

INVITED REVIEW ARTICLE

Slow-slip phenomena in Cascadia from 2007 and beyond: A review

Joan Gomberg[†] and the Cascadia 2007 and Beyond Working Group*

U.S. Geological Survey, University of Washington, Department of Earth & Space Sciences, Box 351310, Seattle, Washington 98195-1310, USA

ABSTRACT

Recent technological advances combined with more detailed analyses of seismologic and geodetic observations have fundamentally changed our understanding of the ways in which tectonic stresses arising from plate motions are accommodated by slip on faults. The traditional view that relative plate motions are accommodated by a simple cycle of stress accumulation and release on “locked” plate-boundary faults has been revolutionized by the serendipitous discovery and recognition of the significance of slow-slip phenomena, mostly in the deeper reaches of subduction zones. The Cascadia subduction zone, located in the Pacific Northwest of the conterminous United States and adjacent Canada, is an archetype of exploration and learning about slow-slip phenomena. These phenomena are manifest as geodetically observed aseismic transient deformations accompanied by a previously unrecognized class of seismic signals. Although secondary failure processes may be involved in generating the seismic signals, the primary origins of both aseismic and seismic phenomena appear to be episodic fault slip, probably facilitated by fluids, on a plate interface that is critically stressed or weakened. In Cascadia, this transient slip evolves more slowly and over more prolonged durations relative to the slip in earthquakes, and it occurs between the 30- and 40-km-depth contours of the plate interface where information was

previously elusive. Although there is some underlying organization that relaxes nearly all the accrued plate-motion stresses along the entirety of Cascadia, we now infer that slow slip evolves in complex patterns indicative of propagating stress fronts. Our new understanding provides key constraints not only on the region where the slow slip originates, but also on the probable characteristics of future megathrust earthquakes in Cascadia. Herein, we review the most significant scientific issues and progress related to understanding slow-slip phenomena in Cascadia and highlight some of their societal implications. We provide a comprehensive review, from the big picture as inferred from studies of regional-scale monitoring data to the details revealed by innovative, focused experiments and new instrumentation. We focus on what has been learned largely since 2007, when several major investments in monitoring and temporary deployments dramatically increased the quality and quantity of available data.

INTRODUCTION

Less than a decade ago, Earth scientists’ view of the ways in which relative plate motions were accommodated consisted of a simple cycle of stress accumulation for hundreds of years or more on “locked” plate-boundary faults. This stress was released quasi-periodically within tens of seconds in major, commonly destructive earthquakes. Over the last decade, a significant investment in instrumentation has resulted in the serendipitous discovery and in-depth exploration of slow-slip phenomena in the Cascadia subduction zone of the Pacific Northwest region of the United States (Fig. 1) and elsewhere. Recognition that these seismically and geodetically observed phenomena are manifestations of a significant mode of fault slip that occurs down-

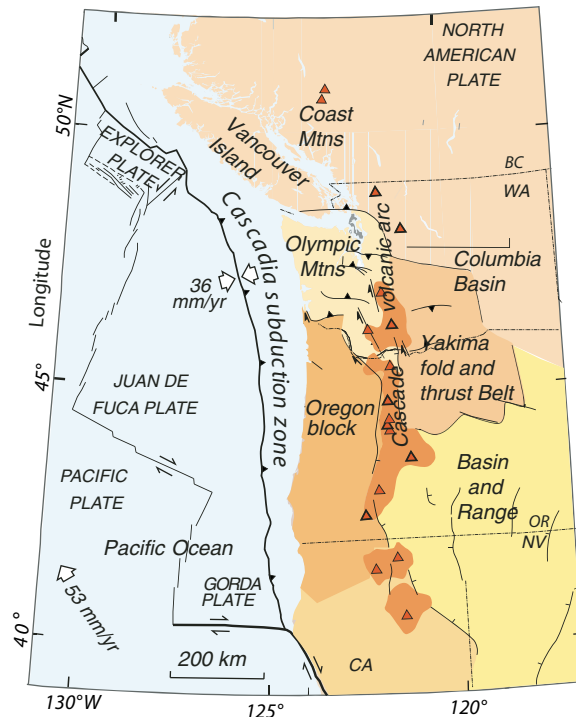
dip of the “locked” zone has radically changed the simple view of plate-boundary stress accumulation and release from that of a decade ago (Beroza and Ide, 2009; Lay, 2009). In addition to exemplifying the excitement that can surround scientific discovery and the challenge of solving new mysteries, findings from studies of slow-slip phenomena in Cascadia have significant practical implications for predicting the magnitude, recurrence, and location of future earthquakes (Chapman and Melbourne, 2009). The purpose of this review is to highlight the progress that has been made in understanding slow-slip phenomena in Cascadia, particularly since 2007, when seismic and geodetic monitoring activities in the region were expanded, made denser, and modernized thanks to the Earthscope program, national and regional investments for earthquake hazard monitoring, and a number of focused field experiments. We begin this review by presenting some background, highlighting the most significant scientific issues and progress related to understanding slow-slip phenomena and their societal implications. In subsequent sections, we provide a more comprehensive review, beginning with the big picture as inferred from studies of regional-scale monitoring data, followed by a review of the observational details that have been revealed since 2007, and noting the innovations and investments that have facilitated their illumination. We then highlight the new insights gained into the responsible physical processes, a few lessons learned from comparisons between Cascadia and Japan, and some of the most interesting outstanding questions. We conclude with speculations about information that the newest and future investments may reveal.

Progress in understanding slow-slip phenomena in Cascadia has revealed a zone of the interface between the subducting Juan de Fuca plate and overlying North American plate where ~50% of the relative plate motion is accommodated by global positioning system (GPS)-detectable

[†]E-mail: gomberg@usgs.gov

*Paul Bedrosian, Paul Bodin, Michael Bostock, Michael Brudzinski, Ken Creager, Herb Dragert, Gary Egbert, Abhijit Ghosh, Joe Henton, Heidi Houston, Honn Kao, Pat McCrory, Tim Melbourne, Simon Peacock, Evelyn Roeloffs, Justin Rubinstein, David Schmidt, Anne Trèhu, John Vidale, Kelin Wang, and Aaron Wech.

Figure 1. Schematic tectonic map of Cascadia. Major provinces (shaded areas and italicized labels) are distinguished by the dominant deformation style and rates, stress fields that drive the deformation, and geologic histories and structures. Arrows straddling major plate boundaries indicate relative plate motions, and other solid curves schematically show trends and types of crustal faults within the North American plate. Triangles denote volcanoes. Figure is modified from Wells and Simpson (2001).



transient slip events (Chapman and Melbourne, 2009). These slip episodes occur with remarkable regularity and are accompanied by distinct, low-frequency seismic signals not seen before in association with tectonic faults (Fig. 2; Dragert et al., 2001; Rogers and Dragert, 2003). The seismic energy release may be significantly smaller, and it tracks the aseismic transient slip that triggers it (Houston, 2007; Aguiar et al., 2009; Chapman and Melbourne, 2009; Wech et al., 2009). The seismic sources, and by proxy the aseismic slip, exhibit patterns indicative of slow-slip fronts that propagate across a heterogeneous surface (Ghosh et al., 2009, 2010). Explanatory models are beginning to emerge that invoke special frictional behaviors (Ide et al., 2007b; Liu and Rice, 2005, 2007, 2009; Ruben, 2008; Segall and Bradley, 2009) and an important role for fluid pressure changes (Brodsky and Mori, 2007; Audet et al., 2009, 2010).

Cascadia refers to the region affected by the convergence of the continental North American and oceanic Juan de Fuca, Gorda, and Explorer plates (Fig. 1). From west to east, the Cascadia subduction zone includes a sediment-filled forearc basin, a well-developed forearc high that extends from Vancouver Island through the Olympic Mountains and southward, a volcanic arc forming the Cascade Range, the Yakima fold-and-thrust belt, and backarc volcanism and extension in the Columbia Plateau and Basin and Range (Wells, 1989; Wells and Simpson, 2001) (Fig. 1). The relatively young (ca. 8 Ma)

crust of the largest oceanic plate, the Juan de Fuca plate, subducts at ~ 4 cm/yr with respect to stable North America. The continentally derived sediments that fill the forearc basin insulate and warm the subducting plate and are partly responsible for weakening the interface as subduction proceeds (Savage et al., 1991; Burgette et al., 2009). These sediments and dehydration of subducting oceanic crust and uppermost mantle rocks supply water to the subduction zone, to depths of ~ 50 km in Cascadia (Peacock, 2009). As described in the following, slow-slip phenomena refer to a class of observations indicative of fault-slip modes intermediate between the end-member behaviors of stick-slip, or seismic slip, and steady creep.

Earthquakes relax accumulated elastic tectonic stresses and strains in stick-slip fashion, in which strain energy stored over millennia while a fault is locked, is relaxed within seconds to minutes as the fault slips fast enough to radiate seismic waves with broad spectral content. The slip in these “fast” earthquakes grows in proportion to the fault rupture dimension, with ratios of $\sim 10^{-4}$ (i.e., ~ 1 m for an ~ 10 -km-long rupture). Because earthquake failure occurs via dynamic shear failure that involves propagating seismic wave fronts, the slip proceeds at seismic wave velocities and thus is completed within seconds (Kanamori and Brodsky, 2004). Earthquake size commonly is measured in terms of its “moment,” which equals the product of the slip, ruptured area, and material rigidity (Aki, 1966);

magnitude is proportional to the logarithm of the moment. Bigger-moment earthquakes radiate larger-amplitude and longer-duration wave trains. At the other extreme, little or no seismic radiation accompanies steady creep because the fault slips at rates comparable to that of the loading deformation, along shear surfaces sufficiently ductile that, at most, only tiny locked patches of fault (“asperities”) may exist to accumulate stresses and fail seismically (Ito et al., 2007; Schwartz and Rokosky, 2007).

We now know that some faults, or portions of faults, relieve stored stresses by slipping in ways intermediate between these end members, i.e., so slowly (over days to months) that the inertial forces involved in dynamic rupture are negligible and little or no seismic wave energy radiates, and so slippage is said to be “quasi-static.” This transient slow slip can be observed geodetically, and GPS data from some subduction zones have revealed repeated and quasi-periodic shear slip events with durations of days to many months at depths of ~ 25 – 40 km (Dragert et al., 2001; Miller et al., 2002). A family of previously unrecognized seismic signals, which differ markedly from those of earthquakes, correlates in space (Fig. 3) and time (Fig. 4) with the geodetically observed slow slip (Rogers and Dragert, 2003). Whether the seismic radiation results directly from the quasi-static slow-slip zone or from some secondary coupled process remains to be determined. A hallmark difference between slow-slip phenomena, both seismic and aseismic, and earthquakes is that for the former, the displacement appears limited to centimeters at most and to a ratio between the slip and fault-dimension (length or width) several orders of magnitude smaller (Brodsky and Mori, 2007).

Relative to the spectra of earthquake signals with similar amplitudes, the spectra of slow-slip seismic signals contain less high-frequency energy and comparable or more low-frequency energy. The most commonly observed of these signals, termed “tremor,” appears much like quasi-continuous, low-amplitude noise in the 1–15 Hz frequency band, with emergent waveforms that have envelopes coherent between recording sites separated by hundreds of kilometers (Fig. 2). These coupled seismic and aseismic phenomena collectively constitute “slow-slip phenomena.” In Cascadia, tremor always accompanies geodetically observed slow slip, although the converse is difficult to prove. The lack of geodetic signal accompanying some tremor likely reflects the detection limitations of the GPS. At its best, GPS data cannot resolve displacements less than several millimeters, which corresponds to the signal amplitude expected from a slow-slip event at typical 35 km depth with moment-magnitude $M_0 \sim 6.3$ (Aguiar

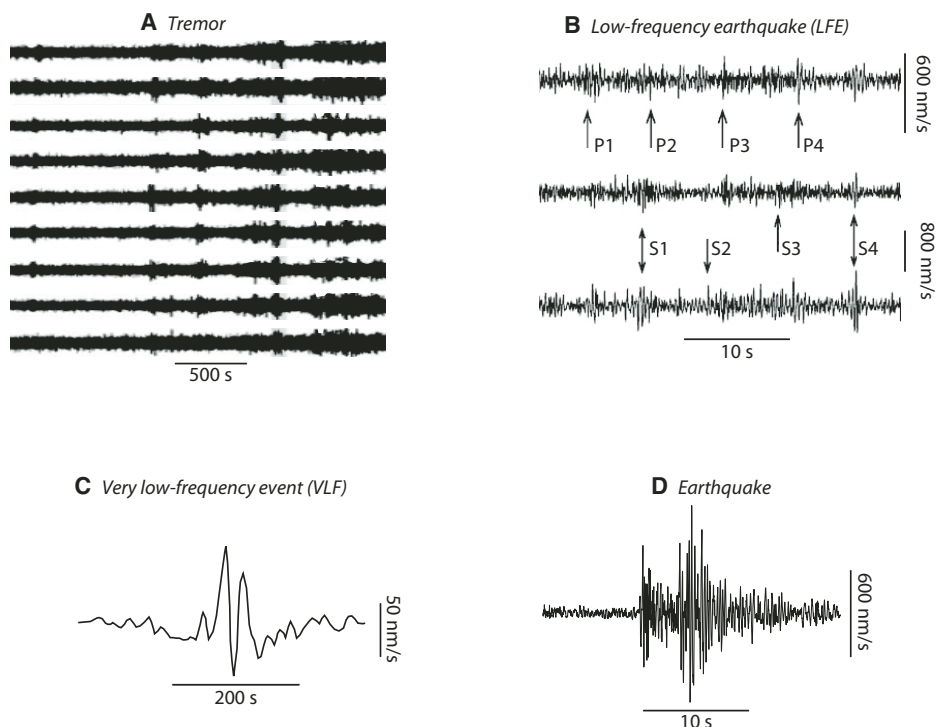


Figure 2. Examples of seismic signals accompanying slow slip. (A) Tremor recorded at seismic stations on Vancouver Island (each line shows ground motions from a different site). The relative timing of coherent bursts of energy can be used to locate the source of the radiation. Amplitudes have been normalized so that all seismograms plot with the same vertical range (from fig. 3 of Kao et al., 2004). (B) Signals from low-frequency earthquakes recorded on a temporary seismic array in western Washington. The three seismograms correspond to vertical, north-south, and east-west ground motions, with first-arriving P-waves most clearly seen on vertical components and later-arriving S-waves on horizontal components. Waves from four low-frequency earthquakes can be identified (labeled). The greater precision that distinct P- and S-wave arrival times can be measured at individual stations, relative to that of time differences of packets of tremor energy at different stations, permits estimation of more accurate source locations (from fig. 3 of LaRocca et al., 2009). (C) Example of a signal from a very low-frequency earthquake originating beneath the Kii Peninsula, Japan, on 28 May 2006, ~02:58, recorded on a vertical component sensor (from fig. 3 of Ide et al., 2008). (D) For comparison purposes, the signal from a M1.9 earthquake recorded in western Washington.

et al., 2009; Chapman and Melbourne, 2009). However, data from borehole strainmeters are more sensitive and have shown that slip accompanies shorter-duration tremor episodes that are below the GPS resolution limit (Wang et al., 2008). In northern Cascadia, the pronounced, longer-duration episodes of coupled tremor and transient aseismic slow slip occur with surprising regularity and have been named “episodic tremor and slip” or ETS (Rogers and Dragert, 2003). While transient slow slip has been documented geodetically in many places globally, the regularity and pronounced correlation with tremors observed in Cascadia are found in only a few subduction zones.

