Non-volcanic Tremor: A Window into the Roots of Fault Zones

Justin L. Rubinstein, David R. Shelly, and William L. Ellsworth

Abstract The recent discovery of non-volcanic tremor in Japan and the coincidence of tremor with slow-slip in Cascadia have made earth scientists reevaluate our models for the physical processes in subduction zones and on faults in general. Subduction zones have been studied very closely since the discovery of slow-slip and tremor. This has led to the discovery of a number of related phenomena including low frequency earthquakes and very low frequency earthquakes. All of these events fall into what some have called a new class of events that are governed under a different frictional regime than simple brittle failure. While this model is appealing to many, consensus as to exactly what process generates tremor has yet to be reached. Tremor and related events also provide a window into the deep roots of subduction zones, a poorly understood region that is largely devoid of seismicity. Given that such fundamental questions remain about non-volcanic tremor, slow-slip, and the region in which they occur, we expect that this will be a fruitful field for a long time to come.

Keywords Tremor · ETS · Slow earthquakes · Slow-slip

Introduction

The analysis of seismic waves provides a direct, high-resolution means for studying the internal structure of the Earth and the geological processes that are actively reshaping it today. The subject of this chapter is a newly discovered source process, non-volcanic tremor, that appears to be most common in the lower crust and upper mantle. Non-volcanic tremor has great potential for shedding light on the dynamics of regions of the Earth that have been difficult to study before.

Beginning in the late nineteenth century, the first seismologists had the challenge of understanding the origin of the feeble motions or "tremors" recorded on the early seismographs. By the turn of the twentieth century, they had established that distant earthquakes, or teleseisms, were the source of the most obvious eventful signals and that the earthquakes were distributed in well-defined belts around the globe. The study of those belts of earthquakes and the nature of the earthquake source itself continued as the principal focus of seismological research during the twentieth century and continues to be a central topic today.

Seismologists were also busy developing a comprehensive understanding of the other signals seen on the seismogram. These ranged from the impulsive signature of nearby earthquakes and explosions to the continuous microseisms generated by waves in the ocean, the ringing of the Earth itself in free oscillations, volcanic tremors created by subterranean movement of magma, sonic booms caused by meteors and aircraft, and cultural noise produced by power plants, highways, trains, etc.

It is in this context that the discovery by Kazushige Obara in 2002 of a new source of seismic waves that had been overlooked for a century took the Earth sciences by surprise. Obara (2002) found that long-duration trains of weak seismic motions were originating from a band following the contour of the subducting Philippine Sea plate beneath southwestern Japan, at depths near 30 km (Fig. 1). These "non-volcanic

J.L. Rubinstein (✉)
United States Geological Survey; Menlo Park, CA 94025, USA
e-mail: jrubinstein@usgs.gov

S. Cloetingh, J. Negendank (eds.), *New Frontiers in Integrated Solid Earth Sciences*, International Year of Planet Earth, DOI 10.1007/978-90-481-2737-5_8,

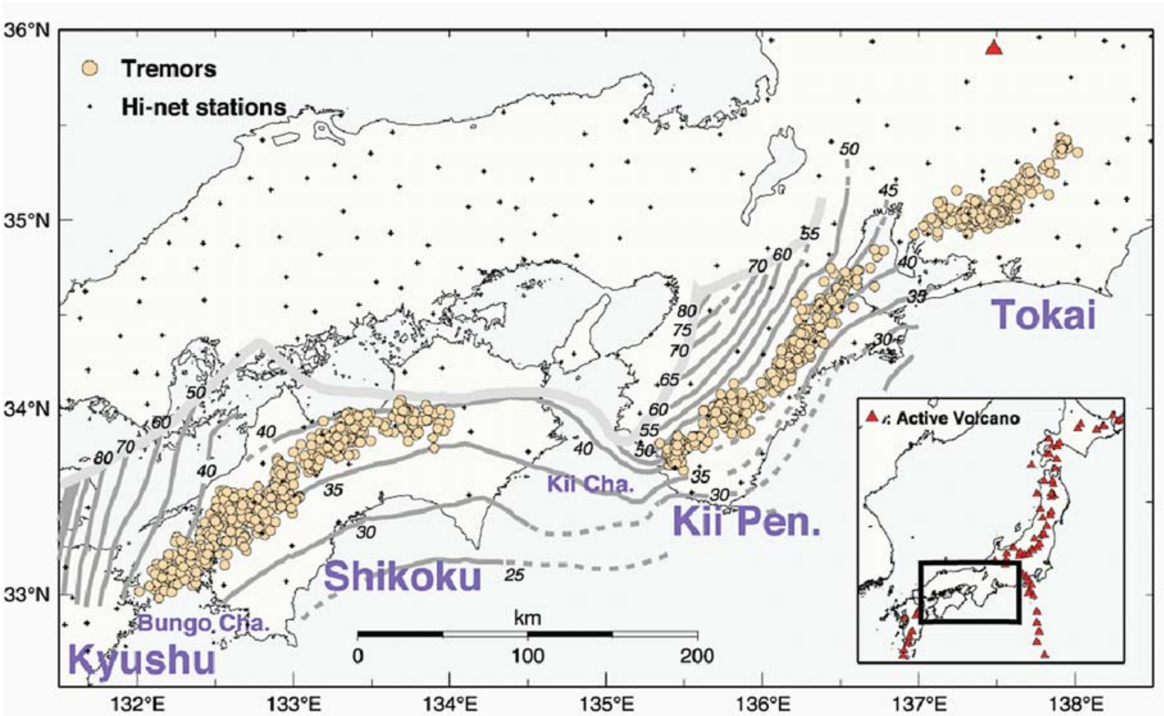

Fig. 1 Map of Southwestern Japan showing distribution of Hi-net stations (*dots*), isodepths of the subduction interface, and locations of tremor (*filled circles*) identified by Obara (2002). Active volcanoes shown by *red triangles*. Figure from Obara (2002)

tremors" were unlike anything previously known to seismologists. Something new had been discovered.

Active periods of non-volcanic tremor lasted for a week or more and migrated laterally along the subduction zone near the seismic-aseismic transition on the plate interface. Individual tremor episodes typically lasted for tens of minutes with dominant seismic frequencies of 1 to 10 Hz. The tremor propagated at the *S* wave velocity, suggesting that these waves were *S* body waves. They typically emerged slowly from the background noise and lacked any regular structure, with peaks in amplitude occurring at random during the episode and without any easily identifiable seismic body wave arrivals. Their appearance is similar to that of a distant train passing by the seismometer.

Episodic Tremor and Slip

The previous year, another important discovery had been made across the Pacific in the Cascadia subduction zone by Dragert et al. (2001). Using Global Positioning Satellite (GPS) measurements, they deduced that a slow earthquake had occurred over a period of about a week beneath Vancouver Island in British Columbia and Puget Sound in Washington State. Slow earthquakes are like regular earthquakes in that they represent slip on a fault, but unlike regular events, they rupture very slowly and take place over many hours/days such that they do not radiate strong high frequency energy like regular earthquakes. Dragert et al. (2001) inferred that event probably ruptured the plate interface downdip of the seismogenic zone. The moment magnitude of the event was 6.7. Miller et al. (2002) expanded this discovery with the remarkable observation that slow earthquakes had been occurring there for years at semi-regular intervals of 13–16 months. But were these transient slip events detected by GPS related in some way to non-volcanic tremor?

Following the report of tremor in southwest Japan by Obara (2002) and the suggestion that it might relate to the prior observation of Cascadia slow slip (Julian, 2002); Rogers and Dragert (2003) demonstrated that tremor indeed accompanied slow slip in Cascadia and that slow-slip and tremor appeared to be coming from the same location. They termed the coupled phenomenon episodic tremor and slip (ETS), showing that each slip event was accompanied by a level of tremor activity far exceeding that observed at other times (Fig. 2). In Cascadia, this phenomenon is highly regular; the coefficient of variation of ETS recurrence times for the 25 year interval (1982–2007) documented by Rogers (2007) is 0.1, far less than 0.5 which is typical for earthquakes (Ellsworth et al., 1999).

Similarly, Obara et al. (2004) reported the discovery of slow slip events accompanying the previously observed tremor in western Shikoku. Presumably because of the smaller size of the events (approximately moment magnitude 6), the slow slip was not readily observed by GPS instruments and was instead recognized using borehole tiltmeters. Slow slip and tremor were observed to migrate together, with a recurrence interval of approximately 6 months between major episodes. Hirose and Obara (2006) reported similar episodes in the Tokai region.

With improving geodetic networks, slow slip events, or periods of accelerated creep, have been recognized to be relatively common in subduction zones. These events can occur at a variety of depths, with durations ranging at least from days (e.g. Obara et al., 2004) to years (Hirose et al., 1999). Strong correlations between tremor and slow slip have now been established in Cascadia and southwest Japan, and evidence is building that similar correlations may exist in Alaska (Ohta et al., 2006; Peterson and Christensen, 2009), Mexico (Payero et al., 2008), and Costa Rica (Brown et al., 2005; Schwartz et al., 2008) (Fig. 3). On the other hand, slow slip has been seen in some places without generating detectible tremor (Delahaye, et al., 2009; Segall et al., 2006; Ozawa et al., 2002, 2003; Kawasaki et al., 1995, 2001; Montgomery-Brown et al., 2009). Thus the two phenomena are not always linked. Slow-slip and non-volcanic tremor were also recently been reviewed by Schwartz and Rokosky (2007).

Consistent, slow-slip at the rates that many of these events are slipping (mm/h) should not radiate the high frequencies (>1 Hz) that we observe associated with tremor. Slip rates that are many orders of magnitude larger are necessary to radiate such high frequencies (Aki and Richards, 2002). The slow slip events that do have tremor associated with them may have brief periods of accelerated slip that produce this tremor, or perhaps have smaller regions within the larger, slow-slipping region that slip much more rapidly. Alternatively, tremor may come from some other process that is coincident with the slow-slip.

New Opportunities

Individually or in concert, non-volcanic tremor and slow-slip represent a great research opportunity for geoscientists. The recent discovery of these phenomena offers many opportunities to learn about the

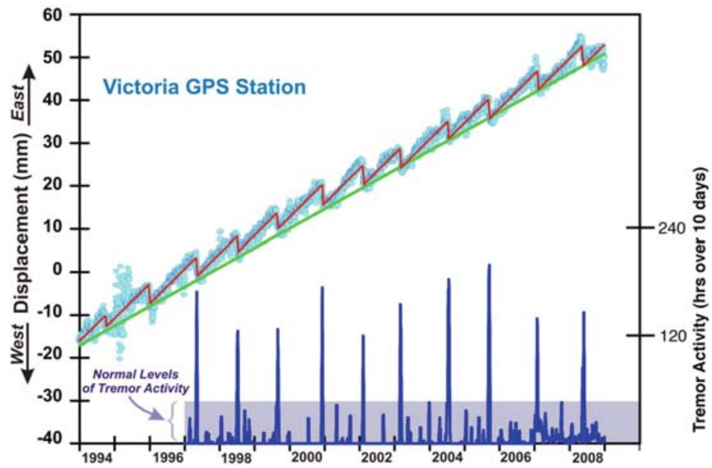

Fig. 2 Comparison of hours of tremor activity to GPS observations of displacement. *Blue dots* represent daily location GPS solutions for the east component of ALBH GPS station. Every ~14 months the trend of the GPS locations reverses direction for ~2 weeks away from the secular trend. These 2 week excursions represent the slow slip events. *Red line* segments represent a linear fit to the trend of the daily locations within periods where the displacement has not reversed. *Green line* represents a linear fit to the entire trend of the displacement. *Black lines* represent hours of non-volcanic tremor recorded in the same region. There is a remarkable correlation between the times when the GPS displacement has a reversal and periods of strong tremor activity. Figure modified from Rogers and Dragert (2003)

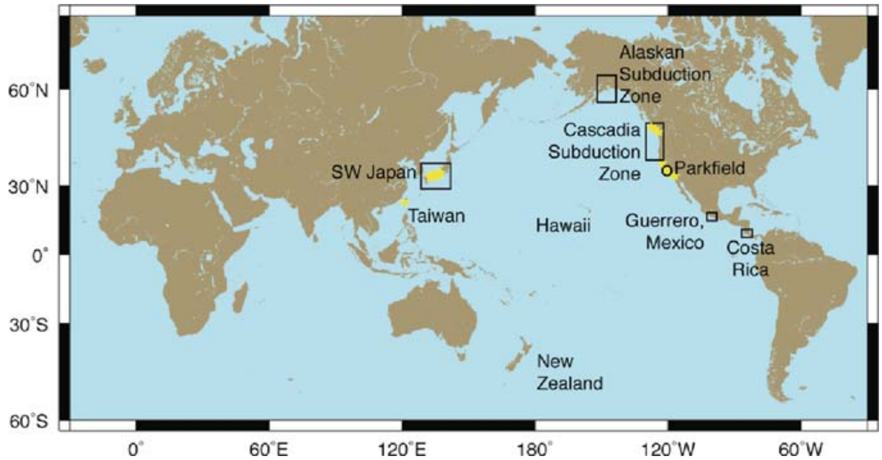

Fig. 3 Map of the world showing regions where non-volcanic tremor and associated slip have been observed. Regions where non-volcanic tremor and coincident slow-slip have observed are indicated by *black boxes*. We note that presently there is only data that indicates there is and episodic relationship between slow-slip in Cascadia and SW Japan, but in the other boxed locations slow-slip and tremor have been observed to be coincident at least once. The *circle* indicates Parkfield, California where non-volcanic tremor has been observed but no associated slow-slip has been observed. The *yellow stars* indicate locations of non-volcanic tremor that was triggered by earthquakes – note that there are many locations where triggered tremor has been observed but spontaneous tremor has yet to be observed. Other locations that are mentioned in the text are also indicated on the map without symbols

rheology and deformation mechanisms in the tremor source region, and it is accelerating research in related topics.

