Cascadia tremor polarization evidence for plate interface slip

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[1] New seismic measurements of the repeated phenomenon of Episodic Tremor and Slip in northern Cascadia indicate identical source processes of tremor and slow slip. Predicted polarization directions of upgoing S-waves radiated from shear slip on the plate interface align with the relative motion between the Juan de Fuca and North American plates. Seismic observations from small-aperture array data on the Olympic Peninsula of the Cascadia subduction zone show uncharacteristically stable linear particle motion coincident with the passage of tremor sources beneath the array. The azimuth of this horizontal ground motion matches expected polarizations from slip on the plate interface. This finding suggests that Cascadia tremor is shear slip on the plate interface, implying that, as in Japan, geodetically observed slow slip and seismically observed tremor are manifestations of the same phenomenon. **Citation:** Wech, A. G., and K. C. Creager (2007), Cascadia tremor polarization evidence for plate interface slip, *Geophys. Res. Lett.*, *34*, L22306, doi:10.1029/2007GL031167.

1. Introduction

[2] Non-volcanic tremor or slow slip has been observed at many subduction zones around the world. The best observations are located along the Nankai Trough in Japan and in northern Cascadia. Marked episodic tremor and slip (ETS) activity in these two areas is likely related to a release of fluids at shallow depths from the downgoing plate caused by the relatively warm, young oceanic lithosphere subducting at slow rates [*Obara*, 2002; *Rogers and Dragert*, 2003; *Hacker et al.*, 2003]. Tilt observations in southwest Japan [*Obara et al.*, 2004; *Hirose and Obara*, 2005] and GPS observations in northern Cascadia [*Rogers and Dragert*, 2003; *Miller et al.*, 2002] provide evidence of periodic, retrograde movements from the ambient direction of relative plate motion between the subducting slabs and overriding continental crusts. These deviatoric motions suggest fault slip along the plate interface [*Dragert et al.*, 2001; *Obara et al.*, 2004]. In both cases, tremor is observed to coincide with this slow slip, having epicentral locations occurring where the plate interface ranges in depth from 30–45 km [*Obara*, 2002; *Kao et al.*, 2005; *McCausland et al.*, 2005; *Obara and Hirose*, 2006] (Figure 1).

[3] There is no clear interpretation, however, of the physical mechanism of tremor that accompanies slow slip in northern Cascadia, nor is the connection between ETS in Cascadia and ETS in Japan yet understood. Here we identify tremor during a Cascadia ETS event exhibiting

stable linear particle-motion polarizations that match those expected from slip on the plate interface. This finding suggests that Cascadia tremor is shear slip on the plate interface. Because slip events increase stress on the locked zone with the potential to trigger a megathrust earthquake [*Dragert et al.*, 2001], this new understanding of tremor could enable near real-time increased seismic hazard estimates using tremor locations from seismic data to track this stress loading on the seismogenic zone with high spatial and temporal resolution.

2. Cascadia and Japan ETS

[4] There are several apparent differences between northern Cascadia and Nankai ETS; some are real, and some may be related to data quality and quantity. Northern Cascadia ETS events occur less frequently with greater deformation than in Japan and are clearly observed with GPS [*Dragert et al.*, 2001; *Miller et al.*, 2002], while Japanese events are not [*Obara et al.*, 2004; *Hirose and Obara*, 2005; *Obara and Hirose*, 2006]. The location of tremor relative to the seismogenic zone also differs. The updip edge of Japanese tremor epicenters is very near the down-dip edge of the locked zone of the subducting plate interface [*Obara*, 2002; *Obara and Hirose*, 2006] compared with a nearly 100-km separation between tremors and the locked zone near the coast in Cascadia [*Kao et al.*, 2005; *McCausland et al.*, 2005; *McCaffrey et al.*, 2007] (Figure 1). Japanese tremor appears to be composed of many low-frequency (2–8 Hz) earthquakes (LFE) [*Shelly et al.*, 2007] and coincident with 20s-period, very-low-frequency (VLF) earthquakes [*Ito et al.*, 2007]. These events allow for phase identifications that yield focal mechanisms and accurate hypocenters, both of which imply that Japanese tremor represents slip on the plate interface [*Shelly et al.*, 2006; *Ide et al.*, 2007b; *Shelly et al.*, 2007; *Ito et al.*, 2007]. In Cascadia, however, neither VLF's nor LFE's have been identified. Cascadia tremor has been located over a range of depths from 10 to 60 km [*Kao et al.*, 2005; *McCausland et al.*, 2005], and there are currently no constraints on tremor focal mechanisms.