Interestingly, although one might expect that slow slip is simply part of a continuous spec-

trum of deformation modes, a popular interpretation is that there is not a continuum but two distinct modes (Ide et al., 2007b). One mode contains slow-slip phenomena, and the other includes the high-speed seismic slip that occurs in earthquakes. A distinguishing feature between these two modes is the manner in which their durations scale with the moment or energy released (Ide et al., 2007b, 2008).

Although only recognized recently, the acquisition and interpretation of signals from seismic monitoring networks operated since the 1980s, and from geodetic networks since the early 1990s, have played a key role in providing new insights into processes occurring along fault zones. Expanded continuous monitoring shows that slow slip having durations of days to

weeks occurs with near-periodicities that differ from one region to the next along the Cascadia margin (e.g., over several hundred kilometers; Brudzinski and Allen, 2007; Holtkamp and Brudzinski, 2010; Boyarko and Brudzinski, 2010). As monitoring has become more uniform throughout the Cascadia subduction zone, it has also become clear that slow-slip phenomena occur with a spatial and temporal coherence, implying an underlying organization extending over nearly the entire length of Cascadia (Fig. 5) (Boyarko and Brudzinski, 2010). Observations from temporary seismic deployments can identify significant temporal variations in tremor propagation patterns and durations on time scales of minutes to weeks (Ghosh et al., 2009, 2010; Kao et al., 2007, 2009).

Documentation and quantification of these behaviors provide strong constraints on the possible failure mechanisms. For example, tremor likely tracks the propagation of slip on the plate interface, and tremor patterns indicate that, rather than the entire plate interface slipping as a rigid block, slip fronts propagate along the strike of the subduction zone, triggering seismic radiation along an irregular, striated surface (Ghosh et al., 2010). Estimates of propagation velocities may permit elimination of some models involving fluid pressures, because fluid diffusion rates may be too slow to be consistent with velocities of small-scale tremor “streaks” and the slip fronts that may drive them (Roland and McGuire, 2009; Ghosh et al., 2009, 2010). However, fluid diffusion rates are consistent with the velocities with which tremor distributions migrate along strike on a regional scale and with ETS event periodicities (Audet et al., 2010).

Most models of aseismic slow-slip events invoke frictional processes on surfaces with properties that are transitional between values that predict stick-slip and steady sliding behaviors (Liu and Rice, 2005, 2007, 2009; Rubin, 2008; Segall and Bradley, 2009). Some studies have suggested that these large-scale events represent the coherent superposition of tiny slips that radiate tremor and other seismic slow-slip signals, but frictional models of the sources that radiate slow seismic signals have not yet emerged. Researchers have proposed that fluids play a role (Shelly et al., 2006; Audet et al., 2009) because signals with similar spectral and temporal characteristics to tremor occur beneath volcanoes and have been linked to the movement of magmas and fluids. A growing body of evidence points to near-lithostatic pore fluid pressures that reduce the effective normal stress on source faults that exhibit slow-slip behaviors (Audet et al., 2009).

The resolution of slow-slip phenomena now achieved in Cascadia has also been achieved in the subduction zone of Japan. Although we

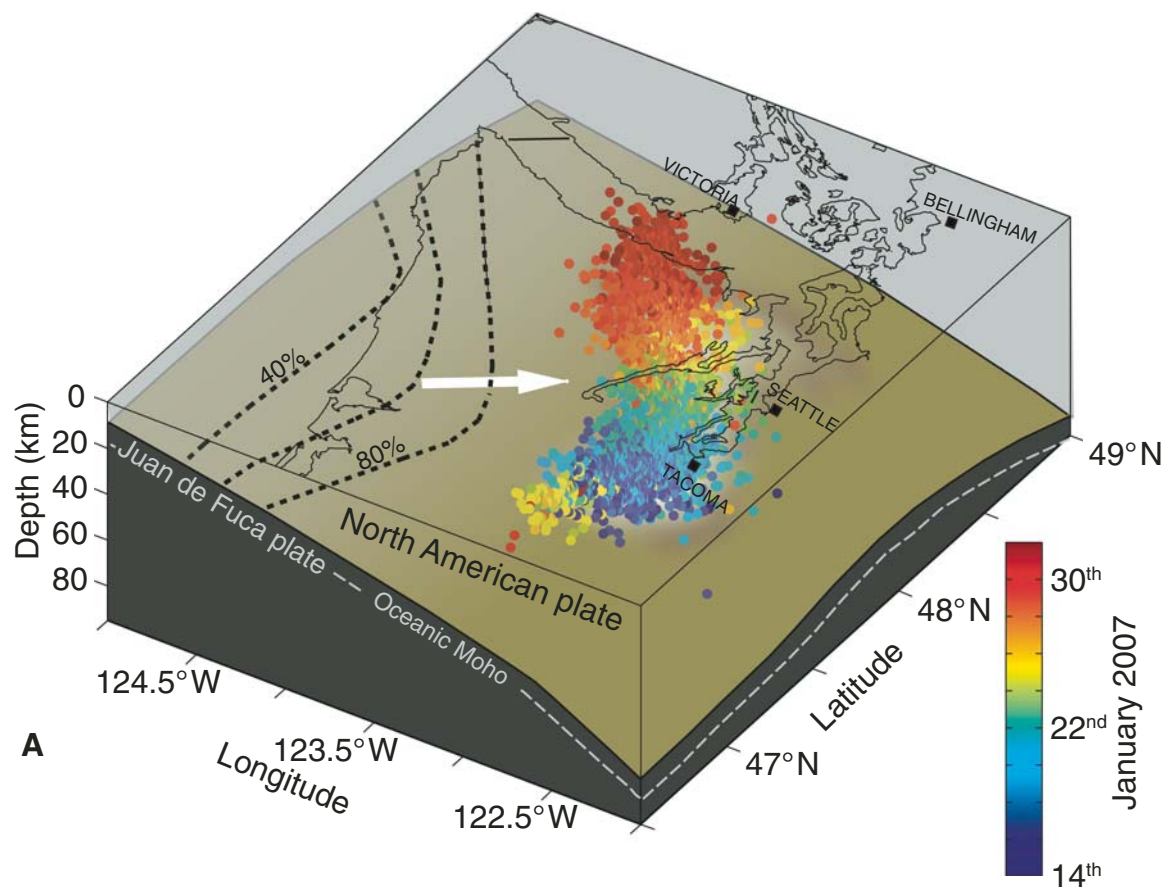


Figure 3 (on this and following page). Illustration of northern Cascadia and 2007 episodic tremor and slip (ETS) event. (A) The oceanic Juan de Fuca plate sinks or subducts beneath the continental North America plate with an ~ 4 cm/yr convergence direction, roughly perpendicular to the coast (white arrow). The plates are coupled at part of their interface (khaki-colored surface) such that relative motion is inhibited or “locked” to varying degree. The location and mechanism by which the locking changes to a freely slipping interface are uncertain. Here, we show one model in which the fraction of relative plate motion is portrayed as continuous aseismic slip that increases downdip from 40% to 80% (dashed contours) (McCaffrey et al., 2007). Inland of the locked zone, tremor epicenters projected onto the plate interface (circles) overlie the area that experienced slow slip (gray area on plate interface) during the last two weeks of January 2007. Color shading of tremor epicenters shows its temporal migration. Figure is from Forsyth et al. (2009).

focus herein on Cascadia, the identification of controlling properties and the testing of models of the processes that generate slow-slip phenomena require observations and inferences from different tectonic settings (e.g., transform, convergent, and extensional boundaries) and with varied temperatures, rock types, etc. For example, comparisons of thermal models and observations of slow-slip phenomena from Cascadia and other subduction zones lead to the conclusion that the phenomena do not require a specific temperature or metamorphic reaction (Peacock, 2009). Another example with both scientific and pragmatic implications considers the connections between slow-slip phenomena and earthquakes. A growing global database of observations is beginning to reveal an anticorre-

lation between these benign and destructive slip modes, respectively (Kao et al., 2009).

Locked subduction-zone plate boundaries are sites of “megathrust” earthquakes, like the devastating 2004 M9.1 Sumatra–Andaman and 2010 M8.8 Chilean earthquakes. Geologic and historic records show that comparable \sim M9.0 earthquakes have occurred in Cascadia, most recently in 1700, and future ones are likely (Dragert, 2007; Olsen et al., 2008). The degree to which sliding along the interface between the subducting and overlying plates keeps up with the rate of convergence, generally referred to as the “coupling” on the fault, varies with depth from being fully locked to slipping steadily. Slow-slip phenomena originate in the region between the locked zone where earthquakes oc-

cur and the freely slipping zone below. Thus, defining the locations and mechanisms by which accumulated plate motion is accommodated by slow-slip phenomena also helps to predict the location and nature of the rupture zone of future great Cascadia earthquakes, particularly their proximity to major urban communities along the Georgia Strait in British Columbia, Puget Basin in Washington, and Willamette Valley in Oregon. Previously, the boundaries of the locked zone were thought to be offshore at plate interface depths of <15 km. This estimate was based largely on GPS, strain gauge, and leveling data, and the assumption that significant locking terminated at the 350 °C isotherm, which was derived from models of the temperature variations with depth along the interface (Hyndman

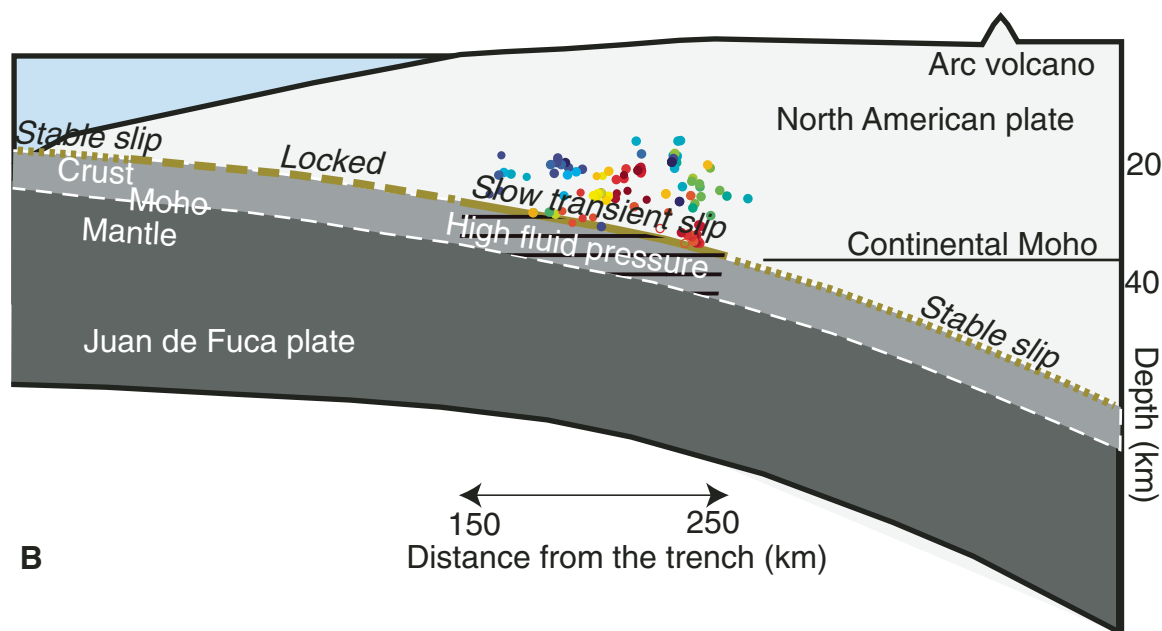


Figure 3 (continued). (B) Schematic cross section of the Cascadia subduction zone modified from Audet et al. (2009, 2010) and Kao et al. (2009). The various slip modes believed to occur along the plate interface are noted (khaki lines), and the corresponding observed phenomena indicative of each are listed below. The zone of high fluid pressure is inferred from seismic images and corresponds to the E-zone, the top of which has been inferred recently to be the plate interface where aseismic slow slip occurs. Tremor source location estimates concentrate at and above the interface (see text), albeit with large uncertainties.

and Wang, 1995; Wang et al., 2001). Only a few direct observations from within the locked zone exist because large or small earthquakes on the interface fault are rare in Cascadia (Trèhu et al., 2008). A simple relationship between locking and the thermal structure now appears questionable, as evident in results of ETS studies (Kao et al., 2005; Wech and Creager, 2008; Boyarko and Brudzinski, 2010) and new thermal models (Peacock, 2009; Kummer and Spinnelli, 2009). New GPS analyses along the entire margin suggest that the depth of coupling varies along strike and perhaps non-monotonically along dip (Holtkamp and Brudzinski, 2010). The most significant practical change resulting from analyses of ETS observations has been the suggestion that the locked zone and future great earthquakes may extend ~60 km farther inland in Washington (Chapman and Melbourne, 2009). In southernmost Oregon, a recent study of tidal and leveling records also infers an inland shift of the transition from locked to decoupled by ~30 km (Burgette et al., 2009).

