the two candidate fault
 89 planes or unambiguously determine rupture area [Bustin et al., 2004]. While
 90 stress-levels are expected to increase with depth, it is unclear if this applies
 91 to the entire thrust interface. It is also unclear what differences in stress
 92 drop should be expected between the interplate and intraslab earthquakes.
 93 To predict the ground motions expected from future earthquakes in slabs
 94 with relatively little historical data, like the Cascadia Subduction zone, it is
 95 key to understand how both rupture velocity and stress drop can be expected
 96 to vary spatially within the subduction zone.

97 Several types of observations indicate that there may be a drastic contrast
 98 in strength between the thrust interface and the faults that regularly rup-
 99 ture in the mantle of the subducting plate. The subducting mantle is likely
 100 governed by the mechanical properties of olivine that are well known from
 101 laboratory studies in both the brittle and ductile regimes [Boettcher et al.,
 102 2007]. The intersection of Byerlee’s law for the brittle regime and the ductile
 103 flow laws indicates that the strength of faults in oceanic lithosphere should

104 reach extremely high values. For instance, the mantle of the incoming Gorda
 105 plate in the southern Cascadia subduction zone should be able to support
 106 differential stresses of 400-600 MPa [Wei and McGuire, 2014]. This model
 107 of high stress levels appears to be born out in at least some large earth-
 108 quakes on oceanic faults which show unusually high levels of apparent stress:
 109 $\sigma_a = \mu E_s / M_0$, where μ is the rigidity, E_s is the radiated energy, and M_0 is the
 110 seismic moment [Wyss and Brune, 1968]. A global survey by Choy and Mc-
 111 Garr [2002] pointed out that the majority of earthquakes with high apparent
 112 stresses (≥ 10 MPa) occur as strike-slip events within oceanic lithosphere,
 113 with many in the Gorda plate. These events have extremely impulsive seis-
 114 mic waveforms resulting from the radiation of a large amount of energy in
 115 a very short time compared to a typical continental earthquake. This “high
 116 stress” scenario is consistent with lab results on dry unaltered rocks and at
 117 least some seismological results. If σ_a reflects overall lithospheric strength
 118 [Choy and McGarr, 2002], then the strength of the Gorda Plate mantle must
 119 be quite high (≥ 400 MPa) as it enters the Cascadia subduction zone.

120 In contrast, the thrust interface in Cascadia, located just a few km away
 121 is thought to be the relatively weak. On a global scale, studies of focal mech-
 122 anism orientations have concluded that subduction zone thrust interfaces
 123 follow the “weak fault in a weak crust” model throughout the seismogenic
 124 zone [Hardebeck, 2015]. In Cascadia, the direction of maximum compressive
 125 stress is oriented almost orthogonal to the subduction direction implying
 126 that the thrust interface operates at extremely low stress levels [Schwartz

127 and Hubert, 1997], perhaps as low as 10 MPa [Geist, 1996]. Observations
 128 of the shallowest part of subduction interfaces indicate that both high fluid
 129 pressures [Tobin and Saffer, 2009] and low friction coefficients [Fulton et al.,
 130 2013] are consistent with this model. Similarly, below the seismogenic zone,
 131 the occurrence of episodic tremor and slip (ETS) events requires near litho-
 132 static pore-pressure and hence extremely low effective stress levels, typically
 133 about 1 MPa, to reproduce in numerical models [Liu and Rice, 2005, Segall
 134 and Bradley, 2012]. These low stress levels are corroborated by the ex-
 135 treme sensitivity of ETS to small dynamic stresses (≤ 50 kPa) from passing
 136 seismic waves [Rubinstein et al., 2007]. Moreover, many large subduction
 137 earthquakes are often followed by deep afterslip of roughly equal moment
 138 implying that the downdip thrust interface is operating at shear stress levels
 139 of at most a few MPa [Montési, 2004]. While the stresses in the seismo-
 140 genic zone are likely somewhat higher (10s of MPa), the weak fault model
 141 of Hardebeck [2015] indicates that they fall well short of the values inferred
 142 for the subducting mantle. Thus, in the particular case of the Cascadia sub-
 143 duction zone there is a surprising juxtaposition. The high strength faults of
 144 the subducting mantle are failing regularly in M6-7 earthquakes while just a
 145 few km away, the weak thrust interface, with a strength that is expected to
 146 be at least an order of magnitude smaller, remains completely locked from
 147 geodetic observations and seismically quiet with almost no microseismicity.

148 The goal of this study is to investigate whether these seemingly large
 149 variations in fault strength over just ten km or so are apparent in the variation

150 of earthquake rupture properties. We will use moderate earthquakes recorded
151 by a dense onshore-offshore seismic network. In contrast to studies of the
152 radiation from great earthquakes, our approach allows for very fine spatial
153 resolution albeit at a lower magnitude of slip (M4s and smaller).

154 **2 Dataset and catalog**

155 The Cascadia Initiative is a multiyear community seismic and geodetic ex-
156 periment that included the deployment of a dense OBS array within the
157 Mendocino Triple Junction (MTJ) region during 2012-2013 [Toomey et al.,
158 2014]. Six-months of usable data spanning September 2012 to February
159 2013 were recovered. We build a regional seismic database by combining the
160 OBS data with the Northern California Seismic Network (NCSN) and Plate
161 Boundary Observatory (PBO) borehole network. We do not use OBSs with
162 significant timing problems for analysis, resulting in a combined dataset of
163 30 stations (Figure 1a). We use Antelope software package to detect and
164 locate earthquakes based on STA/LTA ratio and a 1D velocity model that is
165 appropriate for the MTJ area. We manually check and adjust all the phase
166 picks for these events to ensure the reliability of arrivals. Our final catalog
167 includes 1137 earthquakes, with an estimated magnitude of completeness,
168 $M_c = 1.5$: an approximately 10-fold increase over the 134 earthquakes listed
169 in the NCSN catalog for this time period.

170 After building event catalog using Antelope, we band-pass filter the P and

171 S waveforms between 1 to 10 Hz, and computed waveform cross-correlation
172 differential arrival times for all available event pairs using time windows 0.5s
173 before and 2s after the first arrival picks. Utilizing both the absolute and
174 differential arrival times, we relocated 1137 earthquakes using the TomoDD
175 software package [Zhang and Thurber, 2003, 2006] and a 3D P-velocity model
176 obtained via a prior active source seismic experiment by Hole et al. [2000].
177 We use a V_p/V_s ratio of 1.9 within the accretionary wedge, and 1.732 else-
178 where to compute S-wave arrival times. The higher values in the offshore
179 accretionary prism were required to match S-P travel times on the OBS data
180 (Figure S2). The velocity model used for relocation is shown in Figures S3
181 and S4. Although a full tomographic inversion is beyond the scope of this
182 paper, the inclusion of a high V_p/V_s ratio zone was necessary to minimize
183 the arrival time misfit and hence better constrain source depths. The use
184 of the high V_p/V_s zone resulted in hypocenters that were 1-4 km shallower
185 for events beneath the accretionary wedge compared with a uniform V_p/V_s
186 model (Figure S5). The uncertainty in the S-wave velocity model is the pri-
187 mary contribution to the uncertainty in the hypocenter positions (primarily
188 depth) and is about 2-3 km for most events (Figure S5).

189 Based on the relocated catalog and station coverage, we divide the earth-
190 quakes into three areas for focused spectral analysis: area-1 (within the Gorda
191 Plate); area-2 (between Pacific Plate and Gorda Plate); and area-3 (directly
192 beneath Cape Mendocino). Area-1 is dominated by microearthquakes within
193 narrow magnitude range (from M1 to M2.5, see Figure 1c)); while area-2 and

194 area-3 have wider magnitude ranges up to M4.5 (Figure 1d and 1e). Based
 195 on the geometry of the subducted slab and the depth to the plate interface
 196 [McCorry et al., 2012, Hayes et al., 2012], we define three tectonic blocks for
 197 focused interpretation (see Figure 1): the T-Fault (area-2, the major plate
 198 boundary transform fault); the M-Fault (deeper portions of area-2 and area-
 199 3, representing faults within the oceanic mantle); and the ST-Zone (shal-
 200 lower portions of area-2 and area-3, representing events within proximity of
 201 the thrust interface and the overriding wedge). For spectral deconvolution
 202 and EGF (empirical Green’s function) analysis, we process each area (area-1,
 203 area-2 and area-3) separately, as these events are located within the middle
 204 of the network, and share similar source-receiver paths. For discussion and
 205 interpretation, we focus on the relative variations among the three tectonic
 206 blocks (T-Fault, M-Fault, and ST-Zone), as they are separated based on the
 207 tectonic regimes.