[5] A question that is central to understanding ETS is whether tremor and slow slip are manifestations of the same or different processes. Evidence from Japan ETS events indicates that tremor and slow slip are two observations of the same process [*Shelly et al.*, 2006; *Ide et al.*, 2007b; *Shelly et al.*, 2007]. Their relationship in Cascadia, however, is unsettled. It is widely agreed that slow slip in northern Cascadia represents slip in the direction of relative plate motion on, or near, the plate interface [*Dragert et al.*, 2001], but there is no consensus on how tremor relates to slip. Tremor is characterized by a lack of high frequency content relative to normal earthquakes [*Obara*, 2002], which suggests that it results from a slow, low stress-drop process, probably associated with high fluid pore pressure [*Kao et*

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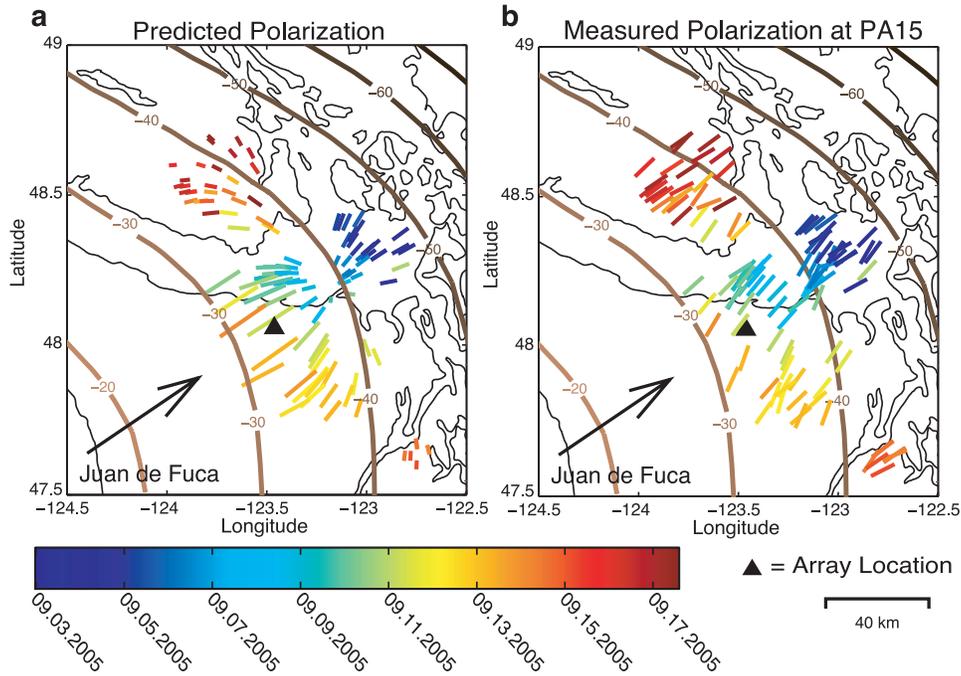
Figure 1. Tremor locations and comparison between (a) predicted and (b) observed polarizations. The locations of the lines show tremor locations, color-coded with time, with the line orientations depicting horizontal polarization directions. Relative predicted amplitudes are represented by the length of lines in Figure 1a. Agreement between predicted (Figure 1a) and observed (Figure 1b) is best where rays are steep and tremor is within 40 km of array. Contours indicate the depth of the plate interface [*McCrory et al.*, 2004].

al., 2005; *Ide et al.*, 2007a]. The apparent wide range of tremor depths in Cascadia and their similarity to volcanic tremor [*Julian*, 1994] favors an interpretation that slow slip induces fluid movement, which is responsible for the seismic tremor signal [*McCausland et al.*, 2005]. Alternatively, this depth distribution has also been attributed to tremor being caused by a reaction to changes in stress induced by the transient slip on the plate interface [*Kao et al.*, 2005]. Both interpretations imply that ETS in Cascadia is fundamentally different from ETS in Japan.