Knowledge of slow-slip phenomena also may improve time-dependent forecasts of damaging earthquakes. Although slow-slip relieves accumulated stresses in a nondestructive manner, it also perturbs the stress field acting on the locked zone of the plate interface, as well

as on the surrounding faults. As observations of both earthquakes and slow-slip phenomena have improved, a physically meaningful spatial and temporal relationship between them has begun to emerge (Kao et al., 2009). At present, the change in probability estimates of a major earthquake on the locked zone of the plate interface resulting from several centimeters of slow slip can be calculated by employing Coulomb stress and frictional models (Mazzotti and Adams, 2004; Beeler, 2009). The readiness and role of such probability estimates in formulating public policy are now being debated.

THE BIG PICTURE

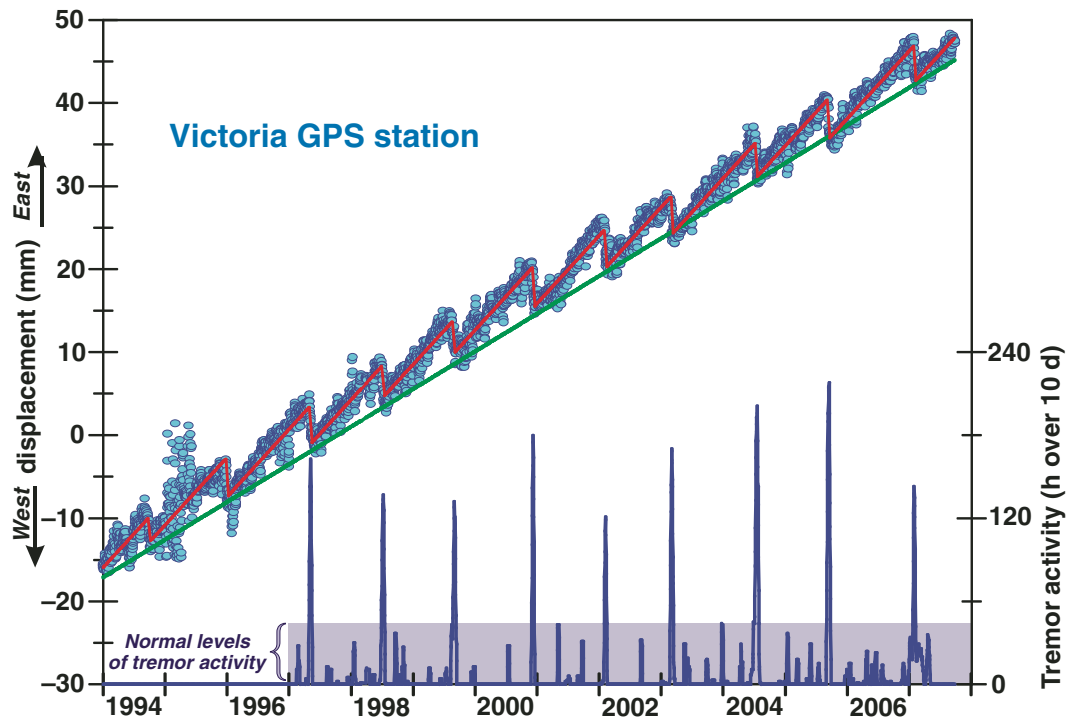
The “big picture” of slow-slip phenomena in Cascadia has been documented using data from regional networks of permanent, continuously recording seismic and GPS stations. In early 2007, the density of seismic stations reached a maximum in northern Cascadia as the Earthscope Transportable USArray rolled through the region, and coverage was supplemented by several temporary arrays. In addition, the Earthscope Plate Boundary Observatory provides continuous data from new permanent GPS stations and, for the first time in Cascadia, from borehole strainmeters and borehole seis-

mometers. In this section, we highlight some of the key elements of the “big picture” illuminated using data from the permanent seismic and GPS installations.

Quasi-static slow-slip events manifest as transient reversals in geodetically observed displacement directions (Figs. 4 and 5A). These aseismic slip events occur regularly along the length of Cascadia, have a characteristic duration of one to five weeks, exhibit uni- or bi-directional along-strike propagation, and typically accrue 5 mm of transient surface displacement. Over three dozen geodetically inferred transient slip events have been observed Cascadia-wide since 1997 (Rogers and Dragert, 2003; Szeliga et al., 2008; Kao et al., 2009; Chapman and Melbourne, 2009; Holtkamp and Brudzinski, 2010; Schmidt and Gao, 2010). Slip models derived from the GPS data show a consistently narrow downdip width of the slow-slip zone—the majority of slip occurs between the 30- and 40-km-depth contours on the plate interface (assuming the widely cited interface depth model of McCrory et al., 2006) (Fig. 6).

Peaks in tremor activity have accompanied all of the quasi-static slow-slip events (Fig. 4), and have been documented on regional seismic networks with similar regularity since 1997 (Kao et al., 2009). In Cascadia, GPS-detected

Figure 4. Iconic display of episodic tremor and slip (ETS) in Cascadia showing the synchronous occurrence of slow slip and vigorous tremor activity. Daily changes in measured east-west displacements of the global positioning system (GPS) station at Victoria, British Columbia, relative to stable North America (blue circles) show that the site moves eastward on average at ~5 mm/yr (green line). It does so in a sawtooth fashion by moving eastward more rapidly than the average for 15 mo and then westward ~4 mm over a period of 2 wk (red line). These transient reversals are caused by slow slip on the deeper plate interface. Tremor activity occurs throughout the year generally for less than a few hours per day, except during slow-slip events, when activity is seen to increase by more than an order of magnitude. Only analog seismic data exist prior to 1996, and the same pattern is apparent in these but is not easily quantified. Figure is updated from Rogers and Dragert (2003).



slow-slip events are always accompanied by tremor (Figs. 4 and 5) and both locate in the same or overlapping regions (Fig. 6), landward of the inferred megathrust seismogenic zone (Wech et al., 2009; Aguiar et al., 2009; Kao et al., 2009; Boyarko and Brudzinski, 2010). Some tremor is located above this to ~25 km depth, and tremor between ETS events appears to be deeper, between 35 and 45 km depth (Aguiar et al., 2009; Kao et al., 2009; Wech et al., 2009; Audet et al., 2010; Boyarko and Brudzinski, 2010). The accompanying tremor distribution appears to be resolvable wider than the region of detectable quasi-static slow slip, systematically extending further inland (Fig. 6), for reasons that can only be speculated about. Measured surface displacements constrain models of causative slip along the subduction plate interface, which have been used to estimate equivalent moment magnitudes ranging between M_w 6.2 and 6.8 for individual slip episodes. Although the data are not as complete south of Washington State, the general coincidence of aseismic slow slip and tremor appears to apply to the entire Cascadia subduction zone, at least on scales of tens of kilometers and several days (Boyarko and Brudzinski, 2010) (Fig. 5).

The similarity in temporal and spatial distributions of tremor activity and quasi-static slow slip and inference that the two phenomena are physically coupled permit quantitative budget-

ing between the accumulation and relaxation of strain energy in Cascadia. This budgeting is an essential ingredient of earthquake forecasts and hazard assessments. Although tremor sources themselves likely accommodate an insignificant fraction of the plate-motion deformation (Houston, 2007), some studies have proposed using tremor as a proxy to monitor quasi-static slow-slip (Aguiar et al., 2009; Hiramatsu et al., 2008; Shelly, 2010) based on the fact that several measures of tremor activity appear proportional to geodetically derived estimates of slip. Assuming this relationship can be extrapolated to smaller slip values, the detection of tremor in the absence of GPS signals makes tremor a more sensitive indicator of small-magnitude quasi-static slow slip. Moreover, tremor sources can be tracked with greater spatial and temporal resolution than geodetically estimated slow slip. Wech et al. (2009) made several important inferences from a tremor catalog from northern Cascadia spanning 2004–2008, derived from Pacific Northwest Seismic Network data. As in Chapman and Melbourne (2009), the continuous cataloging of tremor activity in northern Cascadia by Wech et al. (2009) shows that 45%–65% of the plate convergence rate of 4 cm/yr is accommodated during ~2-wk-long ETS events every ~15 mo, and 55% of all tremor activity occurs during these events when GPS data indicate that 2–3 cm of quasi-static slip oc-

curs on the plate interface. The remaining 45% of all tremor is observed during the ~14.5 mo between ETS events, when slip is presumably too small to be detected by GPS. However, it is uncertain whether these inter-ETS tremors indicate that the remaining convergence is fully accommodated because these tremor sources are located slightly downdip of ETS tremor sources. The Wech et al. (2009) catalog also shows that the updip edge of the distribution of tremor epicenters is very sharp, rising from just detectable to its peak within ~20 km, and this may reflect a change in plate interface properties. Just updip of this boundary, the plate interface may be locked, with all plate convergence accommodated as seismic slip during megathrust earthquakes. Unfortunately, the interpretations based on available GPS measurements are nonunique, and it is also possible that this region accommodates some fraction of plate convergence accommodated by slow slip that is either continuous or varies on time scales longer than the 10 yr since monitoring began. However, such a model requires that such slip occur here without generating any tremor.

THE FINER-SCALE PICTURE

Details of slow-slip phenomena on time scales of days to weeks, and spatial scales of kilometers, add important constraints on the underlying

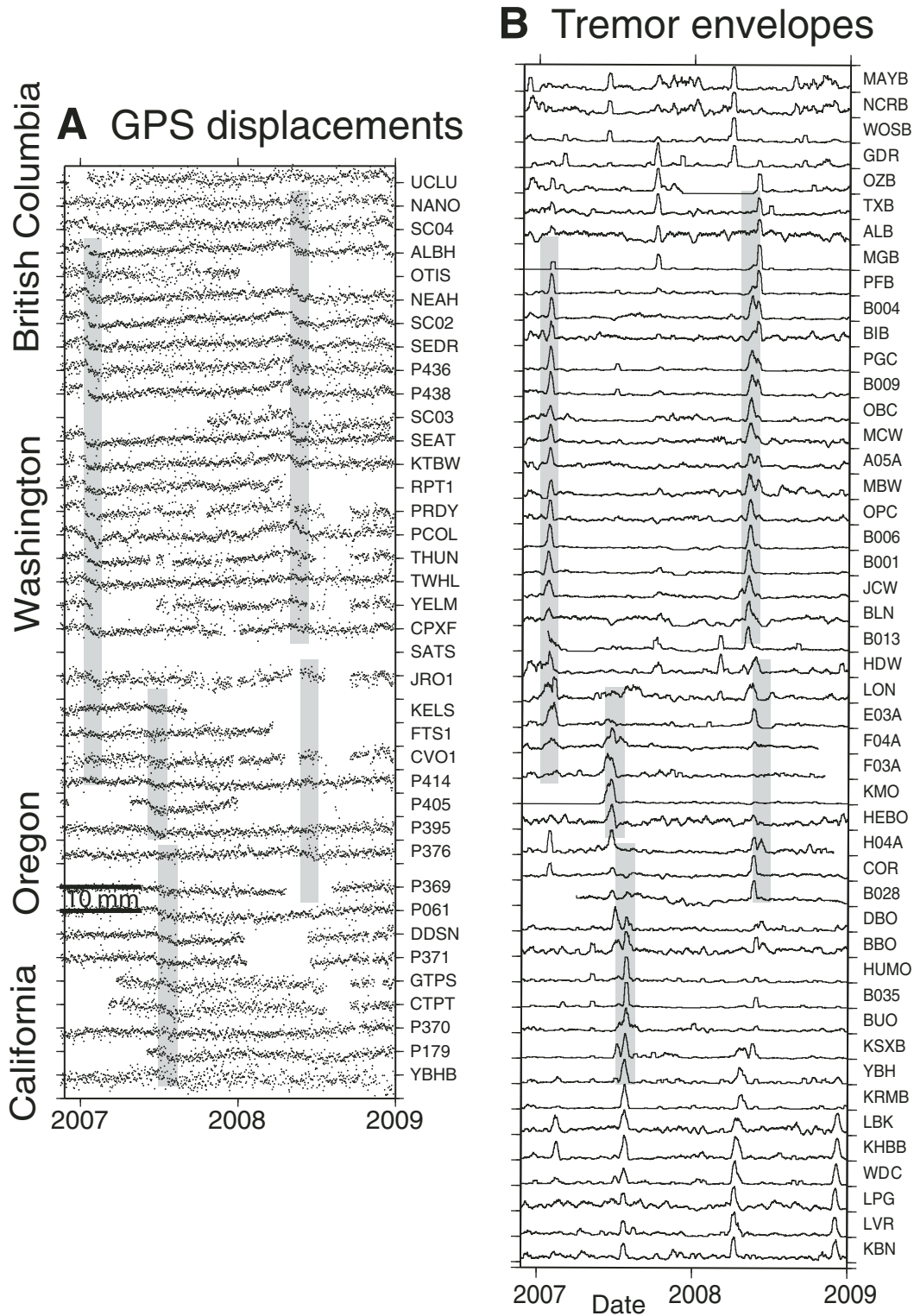
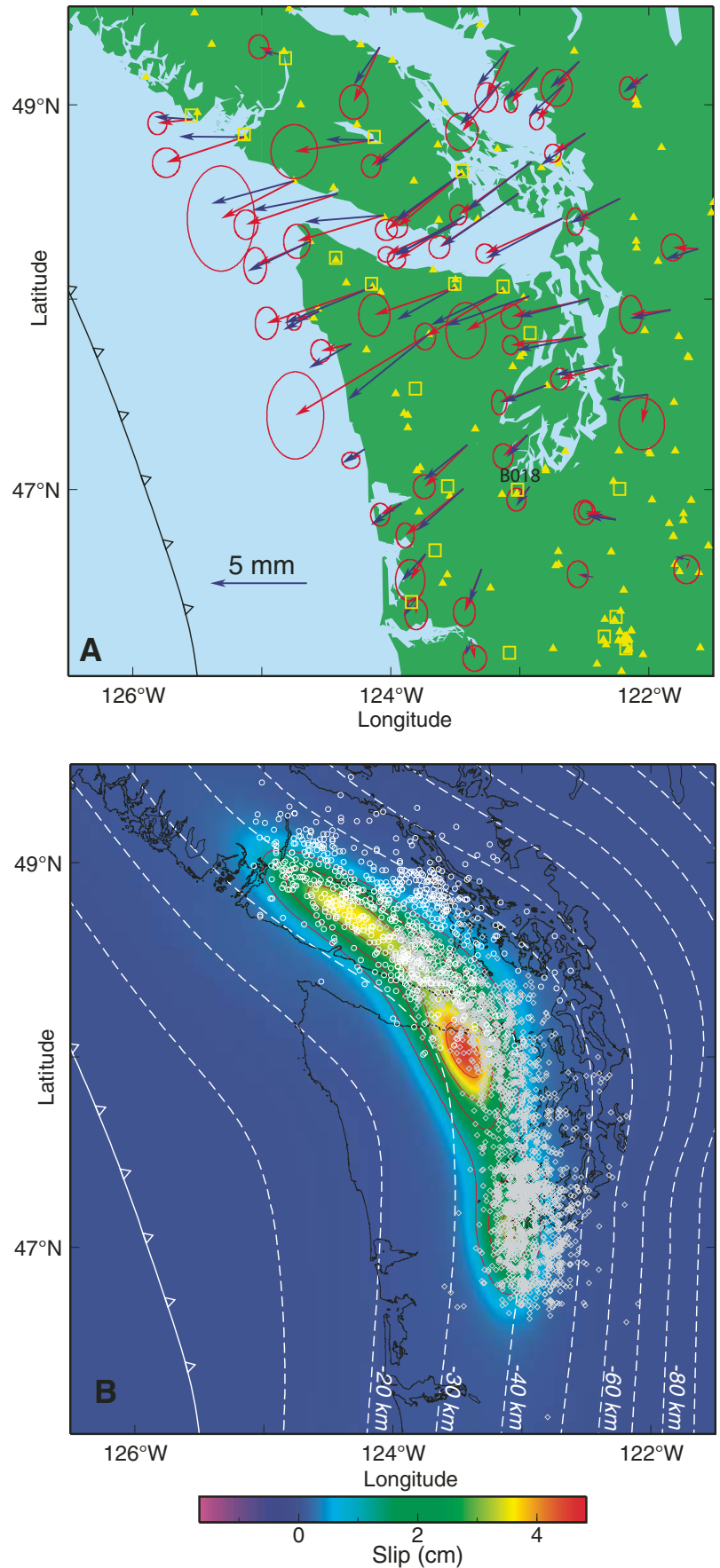