208 **3 Spectral processing**

209 The 30 stations have variable sampling rates: 14 LDEO (Lamont-Doherty
 210 Earth Observatory) stations are sampled at 125 Hz; 5 WHOI(Woods Hole
 211 Oceanographic Institution) and SIO (Scripps Institution of Oceanography)
 212 stations are sampled at 50 Hz; 11 NCSN, BK (Berkeley network) and PBO
 213 stations are sampled at 100 Hz. For easy data archival, we resample all
 214 waveforms to a uniform 100 Hz sampling rate, and assemble them into event-

215 based-binary files. We then apply an automatic clip-detector to remove any
 216 clipped waveforms which are common for $M \geq 3.0$ earthquakes in the OBS and
 217 NCSN data. The displacement spectra are computed based on a 1.28s time
 218 window starting from P- and S-wave picks for each event at each station and
 219 noise spectra using a 1.28s time window before the P and S-wave picks. The
 220 spectral calculation is based on a multi-taper algorithm developed in Prieto
 221 et al. [2009]. We compute the signal-to-noise ratio (STN) at each frequency
 222 by taking the difference between displacement spectra and noise spectra.
 223 Only spectra that have an averaged $STN \geq 3$ at three different frequency
 224 bands: 1-3 Hz, 5-10 Hz, and 10-15 Hz are retained. Although the 5 stations
 225 with low-sampling rate may limit the upper bound of frequency limit to 25
 226 Hz, the majority of the usable spectra are recorded on the 25 stations with
 227 Nyquist frequencies ≥ 50 Hz. Because of the poorly determined orientation
 228 for most of the shallow water OBS stations, we do not attempt to rotate the
 229 horizontal channels to obtain transverse and radial components. Instead,
 230 we average the two horizontal channels to obtain averaged S-wave spectra at
 231 each station. As a result, we only use stations where both horizontal channels
 232 meet the STN requirement. In general, the S-wave STN is much better than
 233 P-wave for this dataset, and we only use S-waves for the spectral analysis.

234 The iterative stacking approach in Shearer et al. [2006] is applied to iso-
 235 late the event term for all events in the entire study region, as well as for the
 236 three areas separately (outlined in Figure 1). Only event terms averaged from
 237 at least five stations are used for spectral analysis. Based on the linear rela-

238 tionship between averaged low frequency spectral amplitude (A_s) (between 1
 239 and 3 Hz), and local magnitude (M_l) estimated using the Antelope software
 240 package, the relative log amplitude is calibrated to seismic moment, with
 241 constraints from NCSN moment magnitude estimates for a few larger events.
 242 The spectral-derived moment may be different from NCSN solutions, how-
 243 ever, we only use the spectral moment estimate for all events for consistency
 244 in the spectral analysis.

245 For each event, we find the nearest 50 events, stack event spectra in each
 246 0.2 magnitude bin, then solve for an empirical correction spectrum (ECS)
 247 by averaging the difference between stacked spectra and computed spectra
 248 based on Brune-source model for all magnitude bins. The choice of 50 events
 249 may appear fewer than previous studies (e.g., [Allmann and Shearer, 2007,
 250 Shearer et al., 2006]), however, the lower number is necessary due to the
 251 relative sparsity of the events. The consistency with time domain analysis
 252 in terms of the expected corner frequency ratio suggests that the choice is
 253 appropriate for this region (see text in the time domain section). With the
 254 corrected event term, we grid-search the best-fitting corner frequency using
 255 a hybrid frequency band: for events with $M \leq 2.7$, we search from 1 to 40
 256 Hz; for events with $M \geq 2.7$, we search from 1 to 20 Hz. Justifications of
 257 fitting choices are discussed in Appendix C in detail. We calculate the stress
 258 drop by assuming a circular rupture [Eshelby, 1957]:

$$\Delta\sigma = \frac{7M_0}{16r^3} = \frac{7M_0}{16(V_r T)^3} = \frac{7M_0 f_c^3}{16(k\beta)^3}, \text{ with } r = \frac{k\beta}{f_c} \quad (1)$$

259 where r is source radius, V_r is rupture velocity, T is source duration, β
 260 is the shear wave velocity and k is a factor derived from source modeling
 261 that depends on wave type, and the assumed model such as: Brune [1970]
 262 or Boatwright [1980]. We use the k value from Kaneko and Shearer [2014]
 263 ($k = 0.26$), as it is more closely related to the input stress level. The apparent
 264 stress is computed by extrapolating spectrum to 0.001 and 2000 Hz based on
 265 an ω^2 model:

$$u(f) = \frac{\Omega_0}{(1 + (f/f_c)^{\gamma n})^{1/\gamma}} \quad (2)$$

266 where Ω_0 is the low-frequency amplitude, proportional to seismic moment
 267 M_0 , and f_c is the corner frequency. And the apparent stress is:

$$\sigma_a = \frac{\mu}{M_0} \frac{1}{4\pi^2 \rho \beta^5} \int_0^\infty |\omega \dot{M}(\omega)|^2 d\omega \quad (3)$$

268 Because the stress drop is a model-dependent parameter, we also directly
 269 compare the averaged spectral shape in order to minimize the influence of
 270 model dependence. We present results from the Brune source model with (n
 271 $= 2$, $\gamma=1$), although the averaged spectral variation patterns do not depend
 272 on the model choice. For more details, refer to Appendix C and Shearer
 273 et al. [2006]. Both stress drop and apparent stress are corrected for the
 274 3D variations in S-wave velocity using the same model as for the tomoDD
 275 locations. In our final result, for events within the three areas, we use the
 276 separately solved event terms; for events outside the three areas, we use the

277 event terms solved from the entire region. Our data analysis is focused on
 278 the lateral and depth variations of stress drop and spectra within the three
 279 different tectonic blocks.

280 4 Spatial variations of stress drop

281 We obtain stress drop estimates for 292 earthquakes in the entire study re-
 282 gion, although we only focus on the 237 earthquakes that are within the
 283 three areas of focus, because they are located within the network coverage
 284 (see Figure 1). Events outside regions of focus have limited azimuthal cov-
 285 erage, and are outside the velocity model coverage. In Figure 2, we compare
 286 the stress drop ($\Delta\sigma$) and apparent stress (σ_a) distributions for the three tec-
 287 tonic blocks. Each region exhibits strong variability, with $\Delta\sigma$ and σ_a ranging
 288 over two orders of magnitude. The median and standard deviation for each
 289 are listed in Table 1. The standard deviations of $\log_{10}(\Delta\sigma)$ and $\log_{10}(\sigma_a)$ are
 290 slightly higher than the standard deviation obtained in Baltay et al. [2013],
 291 and comparable to Allmann and Shearer [2009]. For the Mendocino Trans-
 292 form (T-fault), we observe a bimodal distribution of stress drop. Note, in
 293 our calculation, we correct the stress drop and apparent stress estimates with
 294 expected shear-wave velocity by interpolating the 3D velocity model for each
 295 earthquake location. Without such a correction, the standard deviations in
 296 Table 1 increase by 0.08 for the T-Fault, 0.03 for the ST-Zone, and 0.003
 297 for M-Fault. The relatively small improvement for M-Fault is possibly due

298 to the relatively narrow depth range and small lateral velocity perturbation
 299 at this depth (see Figure S10). The overall reduction in standard devia-
 300 tions suggests that our corrections account for a clear physical effect (higher
 301 V_r with depth). Yet, there remains strong systemic variations along depth
 302 within all three blocks, where the deeper events have higher stress drop and
 303 greater variability (Figure 2). The apparent stress values follow a similar
 304 pattern due to spectral extrapolation, which accounts for 40% to over 90%
 305 of radiated energy depending on the corner frequency (equation S4). As a
 306 result, the apparent stress is highly correlated with stress drop. They result
 307 in a typical Savage-Wood efficiency ($R_{SW} = \frac{\sigma_a}{\Delta\sigma}$) of about 0.1 [Beeler et al.,
 308 2003] - consistent with the prediction from equation 1 and 3 (theoretical
 309 $R_{SW} = \frac{16k^3\pi^2}{35} = 0.08$, if $k = 0.26$), and show no clear magnitude depen-
 310 dence. Because of the high correlation, the following analysis will be focused
 311 on stress drop.

312 In Figure 3A, the map view of stress drop shows that most of the low-
 313 est stress drop events are within ST-Zone, consistent with the histogram in
 314 Figure 2. The individual depth cross-sections show that for ST-Zone and
 315 T-Fault, most of lower stress drop events are located in the vicinity of the
 316 slab interface. For the M-Fault in area-3, a group of high stress drop events
 317 are located at about 25 km depth, east of the center of the cross-section.
 318 This cluster is an aftershock sequence from a Mw 4.4 mainshock, which will
 319 be discussed in details later.