3. Polarization Analysis Method

[6] In this paper we develop a new method using tremor particle motion observations to test whether tremor sources are consistent with slow slip focal mechanisms (Figure 2). In anticipation of the September 2005 ETS event, we deployed an EarthScope temporary array of 10 three-component seismometers with 1-km average spacings on the northern side of the Olympic Peninsula. Using data from this small-aperture array, we estimate the direction of particle motion polarizations during times of active tremor with the power to repeat observations from the same location. The array captured the entire ETS event with a proximity to tremor activity that created a unique opportunity to robustly observe particle motions associated with seismic tremor signals.

[7] To predict polarizations, we need tremor locations. We employ a method, similar to that developed by *Obara* [2002], for locating tremor sources using all available seismic data, including data from Pacific Northwest Seismic Network (PNSN) and EarthScope Transportable Array (TA)

as well as our own array. While tremor can be seen in nearly every hour from September 3 to September 17, we analyze one 15- to 30-minute time window every three to four hours to produce a total of 111 locations. We determine centroid locations of these tremor bursts by cross-correlating pairs of smoothed envelope functions (Figure S1 of the auxiliary material)[1] and performing a 3-D grid search over potential source locations that provide S-wave lag times that optimize the cross correlations (Figure S2). When applied to earthquakes this method produces errors of $1-3$ km in epicenter and $4-17$ km in depth (Table S1).

[8] Tremor began on September 3, 60 km northeast of our array (Figure 1). During the next ten days, tremor bursts migrated to the southwest, stalling beneath our array before bifurcating and heading southeast (ending on September 15) and northwest (ending about September 30). For this paper we located tremor sources through September 17, after which their increasing distance from our array and consequent shallower incident angles increased the noise in the observed polarization directions.

[9] We determine the average particle motion polarization for a given time window by band-pass filtering between 1 and 6 Hz and computing a covariance matrix of 9 dot products from the 3 components of motion [*Jurkevics*, 1988]. Each matrix describes the second moment of the spatial distribution of particle motions. We look for time windows in which one of the eigenvalues dominates over the other two, indicating that the polarization is nearly linear. The eigenvector corresponding to this largest eigen-
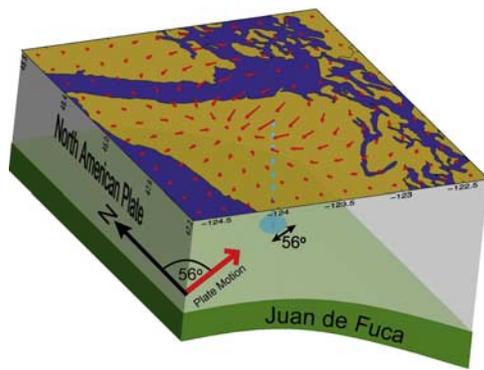
Figure 2. Schematic figure of polarization radiation pattern. The arrows indicate the projected horizontal polarization directions predicted at the arrow base locations—all resulting from a single patch slip on the plate interface.

value is the direction of this nearly linear polarization [*Jurkevics*, 1988]. The degree of linearity is measured by the parameter $L = 1 - (\lambda_2 + \lambda_3)/2\lambda_1$ which varies from 0 (not linear) to 1 (perfectly linear), where λ_i are the eigenvalues ordered from largest to smallest [*Jurkevics*, 1988].