Figure 5. Global positioning system (GPS) and seismic signatures of the 2007 and 2008 episodic tremor and slip (ETS) events. (A) Slow-slip events manifest as coherent reversals in east-west displacements measured at GPS stations (labeled) throughout Washington to northern Vancouver Island (gray lines). Slip beneath Oregon to Mendocino, California, occurred separately in 2007 and is poorly constrained for 2008. (B) Seismic-wave amplitudes recorded at seismic stations similarly distributed along a N-S transect. These have been filtered to exclude frequencies outside the range of tremor signals, normalized, and smoothed over hour-long time windows. The peaks in tremor activity correlate with slow-slip events (in gray). Additional tremor episodes can be seen, particularly in northern California. During a 2 mo period in 2008, tremor is recorded along nearly the entire subduction margin.

Figure 6. Slow-slip inferred from global positioning system (GPS) data. (A) Observed (red arrows with 95% error ellipses) net displacements measured by GPS for the May 2008 episodic tremor and slip (ETS) episode, with modeled displacements (blue arrows) calculated from the slip model in B. Yellow triangles denote seismic stations, and squares denote borehole strainmeters. (B) Slip model derived from the inversion of observed horizontal and vertical displacements, with slip constrained to the plate interface with geometry from McCrory et al. (2006). As a smoothness constraint, the slip was assigned Gaussian autocorrelation lengths of 50 km along strike and 25 km normal to strike of the subduction zone. The direction of slip as determined by the inversion is updip, roughly parallel to plate convergence, but with small spatial variations. Curiously, the greatest density of tremor epicenters (symbols) does not overlie the maximum slip. Different tremor catalogs covering the northern (white circles) and southern (gray diamonds) portions of the region have been merged, but the latter had to be decimated by a third to have roughly the same density where catalogs overlap. This difference in detection rate presumably reflects the differing detection and location methods used to derive each catalog.



processes to the big picture. We illustrate this with examples from the 2008 ETS episode. Permanent network monitoring data reveal that the phenomena migrate steadily along strike at rates of a few to tens of km/d, and tremor activity exhibits halting (i.e., spatially stationary but varying temporally on scales of several days) and/or jumping behavior (i.e., clusters appear as far as 150 km away, without events in between) (Kao et al., 2009; Boyarko and Brudzinski, 2010). The prolonged May 2008 ETS episode spread bidirectionally from beneath the center of the Olympic Peninsula to the northwest and southeast at an average rate of 6–9 km/d, with the northwest migration possibly accelerating from ~5 to 10 km/d (Dragert et al., 2008).

Although revealing processes that are non-stationary in both time and space, the GPS data do not resolve quasi-static slow slip on scales comparable to the resolution of individual tremor sources. In addition, the permanent seismic network data only capture tremor of the emergent, 1–15 Hz flavor and miss tremor when noise levels are high during daytime hours or storms. Strainmeter data and seismic data from small-aperture, dense seismic arrays mitigate these shortcomings and reveal unanticipated findings.

Special data stacking methods, or “array beam-forming” analyses, applied to data from a dense array of many tens of seismometers deployed within an aperture of about a kilometer, yield higher-resolution estimates of tremor source distributions in time and space, albeit with diminishing capability with distance from the array. Analysis results from the first such array experiment in Cascadia find that the tremor epicenter distribution is confined between the 30–40 km plate-interface depth contours projected to the surface, noting that tremor source depths were constrained a priori to lie on the plate interface (Ghosh et al., 2009, 2010). These arrays have resolved several distinct patches on the Cascadia subduction interface that released much of the tremor moment during the 2008 ETS event (Fig. 7). Although unable to resolve heterogeneity at the same scale, it is noteworthy that the slip imaged with GPS appears greatest in an area just adjacent to, but not quite coincident with, the tremor patches. Adjacent, rather than coincident, quasi-static slow-slip and seismic sources have been observed in Japan (Obara et al., 2009), Alaska (Peterson and Christensen, 2010), and Mexico (Song et al., 2009), and they suggest

that the mode of stress relaxation (quasi-static or seismic) depends on permanent characteristics rather than transient or evolving processes.

Array analyses also show streaks of tremor sources in addition to the overall pattern of the slower, ~10 km/d, along-strike migration (Ghosh et al., 2010). A streak manifests as a continuous and steady migration of tremor sources on the time scale of several minutes to hours with velocities up to tens of kilometers per hour, mostly parallel to relative overall relative plate movement (Fig. 7; Vidale et al., 2009). Both the along-strike slower propagation and convergence-parallel rapid streaking of tremor sources are reminiscent of propagating ruptures in regular earthquakes. If tremor is a proxy for quasi-static slip that relaxes the tectonic stress, these observations imply that relaxation occurs progressively as a propagating front. Streaks of earthquakes parallel to the long-term slip direction have been inferred to represent geometric or frictional striations (Waldhauser et al., 2004). Similar tremor streaks have also been observed along the San Andreas fault in California by Shelly et al. (2009), and as noted in that study, future observations are needed to establish whether they recur in the same place, before a definitive interpretation can be made.

The quieter seismic array site and noise reduction achieved by data stacking permits detection of smaller signals. Results from array analyses revealed four times as much tremor as from the permanent network, suggesting a process of continuous chatter in which the volume is raised and lowered (i.e., as the driving aseismic slip increases and decreases). That is, tremor is on all the time, but with fluctuating volume, and is only audible when louder than the detection threshold of the observational system. This is consistent with the aforementioned study of Wech et al. (2009), who noted ~45% of tremor between ETS events and suggested that although not always detectable in GPS data, the interface is always slipping quasi-statically at some rate and effectively serves as the tremor volume control. The array-enhanced seismic detection has also revealed classes of slow-slip events that are well documented in southwest Japan but were previously elusive in Cascadia. These include nearly 100 low-frequency earthquakes (often referred to as LFEs) with magnitudes of 1–2 and durations of 0.1–1 s, much longer than regular earthquakes of comparable magnitude, which are thought to be the “building blocks” of continuous tremor (Katsumata and Kamaya, 2003; Shelly et al., 2006; Brown et al., 2009; LaRocca et al., 2009). Additionally, larger (M 2.5–4), longer-duration (10–100 s), very low-frequency earthquakes (VLFs) (Obara and Ito, 2005; Ito et al., 2009) have now been

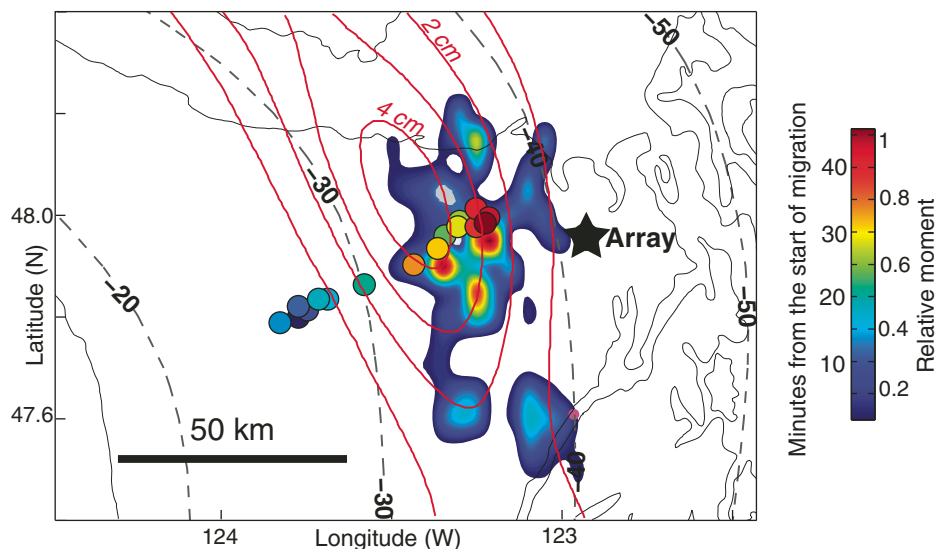


Figure 7. 2008 episodic tremor and slip (ETS) event tremor moment and migration, and global positioning system (GPS)–estimated slip (from Fig. 6 herein and figures in Ghosh et al., 2009). Beam-forming estimates of the distribution of total tremor seismic moment (relative to the maximum in the interval) show that it radiates mostly from three small patches (blue to red shading). The region of maximum slip inferred from GPS data is more poorly resolved (red curves are contours from Fig. 6), but coincides with the tremor patches. Samples of successive individual beam-forming estimates of tremor source locations during a 45 min period (circles) show very rapid, linear migration. The color scale corresponds to both the time spanned by the migrating tremor locations relative to the first source in this particular sequence and to the spatial distribution of seismically estimated moment normalized by the maximum for the entire 2008 ETS event.

detected in Cascadia, also at a much lower rate than in Japan (Sweet et al., 2008). We now know that models applied to explain slow-slip phenomena in Japan can reasonably be applied to Cascadia. In particular, an applicable model proposes that quasi-static slow-slip drives seemingly independent but concurrent failure of multiple tiny sources that radiate tremor signals. Occasionally, low-frequency and very low-frequency earthquakes result as these sources coalesce into organized rupture able to radiate much lower frequency waves (Ito et al., 2007, 2009). It remains to be determined whether the more elusive nature of LFEs and very low-frequency earthquakes in Cascadia reflects physical or observational capability differences.

Strainmeter data are proving much more sensitive to slip propagation and can detect smaller-magnitude strain events compared to GPS. The noise, and thus the resolution of strainmeter and GPS data, depends on signal duration. For strainmeter data and durations of 1 and 15 d, the resolution exceeds that of the GPS by factors of 20 and 5, respectively, corresponding to minimum resolvable strainmeter signals of ~ 5 and 40 nanostrain (Langbein, 2004; Langbein et al., 2006; Smith and Gomberg, 2009). If passing by the strainmeter, the direction of travel and loca-

tion of slip can be tracked, the latter within tens of kilometers. Strainmeter data have resolved signals characteristic of slip on the plate interface for very small (<30 km) asperities (Wang et al., 2008). Strainmeter and tremor observations have confirmed the synchronicity of these localized slip and tremor events (McCausland et al., 2008), within a few hours, of transient strain signals and of the arrival of seismic signals (Fig. 8).