320 We compare the depth distribution of stress drop with predicted rock

321 strength profiles from Wei and McGuire [2014] that were calculated for the
 322 subducting Gorda plate in the vicinity of the M-fault. Figure 3B shows
 323 strength envelopes for both dry and hydrostatic profiles that incorporate
 324 both brittle friction and a viscous flow law for the oceanic mantle. Choy and
 325 McGarr [2002] compiled global patterns of radiated energy and apparent
 326 stress, and found that the depth distribution of apparent stress is consis-
 327 tent with a proposed envelope defined by taking 6 per cent of the predicted
 328 strength profile. Since we are using stress drop instead of apparent stress, we
 329 compare the depth profile with 60 per cent of the predicted strength profile
 330 (accounting for a Savage-Wood efficiency of approximately 0.1). For each
 331 3.5 km depth bin with at least 5 events, we plot the values of $\Delta\sigma$ that ex-
 332 ceed 25%, 50%, 75% and 90% of the earthquakes. For all three blocks, there
 333 are general trends of increase with depth for each percentile profiles. We
 334 also note that the inter-event variability increases at greater depths for the
 335 T-Fault (indicated by the increased separation between the 25% and 90%
 336 profiles). The ST-Zone is consistently lower than the T-Fault through the
 337 entire depth range between 17 and 25 km where they overlap.

338 Since our results have been corrected for potential rupture velocity vari-
 339 ations due to increased S-wave velocity at depth, the overall trend of along-
 340 depth stress drop variations suggests that earthquake source parameters re-
 341 flect the underlying fault strength changes, at least in this environment. Ow-
 342 ing to the bandwidth limitations, it is not clear if our results match the 6%
 343 ratio between strength and apparent stress from Choy and McGarr [2002]. If

344 we had used a Savage and Wood efficiency of 0.2, as found for a M6 event in
 345 the subducted Gorda plate [Wei and McGuire, 2014], the strength envelopes
 346 in Figure 3b would have provided a much closer match to the 90% stress-
 347 drop envelopes in terms of peak values. It should be noted that the stress
 348 drop measurements are derived based on equation 1, which is based on a
 349 simple circular crack source model, and a rupture velocity equal to 90% of
 350 S-wave velocity. The $k = 0.26$ is calculated from dynamic simulation with
 351 cohesive zones, which provides stress drop measurements most close to the
 352 ground-truth values [Kaneko and Shearer, 2014], yet, different “k” value will
 353 change the absolute magnitude of the stress drop estimates. Therefore, we
 354 consider the inter-event variability to be more robust.

355 **5 Systematic variation of source spectra**

356 In order to better understand the variations of source spectra in a way that is
 357 relatively model independent, we directly compare averaged spectra following
 358 Chen and Shearer [2013]: by assuming self-similarity, the corner frequency
 359 changes as a function of $M_0^{-1/3}$. Therefore, for each event, we shift the spec-
 360 trum following $f_c \propto M_0^{-1/3}$ to an hypothetical Mw 3.5 earthquake. After
 361 correcting all individual spectra to the Mw 3.5 earthquake, we compare the
 362 averaged behaviors by stacking spectra according to different grouping op-
 363 tions.

364 We first compare averaged spectra for the three tectonic blocks using all

365 available events within a given block respectively. In Figure 4B, the ST-Zone
 366 has lower high-frequency energy compared with the T-Fault and M-Fault,
 367 consistent with the lower median stress drop from Figure 3B. To estimate
 368 the uncertainty of the averaged spectra, we use a bootstrap resampling test,
 369 and compute 100 mean spectra for each resampled dataset. The vertical
 370 lines in Figure 4B represent the 25% and 75% of the mean spectra at each
 371 frequency.

372 We then examine the averaged along-depth spectra variation. For each
 373 block, we compute averaged spectra for each 3.5 km depth bin, and calcu-
 374 late an averaged corner frequency from the stacked spectra by a grid search
 375 between 1.5 and 10 Hz. In Figure 4A, these averaged corner frequencies gen-
 376 erally increase with depth, consistent with the expected trend of increasing
 377 high frequency energy at deeper depth (Figure S11). The averaged corner
 378 frequency variations combine the effects of rupture velocity and stress drop
 379 variations: following equation 1, f_c scales with $\Delta\sigma\beta^3$. The M-fault and T-
 380 fault corner frequencies are roughly continuous with depth and their averaged
 381 spectra nearly overlap with similar shape and fall-off rate. On the other hand,
 382 the ST zone shows a significantly lower corner frequency than the T-fault in
 383 the depth range where they overlap, and has a much lower spectral ampli-
 384 tudes in the 1-10 Hz range for the same moment compared with the T-fault
 385 and M-fault. All three blocks show a clear increase in corner frequency and
 386 high-frequency energy content with increasing source depth Figure 4C, D,
 387 and E.

388 Combining the averaged spectral variation and the spatial distribution
 389 of stress drops, it is clear that events which occurred at the subducted slab
 390 interface and in its vicinity (i.e., directly above the interface within the over-
 391 riding crust, and directly below the interface within the subducted oceanic
 392 crust), have lower stress drops and are depleted in high frequency energy com-
 393 pared with events a few km away in the oceanic mantle. These systematic
 394 variations suggest that the thrust interface is maintained at low stress level,
 395 consistent with the proposed high pore-pressure and low frictional strength
 396 at plate interface (e.g., [Liu and Rice, 2005] [Segall and Bradley, 2012]).

397 **6 The two M4 earthquakes in the high stress** 398 **drop cluster**

399 To verify the inference of fast rupture velocities for the deeper earthquakes in
 400 our dataset, we focus on a cluster of high stress drop events in the ST-Zone,
 401 located at ~ 25 km depth in the mantle of the subducted Gorda plate (Figure
 402 3E). Most of these events belong to the aftershock sequence of a Mw 4.4
 403 earthquake that occurred on 9/14/2012. We analyze two $M \geq 4$ earthquakes
 404 (the Mw 4.4 mainshock at 11:53 and a Mw 4.0 aftershock at 18:19) that were
 405 located only about 3 km apart. Each had a strike-slip focal mechanism albeit
 406 with slightly different orientation. We use the method developed in McGuire
 407 [2004] to estimate the rupture area and rupture velocity. This method uses an
 408 Empirical Green's Function (EGF) to isolate the station by station variation

in the apparent duration of the earthquake. The two EGF events for these two earthquakes are: M 1.6 earthquake at 09/14/2012,12:01 (for Mw4.4) and M 1.4 at 09/14/2012,19:16 (for Mw4.0), respectively.

6.1 Time-domain Second Moment analysis

To enforce positivity on the derived moment-rate functions in the EGF analysis, we use the Projected Landweber Deconvolution method [Bertero et al., 1997, Lanza et al., 1999]. An example of this approach is described in Supplementary Figure S10. The result of the EGF deconvolution, termed the Apparent Source Time Function (ASTF), for each station and phase (e.g. P or S) is a distorted view of the true moment-rate function. The extent of the distortion depends on whether the rupture propagated towards or away from that particular station as well as the takeoff angle and speed of that particular seismic phase. We quantify the variations in duration of the set of ASTFs for a given earthquake by their misfit (second temporal moment) which, following McGuire [2004], is termed $\mu^{(0,2)}(\underline{\mathbf{s}})$ where $\underline{\mathbf{s}}$ is the slowness vector of the particular phase. The azimuthal variations in the ASTFs are usually discussed as variations in the characteristic duration $\tau_c(\underline{\mathbf{s}}) = 2\sqrt{\mu^{(0,2)}(\underline{\mathbf{s}})}$ which is a weighted average measure of the duration of the earthquake as viewed at a particular station and phase. We determine rupture area, directivity, and velocity using the second moments of the slip distributions as detailed in Appendix C and McGuire [2004].

Figures 5 and 6 show the ASTFs resulting from the deconvolution of the

EGF seismograms from those of the mainshocks. The station locations in Figures 5 and 6 are color coded by the observed values of $\tau_c(\underline{s})$. The first observation from the EGF deconvolutions is that neither event is well described by a simple STF such as those from the Madariaga model. Both show considerable structure with multiple subevents, at least for some azimuths, and one pulse late in the rupture that dominates the moment-release.

For the M4.4 earthquake we measured apparent durations from 5 P-waves and 8 S-waves at 12 unique stations. The longest apparent durations were at stations directly above the earthquake while the shortest were observed at more distant stations with horizontal or downward take-off angles (Figure 5). The inversion produced a lower misfit for the fault plane striking EW compared to the NS auxiliary plane and the second moment estimates correspond to a characteristic rupture length L_c of 550m and a strongly unilateral rupture oriented downdip (Table 2). The rupture velocity constraints are $V_R \geq 4.4$ km/s and $V_R \geq 2.3$ km/s.