4. Polarization Results

[10] This procedure is applied to each station and every 10-minute window from August 15 through October 4, 2005. To exclude noise from a nearby logging campaign, we plot windows only from 5PM to 5AM local time each day (Figure 3). These data show an increase in linearity correlating with stable polarization azimuths from September 5 to 14. A two-dimensional histogram of number of time windows with linearities > 0.7 for each day shows an increase in the number of linear polarization observations starting on September 5, a dramatic increase on September 9, and a sharp fall off after September 14 (Figure 4). These days of strongest linearity correlate with the time that tremor epicenters located within 40 km of the array, producing nearly horizontal polarizations very stable azimuths of N57°E ± 8° (Figure 3 and Figure 4). This direction agrees remarkably well with both the 56° azimuth of relative plate motion between the Juan de Fuca and North American plates [*Riddihough and Hyndman*, 1991] (Figure 1 and Figure 4) and the direction of geodetically measured [*Dragert et al.*, 2001] slow-slip—a match that is predicted for near-vertical raypaths emanating from slip beneath the array (Figure 2).

[11] We quantify the comparison between predicted and observed polarization directions by looking event by event at the located tremor bursts. Our tremor location method has poor depth resolution so we assume tremor occurs on the plate interface below the located epicenter. We further consider a focal mechanism that represents slip on the plate interface with a slip vector in the direction of relative plate motion. Tremor signals are dominated by S-waves [*Obara*, 2002; *Rogers and Dragert*, 2003; *La Rocca et al.*, 2005]. We use ray theory to propagate the S-wave polarization from this source focal mechanism up to our array. Because tremor from a given 15–30 minute window is likely to come from a localized distribution of multiple sources, not a single point source, we compute the average polarization from 7 localized slip locations (6 points circularly distrib-

uted around the epicentral point) on the plate interface given by *McCrory et al.* [2004] (Figure 1) over an area with a 25 km diameter. This is a typical area of tremor sources estimated for a 30-minute window [*Obara*, 2002; *Shelly et al.*, 2007; *McCausland*, 2006].

[12] We apply this model to tremor epicenters within 40 km of our array to compare with polarization measurements from the same time windows used to locate tremor (Figure 1 and Figure S3). Our observations agree with predicted S-wave polarizations generated by slip in the direction of relative plate motion on the plate interface, supporting the similarity of the ETS phenomena in Japan and Cascadia. The mean and standard deviation of the observed minus predicted polarization azimuths for the 136 observations corresponding to station/tremor pairs with linearities > 0.7 and epicentral distances < 40 km is $-8° \pm 24°$ (Table S2). We also look at PNSN and TA stations, and, despite some correlation with predictions, there are too few linear observations within 40 km to enhance our data set. We test this methodology with the only nearby earthquake recorded by our array—a magnitude 2.9 event with an epicentral distance and depth of 48 and 49 km respectively. The mean polarization direction of the entire P, S and coda wave train across our array is N147°E, while the polarization predicted using the focal mechanism reported by the PNSN is N142.6°E (Table S3). This is a good match in a direction nearly orthogonal to the stable tremor polarizations suggesting that local geology is not significantly biasing the tremor polarization observations averaged across our array.

5. Implications

[13] Though the presence of fluids may well be an integral part of ETS, the correlation of the direction of our linear polarizations with the direction of relative plate motion is difficult to interpret in a fluid-flow tremor source paradigm. Our polarization observations are explained by S-wave polarizations produced by slip accommodating relative motion between the subducting Juan de Fuca plate and the North American plate. Our calculations have assumed the tremor source is at the plate interface, but the theoretical polarization varies weakly with source depth for sources at small epicentral distances. Therefore, our observed polarizations are consistent both with tremor emanating from the plate interface and with hypocenters distributed over a range of depths. We suggest tremor is the direct result of a patch of shear fault slip along the plate boundary, though we cannot rule out a series of vertically stratified faults parallel to, but above the plate boundary.