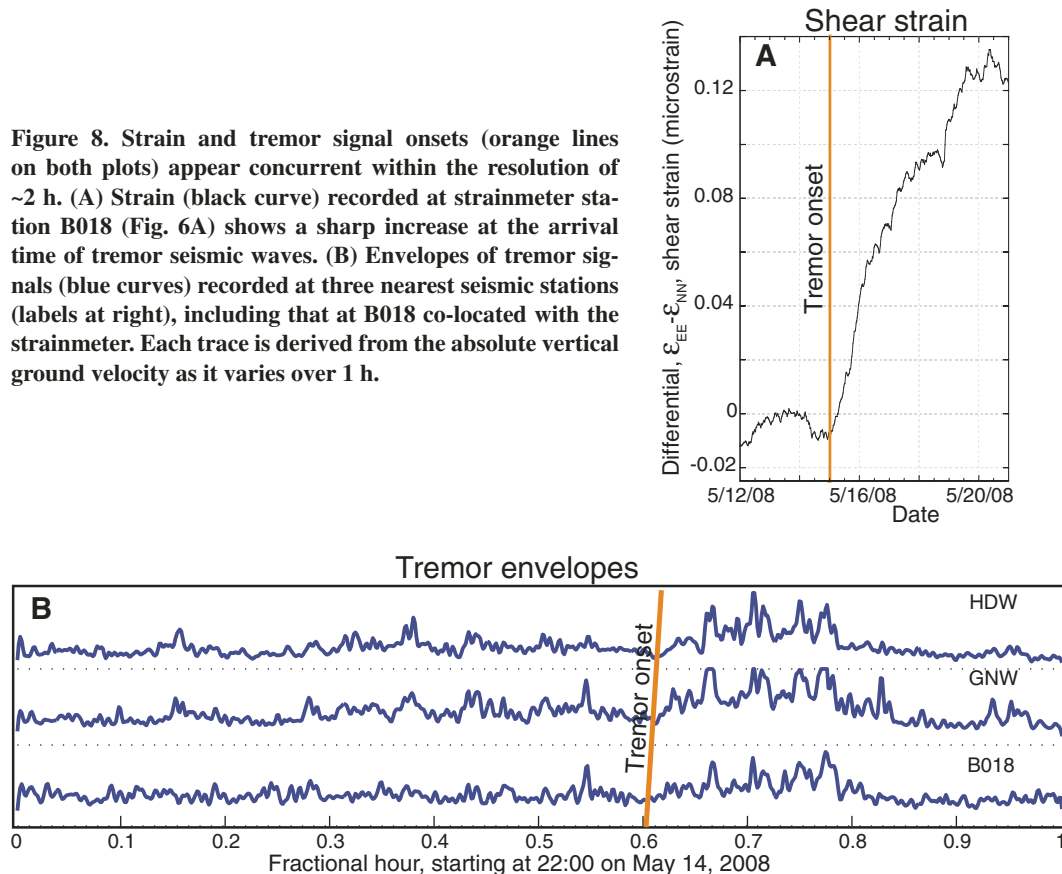
In northern Cascadia, the recent investments in instrumentation and research have significantly improved resolution, and in southern Cascadia, they have answered more elementary questions about whether these phenomena occur and if so, the basic characteristics they exhibit (Brudzinski and Allen, 2007; Boyarko and Brudzinski, 2010). We find ETS is also prevalent in southern Cascadia, with tremor epicenters occurring in a narrow band similar to that in the north (Fig. 9). ETS activity in 2007–2008 over the southern half of Cascadia commonly occurred synchronously with that in the north (Fig. 5B). Moreover, in a few cases, the locations of tremor sources show an intriguing spatial correlation with earthquakes and structural features. The locations of some clusters of tremor sources may correlate spatially with moderate- to micro-earthquakes up-

dip and offshore on the megathrust at depths of 11–16 km, with forearc segmentation, with subducted seamounts offshore (Trèhu et al., 2008), and with variations in the distribution of plate coupling as determined from GPS and leveling data (Burgette et al., 2009).

NEW INSIGHTS INTO PHYSICAL PROCESSES

Mounting evidence points to a plate interface that has very low effective stresses and/or is critically stressed. One set of clues comes from results of theoretical and numerical modeling. Current models of quasi-static slow-slip events almost all involve frictional failure, where fault strength depends on sliding velocity. Numerical experiments invoking rate- and state-dependent friction can produce episodic behavior with repeat times similar to those of Cascadia ETS events, and most seem to require low effective stress to do so (Liu and Rice, 2005, 2007, 2009; Rubin, 2008; Segall and Bradley, 2009). In some models, episodic slow-slip events represent a type of oscillatory behavior between (sliding) velocity weakening at slow-slip speed and velocity strengthening at higher slip speed (Rubin, 2008). Fluids are often involved in these

Figure 8. Strain and tremor signal onsets (orange lines on both plots) appear concurrent within the resolution of ~ 2 h. (A) Strain (black curve) recorded at strainmeter station B018 (Fig. 6A) shows a sharp increase at the arrival time of tremor seismic waves. (B) Envelopes of tremor signals (blue curves) recorded at three nearest seismic stations (labels at right), including that at B018 co-located with the strainmeter. Each trace is derived from the absolute vertical ground velocity as it varies over 1 h.



frictional models, to elevate pore pressures and thus keep effective stresses low. Fluids may also be important in limiting the slip amplitude and/or speed through the mechanism of dilatancy strengthening (Segall and Bradley, 2009; Samuelson et al., 2009).

Evidence of very low effective stresses and/or a critically stressed plate interface in Cascadia also comes from observations of seismically triggered (Rubinstein et al., 2007, 2009) and tidally modulated tremor (Rubinstein et al., 2008; Lambert et al., 2009) and tidally modulated quasi-static slip (Hawthorne and Rubin, 2009). These observations indicate very small stress changes, on the order of a few to a few tens of kPa, can cause slip and increased tremor rates. In the case of seismically triggered tremor, tremor bursts are observed synchronously with the arrival of earthquake-generated seismic wave

peaks. These peaks impart failure-promoting stress changes by enhancing tectonic shear stresses and/or pore-pressure changes (Hill, 2010). Modulated tremor rates and strainmeter signals at the same periods as the dominant Earth and ocean tides also imply a sensitivity to stress changes orders of magnitude smaller than lithostatic stresses (100 MPa) and smaller than even the low stress drops estimated for quasi-static slow-slip events (~10 to ~100 kPa; Brodsky and Mori, 2007). This sensitivity implies very small effective confining stresses and/or a critically stressed state, the former of which is most easily achieved with high pore pressures.

Seismological studies of the Cascadia subduction-zone structure are providing insights into the relation between ETS and plate-boundary processes. The most widely used model of the entire plate interface is a surface defined primarily by the upper limit of seismicity within the subducting plate (in northern Cascadia) and the location of the Moho of the subducted plate (Fig. 3) estimated from P-wave seismic-reflection and -refraction data (McCroly et al., 2006). Several alternative plate-interface models exist for northern Cascadia, depending mostly on the interpretation of the nature of a band of high seismic reflectivity and low seismic velocity called the E-zone. One interpretation is that the E-zone itself is the subducted crust (Audet et al., 2009, 2010; Fig. 3B), which is consistent with interpretations of similar low-velocity zones in southwest Japan and Mexico (Shelly et al., 2006; Song et al., 2009). The inferred interface would then be shallower by several kilometers in the area where ETS occurs, and the tip of the mantle wedge would be 20 km farther landward (Audet et al., 2009, 2010). The reported high Poisson ratio (the ratio of axial to transverse strains) within the E-zone and evidence for serpentinization of the forearc mantle wedge are interpreted to imply a dramatic change in permeability across the plate interface in the vicinity of the wedge corner (Audet et al., 2009; Fig. 3B). A high Poisson's ratio also has been inferred beneath a tremor zone in Nankai, Japan (Shelly et al., 2006).

Fluids also may facilitate tremor and quasi-static slow slip via mechanisms like permeability pumping (Miyazawa and Brodsky, 2008) and hydrofracturing (Wang et al., 2006). As in other subduction zones, Cascadia ETS occurs around the seaward tip of the mantle wedge, where thermal-petrological models argue for the presence of free fluid (Wada et al., 2008) generated by prograde metamorphic reactions in the subducting plate at the depths of ETS (Peacock, 2003). Seismic observations now corroborate these models (Abers et al., 2009; Audet et al., 2009, 2010), and all together led Audet et al. (2010) to propose a feedback process that explains many of

the features of Cascadia ETS, including its nearly periodic recurrence. Dehydration of the subducting oceanic crust releases fluids that are trapped at the plate boundary where it contacts the overlying continental crust but are released where the boundary contacts the mantle wedge. Trapped fluids cause pressure to build to near-lithostatic levels within the oceanic crust beneath the sealed interface, lowering the effective stress on the plate boundary and allowing slip to initiate and grow. As slip propagates, the seal experiences hydrofracturing and becomes more permeable, allowing fluid flow. Pore-fluid pressure then drops, and the effective stress rises, inhibiting further slip. The low permeability seal eventually is restored by precipitation or grain-size reduction due to fracture, and the cycle repeats, much like the fault-valve model proposed to explain recurrent earthquakes (Sibson, 1990). Tracking the presence of moving fluids deep in the Cascadia subduction zone now seems possible, albeit nonunique, based on inferences elsewhere of the temporal evolution of seismic wave velocities (Husen and Kissling, 2001).

LEARNING FROM JAPAN

An accounting of the similarities and differences between Cascadia and other regions provides clues about processes that cause slow-slip phenomena. Japan leads globally in activities focused on studies of slow-slip phenomena, and many of the phenomena observed in Cascadia also have been documented there. Slow-slip phenomena have been observed in southwest Japan where the Philippine plate subducts beneath the Eurasian plate, but not elsewhere where the older Pacific plate is more rapidly subducting. In Cascadia and southwest Japan, relatively young (<20 Ma), warm crust subducts at modest plate convergence rates (~4–6 cm/yr). This association between plate age and convergence rate suggests a role for metamorphic reactions. Although thermal and petrologic models suggest that prograde metamorphic reactions likely occur in the subducting crusts in southwest Japan and Cascadia, the thermal structures differ, and thus the metamorphic reactions that likely facilitate slow-slip phenomena also must differ (Peacock, 2009). The presence and trapping of fluids released during dehydration reactions have been inferred to facilitate transient slip in Japan and Cascadia (Shelly et al., 2006; Kodaira et al., 2004; Audet et al., 2009, 2010). Such inferences come from seismically imaged low material velocities and high Poisson's ratios coincident with tremor zones (Shelly et al., 2006; Audet et al., 2009, 2010).

The patterns of tremor migration, and by inference slow slip, inferred for regions of Japan

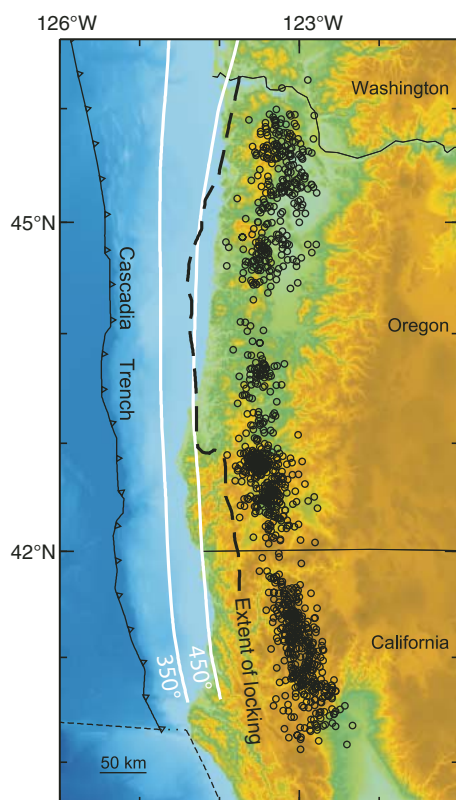


Figure 9. Source locations of tremor bursts in southern Cascadia from late 2005 to late 2007 (from Boyarko and Brudzinski, 2010). Epicenters (circles) are displaced as much as 50 km inland from the thermally defined transition zone (limited by the 350–450 °C interplate isotherms, shown as white lines; Hyndman and Wang, 1995) and one possible model of the downdip edge of interseismic locking inferred from historical leveling data (dashed black line) (Burgette et al., 2009).

are strikingly similar to those in Cascadia. For example, Shelly and Beroza (2007) documented along-strike migration in western Shikoku at rates comparable to those in Cascadia; regional-scale clusters migrate along the strike of the subduction zone at tens of km/d, with streaks that propagate up- and downdip with velocities 20–150 km/h over distances of up to ~20 km. This implies that slow slip is a fundamental process that develops as a propagating rupture front. Another inference common to both Cascadia and southwest Japan is the occurrence of slow slip on surfaces near failure because of near-lithostatic pressures (resulting from the fluid processes discussed previously) and/or at critical stress levels. As already noted for Cascadia, this is also suggested in southwest Japan by the efficacy of very small stress changes to trigger tremor, evident in tidally modulated tremor rates (Shelly and Beroza, 2007; Nakata *et al.*, 2008) and triggered by seismic waves (Miyazawa and Mori, 2005; Miyazawa and Brodsky, 2008; Miyazawa *et al.*, 2008).

In southwest Japan, Obara (2002) first observed tremor radiating from a discontinuous band of sources distributed around the 30-km-depth contour of the plate interface. Subsequently, a simple empirical relationship between tremor activity and aseismic slip established a clear causal connection between these phenomena (Hiramatsu *et al.*, 2008). As in Japan, a similar relationship has been noted in Cascadia (Aguier *et al.*, 2009). Although causally related, quasi-static slow-slip and seismic sources do not appear to be co-located in Japan (Obara *et al.*, 2009), which seems reasonable if the mode of stress relaxation (quasi-static or seismic) depends on permanent properties of the plate interface. Both modes appear to represent shear slip on the plate interface (Shelly *et al.*, 2006; Ide *et al.*, 2007a). In Cascadia, the evidence also indicates that the seismic and aseismic signals result from shear failures, but the source locations of both are more uncertain, and tremor locations are a matter of particular debate, with some estimates indicating a broad distribution of tremor sources extending several tens of kilometers at and above the interface (Kao and Shan, 2004; Kao *et al.*, 2005, 2006) and others on the interface (Wech and Creager, 2007; LaRocca *et al.*, 2009).

Seismic and aseismic slow slip appears to be much more varied in southwest Japan than in Cascadia. In overlapping regions, aseismic “short-term” slow-slip events correlate with tremor activity, and “long-term” events are accompanied only by intermittent tremors. The short-term episodes may last several days, and the long-term events last several years (Hirose and Obara, 2005, 2006). Whatever processes

govern these events, they appear coupled, because the temporal behavior of the short-term events changes when the long-term episodes commence (e.g., a 6 mo periodicity in the former ceases) and vice versa (e.g., the long-term slip accelerates when short-term activity occurs). Low-frequency and very low-frequency seismic signals are much more abundant in southwest Japan (Katsumata and Kamaya, 2003; Shelly *et al.*, 2006; Obara and Ito, 2005; Ito *et al.*, 2007, 2009) relative to Cascadia. The degree to which these regional differences reflect disparities in observational capabilities rather than natural processes remains to be determined, however, noting that Japan benefits from a greater variety and density of instrumentation.

