Deep low-frequency earthquakes in tremor localize to the plate interface in multiple subduction zones

Justin R. Brown, Gregory C. Beroza, Satoshi Ide, Kazuaki Ohta, David R. Shelly, Susan Y. Schwartz, Wolfgang Rabbel, Martin Thorwart, and Honn Kao

Received 10 July 2009; revised 27 August 2009; accepted 3 September 2009; published 14 October 2009.

[1] Deep tremor under Shikoku, Japan, consists primarily, and perhaps entirely, of swarms of low-frequency earthquakes (LFEs) that occur as shear slip on the plate interface. Although tremor is observed at other plate boundaries, the lack of cataloged low-frequency earthquakes has precluded a similar conclusion about tremor in those locales. We use a network autocorrelation approach to detect and locate LFEs within tremor recorded at three subduction zones characterized by different thermal structures and levels of interplate seismicity: southwest Japan, northern Cascadia, and Costa Rica. In each case we find that LFEs are the primary constituent of tremor and that they locate on the deep continuation of the plate boundary. This suggests that tremor in these regions shares a common mechanism and that temperature is not the primary control on such activity. Citation: Brown, J. R., G. C. Beroza, S. Ide, K. Ohta, D. R. Shelly, S. Y. Schwartz, W. Rabbel, M. Thorwart, and H. Kao (2009), Deep low-frequency earthquakes in tremor localize to the plate interface in multiple subduction zones, Geophys. Res. Lett., 36, L19306, doi:10.1029/2009GL040027.

1. Introduction

[2] Deep, non-volcanic tremor was discovered in Japan [Obara, 2002] and found to occur during episodes of slow slip in Cascadia [Rogers and Dragert, 2003] and elsewhere [Schwartz and Rokosky, 2007]. LFEs are small earthquakes (less than magnitude ~2) deficient in amplitude at frequencies above ~10 Hz that occur during episodes of deep tremor in southwest Japan [Katsumata and Kamaya, 2003]. LFEs belong to a newly discovered class of slow earthquakes [Ide et al., 2007b] located above an area of elevated $V_p/V_s$, down-dip from the locked portion of the plate boundary in SW Japan [Shelly et al., 2006]. Tremor generation has been related to fluid release during oceanic slab dehydration [Katsumata and Kamaya, 2003]; however, empirical moment tensor analysis reveals that LFEs are generated by shear slip [Ide et al., 2007a] on the deep extension of the plate boundary [Shelly et al., 2007b]. The spectral characteristics of tremor and LFEs are essentially identical [Shelly et al., 2007a] and comparison of tremor and LFE waveforms indicate that most tremor in southwest Japan is comprised of LFE swarms [Shelly et al., 2007a, 2007b; Brown et al., 2008].

[3] The picture of tremor in other subduction zones is less clear. Studies of non-volcanic tremor in northern Cascadia using a source-scanning algorithm [Kao and Shan, 2004] find that the tremor source has a peak in activity near the plate interface, but that it occurs at depths from about 15–50 km. This suggests the possibility of a different relationship between tremor and slow slip events in Cascadia [Kao et al., 2005]. On the other hand, particle motion analysis of tremor farther south in Cascadia [Wech and Creager, 2007] suggests slip in the direction of relative plate motion, and cross correlation of vertical and horizontal ground motion indicates that tremor originates near the plate boundary [La Rocca et al., 2009]. These discrepancies in the origin of the tremor source need to be resolved if tremor is to be fully understood in other tectonic settings and subduction zones that differ from SW Japan. The signature and frequency characteristics of deep tremor and its association with longer term slow slip events is similar wherever it is well observed, which motivates us to apply techniques we have used to study tremor in Japan more widely.

2. Method

[4] The data consists of three hours of tremor in three different subduction zones (example waveforms in Figure 1). For Japan we use 8 Hi-Net borehole velocity recordings of tremor in western Shikoku from 19:00–22:00 on April 18, 2006, sampled at 100 samples per second. In Cascadia we use tremor data from 00:00–03:00 on September 22, 2005, as recorded at seven 3-component stations on southern Vancouver Island, including both CNSN and POLARIS broadband networks, sampled at 100 samples per second. In Costa Rica we use data from 00:00–03:00 on May 17, 2007, as recorded on 4 broadband STS2 sensors and 4 short period sensors of at least 25 samples per second. Tremor is most clearly recorded in the 1–8 Hz frequency band, which is well within the recording capabilities of all these instruments.