[14] This new understanding that Cascadia tremor and slow slip are manifestations of the same process has significant implications. First, it signifies a strong similarity between Cascadia ETS and ETS in Japan, which were previously thought to be fundamentally different [*Kao et al.*, 2005; *McCausland et al.*, 2005]. This resulting relationship suggests a global process of subduction zone dynamics associated with similar tectonic settings. Second, if LFE's do not exist or do not dominate tremor in Cascadia, this insight suggests a new type of slow shear separate from LFE's, VLF's, and short- or long-term SSE's, reinforcing the new scaling law for slow earthquakes [*Ide et al.*, 2007a].
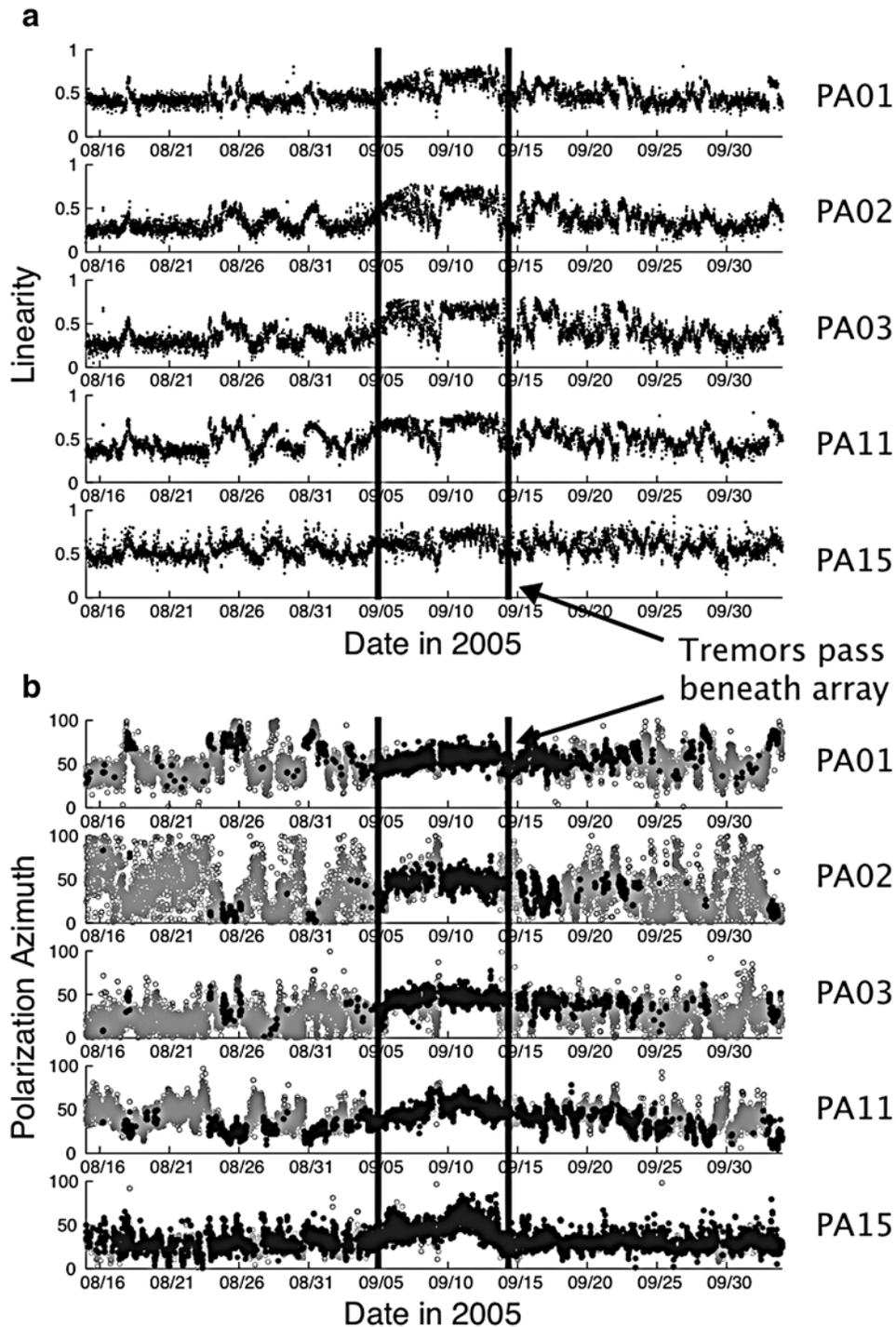
Figure 3. (a) Polarization linearity and (b) azimuth vs. time. The horizontal axis indicates time. Particle-motion linearities (Figure 3a) and polarization azimuths (Figure 3b) measured from overlapping 10-minute windows from 5PM to 5AM local