OUTSTANDING QUESTIONS

Debate continues about whether slip and tremor are confined to the plate interface in Cascadia or are distributed through the overlying plate. Inversion of GPS data has limited resolution and is insensitive to the location of this interface to within 5–10 km (Kao *et al.*, 2009). The accuracy and data-selection criteria of current tremor location procedures still do not permit us to distinguish between models in which tremor sources represent slip only on the Cascadia plate interface (LaRocca *et al.*, 2009) or deformation distributed over a broader zone that extends several tens of kilometers into the overlying forearc crust (Kao *et al.*, 2005, 2009). Rather than providing a constraint on the relative locations of the tremor and interface, Audet *et al.* (2010) used the tremor to select between two otherwise equally plausible plate-interface models, derived using new data and seismic imaging methods. They favored the model that results in a tremor source depth distribution that is more narrow and centered around the plate interface. Thus, in addition to uncertainties in location and mechanism of the tremor sources, we are also left with more fundamental questions about location and nature of the main plate boundary, whether it is a narrow or broad shear zone, and its relationship to the locked fault surface that we expect to host a great earthquake.

One of the most useful aspects of slow-slip phenomena is that they can tell us about the seismogenic potential of the Cascadia plate interface. The most recent estimates find that the strain associated with slow aseismic slip events is approximately equal to the strain accumulated between them, such that there currently appears to be little if any strain buildup downdip of the seismogenic zone (Aguier *et al.*, 2009). Comparisons of source locations for tremor and quasi-static slow slip in Cascadia relative to areas of the plate interface that are experiencing

long-term strain accumulation suggest that the zone of slow-slip phenomena is displaced tens of kilometers inland of the previously inferred main zone of locking (Boyarko and Brudzinski, 2010; Dragert *et al.*, 2008; Wech *et al.*, 2009; Chapman and Melbourne, 2009) (Fig. 10).

Tantalizing evidence of correlations and, thus, possible causal connections, exists between slow slip and tremor and earthquakes in the overlying North American plate. This evidence comes from Vancouver Island in the form of anticorrelation between local small-magnitude seismicity, large earthquakes, and tremor (Kao *et al.*, 2009) (Fig. 10A). Local seismicity and tremor appear to be similarly anticorrelated in southern Cascadia as well (Boyarko and Brudzinski, 2010) (Fig. 10B). This anticorrelation may mean that tremors and earthquakes represent different stress-relieving responses because of different mechanical/rheological conditions. Moreover, these results imply that where tremor is present, stress is not accumulating toward earthquakes, and where it is absent, earthquakes in the relatively recent past have reduced stresses and/or fluid pore pressures, thereby inhibiting tremor activity.

Fortunately, unlike many geologic processes, the time scales over which ETS events vary make rapid progress in understanding feasible, although some questions will take decades of sustained study to answer. For example, the arrival of the most recent northern Cascadia ETS event in March 2009 surprised some, who had expected a late May to June arrival, based on a recurrence interval estimated from the last 12 most well-constrained events and the occurrence time of the penultimate event. However, the timing of the three ETS events in northern Cascadia from 2007 on is consistent with an estimated mean recurrence interval and standard deviation based on 27 yr of available tremor observations. We are still in the initial, exciting discovery phase of these slow-slip phenomena in Cascadia, anticipating that as we better define their spatial and temporal characteristics, a deeper understanding of the underlying causative processes will follow.

CONCLUSIONS

The originally serendipitous discovery of slow-slip phenomena resulted from investments in seismic and geodetic instrumentation in Cascadia, made initially to facilitate monitoring of geologic hazards and studies of Earth structure. Several new initiatives just under way or planned promise more discovery and new questions.

Two of the newest, already in hand, data sets that are just now showing great promise come from Plate Boundary Observatory borehole

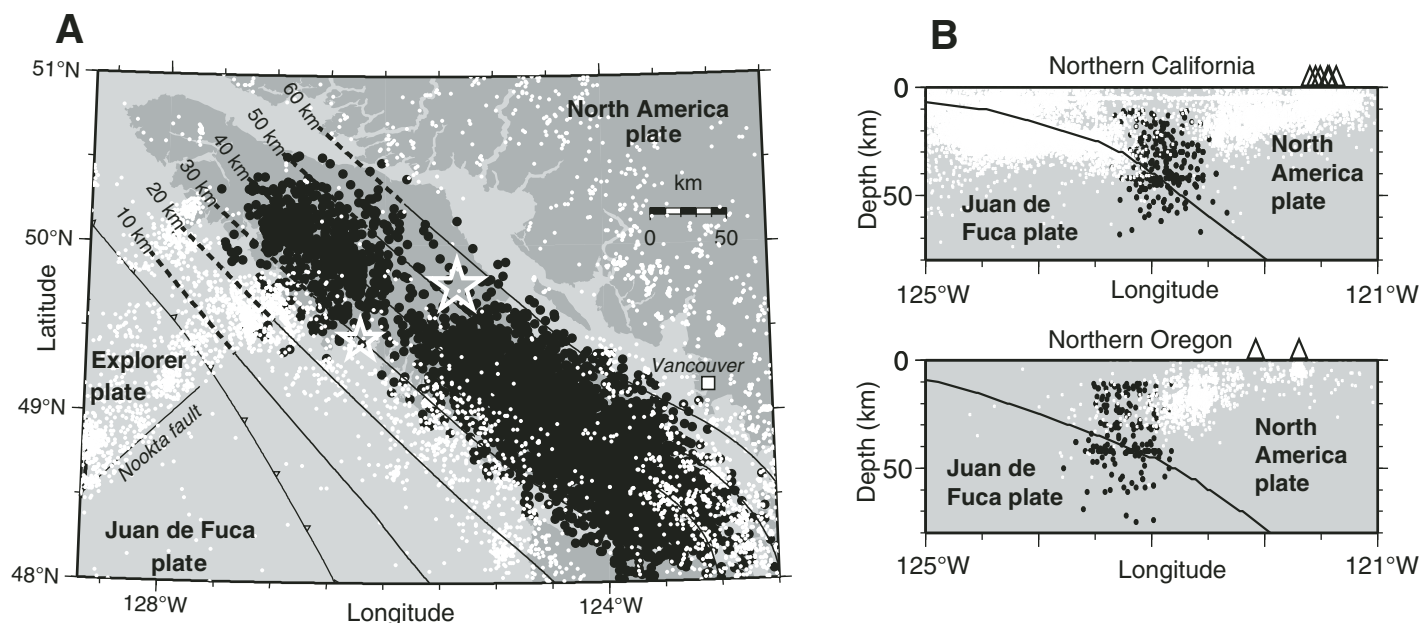


Figure 10. Anticorrelated tremor (black circles) and instrumentally recorded small and moderate earthquakes (white circles). (A) A gap in the density of episodic tremor and slip (ETS) tremors in northern Cascadia between 1997 and 2007 exists in the source area of the two largest crustal earthquakes (stars) in the past 150 yr (M7.3 in 1946, M6.9 in 1918). The northern limit of tremor occurrence is real, whereas the southern extent reflects the use of only Canadian seismic stations to derive this tremor catalogue (figure from H. Kao, S.-J. Shan, H. Dragert, and G. Rogers, Northern Cascadia episodic tremor and slip: A decade of tremor observations from 1997 to 2007, *The Journal of Geophysical Research*, v. 114, B00A12, doi: 10.1029/2008JB006046, 2009, Copyright 2009 American Geophysical Union. Modified by permission of the American Geophysical Union). Instrumentally recorded seismicity during 1990–2007 also appears to be anticorrelated with the tremor distribution. The plate interface (McCrorry et al., 2006) is shown as depth contours, dashed where uncertain. (B) Cross sections through several places along the Cascadia margin show a gap or termination in local seismicity (white) that also coincides with the source area of tremor (black) (figure from D.C. Boyarko, and M.R. Brudzinski, Spatial and temporal patterns of nonvolcanic tremor along the southern Cascadia subduction zone, *The Journal of Geophysical Research*, v. 115 [in press], doi: 10.1029/2008JB006064, 2009, Copyright 2009 American Geophysical Union. Modified by permission of the American Geophysical Union). Seismicity locations are from the ANSS catalog, and tremor locations are based on S-wave envelopes that result in large depth uncertainties. Triangles indicate Cascade volcanoes for reference. Black curve shows the plate interface (McCrorry et al., 2006).

strainmeters and from densely spaced, small-aperture seismic arrays. Most knowledge of aseismic slow slip in Cascadia has come from studies of GPS data, which leave unanswered questions about the spatial and temporal relationships between seismic and aseismic phenomena. Such questions ask whether aseismic deformation always accompanies seismic signals and which phenomenon leads temporally (Aguilar et al., 2009; Wech et al., 2009); answers to these are required to discriminate between cause and effect. Preliminary analyses of strainmeter data indicate that the challenges of calibration and separation of signal from noise can be overcome, and answers will be forthcoming (McCausland et al., 2008; Roeloffs, 2010). To date, data from only a single prototype seismic array experiment have been analyzed, and, as illustrated herein, results demonstrate that seismic slow-slip sources can be tracked at scales of kilometers and minutes. Multiple arrays have just been deployed in northern Cascadia, and

these will ultimately tell us whether the complex migration patterns observed so far repeat locally but differ among locales, and if so, how they vary.

Slow-slip phenomena clearly are integral pieces of the subduction process in Cascadia. Unfortunately, a large percentage of this process occurs offshore, where observations are extremely difficult to make. Two new programs address this shortcoming, providing observations from the seafloor. In late 2009, the Ocean Networks Canada consortium started recording seismic data from the permanent ocean-bottom Neptune network in northernmost Cascadia (see <http://neptunecanada.ca>). In the United States, a new “amphibious” program has just begun, in which several tens of ocean-bottom instrument packages containing seismometers and pressure transducers (for measuring aseismic deformation) will be deployed for several years (see <http://www.nsf-margins.org/Cascadia/09meeting/>). This deployment will be

accompanied by tens of land-based seismic stations augmenting the permanent networks. For the first time, we will have a synoptic direct view of the entire subduction zone. These new observations should be particularly important for testing models of the locked zone (which currently is thought to be mostly offshore) and the transition zones where slow-slip phenomena occur. The amphibious experiment is being accompanied by an upgrade of 232 GPS sites in Cascadia to one sample/s from the current standard daily sampling. These new GPS data will be invaluable for constraining the deformation associated with a very large earthquake in near real-time and facilitate more comprehensive monitoring of rupture processes and volcanic eruptions.

We speculate about what could be learned if new resources became available. Accelerometers have been successfully used throughout Japan to measure transient changes in tilts and thus to infer and characterize quasi-static

slow-slip events (Obara et al., 2004; Obara and Hirose, 2006; Ito et al., 2007). Accelerometers are conventionally used for recording large-amplitude seismic signals, and are more easily deployed than geodetic instruments. Questions remain about whether the paucity of some seismic slow-slip signals (e.g., low-frequency and very low-frequency earthquakes) in Cascadia relative to Japan reflects real differences in fault behaviors. Instead, Japan may benefit from better detection capabilities resulting from having more than six times the number of seismic stations, most of which are buried in boreholes 100 m or more deep where noise levels can be much reduced. Perhaps more of these same seismic signals observed in Japan would be seen in Cascadia with the addition of even a few borehole seismic stations.

We emphasize a few of the societal implications of the current information about slow-slip phenomena in Cascadia, particularly with respect to future, potentially damaging earthquakes. Ground-motion modeling studies show that the intensity of shaking is strongly affected by the geometry and dimensions of the plate interface that is strongly coupled (locked) and thus likely to fail seismically. If the downdip boundary of the locked zone coincides with the updip edge of the region that slips in slow-slip episodes, it may extend tens of kilometers further inland in some places (Chapman and Melbourne, 2009) relative to prior estimates in which it was largely offshore (Hyndman and Wang, 1995; Fluck et al., 1997). Olsen et al. (2008) and Atkinson and Macias (2009) simulated ground motions for plausible M7.5–9.0 earthquakes on the plate interface assuming these earlier models of the locked zone. Atkinson and Macias (2009) showed that increasing the width of a M9.0 rupture by 167% (and decreasing the length to preserve the area and magnitude) would increase the ground motions in the urban centers of Puget Sound and Vancouver Island by more than an order of magnitude, primarily because a greater fraction of the radiated energy travels a shorter distance. Moreover, Chapman and Melbourne (2009) suggested that new coupling models imply a larger potential rupture area and a M9.2 Cascadia earthquake. The impact of slow slip on the timing and probability of occurrence of large earthquakes is more difficult to evaluate. Indeed, during and immediately following transient slow slip, a case can be made for an increase in the probability of a damaging earthquake due to the associated loading on the locked portion of the plate interface (Rubinstein et al., 2010). Numerous ETS events have occurred, and none has been followed by damaging earthquakes, but aseismic slow-slip has triggered small earthquakes in Hawaii (Segall et al., 2006), New

Zealand (Delahaye et al., 2009), Japan (Yoshida et al., 2006), and Mexico (Liu et al., 2007).

The advances needed to address the aforementioned questions will come from improvements in modeling and simply from the passage of time. Ever-more-powerful computers and theoretical developments will improve our understanding, particularly in the extension of frictional models that now predict observed quasi-static slow-slip behaviors but have not addressed the generation of seismic slow-slip signals (Liu and Rice, 2005, 2007, 2009; Rubin, 2008; Segall and Bradley, 2009). Constraints for these models will require a broader spectrum of observations than seismic and geodetic measurements. For example, laboratory measurements of frictional properties of rocks thought to exist along the plate interface are needed and are increasingly more feasible at the appropriate temperature and pressure conditions. A prominent role of fluids, liberated by dehydration reactions, has now become integral to discussions of slow-slip phenomena (see Peacock, 2009, and references therein). Input from petrologists and geochemists, among others, will be invaluable to illuminate these roles. Lastly, quantification of the variability in the occurrence and characteristics of slow-slip phenomena, and of earthquakes, requires sustained monitoring activities for years to come. Fortunately, unlike most geological processes, at least the slow-slip phenomena in Cascadia recur on short enough time scales that the simple passage of time is guaranteed to yield new insights.