For the M4.0 earthquake we measured apparent durations from 4 P-waves and 9 S-waves at 10 unique stations. The longest apparent durations were at stations to the southeast and shortest at stations to the northwest (Figure 6). The inversion produced a lower misfit for the fault plane striking to the NW than the NE and the second moment estimates correspond to a characteristic rupture length, L_c , of 730m and a mostly unilateral rupture to the NW (Table 2). The rupture velocity constraints are $V_R \geq 2.5$ km/s and $V_R \geq 1.8$ km/s.

454 One striking result of the EGF deconvolutions is that the smaller, Mw 4.0
 455 earthquake has a significantly longer duration (about 0.2 s vs 0.12 s) than
 456 the Mw 4.4 earthquake. The Mw 4.0 earthquake has an averaged ASTF
 457 (see Figure 7) that is strongly asymmetric, while the averaged ASTF for the
 458 Mw 4.4 earthquake is nearly symmetric. The Mw 4.0 earthquake clearly has
 459 at least two subevents and while the largest of these is fairly abrupt (~ 0.1 s,
 460 Figure 5), the total duration of the earthquake is about 2-3 times longer. This
 461 variability indicates how a simple circular model is insufficient for describing
 462 the complexity of moderate earthquakes.

463 This source complexity introduces considerable uncertainty into estimates
 464 of stress-drop that are based on mapping observed spectral shapes to geo-
 465 metrically simple dynamic source models. In Table 2 we give the estimates
 466 of rupture area and stress drop derived for both M4 mainshocks from three
 467 different approaches. For the second moment derived estimates of rupture
 468 dimension, we only use L_c because W_c is very poorly constrained owing to
 469 its relatively minimal effect on $\tau_c(\underline{s})$. There is no unique mapping between
 470 L_c and stress drop because a variety of slip vs distance profiles can result
 471 in the same L_c . However, for the circular ruptures in the Madariaga/Brune
 472 model as calculated in Kaneko and Shearer [2014], the value of L_c is nearly
 473 identical to the radius r in equation 1. Therefore, we present Brune models
 474 with 10 MPa and 50 MPa stress drops for the moments of both earthquakes
 475 and compare their radius values to our L_c estimates. For the Mw 4.0 event,
 476 the L_c value is a factor of 2 larger than the radius of a Brune model with

477 a 10 MPa stress drop. While for the Mw 4.4 earthquake the L_c estimate is
 478 similar to a Brune model with a 10 MPa stress drop. Given the complexity
 479 of the moment-rate functions for both the Mw 4.0 and 4.4 earthquakes, their
 480 corner frequency may not have a direct correspondence to the total rupture
 481 area as it does in an idealized circular crack model.

482 **6.2 Comparison with spectral analysis**

483 From the source dimension derived from second moment analysis, these two
 484 earthquakes have very different source parameters. The Mw 4.4 earthquake
 485 has a shorter duration and smaller dimension compared with the Mw 4.0
 486 earthquake, and the stress drop is much higher than the Mw 4.0 earthquake.
 487 From equation 1, the corner frequency $f_c = k\beta/r$. Therefore, given the
 488 parameters in Table 2, we should expect corner frequency ratio of about 1.3
 489 (inverse of source dimension) or 1.6 (inverse of duration) between the M_w 4.4
 490 and the M_w 4.0 earthquakes. We list the source parameter estimated from
 491 spectral analysis in Table 3. We have compared results from the Shearer
 492 et al. [2006] stacking method - SH06 (i.e., using nearest 50 events), and an
 493 individual spectral ratio method using the EGF from the second moment
 494 analysis. Parameters in Table 3 are consistent with values plotted in Figure
 495 3, where we interpolated the 3D velocity model at each earthquake depth, and
 496 computed spectral moment for each event. Note that the rupture velocity
 497 used here is consistent with the lower bound of rupture velocity for M_w
 498 4.4. In Figure 7, we compared the shape of the spectra estimated from

SH06 method, the individual pair method with window length of 1.28s, and
 an averaged ASTF derived from stacking the aligned ASTFs at individual
 stations.

Comparing Tables 2 and 3, the second moment inversions imply significantly lower stress drops and corner frequencies than result from the spectral measurements. This likely results from the influence of high-energy subevents on the source spectrum at frequencies well above the corner frequency. In Figure 7A, the source spectra estimated from SH06 method for the $M_w 4.4$ event clearly shows a secondary bumps at high frequency. It also shows that due to the strong directivity, the azimuthal coverage is very important in obtaining reliable spectral amplitude at all frequencies: the single station spectral ratios have significant inter-station amplitude variations at all frequencies. In Table 3, we note that the SH06 stacking method and individual pair EGF analysis yield similar corner frequency estimates. The corner frequency ratio of 1.3 and 1.48 from the two methods, are consistent with the expected ratio from Table 2, suggesting that the relative inter-event variability is stable, and resolvable independent of the estimation methods. However, the source parameters derived from Brune source model is strongly model dependent. For this reason, as discussed in detail by Silver [1983], the second moments are likely a better measure of source size than estimates derived from the high-frequency asymptote of the source spectrum.

520 7 Discussion

521 Interpreting spatial variations of earthquake source parameters (mostly stress
522 drop and apparent stress) in the context of lithosphere strength variations is
523 an important issue for understanding the key factors that control earthquake
524 rupture mechanics and energy radiation. Along depth variations of stress
525 drop have been observed in several different tectonic regions. In some cases,
526 the increasing trend with depth can be corrected via a 1D velocity profile
527 (e.g., in the Parkfield region [Allmann and Shearer, 2007]); in other cases,
528 velocity-corrected stress drop still exhibit strong depth dependence, for ex-
529 ample, stress drop changes by a order of magnitude at shallow depth versus
530 deeper depth in the Hayward fault region [Hardebeck and Aron, 2009]. The
531 depth variations are generally consistent with increased overburden stress
532 with depth, while lateral variations are often linked with patches with dif-
533 ferent slip behaviors, e.g., creeping section versus locked patches, with the
534 former typically have lower stress drops [Hardebeck and Aron, 2009, Allmann
535 and Shearer, 2007].

536 For the ST-Zone, as most of our earthquakes are small (≤ 4), and lack
537 focal mechanism solutions, we do not distinguish between intraplate and
538 interplate earthquakes, and refer to them as near-interface earthquakes. As in
539 Uchide et al. [2014], we do not find significant differences between interplate
540 and nearby events. The averaged stress drop of 5 MPa in the 17 to 25 km
541 depth range is consistent with the averaged stress drop within a similar depth

542 range in the Japan Trench (2011 Tohoku earthquake region) [Uchide et al.,
543 2014]. Seismological, geological and borehole breakouts suggest that the
544 Tohoku earthquake had very low coseismic friction level, a nearly “complete”
545 stress drop [Lin et al., 2013, Fulton et al., 2013, Ujiie et al., 2013], and that
546 the averaged stress drop approximately represents the expected fault strength
547 at the plate interface.

548 For the T-Fault and M-Fault, the depth profiles we observe are consis-
549 tent with the overall shape of the strength profile predicted from laboratory
550 results on olivine: the stress drop generally increases with depth and the
551 seismicity cuts off in the depth range where viscous deformation mechanisms
552 are expected to rapidly reduce the strength of the plate. The 90% stress drop
553 curve is about a factor of 5 below the theoretical frictional strength with a
554 friction coefficient of 0.6 and hydrostatic pore pressure (Figure 3). The “in-
555 complete” stress drop is consistent with other studies (e.g., [Hardebeck and
556 Aron, 2009]), that find stress drop is only a small fraction of the background
557 stress.

558 The Mendocino triple junction involves an oceanic transform fault, a
559 subduction slab interface, and intraplate faulting in both the overriding plate
560 and the subducting mantle. To elucidate the variations between the different
561 regions, Figure 8 shows the stress drop estimates from Figure 3 but with
562 a smoothing function applied (each event shows the median value of its 10
563 nearest neighbors, the choice of median over mean value is to minimize the
564 effect of outliers). The M and T faults show average stress drop values that

are a factor of 10-20 higher than those for the ST-zone. The two cross sections show variations along the M-fault and along the 20km depth contour of the subducting interface. The events located near the thrust interface show a factor of 10-20 lower stress drops than those immediately below them in the subducted mantle. Overall, in our dataset, the clearest spatial variations in stress drop are between the events within ± 5 km of the thrust interface and those located immediately beneath them in the subducting mantle.