The global distribution of non-volcanic tremor is still being ascertained, but it has been found in a number of different tectonic environments. While most observations of tremor, to date, have been in the Cascadia (e.g., Rogers and Dragert, 2003; Szeliga et al., 2004; McCausland et al., 2005) and the Nankai subduction zones (Obara, 2002), tremor has also been identified in the subduction zones of Alaska (Peterson and Christensen 2009), and Mexico (Payero et al., 2008), but is absent in others including the Japan Trench in northern Honshu (Obara, 2002). It has also been observed along the strike-slip plate boundary in California (Nadeau and Dolenc, 2005; Gomberg et al., 2008), the region beneath the Western Tottori earthquake in Japan (Ohmi et al., 2004), and the central Ranges of Taiwan, within Taiwanese collision zone (Peng and Chao, 2008). Repeating slow-slip events, without tremor, have also been identified in New Zealand (Delahaye et al., 2009; McCaffrey et al., 2008).

Tremor and slow-slip can be used to learn about the conditions in the deep root zone of some major faults. In strike-slip-dominated regions (Gomberg et al., 2008; Nadeau and Dolenc, 2005) and collisional environments (Peng and Chao, 2008) tremor is found below the seismogenic zone, a poorly sampled and understood region. In subduction zones, tremor is typically found in depth ranges with little seismicity (Kao et al., 2005; Shelly et al., 2006), leaving it poorly sampled as well.

In addition to facilitating the study of regions that produce tremor and slow-slip, these phenomena have led to a frenzied development of new techniques and more careful analysis of many data sources. As a result new phenomena are being discovered, including: Very Low Frequency earthquakes (VLF) with durations of ~10 s (Ito et al., 2007) and slow earthquakes with one hundred second duration (Ide et al., 2008). With the continued pace and fervor of research and development in this field, we fully expect that there will be many more breakthroughs in this young and exciting field.

Fundamental Properties of Tremor

Non-volcanic tremor is typically observed as long-duration, low-amplitude, non-impulsive seismic radiation most readily seen in the 1–10 Hz frequency band (Fig. 4, 5). The lack of easily identifiable features makes it difficult to distinguish from cultural or environmental noise and likely prevented an earlier identification of tremor. The unremarkable nature of its

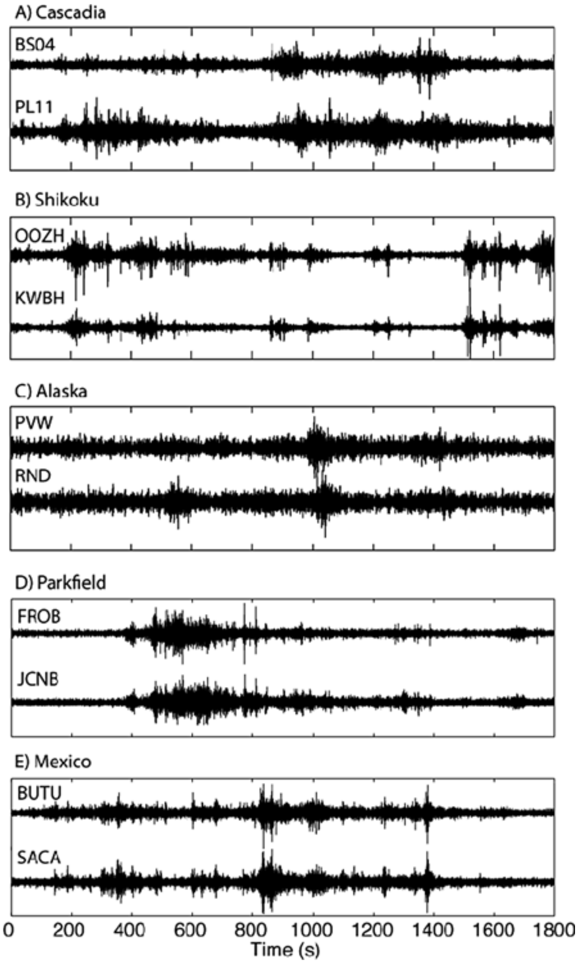

Fig. 4 Recordings of non-volcanic tremor in (**a**) the Cascadia subduction zone (**b**) the Nankai Trough (**c**) the Alaska subduction zone (**d**) Parkfield, California on the San Andreas strike-slip fault and (**e**) the Mexican subduction zone. Records are bandpass filtered at 1–8 Hz. (**b**) is modified from Shelly et al. (2007a)

waveforms poses a challenge for those trying to identify it. Most use very simple methods based on envelope amplitude like those that Obara (2002) used to initially identify tremor, although more complex, automated methods to identify tremor are starting to be developed (Kao et al., 2007a; Wech and Creager, 2008 Suda et al., in press). The absence of easily identified body wave arrivals also contributes to the difficulty in locating non-volcanic tremor. Methods used to locate earthquakes largely depend on the impulsive nature of their body wave phases, rendering them rather ineffective for locating tremor. The issue of tremor location is more fully explored in section "Locating Non-volcanic Tremor".

While non-volcanic tremor usually lacks distinguishable arrivals, impulsive arrivals in Japanese tremor have been observed (Katsumata and Kamaya, 2003). These arrivals are typically *S* waves, but *P* waves have also been found (Shelly et al., 2006). These body wave arrivals are regularly identified and cataloged by the Japanese Meteorological Agency (JMA) as Low Frequency Earthquakes (LFEs). These observations are made primarily on the Hi-Net in Japan, a nationwide network of high-sensitivity borehole seismometers (Obara et al., 2005). The unprecedented density and low noise of the instruments in the Hi-net facilitates the detection of weak signals. LFEs are only rarely identified in regions with tremor outside of Japan (e.g. Kao et al., 2006, Sweet et al., 2008). It is unclear if this difference represents a real variation in tremor activity or simply a limitation in the observation capabilities of networks outside of Japan.

At many time-scales tremor can appear to be very stable, maintaining a fairly constant amplitude for significant amounts of time (Fig. 4) with some waxing and waning of tremor amplitude. At other times, tremor is rather spasmodic, with many bursts that have significantly higher amplitude than the ongoing background tremor (Fig. 4). These bursts can range from less than one minute to tens of minutes. The maximum amplitude of tremor is always relatively small, but appears to vary somewhat from region to region.

Tremor duration is also highly variable. The duration of tremor can range from discrete bursts that last only minutes to ongoing sources that last hours or days (Rogers and Dragert, 2003). During an ETS episode, tremor activity sometimes may continue for days uninterrupted or may also turn on and off erratically throughout the episode. Minor episodes of tremor are routinely observed outside of times of major ETS events. This is also true in California near the town of Parkfield, where correlated slip has not been observed despite excellent detection capabilities provided by borehole strainmeters (Johnston et al., 2006; Smith and Gomberg, in press), in that it is very infrequent that a week goes by without tremor being observed in the Parkfield area.

Watanabe et al. (2007) examined the relationship between duration and amplitude of tremor in southwest Japan, comparing exponential and power law models. They found that the exponential model provided a much better fit, suggesting that tremors, unlike earthquakes, must be of a certain size. As a result, they propose that tremor is generated by fluid processes of a fixed size, or alternatively, that tremor is generated by shear slip on a fault patch of fixed size with variable stress drop.

The spectral content of non-volcanic tremor clearly distinguishes it from earthquakes (Fig. 5), although, at times, non-volcanic tremor can look similar to volcanic tremor. Relative to local earthquakes, tremor is deficient in high frequency energy, in that it has a much steeper drop off of amplitude with increasing frequency. Because of the presence of low-frequency noise and attenuation and smaller source spectra at high frequencies, tremor is most easily identified in a narrow frequency band ranging from approximately 1–10 Hz (Obara, 2002). While energy from tremor undoubtedly extends to a wider frequency range, it is in this frequency range where tremor typically has its highest signal to noise ratio.

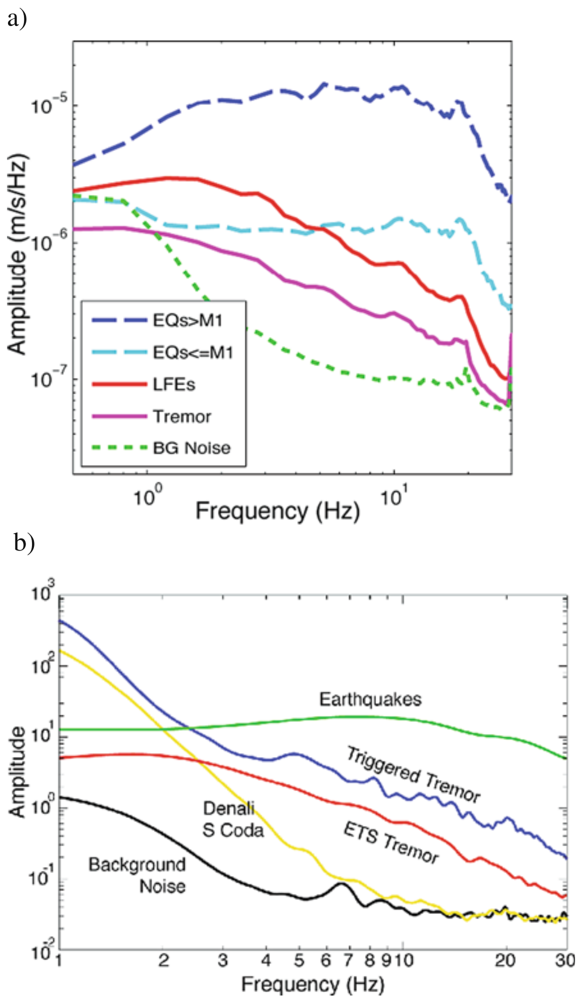

Fig. 5 Velocity spectrum of tremor in Shikoku, Japan (**a**) and Vancouver Island, Canada (**b**). Tremor and local earthquakes have significantly different spectral shape. Triggered tremor (**b**) also has a similar spectral shape as ambient tremor. Figures from Shelly et al. (2007a) (**a**) and Rubinstein et al. (2007) (**b**). We note in (**a**) that the tremor falls below the noise at the lowest frequencies, this is because the noise and tremor were measured at different times and the level of noise during the period of measured tremor was much lower

The tremor wavefield is believed to be dominated by shear waves because it propagates at the S wave velocity and shows higher amplitudes on horizontal components of motion (Obara, 2002; La Rocca et al., 2005). Furthermore, polarization analysis of tremor indicates that tremor is largely composed of shear waves (La Rocca et al., 2005; Wech and Creager, 2007; Payero et al., 2008; Miyazawa and Brodsky, 2008). It seems likely that tremor is generated by a shear source, although fluid based sources can produce shear waves as well (e.g., Chouet, 1988).

Tremor is also highly repeatable with respect to location. Within an individual ETS episode, highly-similar bursts of tremor repeat many times, suggesting that tremor radiates from an individual location many times (Shelly et al., 2007a). From ETS episode to ETS episode, tremor also typically occurs in the same locations (Shelly et al., 2007a; Kao et al., 2006), whereby much of the area where tremor occurs is the same from event to event. Ambient tremor occurring outside ETS events is typically found in these same locations as well.