ACKNOWLEDGMENTS

This paper and the Cascadia 2007 and Beyond Working Group grew out of discussions at an informal (participant and U.S. Geological Survey sponsored) workshop held 2–3 March 2009. The Working Group members would like to thank David Shelly, Roland Bürgmann, Roy Hyndman, Stephen Johnston, Nancy Riggs, and especially Brendan Murphy for their valuable reviews of this paper as it developed.

REFERENCES CITED

- Abers, G.A., MacKenzie, L.S., Rondenay, S., Zhang, Z., Wech, A.G., and Creager, K.C., 2009, Imaging the source region of Cascadia tremor and intermediate-depth earthquakes: *Geology*, v. 37, p. 1119–1122, doi: 10.1130/G30143A.1.
- Aguilar, A.C., Melbourne, T.I., and Scrivner, C.W., 2009, Moment release rate of Cascadia tremor constrained by GPS: *Journal of Geophysical Research*, v. 114, B00A05, 11 p., doi: 10.1029/2008JB005909.
- Aki, K., 1966, Generation and propagation of G waves from the Niigata earthquake of June 16, 1964: 2. Estimation of earthquake moment, released energy, and stress-strain drop from G wave spectrum: *Bulletin of the Earthquake Research Institute, Tokyo*, v. 44, p. 73–88.
- Atkinson, G.M., and Macias, M., 2009, Predicted ground motions for great interface earthquakes in the Cascadia subduction zone: *Bulletin of the Seismological Society of America*, v. 99, p. 1552–1578, doi: 10.1785/0120080147.

- Audet, P., Bostock, M.G., Christensen, N.I., and Peacock, S.M., 2009, Seismic evidence for overpressured subducted oceanic crust and megathrust fault sealing: *Nature*, v. 457, p. 76–78, doi: 10.1038/nature07650.
- Audet, P., Bostock, M.G., Boyarko, D.C., Brudzinski, M.R., and Allen, R.M., 2010, Slab morphology in the Cascadia forearc and its relation to episodic tremor and slip: *Journal of Geophysical Research*, 53 p., doi: 10.1029/2008JB006053 (in press).
- Beeler, N., 2009, Estimated effects of episodic slow-slip on the occurrence and probability of great earthquakes in Cascadia using lab-based failure relations: *Seismological Research Letters*, v. 80, no. 2, p. 295.
- Beroza, G.C., and Ide, S., 2009, Deep tremors and slow quakes: *Science*, v. 324, p. 1025–1026, doi: 10.1126/science.1171231.
- Boyarko, D., and Brudzinski, M., 2010, Spatial and temporal patterns of nonvolcanic tremor along the southern Cascadia subduction zone: *Journal of Geophysical Research*, doi: 10.1029/2008JB006064 (in press).
- Brodsky, E.E., and Mori, J., 2007, Creep events slip less than ordinary earthquakes: *Geophysical Research Letters*, v. 34, p. L16309, doi: 10.1029/2007GL030917.
- Brown, J.R., Beroza, G.C., Ide, S., Ohta, K., Shelly, D.R., Schwartz, S.Y., Rabbel, W., Thorwart, M., and Kao, H., 2009, Deep low-frequency earthquakes in tremor localize to the plate interface in multiple subduction zones: *Geophysical Research Letters*, v. 36, p. L19306, doi: 10.1029/2009GL040027.
- Brudzinski, M.R., and Allen, R.M., 2007, Segmentation in episodic tremor and slip all along Cascadia: *Geology*, v. 35, p. 907–910, doi: 10.1130/G23740A.1.
- Burgette, R.J., Weldon, R.J., II, and Schmidt, D.A., 2009, Interseismic uplift rates for western Oregon and along-strike variation in locking on the Cascadia subduction zone: *Journal of Geophysical Research*, v. 114, p. B01408, doi: 10.1029/2008JB005679.
- Chapman, J.S., and Melbourne, T.I., 2009, Future Cascadia megathrust rupture delineated by episodic tremor and slip: *Geophysical Research Letters*, v. 36, L22301, 5 p., doi: 10.1029/2009GL040465.
- Delahaye, E.J., Townend, J., Reyners, M.E., and Rogers, G., 2009, Microseismicity but no tremor accompanying slow-slip in the Hikurangi subduction zone, New Zealand: *Earth and Planetary Science Letters*, v. 277, p. 21–28, doi: 10.1016/j.epsl.2008.09.038.
- Dragert, H., 2007, Mediating plate convergence: *Science*, v. 315, p. 471–472, doi: 10.1126/science.1137171.
- Dragert, H., Wang, K., and James, T.S., 2001, A silent slip event on the deeper Cascadia subduction interface: *Science*, v. 292, p. 1525–1528, doi: 10.1126/science.1060152.
- Dragert, H., Wang, K., and Kao, H., 2008, Recent findings on the nature of episodic tremor and slip along the northern Cascadia margin: *Eos Transactions of the American Geophysical Union*, v. 89, p. 53, Fall Meeting Supplement, Abstract U33A-0025.
- Fluck, P., Hyndman, R.D., and Wang, K., 1997, Three-dimensional dislocation model for great earthquakes of the Cascadia subduction zone: *Journal of Geophysical Research*, v. 102, no. B9, p. 20,539–20,550, doi: 10.1029/97JB01642.
- Forsyth, D.W., Lay, T., Aster, R.C., and Romanowicz, B., 2009, Grand challenges for seismology: *Eos (Transactions, American Geophysical Union)*, v. 90, p. 361–362, doi: 10.1029/2009EO410001.
- Ghosh, A., Vidale, J.E., Sweet, J.R., Creager, K.C., and Wech, A.G., 2009, Tremor patches in Cascadia revealed by seismic array analysis: *Geophysical Research Letters*, v. 36, p. L17316, doi: 10.1029/2009GL039080.
- Ghosh, A., Vidale, J.E., Sweet, J.R., Creager, K., Wech, A., and Houston, H., 2010, Tremor bands sweep Cascadia: *Geophysical Research Letters*, in press, doi: 10.1029/2009GL042301, 13 p.
- Hawthorne, J.C., and Rubin, A.M., 2009, Is slow-slip in Cascadia tidally modulated?: *Eos (Transactions, American Geophysical Union)*, v. 90, no. 52, Fall Meeting supplement, abstract T21F-05.
- Hill, D.P., 2010, Surface wave potential for triggering tectonic (non-volcanic) tremor: *Bulletin of the Seismological Society of America* (in press).

- Hiramatsu, Y., Watanabe, T., and Obara, K., 2008, Deep low-frequency tremors as a proxy for slip monitoring at plate interface: *Geophysical Research Letters*, v. 35, p. L13304, doi: 10.1029/2008GL034342.
- Hirose, H., and Obara, K., 2005, Repeating short- and long-term slow-slip events with deep tremor activity around the Bungo channel region, southwest Japan: *Earth, Planets, and Space*, v. 57, p. 961–972.
- Hirose, H., and Obara, K., 2006, Short-term slow-slip and correlated tremor episodes in the Tokai region, central Japan: *Geophysical Research Letters*, v. 33, p. L17311, doi: 10.1029/2006GL026579.
- Holtkamp, S., and Brudzinski, M.R., 2010, Determination of slow-slip episodes and strain accumulation along the Cascadia margin: *Journal of Geophysical Research*, 59 p., doi: 10.1029/2008JB006058 (in press).
- Houston, H., 2007, Scaling of the tremor source: *Eos (Transactions, American Geophysical Union)*, v. 88, Fall Meeting supplement, abstract T13F-04.
- Husen, S., and Kissling, E., 2001, Postseismic fluid flow after the large subduction earthquake of Antofagasta, Chile: *Geology*, v. 29, p. 847–850, doi: 10.1130/0091-7613(2001)029<0847:PPFATL>2.0.CO;2.
- Hyndman, R., and Wang, K., 1995, The rupture zone of Cascadia great earthquakes from current deformation and the thermal regime: *Journal of Geophysical Research*, v. 100, p. 22,133–22,154, doi: 10.1029/95JB01970.
- Ide, S., Beroza, G.C., Shelly, D.R., and Uchide, T., 2007a, A scaling law for slow earthquakes: *Nature*, v. 447, p. 76–79, doi: 10.1038/nature05780.
- Ide, S., Shelly, D.R., and Beroza, G.C., 2007b, Mechanism of deep low frequency earthquakes: Further evidence that deep non-volcanic tremor is generated by shear slip on the plate interface: *Geophysical Research Letters*, v. 34, p. L03308, doi: 10.1029/2006GL028890.
- Ide, S., Imanishi, K., Yoshida, Y., Beroza, G.C., and Shelly, D.R., 2008, Bridging the gap between seismically and geodetically detected slow earthquakes: *Geophysical Research Letters*, v. 35, L10305, 6 p.
- Ito, Y., Obara, K., Shiomi, K., Sekine, S., and Hirose, H., 2007, Slow earthquakes coincident with episodic tremors and slow-slip events: *Science*, v. 315, p. 503–506, doi: 10.1126/science.1134454.
- Ito, Y., Asano, Y., and Obara, K., 2009, Very low-frequency earthquakes indicate a transpressional stress regime in the Nankai accretionary prism: *Geophysical Research Letters*, v. 36, p. L20309, doi: 10.1029/2009GL039332.
- Kanamori, H., and Brodsky, E.E., 2004, The physics of earthquakes: Reports on Progress in Physics, v. 67, p. 1429–1496, doi: 10.1088/0034-4885/67/8/R03.
- Kao, H., and Shan, S.-J., 2004, The source-scanning algorithm: Mapping the distribution of sources in time and space: *Geophysical Journal International*, v. 157, p. 589–594, doi: 10.1111/j.1365-246X.2004.02276.x.
- Kao, H., Shan, S., Dragert, H., Rogers, G., Cassidy, J.F., and Ramachandran, K., 2005, A wide depth distribution of seismic tremors along the northern Cascadia margin: *Nature*, v. 436, p. 841–844, doi: 10.1038/nature03903.
- Kao, H., Shan, S.-J., Dragert, S.-J., Rogers, G., Cassidy, J.F., Wang, K., James, T., and Ramachandran, K., 2006, Spatial-temporal patterns of seismic tremors in northern Cascadia: *Journal of Geophysical Research*, v. 111, B03309, doi: 10.1029/2005JB003727.
- Kao, H., Shan, S.-J., Rogers, G., and Dragert, H., 2007, Migration characteristics of seismic tremors in the northern Cascadia margin: *Geophysical Research Letters*, v. 34, p. L03304, doi: 10.1029/2006GL028430.
- Kao, H., Shan, S.-J., Dragert, H., and Rogers, G., 2009, Northern Cascadia episodic tremor and slip: A decade of tremor observations from 1997 to 2007: *Journal of Geophysical Research*, v. 114, doi: 10.1029/2008JB006046.
- Katsumata, A., and Kamaya, N., 2003, Low-frequency continuous tremor around the Moho discontinuity away from volcanoes in the southwest Japan: *Geophysical Research Letters*, v. 30, no. 1, p. 1020, doi: 10.1029/2002GL015981.
- Kodaira, S., Iidaka, T., Kato, A., Park, J.-O., Iwasaki, T., and Kaneda, Y., 2004, High pore fluid pressure may cause silent slip in the Nankai Trough: *Science*, v. 304, p. 1295–1298, doi: 10.1126/science.1096535.
- Kummer, T., and Spinelli, G., 2009, Thermal effects of fluid circulation in the basement aquifer of subducting ocean crust: *Journal of Geophysical Research*, v. 114, p. B03104, doi: 10.1029/2008JB006197.
- Lambert, A., Kao, H., Rogers, G., and Courtier, N., 2009, Correlation of tremor activity with tidal stress in the northern Cascadia subduction zone: *Journal of Geophysical Research*, v. 114, p. B00A08, doi: 10.1029/2008JB006038.
- Langbein, J., 2004, Noise in two-color electronic distance meter measure—837 measurements revisited: *Journal of Geophysical Research*, v. 109, p. B04406, doi: 10.1029/2003JB002819.
- Langbein, J., Murray, J.R., and Snyder, H.A., 2006, Coseismic and initial postseismic deformation from the 2004 Parkfield, California, earthquake, observed by Global Positioning System, electronic distance meter, creepmeters, and borehole strainmeters: *Bulletin of the Seismological Society of America*, v. 96, p. S304–S320, doi: 10.1785/0120050823.
- LaRocca, M., Creager, K.C., Galluzzo, D., Malone, S., Vidale, J.E., Sweet, J.R., and Wech, A.G., 2009, Cascadia tremor located near the plate interface constrained by S minus P wave times: *Science*, v. 323, p. 620–623, doi: 10.1126/science.1167112.
- Lay, T., ed., 2009, *Seismological Grand Challenges in Understanding Earth's Dynamic Systems: Report to the National Science Foundation, IRIS Consortium*, 76 p.
- Liu, Y., and Rice, J.R., 2005, Aseismic slip transients emerge spontaneously in three-dimensional rate and state modeling of subduction earthquake sequences: *Journal of Geophysical Research*, v. 110, p. B08307, doi: 10.1029/2004JB003424.
- Liu, Y., and Rice, J.R., 2007, Spontaneous and triggered aseismic deformation transients in a subduction fault model: *Journal of Geophysical Research*, v. 112, B09404, 23 p.
- Liu, Y., and Rice, J.R., 2009, Slow-slip predictions based on granite and gabbro friction data compared to GPS measurements in northern Cascadia: *Journal of Geophysical Research*, v. 114, B09407, doi: 10.1029/2008JB006142, 19 p.
- Liu, Y., Rice, J.R., and Larson, K.M., 2007, Seismicity variations associated with aseismic transients in Guerrero, Mexico, 1995–2006: *Earth and Planetary Science Letters*, v. 262, p. 493–504, doi: 10.1016/j.epsl.2007.08.018.
- Mazzotti, S., and Adams, J., 2004, Variability of near-term probability for the next great earthquake on the Cascadia subduction zone: *Bulletin of the Seismological Society of America*, v. 94, p. 1954–1959, doi: 10.1785/012004032.
- McCaffrey, R., Qamar, A., King, R.W., Wells, R.W., Khazaradze, G., and Williams, C., 2007, Deformation in the Pacific Northwest: *Geophysical Journal International*, v. 169, p. 1315–1340, doi: 10.1111/j.1365-246X.2007.03371.x.
- McCausland, W.A., Roeloffs, E., and Silver, P., 2008, New insights into Cascadia slow-slip events using Plate Boundary Observatory borehole strainmeters: *Eos (Transactions, American Geophysical Union)*, v. 89, no. 53, Fall Meeting supplement, abstract G21B-0691.
- McCrory, P.A., Blair, J.L., Oppenheimer, D.H., and Walter, S.R., 2006, Depth to the Juan de Fuca slab beneath the Cascadia subduction margin: A 3-D model for sorting earthquakes: *U.S. Geological Survey Data Series DS-91*, 1 CD-ROM, version 1.2.
- Miller, M.M., Melbourne, T., Johnson, D.J., and Sumner, W.Q., 2002, Periodic slow earthquakes from the Cascadia subduction zone: *Science*, v. 295, p. 2423, doi: 10.1126/science.1071193.
- Miyazawa, M., and Brodsky, E.E., 2008, Deep low-frequency tremor that correlates with passing surface waves: *Journal of Geophysical Research*, v. 113, p. B01307, doi: 10.1029/2006JB004890.
- Miyazawa, M., and Mori, J., 2005, Detection of triggered deep low-frequency events from the 2003 Tokachi-oki earthquake: *Geophysical Research Letters*, v. 32, p. L10307, doi: 10.1029/2005GL022539.
- Miyazawa, M., Brodsky, E.E., and Mori, J., 2008, Learning from dynamic triggering of low-frequency tremor in subduction zones: *Earth, Planets, and Space*, v. 60, p. e17–e20.
- Nakata, R., Suda, N., and Tsuruoka, H., 2008, Non-volcanic tremor resulting from the combined effect of Earth tides and slow-slip events: *Nature Geoscience*, v. 1, p. 676–678, doi: 10.1038/ngeo288.
- Obara, K., 2002, Nonvolcanic deep tremor associated with subduction in southwest Japan: *Science*, v. 296, p. 1679–1681, doi: 10.1126/science.1070378.
- Obara, K., and Ito, Y., 2005, Very low frequency earthquakes excited by the 2004 off the Kii peninsula earthquakes: A dynamic deformation process in the large accretionary prism: *Earth, Planets, and Space*, v. 57, p. 321–326.
- Obara, K., and Hirose, H., 2006, Non-volcanic deep low-frequency tremors accompanying slow-slips in southwest Japan subduction zone: *Tectonophysics*, v. 417, p. 33–51, doi: 10.1016/j.tecto.2005.04.013.
- Obara, K., Hirose, H., Yamamizu, F., and Kasahara, K., 2004, Episodic slow-slip events accompanied by non-volcanic tremors in southwest Japan subduction zone: *Geophysical Research Letters*, v. 31, p. L23602, doi: 10.1029/2004GL020848.
- Obara, K., Tanaka, S., and Maeda, T., 2009, Reevaluation of nonvolcanic tremor activity based on the hybrid method: *Eos (Transactions, American Geophysical Union)*, v. 90, no. 52, Fall Meeting supplement, abstract T11C-1835.
- Olsen, K., Stephenson, W.J., and Geisselmeyer, A., 2008, 3D crustal structure and long-period ground motions from a M9.0 megathrust earthquake in the Pacific Northwest region: *Journal of Seismology*, v. 12, p. 145–159, doi: 10.1007/s10950-007-9082-y.
- Peacock, S.M., 2003, Thermal structure and metamorphic evolution of subducting slabs, *in* Eiler, J.M., ed., *Inside the Subduction Factory: American Geophysical Union Geophysical Monograph*, v. 138, p. 7–22.
- Peacock, S.M., 2009, Thermal and metamorphic environment of subduction zone episodic tremor and slip: *Journal of Geophysical Research*, v. 114, B00A07, doi: 10.1029/2008JB005978, 9 p.
- Peterson, C.L., and Christensen, D.H., 2009, Possible relationship between nonvolcanic tremor and the 1998–2001 slow slip event, south central Alaska: *Journal of Geophysical Research*, v. 114, B06302, doi: 10.1029/2008JB006096.
- Roeloffs, E., 2010, Tidal calibration of PBO borehole strainmeters: The role of vertical and shear coupling: *Journal of Geophysical Research*, doi: 10.1029/2009JB006407 (in press).
- Rogers, G., and Dragert, H., 2003, Episodic tremor and slip on the Cascadia subduction zone: The chatter of silent slip: *Science*, v. 300, p. 1942–1943.
- Roland, E., and McGuire, J.J., 2009, Earthquake swarms on transform faults: *Geophysical Journal International*, v. 178, p. 1677–1690, doi: 10.1111/j.1365-246X.2009.04214.x.
- Rubin, A.M., 2008, Episodic slow-slip events and rate-and-state friction: *Journal of Geophysical Research*, v. 113, B11414, 18 p.
- Rubinstein, J.L., Vidale, J.E., Gombert, J., Bodin, P., Creager, K.C., and Malone, S.D., 2007, Non-volcanic tremor driven by large transient shear stresses: *Nature*, v. 448, p. 579–582, doi: 10.1038/nature06017.
- Rubinstein, J.L., LaRocca, M., Vidale, J.E., Creager, K.C., and Wech, A.G., 2008, Tidal modulation of non-volcanic tremor: *Science*, v. 319, p. 186–189, doi: 10.1126/science.1150558.
- Rubinstein, J.L., Gombert, J., Vidale, J.E., Wech, A.G., Kao, H., Creager, K.C., and Rogers, G., 2009, Seismic wave triggering of nonvolcanic tremor, episodic tremor and slip, and earthquakes on Vancouver Island: *Journal of Geophysical Research*, v. 114, p. B00A01, doi: 10.1029/2008JB005875.
- Rubinstein, J.L., Shelly, D.R., and Ellsworth, W.L., 2010, Non-volcanic tremor: A window into the roots of fault zones: *New Frontiers in Integrated Solid Earth Sciences*, p. 287–314, doi: 10.1007/978-90-481-2737-5.
- Samuelson, J., Ellsworth, D., and Marone, C., 2009, Shear-induced dilatancy of fluid-saturated faults: Experiment and theory: *Journal of Geophysical Research*, v. 114, p. B12404, doi: 10.1029/2008JB006273.
- Savage, J.C., Lisowski, M., and Prescott, W.H., 1991, Strain accumulation in western Washington: *Journal of Geophysical Research*, v. 96, p. 14,493–14,507, doi: 10.1029/91JB01274.