Since the OBS data only provides six-months of record, it is important to understand if the spatial pattern could be representative of the long-term behavior. We check the apparent stress for the three largest earthquakes in the 1992 sequence: the two M6 aftershocks have apparent stress of 4.0 MPa and 16.4 MPa, respectively [Choy and McGarr, 2002]; and the M7 mainshock has $E_s = 8.3 \times 10^{14}$ J, and $M_0 = 2.6 \times 10^{19}$ Nm (from USGS PDE catalog), thus the apparent stress $\sigma_a = \mu E_s / M_0 = 0.95$ MPa with $\mu = 30$ GPa. Assuming the Savage-Wood efficiency of 0.1, these apparent stresses correspond to stress drop values of 40 and 164 MPa for the two aftershocks, respectively - consistent with the surrounding small earthquakes in the oceanic mantle (Figure 8); and 9.5 MPa for the M7 mainshock - also consistent with the overall low stress drop of small events within proximity (Figure 8). While we do not have direct comparison along the transform fault, the difference between M-Fault and ST-Zone is consistent with the difference between the two aftershocks and the mainshock.

Equation 1 shows that there is trade-off between assumed rupture ve-

588 locity and the stress drop. Although we corrected the stress drop values
 589 using a simple 1D velocity model, assuming that rupture velocity scales with
 590 shear-wave velocity, it is important to validate the spectral analysis with a
 591 finite source study. Focusing on two largest earthquakes within a high stress
 592 drop cluster at 26 km depth of the ST-Zone (Figure 3), we confirm the high
 593 rupture velocity for the Mw4.4 earthquake. The ratio of corner frequencies
 594 for the two M4 events derived from spectral analysis, is consistent with the
 595 inverse of the ratio of source dimension of the two events, suggesting that
 596 the relative inter-event variability is resolvable and independent of analysis
 597 methods. However, the source dimension estimated from spectral method is
 598 much smaller compared with second moments analysis, suggesting that the
 599 model independent second moments analysis is a better approach to estimate
 600 rupture area.

601 8 Conclusions

602 Using a regional dense OBS array we detected and relocated 1137 earth-
 603 quakes in the Mendocino triple junction area. We performed detailed analysis
 604 of earthquake source parameters via spectral analysis for the entire catalog
 605 and finite source inversions for two larger events via time domain EGF de-
 606 convolutions. Our primary conclusions are:

- 607 • The strongest variations in the source spectra in our dataset are as
 608 function of epicentral depth and tectonic regime (i.e., ST-Zone versus

609 T-Fault and M-Fault).

- 610 • Correcting for known variations of shear wave velocity accounts for
611 some of the variation in our stress drop estimates. The high rupture
612 velocities at depths ≥ 25 km were verified with a finite source inversion
613 for the largest events in our dataset.
- 614 • The corrected stress drop estimates do exhibit both lateral and along
615 depth variations that are consistent with expected fault strength vari-
616 ations. In particular, the stress drop estimates peak in the 20-26 km
617 depth range as would be expected from strength envelopes. Moreover
618 the stress drops are highest on the intraplate faults within the mantle
619 of the subducted plate and lowest in the ST-zone which likely includes
620 earthquakes on the thrust interface and within the over-riding plate.
- 621 • The lower median stress drop estimates in the ST-zone are consistent
622 with models of subduction zones with elevated fluid pressures in the
623 subducted oceanic crust and along the thrust interface.
- 624 • These variations between fault types are verified by a direct comparison
625 of averaged spectral shape which does not depend on a particular source
626 model.
- 627 • The spectral analysis is convenient in analyzing large dataset system-
628 atically, but the time domain analysis is better in estimating detailed
629 source dimension due to the deviations of M4 earthquakes from simple

630 circular crack source models. Both methods can estimate inter-event
631 variability consistently.

632 After these results have been verified with larger and longer datasets, the
633 basic patterns of increasing V_r and stress drop with depth should be incor-
634 porated into seismic hazard models for Cascadia. The 2014 U.S.G.S Na-
635 tional Seismic Hazard maps used different ground motion prediction equa-
636 tions (GMPEs) for earthquakes occurring on the thrust interface than for in-
637 traslab events [Atkinson and Macias, 2009, Abrahamson et al., 2015]. While
638 the GMPEs are purely empirical and reflect both source and path effects, it
639 is likely that these differences reflect first order differences in rupture dynam-
640 ics between events on the low-stress thrust interface versus the higher stress
641 core of the slab that we observe. Elucidating these differences and their fine
642 scale spatial variations should help improve hazard estimates in subduction
643 zones.

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Table 1: Median and standard deviation of $\Delta\sigma$ and σ_a for the three regions.

	median $\Delta\sigma$ (MPa)	std ($\log_{10}\Delta\sigma$)	median σ_a (MPa)	std ($\log_{10}\sigma_a$)
M-Fault	16.15	0.58	1.38	0.57
T-Fault	18.11	0.52	1.62	0.51
ST-Zone	6.52	0.38	0.61	0.37

Table 2: Estimated source parameters L_c , τ_c , f_c from the second moments compared with Brune models. All radius and stress drop values assume $k = 0.26$ and $\beta = 4.480$ km/s as in equation 1. The Brune models use the seismic moment estimates from the Berkeley catalog (M_w 4.0 and M_w 4.4).

M_w	M_0 (Nm)	$L_c(m)$	$\tau_c(s)$	Brune 10 MPa r (m)	Brune 10 MPa f_c (Hz)	Brune 50 MPa r (m)	Brune 50 MPa f_c (Hz)
4.0	1.078e+15	730	0.2	361	3.2	211	5.5
4.4	5.438e+15	550	0.12	621	1.9	363	3.2

Table 3: Estimated source parameters f_c (Hz), $\Delta\sigma$ (MPa) (and σ_a), and r (m) from the source spectrum analysis comparing Shearer et al. [2006] method (SH06) and spectral ratio analysis (SR) with different time windows. All radius and stress drop values assume $k = 0.26$ as in equation 1. The seismic moments estimates are derived from the low-frequency spectral amplitude calibration. The S-wave velocity for Mw4.4 is 4132.6 m/s, and for Mw4.0 is 4210 m/s.

M_w	M_0 (Nm)	SH06				SR, tw=1.28s		
		f_c	r	$\Delta\sigma$	σ_a	f_c	r	$\Delta\sigma$
4.0	4.12e+14	6.8	160.97	42.8	3.69	5.99	182.7	29.2
4.4	2.24e+15	8.9	120.7	550	48.19	8.88	121.0	546.3

Figure 1: Overview of earthquakes in the study area. (A) Map view of detected seismicity. Circle sizes and colors correspond to earthquake magnitude (see the colorbar). The three outlines mark the three areas selected for focused analysis: ST-Zone (green), M-Fault (blue), T-Fault (red). The lines and stars with corresponding colors indicate the cross sections in C, D, and E and their midpoints respectively. Black triangles represent stations used in the analysis, grey triangles represent stations in the network that were not used. (B) Larger map of regional tectonics around the study area. The earthquakes and stations are the same as in (A). The background is global bathymetry model from Becker et al. [2009]. (C), (D) and (E), depth profiles for different regions, the solid lines represent the top of the subducting slab from the slab model of McCrory et al. [2012]. The cross-section in (C), (D) and (E) are produced with GMT "project" command for both earthquakes and slab geometry points in the same polygon.

Figure 2: Comparison of the three regions. Color scheme is the same as Figure 3: ST-Zone (green), M-Fault (blue), T-Fault (red). (A) Histograms of the stress drop estimates within each region. (B) Histograms of the apparent stress estimates within each region. (C) Stress drop versus depth for each region. (D) Apparent stress versus depth for each region.

Figure 3: Overview of earthquake stress drop distributions. (A) Map view of individual earthquake stress drop measurements (circles) with the colors corresponding to stress drop. The three outlines mark the three areas subset for spectral deconvolution: Area-1, Area-2, Area-3. The lines with corresponding colors indicate the depth profiles in panels C, D, and E, and the black stars mark the center of the profiles. (B) Depth distribution of stress drop for different areas, and comparison with $0.6 \times$ predicted strength envelopes (the solid and dashed black lines denote the modelled strength envelopes for dry rock and hydrostatic pore-pressure profiles from Wei and McGuire [2014]). The solid lines denote the median stress drop for each 3.5 km depth bin. The dashed lines denote the value of the 25th, 75th, and 90th percentiles of the stress drop distribution within each depth bin. The stress drop is plotted on a log scale. (C), (D) and (E), depth profiles for the 3 areas, the solid lines represent the top of the subducting slab from the slab model. The depth ranges for the M-Fault and the ST-Zone within Area-1 and Area-3 are outlined by the rectangular boxes. Note that the color scale in (A), (C), (D) and (E) is from $10^{-0.25}$ MPa to $10^{2.25}$ MPa, events with $\Delta\sigma \leq 10^{-0.25}$ are fixed at $10^{-0.25}$, events with $\Delta\sigma \geq 10^{2.25}$ are fixed at $10^{2.25}$. This is purely for visualization purpose due to the wide range of stress drop, and does not affect the variations discussed in the main text.