Most tremor episodes occur spontaneously, but it also can be triggered when the source region is being dynamically stressed by large amplitude teleseismic surface waves (e.g., Miyazawa and Mori, 2005, 2006; Rubinstein et al., 2007; Gomberg et al., 2008). While triggered tremor has been frequently identified in regions where ambient tremor exists, e.g., Parkfield, Vancouver Island, and Japan, it also has been identified in regions where tremor has not previously been identified, e.g., Taiwan and Southern California. It should be noted however, that the existence of ambient tremor in these regions cannot be ruled out because the appropriate studies have not yet been conducted. Similarly, ambient tremor has been found in many regions where triggered tremor has yet to be seen. These incongruities may imply that there are fundamental differences between these regions or processes, or simply that the data in these regions has yet to be thoroughly analyzed.

Locating Non-volcanic Tremor

The very features of the tremor wavefield that make it such a rich phenomena – including the long duration of the source process and absence of distinct body wave arrivals in the seismogram – also make it very difficult to determine where these waves originate. Standard earthquake location methods, like those described below, rely on picking body wave arrivals and most often cannot be used because impulsive arrivals are difficult to find within tremor. Thus, a wide and sometimes novel suite of techniques to locate the tremor source has been developed to exploit some of the unique characteristics of the tremor wave field. These methods largely reproduce the same epicentral locations for tremor, but often have significant differences in the depths (Hirose et al., 2006), whereby some methods suggest that tremor is largely confined to the plate interface in Japan (e.g., Shelly et al., 2006) and other methods indicate that tremor is distributed within a volume of more than 40 km depth in Cascadia (e.g., Kao et al., 2005). The drastic difference in depth distributions of tremor produced by these methods requires significantly different mechanical models to produce tremor in Cascadia and Japan. Thus, precise location of the tremor source in both space and time is a critical step in understanding the mechanics of tremor generation. Doing this will allow us to determine the appropriate physical model for tremor and whether the differences in depth distribution of tremor are real or if they are driven by differences in methodology or data quality.

In general, we can describe the observed seismogram as a convolution of the source process in both space and time with the impulse response of the earth (Green's function) that connects the source positions with the receiver. The resulting seismogram contains a mix of direct body wave arrivals, converted phases and waves scattered by the complex 3D structure of the earth. If the source process has an impulsive beginning it is usually possible to measure the arrival time of the direct P- and S-waves on the seismogram. For earthquakes, this is typically the case and it is then straightforward to estimate the location of the waves' source as is the point that yields the smallest discrepancy between the observed arrival times and those predicted by an appropriate earth model. This is the location of the initial rupture, or hypocenter. Essentially all earthquakes are located in this manner. Commonly, this is done using an iterative least-squares algorithm based on "Geiger's method", the Taylor series expansion of the travel time about a trial hypocenter (Shearer, 1999). This method is attractive, as it only depends on travel time calculations which can be done quickly and efficiently using ray theory. Typically this method cannot be applied to tremor because it often does not have impulsive arrivals that coherently observed at many stations. At the Japan Meteorological Agency, analysts have sometimes been successful in identifying S-waves (and occasionally P-waves) from "low frequency" earthquakes (LFEs) embedded in tremor episodes and locating their hypocenters using these standard methods (Katsumata and Kamaya, 2003).

Waveform Envelope Location Methods

One of the most successful and widely used approaches to locate tremor uses the envelope of the tremor signal to determine the relative arrival times of the waves across a network of stations. First employed by Obara (2002), this method takes advantage of the station to station similarity of smoothed waveform envelopes of high-pass filtered tremor seismograms. Using cross-correlation, one can compute the delay between the envelopes at a pair of stations. The relative arrival times across the network can then be used to locate the tremor source. The errors in the envelope correlation measurements are typically larger than those involved in picking arrival times of earthquakes. Consequently, the location uncertainty is fairly large, particularly for the focal depth, which can exceed 20 km. This method and variants on it are the most commonly used methods to locate non-volcanic tremor (e.g., McCausland et al., 2005; Wech and Creager, 2008; Payero et al., 2008).

Amplitude Based Location Methods

Envelope cross correlation works because the energy output of the tremor source varies with time, waxing and waning on time scales that vary from seconds to minutes. It is reasonable to consider that short-duration periods of high amplitude represent either the constructive interference of waves being radiated from multiple locations in the tremor source or particularly strong radiation from a specific location. In the latter case, it should be possible to exploit both the arrival time and amplitude information to localize the source. Kao and Shan (2004) developed a "source scanning algorithm" to determine the hypocenter by back projection of the observed absolute amplitudes onto the source volume. When the summed wave amplitudes from a network of stations achieve a maximum at a particular location in both space and time, the event hypocenter has been found. The method is closely related to the back projection reconstruction of rupture kinematics of Ishii et al. (2005) used to image the 2004 Sumatra-Andaman Island earthquake. Kao and Shan (2004) have shown that the method compares favorably with conventional methods for locating earthquakes. Since the source scanning algorithm only requires the computation of travel times, and not their partial derivatives, it can be readily implemented in 3D velocity models using an eikonal solver (Vidale, 1988). The epicentral locations computed using this method are similar to those from other methods, with the majority of tremor in Cascadia lying between the surface projections of the 30 and 45 km depth contours of the subduction interface (Kao et al., 2005). They also find tremor at a wide range of depths (>40 km), with errors estimated to be on the order ±3 and ±5 km for the epicenters and depth.

Small Aperture Seismic Array Based Location Methods

Seismic arrays (Capon, 1969; Filson, 1975; Goldstein and Archuleta, 1987) offer an attractive alternative to regional seismic networks for making use of the phase and amplitude information in the wavefield to study the tremor source as they have been used to locate earthquakes and study earthquake rupture propagation (Spudich and Cranswick, 1984; Fletcher et al., 2006). Following this logic, many seismic arrays have been deployed to record non-volcanic tremor. The ETS episode of 2004 was well recorded by three small arrays deployed above the tremor source region in the northern Puget Sound region in British Columbia and Washington (La Rocca et al., 2005, 2008). Even with just 6 or 7 stations, the arrays proved capable of measuring the backazimuth and apparent velocity of the dominant signal in the 2–4 Hz band. Triangulation for the source location using the 3 arrays provided rough estimates of the source position that were comparable to those determined from envelope correlation (McCausland et al., 2005). Significantly, P-wave energy was also detected on the arrays arriving at different velocities than the S-wave energy.

Phase Based Location Methods

If discrete phase arrivals could be identified in the tremor seismogram and correlated across a network of seismic stations, it would be possible to apply standard earthquake location methods (e.g., Geiger's method) to locate the tremor source. Using LFEs that have some phase picks, Shelly et al. (2006) improved the LFE locations in southwestern Japan using waveform cross-correlation with a double-difference technique. These well-located events were then used as templates in a systematic cross-correlation-based search of tremor episodes in southwestern Japan (Shelly et al., 2007a). These authors found that a significant portion of the tremor seismogram could be explained by multiple occurrences of LFEs. This result is discussed in greater detail in section "Low Frequency Earthquakes". This procedure of cross correlating a known event with another time interval has also been used with great success in studying earthquakes (Poupinet et al., 1984) and has led to the recognition that many earthquakes are "doublets" or repeating earthquakes (e.g. Nadeau et al., 2004; Waldhauser et al., 2004; Uchida et al., 2007). It should be noted that imperfect matches are still useful, as the relative delay between the reference event and match across the network of stations can be used to locate the two events relative to one another (see Schaff et al., 2004), potentially providing a very high resolution image of the tremor source region. The search for template events outside of Japan is an area of ongoing effort by a number of research groups. As of this writing, these efforts have met with limited success. We should note that current templates do not explain all of the tremor signals in Japan either. Brown et al. (2008) has worked to address these limitations using an autocorrelation technique to identify repeating tremor waveforms to use as templates.

Another opportunity to improve tremor locations is to identify *P* waves or compute *S-P* times, as most methods purely use *S* wave arrivals. La Rocca et al. (2009) retrieve *S-P* times by cross-correlating the vertical component of recordings of tremor against the horizontal components. This method relies on the assumption that the tremor arrives at near-vertical incidence so that the *P* waves are predominantly recorded on the vertical component and the *S* waves are predominantly on the horizontal component. Using these newly computed *S-P* times, La Rocca et al. (2009) dramatically improve the vertical resolution of tremor locations in Cascadia. For the events that they locate, tremor appears to lie on or very close to the subduction interface.

The Future of Tremor Location

Despite the progress being made in localizing the tremor source, much work remains to be done. With the exception of locations based on template events and *S-P* times, the location uncertainties are currently much larger than those routinely achieved for earthquakes. In general, the tremor epicenters are much better determined than the focal depths, but even epicentral estimates provided by the different methods do not necessarily agree. Other opportunities would include trying to locate tremor as a line or areal source. While much remains to be done, there are ample opportunities for improving upon the existing analysis methods, implementing new techniques, and gathering data in better ways.

Ideally, we would like to image the tremor source process in both space and time as is now commonly done for earthquakes (Hartzell and Heaton 1983). However, the use of the full waveform for studying the tremor source process is hampered by inadequate knowledge of the path Green's function at the frequencies represented in non-volcanic tremor. Knowledge of this information would allow correcting for the Green's function and determining the true source-spectrum of tremor. Learning about the true source spectrum, would undoubtedly teach us a lot about the source processes of non-volcanic tremor.

Developing a Physical Model for Tremor

In this section, we aim to elucidate the physical processes underlying non-volcanic tremor. There are two predominant models to explain the mechanics of non-volcanic tremor: (1) tremor is a result of fluid-flow and fluid processes at the plate interface and within the overlying plate; and (2) tremor is a frictional process that represents failure on a fault with rupture speeds that are much lower than earthquakes. In the following section we will first discuss the evidence for the fluid based model for non-volcanic tremor. We then present two case studies, examining where and why tremor occurs. The evidence from these case studies suggests that the frictional model, explains some attributes of non-volcanic tremor that the fluid-flow model does not. We note that the frictional models, often still appeal to high fluid pressures and the presence of fluids to explain their observations.

In the first case study, we focus our attention on Japan, where diverse and active subduction along with high-quality data has provided an excellent natural laboratory. These conditions have helped lead to the identification and location of tremor and other slow events on a variety of times scales in southwestern Japan. Growing evidence suggests that these events represent plate convergence shear failure on the subduction interface in the transition zone.

In the second case study, we examine tremor activity triggered by tiny stress perturbations from tides and distant earthquakes. These observations can tell us about the conditions under which tremor occurs, and they indicate a sensitivity to stress far beyond what is seen for earthquakes at comparable depths. This argues that tremors probably occur on faults that are very close to failure, which might be achieved if expected high confining pressures are mitigated by near-lithostatic pore fluid pressures.

The Fluid Flow Model for Non-volcanic Tremor

At the time he discovered non-volcanic tremor, Obara (2002) argued that tremor might be related to the movement of fluid in the subduction zone. The depths at which tremor is believed to occur is consistent with depths where significant amounts of subduction related dehydration from basalt to eclogite is occurring (Peacock and Wang, 1999; Julian 2002; Yoshioka et al., 2008), so large amounts of fluid could be present at or near the plate interface. High fluid pressures could then change the fracture criterion of the rock, thus causing hydraulic fracturing, which would radiate the tremor (Obara, 2002). Obara (2002), then goes on to suggest that long-durations of tremor could be a sequence of fractures that are opening as a chain reaction. Other work, examining the stress regime in which tremor is occurring supports the notion that tremor is a product of hydraulic fracturing (Seno, 2005). Others have argued that non-volcanic tremor is caused by brine resonating the walls of fluid conduits near the plate interface (Rogers and Dragert, 2003). This is quite similar to fluid oscillation models for tremor seen at volcanoes (Chouet, 1988; Julian, 2000). Considering the similarities between non-volcanic and volcanic tremor, we expect that much can be learned by comparing the two processes.

Focal mechanism analysis of one burst of non-volcanic tremor in Japan showed that the tremor appeared to be the result of a single-force type source mechanism, which is consistent with fluid flow and not frictional slip (Ohmi and Obara, 2002). This is in contrast with studies of low frequency earthquakes that indicate that tremor appears to be a double-couple source (i.e. shear on a plane) (Ide et al., 2007a; Shelly et al., 2007a).

Additional evidence that non-volcanic tremor is related to fluid flow comes from the distribution of depths where tremor is identified. Studies from both Japan and Cascadia have determined that tremor depths range more than 40 km (e.g, Kao et al., 2005; Nugraha and Mori, 2006). The locations where the tremor is generated in Cascadia correspond well with high-reflectivity regions believed to have fluids (Kao et al., 2005). If tremor is distributed at this wide range of depths, fluid movement seems a much more viable mechanism to produce tremor than slip, as it seems much more likely for there to regions of fluid distributed widely than regions of slip. As discussed in section "Locating Non-volcanic Tremor" and later in section "Tremor Locations: A Broad Depth Distribution in Some Areas?", other studies suggest that tremor is being radiated from the plate interface and does not have a large depth distribution (La Rocca et al., 2009; Shelly et al., 2006; Brown et al., in press). Clearly, precisely determining tremor locations is critical for our understanding of the source processes of tremor.