- Schmidt, D.A., and Gao, H., 2010, Source parameters and time dependent slip distributions of slow-slip events on the Cascadia subduction zone from 1998–2008: *Journal of Geophysical Research* (in press), doi: 10.1029/2008JB006045.
- Schwartz, S.Y., and Rokosky, J.M., 2007, Slow-slip events and seismic tremor at circum-Pacific subduction zones: *Reviews of Geophysics*, v. 45, p. RG3004, doi: 10.1029/2006RG000208.
- Segall, P., and Bradley, A.M., 2009, Numerical models of slow-slip and dynamic rupture including dilatant stabilization and thermal pressurization: *Eos (Transactions, American Geophysical Union)*, v. 90, no. 52, Fall Meeting supplement, abstract T22B–08.
- Segall, P., Desmarais, E., Shelly, D., Miklius, A., and Cervelli, P., 2006, Earthquakes triggered by silent slip events on Kilauea volcano, Hawaii: *Nature*, v. 442, p. 71–74.
- Shelly, D.R., 2010, Migrating tremors illuminate complex deformation beneath the seismogenic San Andreas fault: *Nature*, v. 463, p. 648–652, doi: 10.1038/nature08755.
- Shelly, D.R., and Beroza, G.C., 2007, Complex evolution of transient slip derived from precise tremor locations in western Shikoku, Japan: *Geochemistry, Geophysics, Geosystems*, v. 8, 18 p., doi: 10.1029/2007GC001640.
- Shelly, D.R., Beroza, G.C., and Ide, S., 2006, Low-frequency earthquakes in Shikoku, Japan, and their relationship to episodic tremor and slip: *Nature*, v. 442, p. 188–191, doi: 10.1038/nature04931.
- Shelly, D.R., Ellsworth, W.L., Ryberg, T., Haberland, C., Fuis, G.S., Murphy, J., Nadeau, R.M., and Bürgmann, R., 2009, Precise location of San Andreas fault tremors near Cholame, California, using seismometer clusters: Slip on the deep extension of the fault?: *Geophysical Research Letters*, v. 36, p. L01303, doi: 10.1029/2008GL036367.
- Sibson, R.H., 1990, Rupture nucleation on unfavorably oriented faults: *Bulletin of the Seismological Society of America*, v. 80, p. 1580–1604.
- Smith, E.F., and Gomberg, J., 2009, A search in strainmeter data for slow-slip associated with triggered and ambient tremor near Parkfield, California: *Journal of Geophysical Research*, v. 114, p. B00A14, doi: 10.1029/2008JB006040.
- Song, T.A., Helmlinger, D.V., Brudzinski, M.R., Clayton, R.W., Davis, P., Pérez-Campos, X., and Singh, S.K., 2009, Subducting slab ultra-slow velocity layer coincident with silent earthquakes in southern Mexico: *Nature*, v. 324, p. 502–506.
- Sweet, J.R., Creager, K.C., Vidale, J.E., Gosh, A., Nichols, M.L., and Pratt, T.L., 2008, Low-frequency earthquakes in Cascadia using a Texan array: *Eos Transactions of the American Geophysical Union*, v. 89, no. 53, Fall Meeting Supplement, Abstract U33A-0024.
- Szeliga, W., Melbourne, T., Santillan, M., and Miller, M., 2008, GPS constraints on 34 slow-slip events within the Cascadia subduction zone, 1997–2005: *Journal of Geophysical Research*, v. 113, p. B04404, doi: 10.1029/2007JB004948.
- Trèhu, A.M., Braunmiller, J., and Nabelek, J.L., 2008, Probable low-angle thrust earthquakes on the Juan de Fuca–North America plate boundary: *Geology*, v. 36, p. 127–130, doi: 10.1130/G24145A.1.
- Vidale, J.E., Gosh, A., Sweet, J.R., Creager, K.C., Wech, A., and Houston, H., 2009, Streaking tremor in Cascadia: *Eos (Transactions, American Geophysical Union)*, v. 90, no. 52, Fall Meeting supplement, abstract T21F–07.
- Wada, I., Wang, K., He, J., and Hyndman, R.D., 2008, Weakening of the subduction interface and its effects on surface heat flow, slab dehydration, and mantle wedge serpentinization: *Journal of Geophysical Research*, v. 113, p. B04402, doi: 10.1029/2007JB005190.
- Waldhauser, F., Ellsworth, W.L., Schaff, D.P., and Cole, A., 2004, Streaks, multiplets, and holes: High-resolution spatio-temporal behavior of Parkfield seismicity: *Geophysical Research Letters*, v. 31, p. L18608, doi: 10.1029/2004GL020649.
- Wang, K., He, J., Dragert, H., and James, T.S., 2001, Three-dimensional viscoelastic interseismic deformation model for the Cascadia subduction zone: *Earth, Planets and Space*, v. 53, p. 295–306.
- Wang, K., Dragert, H., Kao, H., and Roeloffs, E., 2008, Characterizing an “uncharacteristic” ETS event in northern Cascadia: *Geophysical Research Letters*, v. 35, p. L15303, doi: 10.1029/2008GL034415.
- Wang, Z., Zhao, D., Mishra, O.P., and Yamada, A., 2006, Structural heterogeneity and its implications for the low frequency tremors in southwest Japan: *Earth and Planetary Science Letters*, v. 251, p. 66–78, doi: 10.1016/j.epsl.2006.08.025.
- Wech, A.G., and Creager, K.C., 2007, Cascadia tremor polarization evidence for plate interface slip: *Geophysical Research Letters*, v. 34, p. L22306, doi: 10.1029/2007GL031167.
- Wech, A.G., and Creager, K.C., 2008, Automatic detection and location of Cascadia tremor: *Geophysical Research Letters*, v. 35, p. L20302, doi: 10.1029/2008GL035458.
- Wech, A.G., Creager, K.C., and Melbourne, T.I., 2009, Seismic and geodetic constraints on Cascadia slow-slip: *Journal of Geophysical Research*, v. 114, p. B10316, doi: 10.1029/2008JB006090.
- Wells, R., 1989, Mechanisms of Cenozoic tectonic rotation, Pacific Northwest convergent margin, U.S.A., in Kissel, C., and Laj, C., eds., *Paleomagnetic Rotations and Continental Deformation: NATO Advanced Study Institute, v. C254: Netherlands, Kluwer Academic Publishers*, p. 313–325.
- Wells, R., and Simpson, R., 2001, Northward migration of the Cascadia forearc in the northwestern U.S. and implications for subduction deformation: *Earth, Planets, and Space*, v. 53, p. 275–283.
- Yoshida, A., Hosono, K., Tsukakoshi, T., Kobayashi, A., Takayama, H., and Wiemer, S., 2006, Change in seismic activity in the Tokai region related to weakening and strengthening of the interplate coupling: *Tectonophysics*, v. 417, p. 17–31, doi: 10.1016/j.tecto.2005.08.031.

MANUSCRIPT RECEIVED 5 MARCH 2010
 REVISED MANUSCRIPT RECEIVED 5 MARCH 2010
 MANUSCRIPT ACCEPTED 5 MARCH 2010

Printed in the USA