Figure 4: (A) Corner frequency from averaged-shifted spectra versus depth (see text for details). (B) Averaged-shifted spectra for all earthquakes within each region (M-Fault, T-Fault, ST-Zone). (C), (D) and (E) Averaged magnitude corrected spectra for each depth bin for the three regions. Colors correspond to depth, at each 3.5 km depth interval, spectra from all events within each depth bin are shifted to M_w 3.5 based on $f_c \propto M_0^{-1/3}$. In (B), (C), (D) and (E), the vertical bars indicate the uncertainty of the mean spectra estimation based on 100 bootstrap resampling for each appropriate group.

Figure 5: Apparent Source Time Function measurements and 2^{nd} moment inversion results for the Mw 4.4 earthquake at 11:53:16 on 09/14/2012 in the mantle of the downgoing Gorda plate. The focal mechanism in (B) is from Northern California Earthquake Center. (A) Map of the earthquake (gray circle) and station locations (triangles) used in the second moments inversion. Each triangle is color-coded based on the $\tau_c(\vec{s})$ value of its ASTF. The color scale is this apparent duration in seconds. The ASTFs are shown immediately adjacent to their station location and similarly color-coded based on $\tau_c(\vec{s})$. A few stations have two ASTFs corresponding to both a P and S-wave measurement. (B) The apparent durations predicted by the optimal values of the second moments (table 2) for the nodal plane with strike 281 and dip 85. (C) shows two example P-wave ASTFs for stations KCT and B933. The B933 ASTF is elongated and corresponds to a $\tau_c(\vec{s})$ of 0.2 s while the KCT ASTF corresponds to 0.14s. (D) The amplitude spectra for both ASTFs are shown along with a Brune spectrum with a f_c of 5 Hz.

Figure 6: Apparent Source Time Function measurements and 2nd moment inversion results for the Mw 4.0 earthquake at 18:19:50 on 09/14/2012 in the mantle of the downgoing gorda plate. The focal mechanism in (B) is from Northern California Earthquake Center. (A) Map of the earthquake (gray circle) and station locations (triangles) used in the second moments inversion. Each triangle is color-coded based on the $\tau_c(\vec{s})$ value of its ASTF. The color scale is this apparent duration in seconds. The ASTFs are shown immediately adjacent to their station location and similarly color coded based on $\tau_c(\vec{s})$. A few stations have two ASTFs corresponding to both a P and S-wave measurement. Short durations are observed to the NW while longer durations are observed to the SE. (B) The apparent durations predicted by the optimal values of the second moments (table 2) for the nodal plane with strike 325 and dip 84. (C) shows two example P-wave ASTFs for stations KCT and B933. The B933 ASTF is elongated and corresponds to a $\tau_c(\vec{s})$ of 0.31 s while the KCT ASTF corresponds to 0.15s. (D) The amplitude spectra for both ASTFs are shown along with a Brune spectrum with a f_c of 5 Hz.

Figure 7: Comparison between different methods for the Mw 4.4 and Mw 4.0 events in Figures 5 and 6. (A) and (B) are spectrum for the two events. Black solid lines denote event source spectra estimated using the empirical correction function from the nearest 50 events. Grey solid lines denote the averaged spectral ratio between each event and its EGF from Figures 5 and 6. Dashed grey lines denote Brune's source model fitted to the spectrum with corner frequencies noted in each figure. Colored solid lines are spectral ratio at each individual station, colored by station azimuth. (C) and (D) are apparent source time function at each individual stations and averaged STF for each event. Colors correspond to station azimuth.

Figure 8: Map and cross-section views of stress drop estimates smoothed by closest 10 events in 3D coordinates. (A) Map view of smoothed stress drop (color scale is in MPa). The four beachballs are four $M \geq 6.5$ earthquakes before 2013. The two cross-section profiles are centered at the three largest earthquakes in 1992, respectively. (B) Cross-section view of profile X-X', centered at the averaged location of the two largest aftershocks of the 1992 Mendocino sequence (marked by orange star), and the mainshock marked by blue star. (C) Cross-section view of profile Y-Y', centered at the 1992 mainshock (blue star), and the averaged location of the two largest aftershocks (orange star).

859 **A Introduction of supplemental materials**

860 The supplemental materials include four sections: velocity model used in
861 relocation; detailed processing procedure; along depth variation of theoretical
862 source spectra; and constant Q models for each of the three regions. The
863 earthquake catalog is attached as a separate txt file.

864 **B The velocity model used in the relocations**

865 We use the 3D P-wave tomography model from Hole et al. [2000]. We initially
866 tried a constant V_p/V_s ratio of 1.732 throughout the study region, however,
867 even a simple plot of the median value of the ratio between S and P travel
868 times (Figure S2) demonstrates that there is substantial spatial variations of
869 the average V_p/V_s ratio. Based on the coherent spatial pattern in Figure S2,
870 we set the V_p/V_s ratio to 1.9 in a 40 km by 40 km box in the accretionary
871 wedge (see Figure S3), and use V_p/V_s ratio of 1.732 elsewhere (see Figure
872 S4). The high V_p/V_s ratio extends to 9 km depth. Accounting for the high
873 V_p/V_s ratios in the accretionary wedge has a direct effect on the epicentral
874 depth estimates. A full tomographic S-wave inversion will be necessary to
875 further increase the accuracy of the depth estimates. There is not yet a
876 full-scale tomography image of both P and S wave velocity structures in
877 this region that we can validate the V_p/V_s ratio variations, however, the
878 high V_p/V_s ratio in the accretionary wedge is consistent with observations
879 of reduced S-wave velocity elsewhere (e.g., [Reyners et al., 2006, Eberhart-

Phillips et al., 2005]) and our own unpublished studies of seafloor compliance in the Cascadia forearc.

C Methods

C.1 Spectral analysis

Following Shearer et al. [2006], the displacement ($\mathbf{D}$) recorded at each station for each earthquake can be represented by convolution of seismic source term $\mathbf{E}$, station response $\mathbf{S}$ (includes instrument response and site amplification), travel path term between source and station $\mathbf{T}$:

$$\mathbf{D}_{ij}(\mathbf{t}) = \mathbf{E}_i(\mathbf{t}) * \mathbf{S}_j(\mathbf{t}) * \mathbf{T}_{ij}(\mathbf{t}) \quad (\text{S1})$$

where $*$ indicates convolution and $\mathbf{D}_{ij}$, $\mathbf{E}_i$, $\mathbf{S}_j$, and $\mathbf{T}_{ij}$ are time series. In the frequency domain, this equation becomes a multiplication of the Fourier-transform of these terms, which then simplifies in the log-domain to the summation:

$$d_{ij} = e_i + s_j + t_{ij} + r_{ij} \quad (\text{S2})$$

where r_{ij} is the residual [Shearer et al., 2006]. If two events occurred at the same location (i) and are recorded by the same station (j), they share the same (t_{ij}) and (s_j), which can be canceled by spectral ratio, forming the basis for the individual-pair EGF method. Following Shearer et al. [2006],

we can sequentially solve for path term (t_{ij}), station term (s_j) and event term (e_i) using many event and station pairs. The event source spectrum is typically represented by Brune type models (equation 2). And the spectral ratio is:

$$\frac{u_1(f)}{u_2(f)} = \frac{(1 + (f/f_{c2})^{\gamma n})^{1/\gamma}}{(1 + (f/f_{c1})^{\gamma n})^{1/\gamma}} \quad (\text{S3})$$

where f_{c1} and f_{c2} are corner frequencies for two events, respectively. With ($n = 2, \gamma=1$), this is the model proposed by Brune [1970], while with ($n=2, \gamma=2$), this is the model proposed by Boatwright [1980]. Assuming a circular source, the source dimension, r , can be calculated from the corner frequency, and then the stress drop can be inferred using equation 1.

The radiated energy can be estimated from the velocity spectrum squared:

$$E_s = \frac{1}{4\pi^2\rho\beta^5} \int_0^\infty |\omega \dot{M}(\omega)|^2 d\omega \quad (\text{S4})$$

and the apparent stress is:

$$\sigma_a = \mu E_s / M_0 \quad (\text{S5})$$

We extrapolate the source spectra to 0.001 Hz and 2000 Hz based on the Brune model to estimate the radiated energy and apparent stress. For our events, the fraction of energy extrapolation ranges from over 90% for smaller events to about 40% for larger events. The apparent stress is also corrected for the 3D variations in S-wave velocity.