Case Study I: Non-volcanic Tremor in Japan

Since its discovery in southwest Japan (Obara, 2002), non-volcanic tremor has been extensively studied using high-quality data from the Hi-net borehole seismic network, operated by the National Research Institute for Earth Science and Disaster Prevention (NIED) (Obara, 2005). Hi-net data is supplemented by numerous surface stations operated by the Japan Meteorological Agency (JMA), individual universities, and other agencies. Using Hi-net data, Obara (2002) located the tremor source by waveform envelope cross-correlation and found that the epicenters occurred in a band corresponding to the 35–45 km depth contours of the subducting Philippine Sea Plate in the Nankai Trough (Fig. 1). This band extends from the Bungo Channel in the southwest to the Tokai region in the northeast. Gaps in this band, such as that beneath the Kii Channel, may correspond to where a fossil ridge is being subducted resulting in an area that lacks hydrated oceanic crust (Seno and Yamasaki, 2003).

Following the discovery of ETS in Cascadia (Rogers and Dragert, 2003), Obara et al. (2004) established a similar relationship between tremor and slow slip in Nankai Trough using precise measurements of tilt (Obara et al., 2004). Based on these measurements, slow slip events were modeled to occur on the plate interface, downdip of the seismogenic zone, with durations of ~1 week and equivalent moment magnitudes near 6.0. The locations of slip matched with epicentral locations of tremor, but it was not clear whether the depth of the tremor source matched the depth of slow slip.

Low Frequency Earthquakes

The discovery of low-frequency earthquakes (LFEs) in Southwest Japan (Katsumata and Kamaya, 2003) has led to significant progress in our understanding of tremor processes, including markedly reducing the uncertainty in tremor depths. In Japan, LFEs are routinely identified by the JMA and included in the seismic event catalog. Although some of these events are volcanic, many come from regions far from active volcanoes and are, in fact, relatively strong and isolated portions of non-volcanic tremor. Using mostly S-wave arrival times (few P-wave arrivals are determined for LFEs), JMA estimates the hypocenter and origin time for each event, although the locations generally have large uncertainty, especially in depth. Based on these catalog locations, it was unclear whether the tremor was emanating from the megathrust, within the Wadati-Benioff zone immediately below, or within the upper plate. Drawing from analogies with volcanic

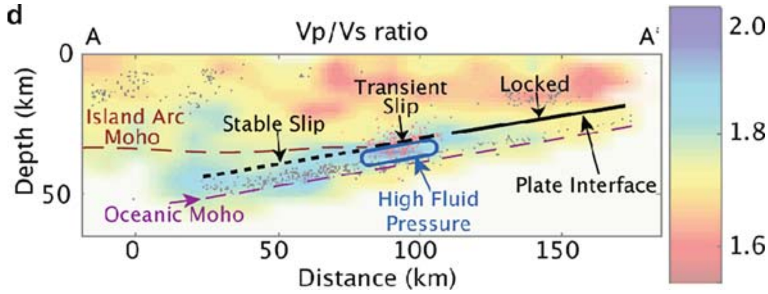

Fig. 6 Cross-section showing hypocenters, Vp/Vs ratios, and structures in western Shikoku. *Red dots* represent LFEs while *black dots* are regular earthquakes. Figure from Shelly et al. (2006)

tremor, initial models of tremor generation proposed that tremor and LFEs might be due to fluid flow near the upper plate Moho (Julian, 2002; Katsumata and Kamaya, 2003; Seno and Yamasaki, 2003).

Shelly et al. (2006) located LFEs and tectonic earthquakes in western Shikoku using waveform cross-correlation and double-difference tomography (Zhang and Thurber, 2003). They found that waveform similarity among LFEs was strong enough to provide accurate differential time measurements, and thus very good focal depth determinations in this region. These locations showed LFEs occurring in a narrow depth range, approximately on a plane dipping with the expected dip of the subducting plate (Fig. 6). These events located 5–8 km shallower than the Wadati-Benioff zone seismicity, and were interpreted as occurring on the megathrust. Based on these locations and the observed temporal and spatial correspondence between tremor and slow slip, Shelly et al. (2006) proposed that LFEs were likely generated directly by shear slip as part of much larger slow slip events, rather than being generated by fluid flow as had been previously suggested.

Support for this hypothesis was provided by Ide et al. (2007a), who determined a composite mechanism for LFEs in western Shikoku using two independent methods. Although the small size of LFEs would normally prevent such an analysis, Ide et al. (2007a) stacked LFE waveforms to improve the signal-to-noise ratio and also utilized waveforms of intraslab earthquakes of known mechanism. Results from an empirical moment tensor using S-waves as well as the mechanism from P-wave first motions both showed motion consistent with slip in the plate convergence direction (Fig. 7). Thus, the kinematics of LFEs appeared to be very similar to regular earthquakes.

Although the above analyses provided strong evidence for the mechanism of LFEs, the relationship between LFEs and continuous tremor was uncertain. Shelly et al. (2007a) argued that the extended duration of tremor could be explained by many LFEs occurring in succession. To identify this correspondence, they used waveforms of catalog LFEs as templates in a matched filter technique applied simultaneously

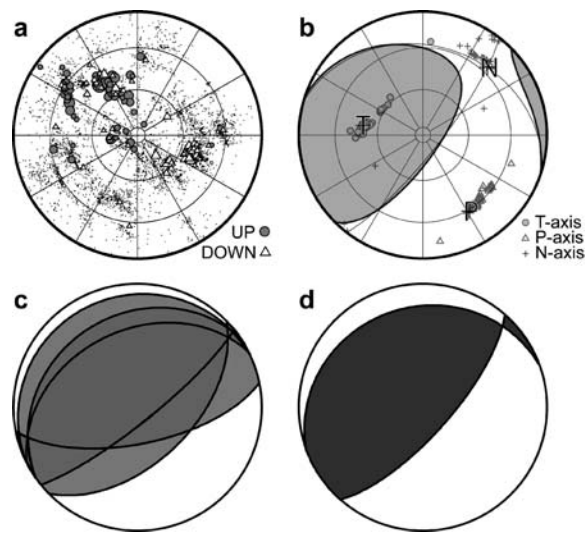

Fig. 7 Comparison of LFE, slow slip event, and megathrust earthquake mechanisms. (**a**) P-wave first motions determined by Ide et al. (2007a) for low frequency earthquakes by cross correlation-based first motion determination. Solid circles and open triangles indicate compressional and dilatational first motions for LFE *P* waves, respectively. SNR for most observations (*small dots*) is too low to determine the polarity. (**b**) Moment tensor inversion results from empirical Green's function analysis of LFE *S* waves. T-, P-, and N-axes are shown together with symbols showing uncertainty and corresponding *P*-wave first motion distribution. (**c**) Overlay of the mechanism for three slow slip events near the study area. (**d**) Mechanism of the 1946 Nankai earthquake, which is the most recent megathrust earthquake in this region and representative of relative plate motion between the Philippine Sea Plate and the overriding plate on the dipping plate interface of the Nankai Trough subduction zone. All these figures are shown in equal area projection of lower focal hemisphere. Figure from Shelly et al. (2007a)

across multiple stations and components (Gibbons and Ringdal, 2006). They found that significant portions of tremor could be matched by the waveforms of a previously recorded LFE. They concluded that, like LFEs, continuous tremor in southwest Japan is also generated directly by shear slip as a component of the larger slow slip events. Importantly, this technique also provided a means to locate this tremor more precisely in space and time.

The successful matching of LFE and tremor waveforms implies that tremor recurs in the same location (or very nearby) during a single ETS episode. Analyzing a two week long ETS episode in western Shikoku, Shelly et al. (2007b) showed that even during a given episode, tremor is generated repeatedly in roughly the same location. In particular, certain patches of the fault, where clusters of LFEs locate, appear to radiate strong tremor in intermittent bursts. The authors suggested that the region of the fault surrounding these patches may slip in a more continuous fashion during an ETS event, driving the LFE patches to repeated failure in a model somewhat analogous to that proposed for repeating earthquakes (Schaff et al., 1998; Nadeau and McEvilly, 1999).

Tremor Migration

Several studies have examined the spatial and temporal evolution of tremor in southwest Japan and found that systematic migration is common. Obara (2002) reported migration of the tremor source along the subduction strike direction at rates of 9–13 km/day, over distances approaching 100 km. Tremor and slip were later seen to migrate together along strike, always at rates of ~10 km/day (Obara et al., 2004; Hirose and Obara, 2005). Along-strike migration directions do not appear to be consistent and migration sometimes occurs bilaterally or activity appears to stall or jump. Similar along-strike migration characteristics have also been reported in Cascadia (Dragert et al., 2004; Kao et al., 2007b).

In addition to relatively slow, along-strike migration, a much faster tremor migration, occurring primarily in the subduction dip direction, was reported by Shelly et al. (2007a, b). Locating tremor by the template LFE method (described above) greatly improved the temporal resolution of tremor locations, allowing locations on a timescale of seconds. Activity was seen to repeatedly migrate up to 20 km at rates of 25–150 km/h, orders of magnitude faster than the observed along-strike migration rates, yet still orders of magnitude slower than typical earthquake rupture velocities. As with the along-strike migration, no preferential direction was observed for along-dip migration. Tremor activity could be seen to propagate updip, downdip, and bilaterally. The downdip migration examples, coupled with relatively fast migration rates, make it unlikely that fluid flow accompanies the tremor. Although it is unclear what generally prevents similar migration velocities in the along-strike direction, a subtle segmentation of the plate boundary, perhaps due to a corrugation in the slip direction, was suggested as a possibility (Shelly et al., 2007b). A similar hypothesis has been proposed to explain streaks of seismicity on faults (Rubin et al., 1999).

A Wide Range of Slow Events

Ito et al. (2007) discovered another new source process occurring along the southwest Japan subduction zone using long period, 20–50 s waveforms. These events, with estimated durations of ~10 s and seismic moment magnitudes of 3.1–3.5, were termed very low frequency (VLF) earthquakes. Timing of these events corresponded with tremor and slow slip. In fact, each VLF was accompanied by a tremor burst in the 2–8 Hz frequency band, but not all tremor bursts were accompanied by detectible VLF events. Focal mechanisms showed thrust faulting, leading to the conclusion that VLFs were also generated by shear slip in the plate convergence direction.

Given the growing number of kinds of shear slip events that occur in the transition zone in southwest Japan (Fig. 8), Ide et al. (2007b) proposed that these events, ranging in duration from ~1 s (LFEs) to years (long-term slow slip), belonged to a single family. This family was unified by a scaling law in which moment scales linearly with duration, rather than as duration cubed as for ordinary earthquakes (Fig. 9). While observations constrain the region between slow events and ordinary earthquakes to be essentially empty, events slower than the proposed scaling relation for a given magnitude might exist beyond the current limits of detection. After this relation was proposed, Ide et al. (2008) detected events predicted by

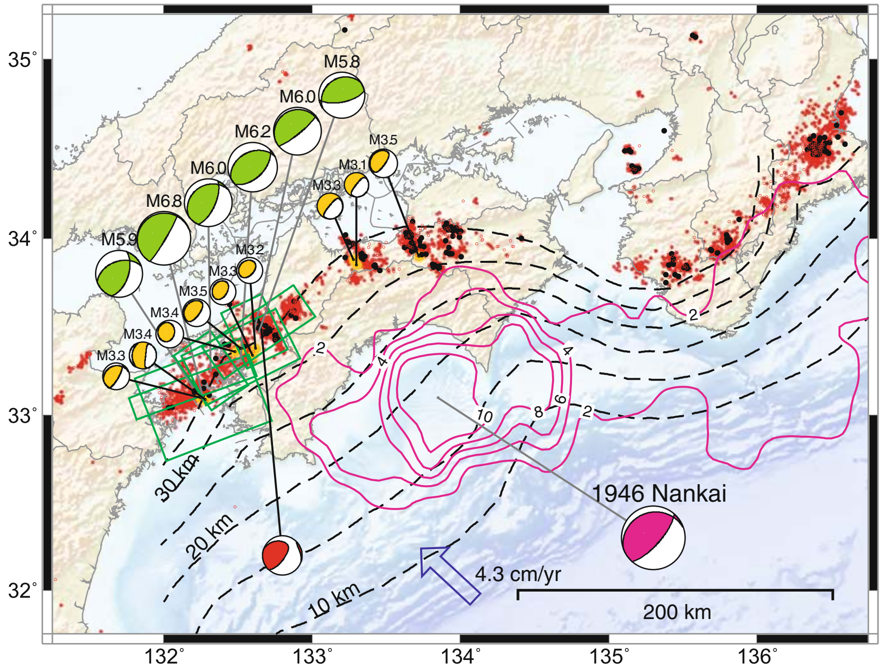

Fig. 8 Various types of earthquakes and their mechanisms along the Nankai Trough, western Japan. *Red dots* represent LFE locations determined by Japan Meteorological Agency. *Red and orange beach balls* show the mechanism of LFEs and VLFs, respectively. *Green rectangles* and *beach balls* show fault slip models of SSE. *Purple contours* and the *purple beach ball* show the slip distribution (in meters) and focal mechanism of the 1946 Nankai earthquake (M8). The top of the Philippine Sea Plate is shown by *dashed contours*. *Blue arrow* represents the direction of relative plate motion in this area. Figure from Ide et al. (2007b)