[5] We apply a running, network autocorrelation [Brown et al., 2008] to detect LFEs within tremor for each of these subduction zones. We use this approach to detect waveforms that nearly repeat as observed across a seismic network in each region. Our approach is closely related to that used to associate LFEs with tremor [Shelly et al., 2007a], however in our case, the origin times and locations of potential LFEs are unknown. This means that we have to search for similarity at all possible lags. Consider a network...
of $n$ station components (e.g., 3-component Hi-Net stations, 8 stations corresponds to $n = 24$) on which we record ground motion at windows represented by the vector $u$ at time $t_i$ and $t_j$. The corresponding network correlation coefficient (CC) sum, $a_{ij}$ can be written as:

$$a_{ij} = \sum_n CC_n = u(t_i) \cdot u(t_j)$$

i.e., the sum of the normalized CC across the network. Summing across the network allows us to search for times when the entire network exhibits waveform similarity, and greatly enhances the ability to distinguish signal from noise. We detect on the statistics of $a_{ij}$ relative to that of all other lags and use the median absolute deviation (MAD) to set a detection threshold [Shelly et al., 2007a; Brown et al., 2008]. The MAD ensures that the detection statistics are not adversely affected by windows with high values (positive detections). Time lags exhibiting very strong similarity represent either repeats, or near-repeats, of LFEs within the tremor.

We save all window pairs that exceed our detection threshold of 6 times MAD, and verify them as LFEs based on the comparison of stacks of aligned events by re-applying the autocorrelation with a detection threshold of 9 times MAD for Japan and Cascadia, and 10 times MAD for Costa Rica data. The detection criterion for Costa Rica is set higher due the lower sampling rate of the data compared to SW Japan and Cascadia. We are able to apply sample precision waveform cross-correlation to measure relative arrival times for all windows that pass the event detection statistic. Because we have both horizontal and vertical components of ground motion for the tremor, we search for S-waves and P-waves on these components, respectively. We search for S-waves via cross-correlation, flag the arrival, and search behind the S arrival for a P-wave on the vertical component. Figures 2a–2c and 2d–2f show alignments on the S- and P-waves, respectively. Both P and S wave pick errors are between 0.1 and 0.5 seconds in all three locations. Starting locations for the LFE hypocenters are estimated using the S-P times for all events.

We use a combination of tomoDD [Zhang and Thurber, 2003] and the summed network correlation coefficients [Ohta and Ide, 2008] to estimate accurate event locations assuming a fixed, 3-D velocity model from tomographic studies in each locale [Shelly et al., 2006;...]

Figure 1. Non-volcanic tremor recordings at three subduction zones. Tremor in all three subduction zones is characterized as a long-duration low-amplitude signal resembling volcanic tremor in some respects.

Figure 2. LFE detections within tremor from three subduction zones on horizontal components. Shown are stacks and alignments of similar LFEs recorded at a common station. Events are aligned on the (a–c) S-wave arrival after cross-correlation and (d–f) P-wave arrivals. (top) Individual events in blue and the normalized stack in red. (bottom) A grayscale plot of aligned seismograms. Positive values are white, and negative values are black.
Figure 3. LFE locations in SW Japan, Cascadia, and Costa Rica. In SW Japan, the LFEs are located in western Shikoku between 30 and 35 km depth. In northern Cascadia the LFEs are located in southern Vancouver Island between 30 and 40 km depth on the plate interface. Light green hypocenters represent LFEs away from the main cross section. In Costa Rica, the LFEs are beneath the Nicoya Peninsula between 30 and 40 km depth. The locked portions of these subduction zones and moho are inferred from previous tomography studies [Shelly et al., 2006; Audet et al., 2009, DeShon et al., 2006] are shown in amber and blue, respectively. Black dashed lines correspond to the deep extent of the plate interface. Ordinary micro-earthquakes are shown in violet, and these events occur in the locked portion of the subduction zone. LFEs localize to the plate boundary in all three subduction zones.

Ramachandran, 2001; DeShon et al., 2006]. Grid space sizes in both SW Japan and Cascadia were designed at 20 by 20 by 5 km grid spacing and 10 by 10 by 5 grid spacing in Costa Rica. The aforementioned tomographic studies are appropriate to use for location due to the presence of recently modeled plate interfaces. The cross-correlation derived relative arrival time measurements are used as input data for tomoDD. The summed network correlation coefficient (NCC) is used as a basis for weighting event-pair differential times for all P- and S-wave recordings. This allows events with stronger network correlation coefficients, viz. those with stronger waveform similarity across the network, to have greater weight in the solution. We use the summed NCC for data weighting for the tomoDD location routine, and retain all LFE detections in each study region. This reduces the potential negative impact on the estimated locations of the more uncertain measurements from weaker LFEs that have low network correlation coefficients [Ohta and Ide, 2008], and yield LFE locations that should be more precise than Brown et al. [2008].