912 In the spectral fitting process, we resample the spectra at even log-
 913 frequency intervals, in order to downweight the high-frequency amplitude
 914 relative to the low-frequency points. We solve for the correction spectrum
 915 (ECS) using magnitude bins from 1.3 and 2.5, and fit between 1 and 40 Hz.
 916 After correction of ECS, we search for best-fitting corner frequency using the
 917 Brune's source model. For earthquakes with $m \leq 2.7$ on the spectral magni-
 918 tude scale, we search for f_c between 1 and 40 Hz, while for larger events, we
 919 search for f_c between 1 and 20 Hz. The adoption of $M=2.7$ is based on an ob-
 920 served corner frequency scaling shift at this magnitude range, and is purely
 921 empirical. However, fitting all events between 1-20 Hz finds much greater
 922 scatter for small events, potentially due to insufficient bandwidth; fitting all
 923 events between 1-40 Hz can not resolve the differences between the two M4
 924 events, potentially due to existence of high-frequency secondary events (see
 925 discussion in the maintext) and causes changes in corner-frequency scaling
 926 at about M2.7. For area-1, we use total of 521 spectra for 128 events, 44
 927 events are resolved from 5 or more stations, with magnitudes ranging from
 928 1.3 to 2.5; for area-3, we use 1067 spectra for 250 events, 91 events are re-
 929 solved from 5 or more stations, with magnitudes ranging from 1 to 4.5; and
 930 for area-2, we use 1056 spectra for 224 events, 102 events are resolved from
 931 5 or more stations, with magnitudes ranging from 1 to 4. For all regions,
 932 there are 3518 spectra for 848 events, which includes some events outside the
 933 regions of focus, 292 events are resolved from 5 or more stations.

934 Due to the small number of events (≤ 50) in area-1, part of the M-

935 Fault, the difference between solving all events within the entire region versus
936 solving events within the focused subset area (i.e., the three regions) has
937 strongest influence on the results within this area. This also affects the
938 influence of inter-region variability. Therefore, without pre-subsetting study
939 regions, it is inevitable to include events outside this area in the spectral
940 analysis process. Thus, we think that results discussed in the main text
941 represent the real variability among the three sub-regions compared to results
942 obtained without pre-selecting the study areas.

943 In Figure S6, we compare the results (raw results before applying the
944 rupture velocity correction) from different fitting choices. In Figure S6, the
945 inter-region variability is consistent with all three different choices of fitting
946 frequency bands: 40 Hz, 20 Hz, and a hybrid (20 Hz for $M \geq 2.7$ and 40 Hz
947 for $M \leq 2.7$). For 20Hz fitting, the scatter for small events is greater, and
948 the fraction of energy extrapolation is also greater.

949 For small events, their corner frequencies can be above the Nyquist fre-
950 quency limit of this network. To check whether our main results are biased by
951 the frequency band, we replicated our comparison with the subset of events
952 with $M \geq 2$. While this dataset is much smaller, all of the corner frequencies
953 are below 50 Hz (see Figure S7). Figure S8 shows that with only $M \geq 2$
954 events, the difference among the three regions is still very clear. Moreover it
955 is consistent with our results from the larger event set shown in Figure S6
956 and Figure 2. For a discussion of the resolution of the smallest event with
957 corner frequencies larger than the Nyquist, see [Allmann and Shearer, 2007].

958 They demonstrate that these events can be resolved albeit with greater un-
959 certainties than the larger events. We have chosen to present the results
960 from the larger subset of events (e.g. all of those whose spectral fits pass
961 our criteria) in the main text because the larger dataset allows a finer spatial
962 regionalization for the propagation corrections. Compared to the resolution
963 limit for corner frequencies, the average-stacked spectra from different regions
964 are relatively model independent, and do not depend on the frequency limit
965 as corner frequencies. In Figure S9, we also compare the averaged spectra
966 difference for only $M \geq 2$ events, which shows similar differences among the
967 three regions. Therefore, our main fault mechanics conclusions would not be
968 affected by the frequency limit in this study.

969 C.2 Time-domain second moment analysis

970 Our approach for determining rupture area, directivity, and velocity involves
971 estimating the higher-order moments of the slip distribution as detailed in
972 McGuire [2004]. Unlike finite fault inversions with dozens to hundreds of pa-
973 rameters and assumptions, this approach estimates integral quantities that
974 are well constrained (only 5-6 independent parameters) and are free of the
975 typical assumptions in finite fault inversions (such as a uniform rupture veloc-
976 ity and damping parameters). Moreover, following Silver [1983] and McGuire
977 [2004], $\mu^{(0,2)}(\underline{s})$ is linearly related to the true second moments of the source
978 by:

$$\mu^{(0,2)}(\underline{\mathbf{s}}) = \hat{\mu}^{(0,2)} - 2\underline{\mathbf{s}} \cdot \underline{\hat{\mu}}^{(1,1)} + \underline{\mathbf{s}} \cdot \underline{\underline{\hat{\mu}}}^{(2,0)} \cdot \underline{\mathbf{s}} \quad (\text{S6})$$

979 where $\hat{\mu}^{(0,2)}$, $\underline{\hat{\mu}}^{(1,1)}$, and $\underline{\underline{\hat{\mu}}}^{(2,0)}$ are the second moments (defined below) that
 980 quantify the duration, directivity, and rupture area of the source respectively.
 981 Thus, by measuring $\mu^{(0,2)}(\underline{\mathbf{s}})$ for P and S waves at a variety of takeoff angles
 982 and azimuths (e.g. stations), we have the data necessary for a well-posed
 983 inverse problem to constrain the rupture velocity of small earthquakes. The
 984 inverse problem for equation S6 is solved using a convex optimization tech-
 985 nique so that the physical constraint that the source region have non-negative
 986 volume can be enforced [McGuire et al., 2001]. This constraint reduces the
 987 non-uniqueness in the inversion. Additionally, we assume the rupture was
 988 confined to a 2-dimensional fault-plane reducing the second moments to only
 989 six independent quantities. We perform the inversion for both fault planes
 990 of the earthquakes focal mechanism and choose the one with the lower misfit
 991 as the true plane. The misfit is defined as:

$$misfit = \sum_i (\mu^{(0,2)}(\underline{\mathbf{s}}) - \hat{\mu}^{(0,2)}(\underline{\mathbf{s}}))^2 / \sum_i \mu^{(0,2)}(\underline{\mathbf{s}})^2 \quad (\text{S7})$$

992 where $\hat{\mu}$ denotes the predicted value based on the second moment esti-
 993 mates. The uncertainty in the derived second moment quantities is estimated
 994 via a jackknife calculation where groups of stations of particular azimuth
 995 ranges (typically 45 degrees) are deleted to produce subsets of the dataset
 996 for the jackknife [McGuire et al., 2001].

997 The second moments of the slip-rate distribution for any earthquake de-
 998 scribe the spatial and temporal extent of the rupture as well as its propa-
 999 gation, all of which contribute to the duration observed in any particular
 1000 far-field phase. They are defined as:

$$\begin{aligned}
 \hat{\mu}^{(2,0)} &= \int \int \dot{f}(\mathbf{r}, t)(\mathbf{r} - \mathbf{r}_0)(\mathbf{r} - \mathbf{r}_0)dV dt \\
 \hat{\mu}^{(0,2)} &= \int \int \dot{f}(\mathbf{r}, t)(t - t_0)(t - t_0)dV dt \\
 \hat{\mu}^{(1,1)} &= \int \int \dot{f}(\mathbf{r}, t)(\mathbf{r} - \mathbf{r}_0)(t - t_0)dV dt
 \end{aligned} \tag{S8}$$

1001 where $\dot{f}(\mathbf{r}, t)$ is a scalar function that describes the spatial and temporal
 1002 distribution of moment release along the fault [McGuire et al., 2001], $\mathbf{r}_0$ and
 1003 t_0 denote the centroid location and time (i.e. the first moments) respectively.
 1004 The hat denotes that these are central moments taken about the centroid.
 1005 The integrals are taken over the entire source volume and earthquake du-
 1006 ration [Backus, 1977b,a, McGuire et al., 2001]. When $\dot{f}(\mathbf{r}, t)$ is integrated
 1007 over the volume of the source, it is known as the source time function (STF)
 1008 $\dot{f}(t)$. The second spatial moment $\hat{\mu}^{(2,0)}$, is related to the spatial extent of the
 1009 rupture area, the second temporal moment $\hat{\mu}^{(0,2)}$ is related to the duration
 1010 of rupture, and the mixed moment $\hat{\mu}^{(1,1)}$ is related to rupture propagation.
 1011 In particular, a characteristic rupture duration τ_c , rupture length L_c , and
 1012 average propagation velocity of the instantaneous spatial centroid $\mathbf{v}_0$ are de-
 1013 fined following Backus and Mulcahy [1976], Backus [1977b], Silver [1983] and
 1014 McGuire et al. [2001]:

$$\begin{aligned}
x_c(\hat{\mathbf{n}}) &= 2\sqrt{\hat{\mathbf{n}}^T \hat{\boldsymbol{\mu}}^{(2,0)} \hat{\mathbf{n}}} \\
\tau_c &= 2\sqrt{\hat{\mu}^{(0,2)}} \\
v_c &= \mathbf{L}_c / \tau_c \\
\mathbf{v}_0 &= \hat{\mu}^{(1,1)} / \hat{\mu}^{(0,2)}
\end{aligned} \tag{S9}$$

1015 where L_c is the maximum value of x_c (i.e. corresponding to the largest
1016 eigenvalue of $\hat{\mu}^{(2,0)}$).