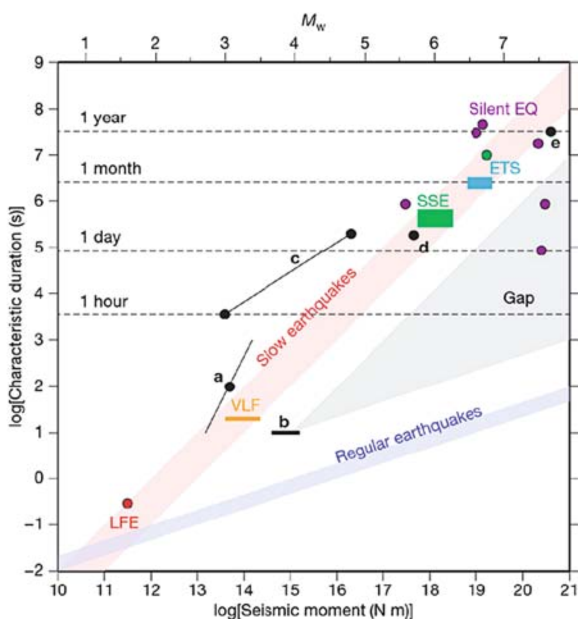

Fig. 9 LFE (*red*), VLF (*orange*), and SSE (*green*) occur in the Nankai trough while ETS (*light blue*) occur in the Cascadia subduction zone. These follow a scaling relation of M_0 proportional to t, for slow earthquakes. *Purple circles* are silent earthquakes. Black symbols are slow events. *a* Slow slip in Italy, representing a typical event (*circle*) and proposed scaling (*line*). *b*, VLF earthquakes in the accretionary prism of the Nankai trough. *c*, Slow slip and creep in the San Andreas Fault. *d*, Slow slip beneath Kilauea volcano. *e*, Afterslip of the 1992 Sanriku earthquake. Typical scaling relation for shallow interplate earthquakes is also shown by a *thick blue line*. Figure from Ide et al. (2007b)

the scaling law with a source duration of 20–200 s and moment magnitude 3–4 under the Kii Peninsula. Such events at these long durations may be common but are difficult to detect due to noise levels and the domination of near-field terms that decay with squared distance. These ~100 s events exhibit a close correspondence between moment rate and high-frequency radiated energy, providing a link between the larger, longer-duration events detected geodetically and smaller shorter-duration events detected seismically.

Case Study II: Stress Interactions of Tremor with Other Earth Processes

Since the discovery of non-volcanic tremor, authors have been interested in the stress interactions between non-volcanic tremor and other earth processes. The periodic nature of ETS makes it easy to connect earth processes to it. For example, the 14-month periodicity of ETS in Northern Cascadia has the same periodicity as the Chandler Wobble (also called the pole-tides). Based on this connection, some have argued that the small gravitation changes associated with the Chandler Wobble are responsible for the periodicity of ETS in Cascadia (Miller et al., 2002; Shen et al., 2005). Similar claims have been made for ETS in Mexico and

Japan, where climatic loading has been argued as the source of the ~12 and ~6 month periodicities of ETS in those locations respectively (Lowry, 2006). However the wide range of dominant ETS periods, from 3 to 20 months in different regions, suggests that outside forcing is, at most, a secondary factor.

A much clearer impact on tremor activity results from small stress changes from distant and local earthquakes as well as the earth and ocean tides. With the aim of elucidating the physical processes underlying non-volcanic tremor, we examine these weak stress perturbations and their effect upon non-volcanic tremor and ETS activity.

Earthquakes Influencing Tremor

Strong evidence suggests that non-volcanic tremor can be influenced by local and distant earthquakes both dynamically, where it is instantaneously triggered by the passage of seismic waves, and in an ambient sense, where periods of active tremor appear to be started or stopped by an earthquake.

Along with the discovery of non-volcanic tremor, Obara (2002) identified the interaction of self-sustaining tremor and local earthquakes. Specifically, periods of active tremor are observed to both turn on and turn off shortly following local and teleseismic earthquakes (Obara, 2002, 2003). An increase in tremor rates is also seen following two strong earthquakes in Parkfield, CA (Nadeau and Guilhem, 2009). A similar observation has been made in Cascadia, where ETS episodes that are "late" appear to be triggered by teleseismic earthquakes (Rubinstein et al., 2009). The interpretation of these observations is complex. For local and regional events, the change in the static stress field caused by the earthquake could be large enough to either start or stop a period of enhanced tremor activity. For teleseismic events, the changes in static stress will be negligible, such that the dynamic stresses associated with them must somehow start or stop a period of enhanced tremor. Rubinstein et al. (2009), propose that when a region is particularly loaded, the small nudge that the dynamic stresses from a teleseismic earthquake provide are enough to start an ETS event going. No satisfactory model has been proposed to explain how a teleseismic event might stop a period of active tremor.

The other mode in which tremor can be influenced by earthquakes is instantaneous triggering by the strong shaking of an earthquake. The first observations of instantaneous triggering of tremor come from Japan, where high-pass filtering broadband records of teleseismic earthquakes showed that there is tremor coincident with the large surface waves (Obara, 2003). Further study identified that tremor was instantaneously triggered by a number of different earthquakes in Japan (Miyazawa and Mori, 2005; 2006). Most observations of triggered tremor are triggered by surface waves, but in at least one case tremor has been observed to have been triggered by teleseismic P waves (Ghosh et al., in press(a)). While triggered tremor is typically larger than self-sustaining tremor, the spectrum of triggered tremor is very similar to that of regular tremor, suggesting that they are the same process (Rubinstein et al., 2007; Peng et al., 2008).

Careful analysis of the phase relationship between the surface waves from the Sumatra earthquake and the tremor it triggered in Japan shows that the tremor is very clearly modulated by surface waves. The tremor turns on when there are positive dilatations associated with the Rayleigh waves and turns off when the dilatation is negative (i.e. during compression) (Miyazawa and Mori, 2006) (Fig. 10). Miyazawa and Mori (2006) interpret this to mean that tremor is related to pumping of fluids from changes in pore space, which might induce brittle fracture and thus generate tremor. Observations of tremor on Vancouver Island triggered by the

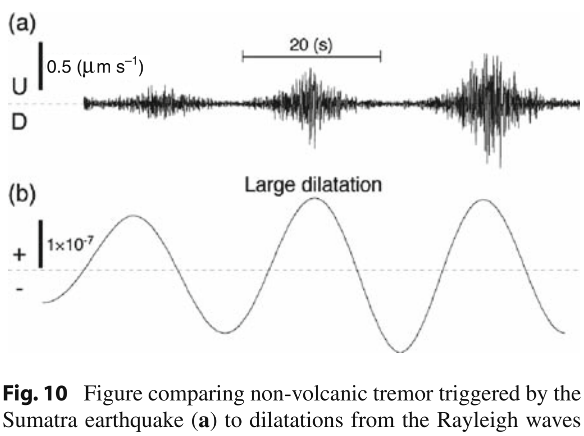

Fig. 10 Figure comparing non-volcanic tremor triggered by the Sumatra earthquake (**a**) to dilatations from the Rayleigh waves from that same earthquake (**b**). Traces have been adjusted to reflect the timing and cause and effect relationship between the surface waves and the tremor. Figure modified from Miyazawa and Mori (2006)

Denali earthquake show instead that tremor is clearly triggered by the Love waves, which have no dilatational component (Rubinstein et al., 2007) (Fig. 11). Rubinstein et al. (2007) offer an alternative explanation for this process, that increased coulomb failure stress from the teleseismic waves promotes slip on the plate interface. They show that when shear stress from the Love waves encourages slip on the plate interface, tremor turns on and when it discourages slip, tremor turns off. This also is supported by the apparent modulation of the triggered tremor amplitude by the shear stress amplitude (Rubinstein et al., 2007), which is predicted by modeling of Coulomb based triggering (Miyazawa and Brodsky, 2008). This behavior is not observed for all observations of triggered tremor (e.g., Peng et al., 2008; Rubinstein et al., 2009). Rubinstein et al. (2007) also argue that this model can explain the observations of tremor being modulated by dilatation (Miyazawa and Mori, 2006), in that increased dilatation also results in a reduction of the Coulomb Failure Criterion on the fault, and should thus encourage slip. Further study of the tremor triggered by the Sumatra earthquake in Japan shows that either model, frictional failure or pumping of fluids can explain the phasing of the tremor with the surface waves (Miyazawa and Brodsky, 2008). It has further been suggested that the difference in triggering behaviors in Cascadia and Japan may be related to the effective coefficient of friction, implying that fluid pressure may be higher in Cascadia than in southwest Japan (Miyazawa et al., 2008)

Tremor triggered at teleseismic distances by large earthquakes offers a powerful tool for identifying additional source regions. For example, tremor was triggered in 7 locations in California by the 2002 Denali earthquake (Gomberg et al., 2008) and underneath the Central Range in Taiwan by the 2001 Kunlun earthquake (Peng and Chao, 2008). With the exception of the tremor triggered in the Parkfield region of California, these observations of triggered tremor are in locations where tremor had never been observed previously. Notably, none of these source regions are in subduction zones. This suggests that tremor is much a much more common process than previously thought and is not limited to subduction zones. Furthermore, these findings indicate that the necessary conditions for producing non-volcanic tremor must exist in a wide variety of tectonic environments.

Computations of shear stress change imparted by teleseismic earthquakes are on the order of tens of kPa (Hill, 2008), or about 10^5 times smaller than the expected confining pressures at these depths. Based on this, some have argued that tremor, at least in its triggered form, occurs on faults that are extremely close to

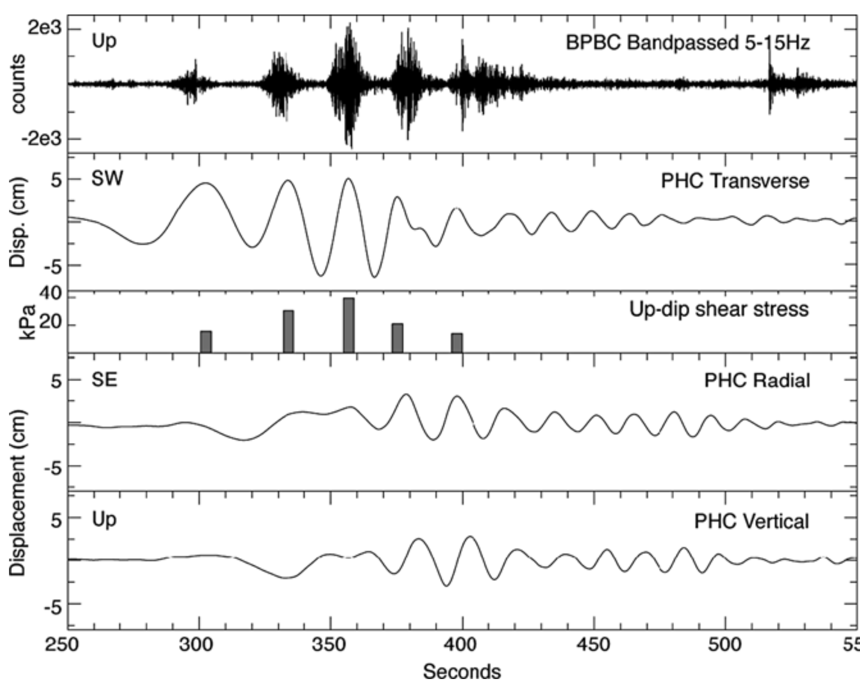

Fig. 11 Figure comparing non-volcanic tremor triggered by the Denali earthquake (a) to surface waves from that same earthquake (b, d, e). Traces have been adjusted to reflect the timing and cause and effect relationship between the surface waves and the tremor. Middle panel reflects the approximate peak shear stress on the plate interface enhancing slip in a subduction sense from the five largest Love wave pulses. Figure modified from Rubinstein et al. (2007)

failure, possibly because of near lithostatic fluid pressures (Miyazawa and Mori, 2006; Rubinstein et al., 2007; Peng and Chao, 2008).