time are shown for 5 array stations. Polarization azimuths (Figure 3b) with corresponding linearities > 0.5 are shown in black, with others in gray. Polarization ranges from 0 to 100 degrees in Figure 3b because polarizations between 100 and 180 with linearity above 0.5 were seldom observed during this time.
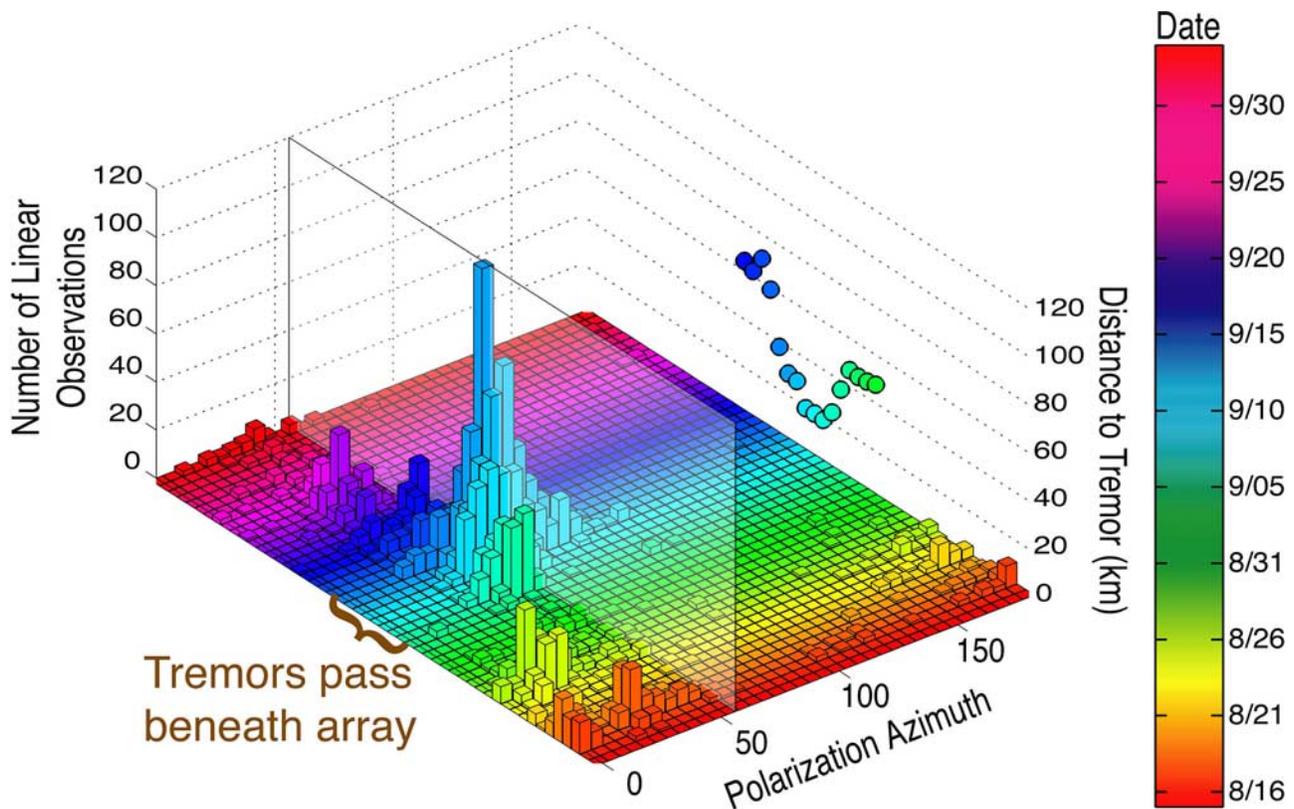
Figure 4. Daily histograms of all polarization observations with linearities > 0.7 are color coded in time. Daily median epicentral distance (dots on back plane) show when tremor was near array. The window-pane points out the predicted polarization direction for tremor close to the array—aligning with the direction of plate motion.

Finally, because geodetic slip detection often occurs after (or late into) an ETS event, the correlation establishes tremor as a valuable tool for monitoring spatial and temporal occurrences of stress loading on the seismogenic megathrust zone.

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