3. Results

We detect abundant low-frequency earthquake activity during tremor in all three subduction zones comprising of 298 LFEs in SW Japan, 331 LFEs in northern Cascadia, and 232 LFEs in Costa Rica and find the occurrence of LFEs to be almost continuous. Figures 3a–3c show the locations of detected LFEs in the three regions. There are differences in the distribution of LFEs. In SW Japan and Costa Rica, the LFEs are very strongly clustered; whereas, in Cascadia they are more widespread. In each case the LFEs locate down-dip of the locked seismogenic zone of the plate boundary defined by: the down-dip extent of the last great earthquake (SW Japan) [Yoshioka et al., 2008], geodetic observations (Cascadia) [Hyndman and Wang, 1995], and interplate seismicity (Costa Rica) [DeShon et al., 2006]. We also detect six ordinary micro-earthquakes located on the plate interface, up-dip from the LFEs in a region of abundant interplate micro-earthquake activity [DeShon et al., 2006] in Costa Rica, which suggests that the autocorrelation approach is useful for analyzing earthquake swarms as well. Based on the grid spacing and pick uncertainty, location error is no more than ±5 km in the vertical direction. It is worth noting that the plate interface modeled by Ramachandran [2001] is ~5 km deeper than the location of our LFEs. Recent studies of the plate interface in northern Cascadia reveal a slightly shallower plate interface [Audet et al., 2009], consistent with our LFE locations. Our results suggest that the plate interface is shallower and/or broader and more complex than previously envisioned.

The existence of LFEs during tremor in each of these subduction zones, and their locations, both support the hypothesis that non-volcanic tremor in subduction zones is generated by intermittent shear-slip in the vicinity of the plate interface. Our three-hour sample is not long enough to explore many interesting questions, such as possible along-strike variations in tremor activity or other factors that might
control tremor/LFE occurrence. Nonetheless, all of the LFEs we detect locate within 5 km of the plate interface.

[10] Although our results support the hypothesis that deep tremor is generated by shear slip, evidence exists that fluids play an enabling role in tremor as supported by tomographic and seismic reflection studies in both Cascadia [Kao and Shan, 2004; Audet et al., 2009] and Japan [Shelly et al., 2006] indicating high fluid pressure in the vicinity of LFEs. Although less definitive, seismic tomography studies in Costa Rica also suggest elevated \(V_p/V_s\) at the plate interface [DeShon et al., 2006], co-located with the LFEs we detect in this study.

[11] Given network coverage of the same density used for local earthquake monitoring, we can detect tremor and extract the LFEs that comprise it. This capability should help us obtain a deeper understanding of what controls the distribution of tremor. Table 1 shows the variations in incoming plate age, convergence rate, and modeled temperatures on the plate interface in the LFE depth range for the locations we studied. The Cascadia and SW Japan subduction zones are relatively hot with strongly locked shallow seismogenic zones displaying little or no inter-plate seismicity. LFEs in these locations occur between 30–45 km where modeled temperatures in the subducting oceanic crust reach 400–500°C [Yoshioka et al., 2008; Wada et al., 2008]; conditions favorable for dehydration of hydrous minerals in oceanic basalt [Hacker et al., 2003]. LFEs in the relatively cold and seismically active northern Costa Rica subduction zone occur at comparable depth but where temperatures in the down-going oceanic crust do not exceed 250°C [Hacker et al., 2003]. The cooler Costa Rica plate interface precludes the same dehydration reactions that likely occur in SW Japan and Cascadia; however, fluids may be liberated through dehydration of lower temperature hydrous phases [Hacker et al., 2003]. Despite significant variations in both thermal structure and degree of seismic coupling, all three subduction zones record LFE swarms during tremor that localize to the deep extension of the plate interface (between 30 and 45 km depth). Other parameters may also play a role in controlling the distribution of LFEs, such as the thickness of the over-riding plate, pressure, or topographic load due to the overriding plate [Brudzinski and Allen, 2007].

[12] Tremor is widely observed and offers new opportunities to understand fault behavior. Improved locations will help in monitoring possible time-dependent changes in tremor, and may provide important constraints on the role of the deep extension of faults in the earthquake process.

**References**


G. C. Beroza and J. R. Brown, Department of Geophysics, Stanford University, Stanford, CA, 94305-2215, USA. (beroza@stanford.edu; jrbrown5@stanford.edu)
S. Ide and K. Ohta, Department of Earth and Planetary Science, University of Tokyo, Tokyo 113-0033, Japan. (ide@eps.s.u-tokyo.ac.jp; ohta@eps.s.u-tokyo.ac.jp)
H. Kao, Geological Survey of Canada, Pacific Geoscience Centre, 9860 West Saanich Road, Sydney, BC V8L 1B9, Canada. (hkao@nrcan.gc.ca)
W. Rabbel and M. Thorwart, Institute of Geosciences, Christian-Albrechts-Universität zu Kiel, D-24118 Kiel, Germany. (rabbel@geophysik.uni-kiel.de; thorwart@geophysik.uni-kiel.de)
S. Y. Schwartz, Department of Earth and Planetary Sciences, University of California, Santa Cruz, CA 95064, USA. (susan@pmc.ucsc.edu)
D. R. Shelly, U.S. Geological Survey, 345 Middlefield Road, Menlo Park, CA 94025, USA. (dshelly@usgs.gov)