Despite the incremental stresses associated with teleseismic earthquakes, the presence of triggered tremor appears to be strongly controlled by the amplitude of the triggering waves in Parkfield (Peng et al., 2009) and less so on Vancouver Island (Rubinstein et al., 2009). While the amplitude of triggering waves is clearly important in determining whether tremor will be triggered, many other factors are likely important. These include the presence of an ongoing ETS episode or elevated levels of tremor (Rubinstein et al., 2009), frequency content, and azimuth of the earthquake. We also note that while amplitudes and therefore dynamic stresses associated with local and regional, medium-magnitude events are of similar amplitude as those from teleseismic earthquakes, tremor triggered by these events, if it occurs, cannot be observed easily as it is obscured by the larger, high frequency energy associated with the body waves and coda from these events (Rubinstein et al., 2009).

The Tides Influencing Tremor

The periodic changes in gravitation caused by the moon and the sun (the lunar and solar tides) are frequently employed by the earth science community as a way to better understand earth processes. It seems quite logical that when the small stresses associated with the tides encourage slip on faults, seismicity should increase and conversely it should decrease when the tidal stresses discourage slip on these same faults. While a number of studies have identified a very weak correlation between the tides and seismicity rates in particularly favorable conditions (e.g., Tanaka et al., 2002; Cochran et al., 2004, Wilcock, 2001), careful studies of large data sets find no significant correlation of the tides and earthquakes (e.g. Vidale et al., 1998, Cochran and Vidale, 2007).

In contrast to the results from earthquakes, non-volcanic tremor in Japan, Cascadia and Parkfield have been seen to respond strongly to tidal forcing. A comparison of the hourly tremor durations in eastern Shikoku for two ETS events shows that tremor duration is strongly periodic at the two strongest tidal forcing periods of 12.4 and 24 h (Nakata et al., 2008). Examining LFEs in the same location, Shelly et al. (2007b) also determined that non-volcanic tremor is strongly periodic with the lunar tide of 12.4 h and more weakly periodic with the lunisolar tide of 24–25 h. Similarly, a study of non-volcanic tremor in Cascadia showed that the amplitude of tremor in three ETS episodes was strongly periodic at both the 12.4 and 24–25 h tidal periods (Rubinstein et al., 2008). Nadeau et al. (2008), also identify a periodicity to non-volcanic tremor in Parkfield that indicates that it is influenced by the tides.

The periodicity of tremor is such that in both Japan and Cascadia, it is more energetic with high water (Shelly et al., 2007b; Rubinstein et al., 2008). Lambert et al. (2009) similarly show that tremor levels in Cascadia are highest when the normal stress on the plate interface is highest, although this time also corresponds to the time where shear stresses encouraging thrust slip are largest. Neither of the papers that identify this correlation of tremor with water level compute the specific stresses on the fault plane, but they comment that the stresses induced by the tides are miniscule compared to the confining stress of the overburden. Rubinstein et al. (2008) estimates the confining pressures to be approximately 10^5 times larger than ~10 kPa stresses induced by the tides. Nakata et al. (2008) estimate the peak change in Coulomb stress from the solid-earth tides to be ~1 kPa with a maximum rate of ~10 kPa/day, assuming that it occurs as shear slip on the plate interface. Using this computation they find that the temporal behavior of tremor strongly parallels the predictions of a rate-and-state model that predicts seismicity rate changes given a changing stress field. They also note that the stressing rate from the slow-slip event is comparable to the stressing rate from the tides and argue that the tides only should affect tremor if the slow-slip stressing rate is similar in amplitude as the tidal stressing rate.

All of these observations support the argument that tremor is being produced by faults that are very close to failure because they are extremely weak or under near-lithostatic fluid pressures (Nakata et al., 2008, Shelly et al., 2007b, Rubinstein et al., 2008). This parallels the observation that the faults that produce tremor must be at least an order of magnitude more sensitive to triggering than regions where earthquake swarms are produced (Nakata et al., 2008).

Theoretical Models of Slow Slip (and Tremor)

Several studies have attempted to model subduction zone slow slip using a variety of theoretical models in order to constrain the underlying physical mechanisms. Most of these studies do not attempt to model tremor, but rather focus on simulating slow slip. In order for the event to remain slow, the frictional resistance of the sliding surface must increase as the slip velocity increases. In other words, some form of velocity-strengthening friction must put the brakes on slip to keep it from accelerating and becoming an earthquake. The models discussed below all simulate slow slip behavior, but do so through different mechanisms.

Yoshida and Kato (2003) reproduced episodic slow-slip behavior using a two-degree of freedom block-spring model and a rate-and-state friction law. They argue that temporal and spatial variation in stress and frictional properties are necessary conditions for ETS. Similarly, Kuroki et al. (2004) and Hirose and Hirahara (2004) use more complex numerical simulations of slip, and find that they require spatial heterogeneity in frictional properties to be able to reproduce ETS.

Shibazaki and Iio (2003) imposed a rate and state dependent friction law with a small cutoff velocity, such that behavior in the transition zone is velocity weakening at low slip velocity and velocity strengthening at high slip velocity. Such a slip law naturally generates slow slip behavior and may be supported by laboratory data for halite (Shimamoto, 1986) and quartz gouge (Nakatani and Scholz, 2004), under certain conditions. However, it's unclear whether more realistic lithologies, temperatures, and slip speeds behave the same way (Liu and Rice, 2005). Shibazaki and Shimamoto (2007) used a similar approach to specifically model short-term slow slip events. They successfully reproduced slow slip events with propagation velocities of 4–8 km/day, similar to what is observed in Cascadia and southwest Japan and find that this propagation velocity scales linearly with slip velocity in their models.

Liu and Rice (2005, 2007) took a somewhat different approach to achieving slow slip behavior in their rate-and-state-based models. They were able to reproduce transients with a recurrence interval of about a year using laboratory-based friction values with temperature dependence and inserting a region of width W with very high pore pressures updip from the stability transition. They found that the slip behavior primarily depended on the value of the parameter W/h^*, where h^* represents the maximum fault size that produces stable sliding under conditions of velocity-weakening friction. This is similar to the findings of Hirose and Hirahara (2004), who are able to produce slow-slip in a rate and state based model and find a dependence of the slip behavior on the ratio of the width of the slipping region to its lateral dimension. In modeling of Liu and Rice (2007), the recurrence interval of slow slip events decreases with increasing effective normal stress. An effective stress of ~2–3 MPa produces a recurrence interval of 14 months, corresponding to that observed in northern Cascadia.

Another alternative is that dilatant stabilization may play an important role in regulating slow slip and/or tremor behavior, as proposed by Segall and Rubin (2007) and Rubin (2008). They argue that the fault size constraints of the model of Liu and Rice (2005, 2007) may be too specific given the apparent abundance of slow slip events in subduction zones (Rubin and Segal, 2007). Dilatancy that accompanies shear slip will tend to create a suction and thus reduce pore fluid pressure in the fault zone. Depending on the slip speed and permeability, dilatant strengthening could allow slip to occur at slow speeds but prevent it from reaching dynamic speeds typical of earthquakes. If pore fluid pressures in the fault zone approach lithostatic, as has been suggested, the effect of dilatancy becomes relatively more important in controlling slip behavior. In this model, regions of particularly high permeability could slip faster than those with lower permeability, potentially generating tremor.

Many properties of slow slip and tremor can also be explained with a Brownian walk model, where the radius of a circular fault expands and contracts according to this random process (Ide, 2008). Although this model does not address the underlying physical mechanisms, it successfully reproduces the observed frequency content, migration, and scaling of tremor and slow slip, predicting a slight modification to the scaling law proposed by Ide et al. (2007b).

Few laboratory experiments designed to simulate tremor and slow slip have been performed thus far. One recent study by Voisin et al. (2008) examined the effect of cumulative slip on a NaCl sample designed to emulate the frictional conditions in a subduction zone. Although it's unclear how closely this analog represents real conditions of a subduction zone, the experiment succeeded in producing a transition from stick slip behavior, to slow slip, and finally to steady-state creep with increasing cumulative displacement. In addition, they recorded a seismic signal that was qualitatively very tremor-like. They note that the change in behavior with the evolution of their sample is consistent with some features of the modeling discussed above, namely near-neutral stability (Yoshida and Kato, 2003; Liu and Rice, 2005) and a large slip-weakening distance (Shibazaki and Iio, 2003; Kuroki et al., 2004). Tremor-like signals have also been observed in dehydration experiments (Burlini et al., 2009), suggesting that tremor may arise from fluid induced micro-crack propagation and fluid interaction with crack walls, or that metamorphic dehydration reactions supply the fluid necessary to reduce effective pressure and allow tremor to occur.

Clearly additional laboratory experiments specifically designed to study non-volcanic tremor would be important. Considering that laboratory studies have helped reveal many new facets of earthquakes and brittle failure, we expect that laboratory studies will also allow for great insight into the physical processes underlying non-volcanic tremor and slow-slip. Particularly useful will be laboratory simulations that explore the varying conditions expected where tremor is generated (lithology, temperature, pressure, fluid pressure). This parameter space hasn't been thoroughly explored because earthquakes are not abundant in these conditions.

Discussion and Outstanding Questions

We are only beginning to understand the mechanism and environment that produces tremor. Many questions remain unanswered. Following is a discussion of some of the outstanding issues that are topics of ongoing research.

Understanding Why Tremor Occurs in Certain Places

By now, we are beginning to constrain where tremor does and does not occur. By examining the physical conditions in each of these regions including the depth, temperature, mineralogy and metamorphic state, we may succeed in deducing those conditions that are essential for tremor and thereby learn about the source process. We first compare two different tectonic environments where tremor is observed (subduction and strike-slip faults). We then compare the two similar tectonic environments – southwest and northeast Japan – one where tremor is observed and the other where it is not.

An interesting comparison can be made between strike-slip and subduction tremor-hosting environments. The best-documented strike-slip examples are beneath the San Andreas Fault near Parkfield in central California (Nadeau and Dolenc, 2005) and rare instances of activity beneath the source region of the 2000 Western Tottori earthquake in southwest Japan (Ohmi and Obara, 2002; Ohmi et al., 2004). Tremor triggered by teleseismic waves from the Denali earthquake has been observed in several places in California in addition to Parkfield (Gomberg et al., 2008), as discussed above, but tremor has not yet been investigated at other times at these other locations.

Although the subduction and strike-slip environments that generate tremor may appear quite different, some common features are clear. In each case tremor activity occurs below the crustal seismogenic zone of the major fault. These regions appear to correspond to the transitions from stick slip (earthquake-generating) to stable sliding portions of the fault. In subduction zones, the region of tremor and slow slip corresponds to depths where fluids are expected to be liberated from the subducting slab through metamorphic reactions (e.g. Hacker et al., 2003; Yamasaki and Seno, 2003), although varying thermal structures between different regions suggests that tremor does not correspond to a single metamorphic reaction (Peacock, 2009). Seismic studies support the existence of elevated fluid pressures near the tremor in southwest Japan (Kodaira et al., 2004, Shelly et al., 2006, Nugraha and Mori, 2006, Wang et al., 2006, Matsubara et al., 2009), Cascadia (Audet et al., 2009), and Mexico (Song et al., 2009). Furthermore, some have argued that two prominent gaps in tremor in Japan are due to the lack of dehydration reactions and the associated high fluid pressures above them (Seno and Yamasaki, 2003; Wang et al., 2006). Indeed, numerical models of slow slip (see below) often invoke near-lithostatic fluid pressures and thus very low effective stress. Unlike subduction zones, strike-slip faults do not necessarily have an obvious source of fluids. At least for the San Andreas Fault, however, Kirby et al. (2002) have proposed that the fossil slab from previous subduction in this region-may still provide a fluid source. Although fluids might be a necessary condition, they do not appear to be sufficient. For example, no tremor has been reported in hydrothermal areas such as the Geysers, California, Long Valley, California, and Coso Geothermal Field, California.