1017 Owing to the weighted average nature of the second moments, the charac-
1018 teristic dimensions of a rupture are always smaller than its total dimensions,
1019 but they provide an estimate of the region that contributed significantly to
1020 the moment release. Of particular importance for this study is the differ-
1021 ence between $v_c = L_c / \tau_c$ and the magnitude of $\mathbf{v}_0$. The magnitude of $\hat{\mu}^{(1,1)}$
1022 depends strongly on the style of rupture. For a nearly symmetric bilateral
1023 rupture, the contributions to this integral from positive values of $r - r_0$ will
1024 largely cancel those from negative values resulting in a very small magnitude
1025 of $\hat{\mu}^{(1,1)}$ (and v_0). In contrast, for a unilateral rupture these contributions will
1026 add constructively yielding a large magnitude for $\hat{\mu}^{(1,1)}$ and v_0 . In fact, for a
1027 uniform slip unilateral rupture with instantaneous rise-time, v_0 is equal to its
1028 upper bound V_R , while for a perfectly symmetric bilateral rupture $v_0 = 0$ and
1029 $v_c = 2V_r$. Thus, by estimating the second moments we constrain not only
1030 the source's spatial extent (L_c , W_c) but also place two lower bounds on the
1031 rupture velocity $V_R \geq v_0$, and $V_R \geq v_c/2$. One important factor for evaluat-

1032 ing the second-moment time-domain inversion is to check the goodness-of-fit
1033 to the observed waveform. Below (Figure S10) is an example for the Mw4.0
1034 event recorded at a borehole station, which shows overall good fit to the
1035 observed waveform.

1036 **D Spectral variation along depth**

1037 To illustrate the effect of rupture velocity on the shape of the source spectra,
1038 we generate synthetic Brune-type spectra for an hypothetical Mw 3.5 earth-
1039 quake with stress drop of 3 MPa, located at different depths between 0 and
1040 35 km. The S-wave velocity is drawn from the 1D velocity profile averaged
1041 for the entire study region, and the rupture velocity is 0.9 times the corre-
1042 sponding S-wave velocity. We calculate corner frequency from equation 1,
1043 and compute the spectrum from equation 2. Figure S11 shows that the high
1044 frequency energy gradually increases with depth.

1045 **E Constant Q model for the three regions**

1046 The estimated event terms trade off with the path terms, hence, if there
1047 exists heterogeneous attenuation structure, it will affect the high-frequency
1048 spectral amplitude in the event terms. In the iterative stacking method,
1049 the travel-time path term contains averaged attenuation information of the
1050 study region. We evaluate the performance of the deconvolution by testing

1051 whether we can obtain a reasonable attenuation factor (Q_s) value for each
 1052 region. For each of the three regions of focus, we compute a constant Q
 1053 from the travel-time path terms (Figure S12). We assume the theoretical
 1054 spectrum from attenuation is:

$$A_k = A_0 e^{-\omega t(k)/2Q_s} \quad (\text{S10})$$

1055 where A_k is the amplitude at a given travel-time index $t(k)$, A_0 is the
 1056 amplitude scale, ω is angular frequency, Q_s is the quality factor. In the log
 1057 domain, the spectrum has a linear relationship with frequency. We solved
 1058 for the travel-time terms by discretizing the travel-time into bins of 0.5s (1s
 1059 for M-Fault) interval, and grid search for best-fitting Q_s for the entire region.
 1060 Although these curves are not as smooth as results in Shearer et al. [2006] due
 1061 to the smaller number of events, the variations in the slope of the travel-time
 1062 terms between the three regions is clear. As in Figure S12, the averaged Q_s
 1063 is 610 for T-Fault, 380 for ST-Zone, and 1270 for M-Fault. Although there
 1064 is no published attenuation model to compare to, the low Q for ST-Zone
 1065 is consistent with expected high attenuation in the overriding wedge [Abers
 1066 et al., 2006].

Figure S1: This figure shows the travel-time residual changes as a function of iteration in the tomoDD inversion. The mean and median values of absolute residual at each iteration are shown by red and magenta circles, respectively; and the standard deviation of residual at each iteration is shown by green circles.

Figure S2: This figure shows the median value of the ratio of the S-wave travel time to the P-wave travel time for all first-arrival picks that were used in the tomoDD relocation. The median value for each station is plotted at the location of that station. It includes paths from the entire relocated dataset and is not a tomographic image. It shows that the V_p/V_s ratio in the accretionary wedge is likely substantially higher (≥ 1.9) than in the surrounding crust (~ 1.7 to 1.75).

Figure S3: This shows the S-wave velocity model at 6 km depth below sea level in km/s. The V_p/V_s ratio has been set at 1.9 in a 40 km by 40 km box that includes the OBSs above the accretionary wedge.

Figure S4: This shows the S-wave velocity model at 9 km depth below sea level in km/s. It is simply the V_p model of Hole et al. [2000], divided by a constant V_p/V_s ratio of 1.73. Seismometer locations shown as white triangles.

Figure S5: A histogram of differences in epicenter depths between a tomoDD inversion with a uniform V_p/V_s ratio of 1.7 versus our preferred inversion using the velocity model in Figure S3. While most earthquakes are unaffected, there are a considerable number of events beneath the OBS network that result in epicenters 2-4 km shallower with the modified S-wave velocity model.

Figure S6: Comparison of the inter-region variability by solving all events with pre-subsetting, and fitting with different frequency bands. The stress drop results are without velocity correction. Column (a), fitting with 1 to 20 Hz; column (b), 1-40 Hz; column (c), hybrid fitting with 1-20 Hz for $M \geq 2.7$, and 1-40 Hz for $M \leq 2.7$. In all figures, red: T-Fault; green: ST-Zone; blue: M-Fault. For each column: (1) is the stress drop distribution on log scale for each of the three regions; (2) Stress drop versus depth, cut-off of upper stress drop limit at 50 MPa for plotting purposes.

Figure S7: Comparison of the results with only $M \geq 2$ events (left column) and all events (right column). The stress drop results are without velocity correction. (a) and (b) are corner frequency versus magnitude, the black line shows a constant scaling for $f \propto M_0^{-1/3}$. (c) and (d) are stress drop versus magnitude, the black dashed lines shows the corner frequency limits of 5 and 50 Hz. The colors are the same with all other figures, red: T-Fault; green: ST-Zone; blue: M-Fault.

Figure S8: Comparison of the three regions. Similar to Figure 2, but with only $M \geq 2$ events and without velocity correction. Color scheme is the same as Figure 3: ST-Zone (green), M-Fault (blue), T-Fault (red). (A) Histograms of the stress drop estimates within each region. (B) Histograms of the apparent stress estimates within each region. (C) Stress drop versus depth for each region. (D) Apparent stress versus depth for each region.

Figure S9: Averaged-shifted spectra for all earthquakes within each region (M-Fault, T-Fault, ST-Zone). Similar to Figure 4b, but only with $M \geq 2$ events - for which the corner frequencies are ≤ 50 Hz (see Figure S7).

Figure S10: Example of the EGF deconvolution measurements from the M4.0 mainshock S-wave recorded at station B049. The top left panel shows the velocity seismogram for the M4.0 (blue) and an EGF event (red) with normalized amplitude. Also plotted in green is the tradeoff curve for waveform misfit versus source time function duration with the triangle denoting our pick for the duration of the moment-rate function as seen by this particular S-wave. The upper right panel show the resulting moment-rate function (in units of the EGF events moment). The lower right panel shows the fit to the mainshock seismogram. The lower left panel shows the amplitude spectrum of the ASTF compared to a Brune model with a corner frequency of either 4 or 10 Hz.

Figure S11: (A) Averaged 1-D S-wave velocity profiles for the three regions, and the entire region (with $V_p/V_s = 1.732$). (B) Predicted spectral shape for an M3.5 earthquake with 3MPa stress drop at different depths from 0 to 35 km.

Figure S12: This is fitting of a constant Q to each of the three regions. Black lines are resolved travel-time terms for different travel time bins indicated by the numbers. Red lines are fitting of travel-time terms with a constant Q indicated in the sub-figures.