Indeed, identifying where tremor does not occur is equally important for understanding the underlying mechanisms. While tremor is widespread in the Nankai Trough subduction zone of southwest Japan, it is demonstrably absent at similar levels in the Japan Trench subduction zone of northeastern Japan. Despite the lack of tremor, slow slip is sometimes observed in northeast Japan, often as a large afterslip following an interplate earthquake (e.g. Heki et al., 1997). A major difference between NE and SW Japan subduction is the thermal structure of the subducting plate. In the southwest, the relatively young Philippine Sea Plate subducts at a moderate rate, while in the NE, the much older Pacific plate subducts at a faster rate. Thus the conditions are much colder at a given depth in NE Japan than they are in the SW. This difference significantly influences the seismicity of these regions (Peacock and Wang, 1999); intraslab earthquakes extend to 200 km in NE Japan and only to 65 km depth in the SW. It seems probable that this variability would affect tremor generation as well. If fluids from metamorphic reactions are important in the tremor generation process, they would be released at much greater depth in the NE than in SW. This effect, though, could be negated by advection of fluids to the depths where tremor is believed to originate. Studies of b-values in Tohoku – a region devoid of tremor – suggest that this indeed has happened, leaving the region of 40–70 km depth low in fluids (Anderson, 1980) and less likely to produce tremor. In SW Japan, it has been suggested that the downdip limit of tremor may correspond to where the downgoing slab intersects the island arc Moho, possibly due to the ability of the mantle wedge to absorb fluids though serpentinization (Katsumata and Kamaya, 2003). In NE Japan, however, similar fluid-releasing reactions would take place at a depth of approximately 100 km, long after the slab was in contact with the island arc mantle, preventing the fluids from rising to the depths where tremor is generated. Others have suggested that the segmentation of tremor distribution in Japan is due to stress conditions there, arguing that the stress state of the forearc mantle wedge in NE is compressional and prevents tremor, while in SW Japan the mantle wedge is in tension allowing for hydro-fracture, which they believe to be responsible for tremor (Seno, 2005).

Although new reports come in frequently, thus far only limited locations and times have been searched for tremor. New observations are enabled both by new analyses and by new instrumentation. How does the currently reported distribution of tremor relate to the "true" distribution?

One factor arguing that tremor is widespread, but at levels at or near the noise level, is the variation in the strength of tremor in the currently-identified regions. Some of the strongest tremor may be generated in western Shikoku, where LFEs can be identified and located using methods similar to those for regular earthquakes. Although the Hi-net borehole network certainly assists in this, fewer LFEs are identified in other parts of southwest Japan despite similar station quality and density.

While the maximum amplitude of tremor varies from place to place, it is clearly limited to be relatively small. It is very likely that tremor occurs at or below the noise level of current instrumentation in many places and may evade detection. In other words, the currently recognized distribution of tremor sources should probably be thought of not as the regions that generate tremor, but rather the regions that generate *strong* tremor. Improved seismic instrumentation, increased seismometer density, and addition of low noise seismic sites (e.g., boreholes) would greatly help in identifying tremor in new locations, as well assist in characterizing tremor in locations where tremor has already been seen.

Tremor Locations: a Broad Depth Distribution in Some Areas?

The locations of tremors are fundamental to understanding the underlying processes. A broad distribution of tremor has been reported by several sources for Cascadia (McCausland et al., 2005; Kao et al., 2005; 2006). A similar result has been reported for Mexico (Payero et al., 2008). Although previous studies have argued that tremor is distributed in depth in Japan (e.g., Nugraha and Mori, 2006), these findings and those from other subduction zones contrast with recent results showing that tremor in Japan is concentrated in depth at the plate interface (Shelly et al., 2006; Ohta and Ide, 2008). Does this difference represent a real variation?

A broad depth distribution in tremor would be most easily explained by a fluid-flow mechanism. However, a moment tensor solution in southwest Japan (Ide et al., 2007a) and polarization analysis in Cascadia (Wech and Creager, 2007) argue strongly that tremor is generated by shear slip in both locations. In this case, the broad depth distribution might represent shear slip distributed in depth (Kao et al., 2005). While it's possible to imagine multiple slip interfaces in the subduction zone (e.g. Calvert, 2004), it's perhaps more difficult to imagine these slip zones distributed over a depth range of several 10s of kilometers.

One possibility that must be considered is that not all tremor is generated by the same process. Since "tremor" describes any low-amplitude, extended duration seismic signal, there is no requirement that all tremor be alike. In this scenario, the broad depth distribution and polarization results from Cascadia could be explained if most tremor is generated by shear slip on the plate interface and a smaller component generated at shallower depth by fluid flow (or some other mechanism) in the overlying crust. These tremor sources, while they could be distinct, would still need to be linked as they happen synchronously in episodes of ETS. Although volcanic tremor is believed to arise from multiple processes (McNutt, 2005) so far no evidence has been reported suggesting distinct types of non-volcanic tremor.

Another possibility is that location uncertainty and/or selection bias of different may explain depth discrepancies. No tremor location method locates every part of the signal – to varying degrees, methods either locate only part of the signal or obtain some sort of average over longer periods of time. Methods like source scanning (Kao and Shan, 2004) and LFE location (Shelly et al., 2006) fall into the former category, locating only relatively impulsive events within tremor, while waveform envelope methods fall into the latter category, obtaining some average location over a longer time period. This difference might in part explain the lack of consistency in depth determinations using different location methods in Cascadia (Royle et al., 2006; Hirose et al., 2006). However, the broad tremor depth distributions could also be the result of large location uncertainties. In particular, amplitude-based methods such as source-scanning could be strongly affected by multiple simultaneous sources, as interference of waves from multiple sources could alter the timing of amplitude peaks. This uncertainty would most strongly affect depth estimation. New locations from Cascadia based on *S-P* times (La Rocca et al., 2009) as well as locations from Cascadia and Costa Rica based on waveform cross-correlations (Brown et al., in press) show events localized near the plate interface. This may indicate that, as in southwest Japan, tremor in these areas tracks the plate interface, although again, selection bias must be considered.

Clearly, further studies are needed in order to reduce location uncertainty and resolve this debate, confirming either a broad or narrow tremor depth distribution. One promising avenue for improved locations is the use of seismic arrays. We could learn a great deal about tremor from the installation of multiple large seismic arrays, like the one installed in Washington to record an ETS episode in 2008 (Ghosh et al., in press(b)). Besides providing greatly improved signal-to-noise, such arrays would be capable of distinguishing and locating multiple simultaneous sources, decomposing the complex wavefield in a way that has not thus far been possible.

Relationship Between Tremor and Slow Slip

The precise relationship between slow slip and tremor is still uncertain. Mounting evidence suggests that where tremor is generated by shear failure at the plate interface in the plate convergence direction its distribution in space and time is closely tied to slow slip. Even within this framework, multiple models can be envisioned. One end member would be the idea that slow slip is simply the macroscopic sum of a great many small tremor-generating shear failures (e.g. Ide et al., 2008). In this model, slow slip cannot occur without tremor. This idea may be supported by the linear relationship observed between hours of tremor and slow slip moment (Aguiar et al., 2009), the close correspondence between moment rate and tremor energy for 100 s events (Ide et al., 2008), and the linear relationship between cumulative tremor amplitude measured in reduced displacement and moment measured from strain records of slow-slip events (Hiramatsu et al., 2008). However, this model fails to explain regions that exhibit slow slip without tremor, and the energy radiated through tremor appears to be extremely low compared to the geodetic moment (and slip) of the slow slip events (Ide et al., 2008). Additionally, having many small sources poses a problem of coherence for generating low-frequency energy. An alternative model might be that tremor is only generated at limited locations on the plate boundary, where changes in frictional properties (as a result of geometric, petrologic, or pore pressure heterogeneity) lead to locally accelerated rupture and radiation of seismic waves above 1 Hz while the slow-slip is accommodated elsewhere on the plate boundary. A third, intermediate model might have tremor accompanying slow slip everywhere, with its amplitude varying according to local frictional properties, so as to be undetectable in many locations.

At least in southwest Japan, slow slip events of a week or so are accompanied by (or composed of) slow shear slip events of a range of sizes and durations, at least from tremor/LFEs (~1 s duration) to VLFs (~10 s) (Ito et al., 2007) to 100 s events (Ide et al., 2008) and possibly 1000 s events (Shelly et al., 2007b). While it is clear that these events all contribute to the weeklong slow event, more work needs to be done to clarify their relationships and interactions.

While a clear deformation signal has been observed associated with tremor in the Cascadia and southwest Japan subduction zones, no deformation has yet been detected associated with tremor beneath the San Andreas Fault near Parkfield (Johnston et al., 2006). This could argue for a different mechanism of tremor in this region. However, recent results suggest that at least a portion of the tremor in this zone occurs on the deep extension of the fault, similar to southwest Japan (Shelly et al., 2009). Likewise, based on correlations with small seismic velocity variations following the 2004 Parkfield earthquake, Brenguier et al. (2008) suggest that the Parkfield tremor relates to slow slip at depth. Therefore it's plausible that the basic mechanism is the same as in subduction zones, but the deformation signal is too small to resolve with current instrumentation. A major difference between Cascadia/Japan and Parkfield is the distribution of tremor in time. While the majority of tremor in these subduction zones is concentrated in episodes of relatively intense activity lasting one to a few weeks, tremor near Parkfield appears more diffusely in time. In Parkfield, there are still periods of intense activity, but their intensity relative to the background rate is much smaller than that observed for periods of ETS in Cascadia and Japan. In Cascadia and Japan, a deformation signal is usually not detected until after a few days of active tremor (e.g. Szeliga et al., 2008; Wang et al., 2008). If the tremor and slip beneath the San Andreas are occurring relatively continuously, the associated deformation could be absorbed into the normal interseismic strain signal. Nevertheless, work is ongoing to detect a geodetic complement to tremor in this region; recently, a long-baseline strainmeter has been installed that should offer improved resolution over current instrumentation.

Seismic Hazard Implications

Another important avenue of research is to understand the seismic hazard implications of non-volcanic tremor and ETS. It has been argued that the seismic hazard during an ETS is higher than it is during periods that are quiescent (e.g., Rogers and Dragert, 2003). This is frequently used as a practical justification as to why ETS and tremor should be studied, although we have yet to see a great subduction zone earthquake preceded by an ETS event. Whether the conjecture that ETS elevates seismic hazard is correct is dependent upon the relationship between the area slipping in slow-slip and the seismogenic zone (Iglesias et al., 2004); if the slow-slip event extends into the seismogenic zone, one would expect it to bleed off some of the accumulated strain energy and therefore decrease the hazard (e.g., Yoshioka et al., 2004; Kostoglodov et al., 2003; Larson et al., 2007; Ohta et al., 2006), but if the slow-slip event terminates below the down-dip extent of the seismogenic zone it would effectively load the region (e.g., Brudzinski et al., 2007; Dragert et al., 2001; Lowry et al., 2001). In the loading case, the affect of ETS on seismic hazard may be negligible as the stresses will be quite small. The utility of this information has yet to be fully realized. Indeed, Mazzoti and Adams (2004) used statistical methods to estimate that the probability of a great earthquake is 30 to 100 times higher during an ETS episode than it is at other times of the year, but it is difficult to see how this could be used by emergency managers or the general public as this happens every 14 months in the Seattle region and more frequently elsewhere. If we further consider that any plate boundary where ETS is occurring has more than 1 ETS generating region on it (at least 7 on the Cascadia boundary (Brudzinski and Allen, 2007)), we find hazard estimation even more difficult as all of the ETS generating regions would contribute to the hazard on the entire subduction zone at different times. The problem of hazard estimation based on ETS is further complicated by our poor understanding of the physical and frictional properties at these depths. Knowledge of the physical and frictional properties of the subduction zone is necessary to understand how ETS will affect the earthquake producing region up-dip of it.

There is other information we have learned from tremor and slow-slip which has been useful for better characterizing hazard in subduction zones. Prior to the discovery of non-volcanic tremor and slow-slip, hazard models for subduction zones typically determined the area of the locked zone (i.e. the region expected to slip in a megathrust earthquake) using temperature profiles for the subducting slab as a guide to when it will slip in stick-slip vs creep-slip. Slow-slip events provide a new tool to map the strength of coupling on the plate interface, which in turn can be used to estimate seismic potential (Correa-Mora et al., 2008). Meade and Loveless (2009) offer an alternative interpretation of coupling, suggesting that observations of apparent, partial elastic coupling may actually indicate that an ongoing $M_W >= 8$ slow earthquake is occurring with a duration of decades to centuries. Similarly, McCaffrey et al. (2008) used slow-slip events and the geodetically observed transition from fault locking to free slip at the Hikurangi subduction zone in New Zealand to show that the locked/partially-locked region in this subduction zone is much larger than predicted. Similar work in Cascadia has shown that the locked zone in the Cascadia subduction zone is both larger than expected by thermal models, but also closer to and therefore more dangerous to the major population centers of the region (e.g., Seattle and Vancouver) (McCaffrey, 2009; Chapman, 2009). This method can easily be applied to any subduction zone with slow-slip event and geodetic coverage, which allow seismologists to better characterize the region that will slip in a major earthquake and the hazards associated with it.

There is additional evidence that hazard assessment based on slow-slip is promising. Specifically, we note that slow-slip events in Hawaii (Segall et al., 2006; Brooks et al., 2006, 2008; Wolfe et al., 2007), New Zealand (Delahaye et al., 2009; Reyners and Bannister, 2007), Tokai (Yoshida et al., 2006), and Mexico (Larson et al., 2007; Liu et al., 2007) do appear to have triggered earthquakes. While none of the triggered earthquakes were large enough to pose a hazard to people, the fact that events were triggered demonstrates that the stresses associated with the slow-slip events are large enough to influence earthquakes and therefore affect seismic hazard. While this clearly indicates that there is a relationship between slow-slip events and earthquakes, this is still a difficult problem, as recurrence times of large earthquakes are quite long and therefore makes testing the significance of any prediction very difficult. Another avenue which may be promising is the suggestion of frictional models that the behavior of ETS in a region may change as the region gets closer to catastrophic failure, as hinted by some numerical models (e.g. Liu and Rice, 2007; Shibazaki and Shimamoto, 2007). Similary, Shelly (in press) suggested that changes in tremor migration patterns near Parkfield in the months before the 2004 M 6.0 earthquake might have reflected accelerated creep beneath the eventual earthquake hypocenter. If further observations solidify these hints of a connectionn between ETS and earthquakes, measurements of tremor and slow slip could become powerful tools to forecast large earthquakes.

An additional complication with the earthquakes triggered by the slow slip events in Hawaii and New Zealand is the question as to whether the slow-slip is the same in these events as they are in ETS. The slow-slip in Hawaii and New Zealand that triggers earthquakes occurs in the demonstrable absence of strong tremor, which may imply that different physical processes are occurring. This is another important avenue of future research, clarifying whether the slow-slip events in Hawaii and New Zealand are members of the same family of events that ETS based slow-slip events are. It is certainly possible that these events are producing tremor, only very weakly. Further study of these events and the physical conditions in which these occur should help understand the physics of ETS and slow-slip.

Because very little is known about tremor in continental regimes, it's hazard implications are poorly understood at present, but it does stand to reason that if there is slow-slip associated with the tremor seen in continental regions, that tremor would raise the likelihood of earthquakes. As we learn more about tremor in continental regions and subduction zones we expect that more can be said about the hazard is poses in continental regions.

Summary

We have already learned a great deal about non-volcanic tremor, but the field is still in its infancy. Investigation up to this point has mostly concentrated on understanding the tremor source. No doubt much work remains, but as our understanding of the source progresses, we will begin to find tremor to be an effective tool to study the conditions of deep deformation at various locations in the earth. Through new instrumentation and analysis, and well as new modeling and laboratory experiments, we expect progress to continue at a rapid pace. While tremor and other slow-slip processes may occur in the deep roots of fault zones, we expect that these discoveries will add to our knowledge of tectonic processes in a broad sense, eventually feeding back to aid our understanding of earthquakes.

Acknowledgements The authors would like to thank Roland Burgmann, Joan Gomberg, Jeanne Hardebeck, Stephanie Prejean, Tetsuzo Seno, John Vidale, and an anonymous reviewer for their thorough reviews. We also thank Chloe Peterson, Doug Christensen, Xyoli Perez-Campos, and Vladimir Kostoglodov for their help in procuring sample tremor data for Fig. 4. For Fig. 4: data from Mexico was part of the MesoAmerican Subduction Experiment (MASE) project; data from Alaska comes from the Broadband Experiment Across Alaskan Ranges (BEAAR) experiment; data from Parkfield comes from the High Resolution Seismic Network (HRSN); data from Cascadia comes from the Cascadia Arrays For Earthscope experiment (CAFE); and the data from Shikoku, Japan is from the High Sensitivity Seismic Network (Hi-Net).

References

Aguiar, A.C, T.I. Melbourne, and C.W. Scrivner (2009), Moment release rate of Cascadia tremor constrained by GPS, *J. Geophys. Res.,* 114, B00A05, doi:10.1029/2008JB005909.

Aki, K. and P.G. Richards (2002), Quantitative Seismology, 2nd Edition, University Science Books, Sausalito.

Anderson, R.N. (1980), Phase changes and the frequency-magnitude distribution in the upper plane of the deep seimic zone beneath Tohoku, Japan, *J. Geophys. Res.*, 85, 1389–1398.

Audet, P., M.G. Bostock, N.I. Christensen, and S.M. Peacock (2009), Seismic evidence for overpressured subducted oceanic crust and megathrust fault sealing, *Nature*, 457, 76–78, doi:10.1038/nature07650.

Brenguier, F, M. Campillo, C. Hadziioannou, N.M. Shapiro, R.M. Nadeau, and E. Larose (2008), Postseismic relaxation along the San Andreas fault investigated with continuous seismological observations, *Science*, 321, 1478–1481.

Brooks, B.A., J.H. Foster, M. Bevis, L.N. Frazer, C.J. Wolfe and M. Behn, (2006), Periodic slow earthquake on the flank of Kilauea volcano, Hawai'I, *Earth Planet. Sci. Lett.*, 246, 207–216.

Brooks, B.A., J. Foster, D. Sandwell, C.J. Wolfe, P. Okubo, M. Poland, and D. Myer (2008), Magmatically triggered slow slip at Kilauea Volcano, Hawaii, *Science*, 321, 1177, doi:10.1126/science.1159007.

Brown, J.R., G.C. Beroza, and D. R. Shelly (2008), An autocorrelation method to detect low frequency earthquakes within tremor, *Geophys. Res. Lett.*, 35, L16305, doi:10.1029/2008GL034560.

Brown, J.R., G.C. Beroza, S. Ide, K. Ohta, D.R. Shelly, S.Y. Schwartz, W. Rabbel, M. Thorwart, and H. Kao, Deep low-frequency earthquakes in tremor localize to the plate interface in multiple subduction zones, *Geophys. Res. Lett.*, in press.

Brown, K.M., M.D. Tryon, H.R. DeShon, L.M. Dorman, S.Y. Schwartz (2005). Correlated transient fluid pulsing and seismic tremor in the Costa Rica subduction zone, *Earth Planet. Sci. Lett.*, 238, 189–203.

Brudzinski, M.R. and R.M. Allen (2007), Segmentation in episodic tremor and slip all along Cascadia, *Geology*, 35. 907–910.

Brudzinski, M., E. Cabral-Cano, F. Correa-Mora, C. Demets, and B. Marquez-Azua (2007), Slow slip transients along the Oaxaca subduction segment from 1993 to 2007, *Geophys. J. Intl.*, 171, 523–538, doi:10.1111/j1365-246X.2007.03542.x.

Burlini, L., G. Di Toro, and P. Meredith (2009), Seismic tremor in subduction zones, *Rock Phys. Evidence, Geophys. Res. Lett.,* 36, L08305, doi:10.1029/2009GL037735.

Calvert A. (2004) Seismic reflection imaging of two megathrust shear zones in the northern Cascadia subduction zone, *Nature*, 428, 163–167.

Capon, J. (1969). Investigation of long-period noise at the large aperture seismic array, *J. Geophys*. Res., 74, 3182–3194.

Chapman, J. (2008), M.S. Thesis, Central Washington University, Ellensburg, Washington.

Chouet, B. (1988). Resonance of a fluid driven crack: Radiation properties and implications for the source of long-period events and harmonic tremor, *J. Geophys. Res.*, 93, 4375–4400.

Cochran, E.S., J.E. Vidale, and S. Tanaka (2004). Earth tides can trigger shallow thrust fault earthquakes, *Science*, 306, 1164–1166.

Cochran, E.S. and J.E. Vidale (2007), Comment on tidal synchronicity of the 26 December 2004 Sumatran earthquake and its aftershocks, *Geophys. Res. Lett.*, 34, l04302, doi:10.1029/2006GL028639.

Correa-Mora, F., C. DeMets, E. Cabral-Cano, O. Diaz-Molina, and B. Marquez-Azua (2008), Interplate coupling and transient slip along the subduction interface beneath Oaxaca, Mexico, *Geophys. J. Int.* 175, 269–290, doi:10.1111/j.1365-246X.2008.03910.x

Delahaye, E.J., J. Townend, M.E. Reyners, and G. Rogers (2009), Microseismicity but no tremor accompanying slow slip in the Hikurangi subduction zone, New Zealand, *Earth Planet. Sci. Lett.,* 277, 21–28.

Dragert, H., K. Wang, and T.S. James (2001), A silent slip event on the deeper Cascadia subduction interface. *Science*, 292, 1525–1528.

Dragert, H., K. Wang and G. Rogers (2004), Geodetic and seismic signatures of episodic tremor and slip in the northern Cascadia subduction zone, *Earth Planet. Space*, 56, 1143–1150.

Ellsworth, W.L., M.V. Matthews, R.M. Nadeau, S.P. Nishenko, P.A. Reasenber, R.W. Simpson (1999), A Physically-based earthquake recurrence model for estimation of long-trerm earthquake probabilities, *U.S. Geol. Surv. Open-File Rept*. 99–522.

Filson, J. (1975), Array seismology. *Ann. Rev. Earth Plaet. Sci.*, 3, 157–181.

Fletcher, J.B., P. Spudich, and L. Baker (2006), Rupture propagation of the 2004 Parkfield, California earthquake from observations at the UPSAR, *Bull. Seismol. Soc. Am.*, 96, 129–142.

Ghosh, A., J.E. Vidale, Z. Peng, K.C. Creager, and H. Houston, Complex non-volcanic tremor near Parkfield, California, triggered by the great 2004 Sumatra earthquake, *J. Geophys. Res.,* in press (a).

Ghosh, A., J.E. Vidale, J.R. Sweet, K.C. Creager, and A.G. Wech, Tremor patches in Cascadia revealed by seismic array analysis, *Geophys. Res. Lett.* (in press (b)).

Gibbons, S.J. and F. Ringdal (2006), The detection of low magnitude seismic events using array-based waveform correlation, *Geophys. J. Int.,* 165, 149–166.

Goldstein, P. and R.J. Archuleta (1987), Array analysis of seismic signals, *Geophys. Res. Lett.*, 14, 13–16.

Gomberg, J., J.L. Rubinstein, Z. Peng, K.C. Creager, J.E. Vidale, and P. Bodin, (2008) Widespread triggering on non-volcanic tremor in California, *Science*, 319, 713.

Hacker, B.R., Peacock, S.M., Abers, G.A., and Holloway, S., (2003), Subduction Factory 2. Intermediate-depth earthquakes in subducting slabs are linked to metamorphic dehydration reactions. *J. Geophys. Res.*, 108, doi:10.1029/2001JB001129.

Hartzell, S.H. and T.H. Heaton (1983), Inversion of strong ground motion and teleseismic waveform data for the fault rupture history of the 1979 Imperial Valley, California, earthquake, *Bull. Seismol. Soc. Am.*, 73, 1553–1583.

Heki, K, S. Miyazaki, and H. Tsuji (1997), Silent fault slip following an interplate thrust earthquake at the Japan Trench, *Nature*, 386, 595–598.

Hill, D.P (2008), Dynamic stress, coulomb failure, and remote triggering, *Bull. Seismol. Soc. Am.*, 98, 66–92.

Hiramatsu, Y., T. Watanabe, and K. Obara (2008), Deep low-frequency tremors as a proxy for slip monitoring at plate inteface, *Geophys. Res. Lett.*, 35, L13304, doi:10.1029/2008GL034342.

Hirose, H. and K. Hirahara (2004). A 3-D quasi-static model for a variety of slip behaviors on a subduction fault, *PAGEOPH* , 161, 2417–2431.

Hirose, H. and K. Obara (2005), Repeating short- and long-term slow slip events with deep tremor activity, around the Bungo channel region, southwest Japan, *Earth Planet. Space*, 57, 961–972.

Hirose, H. and K. Obara (2006), Short-term slow slip and correlated tremor episodes in the Tokai region, central Japan, *Geophys. Res. Lett.*, 33, L17311, doi:10.1029/2006GL026579.

Hirose, H., K. Hirahara, F. Kimata, N. Fujii, and S. Miyazaki (1999), A slow thrust slip event following the two 1996 Hyuganada earthquakes beneath the Bungo Channel, southwest Japan. *Geophys. Res. Lett.* 26, 3237–3240.

Hirose, H., H. Kao, and K. Obara (2006). Comparative study of nonvolcanic tremor locations in the Cascadia subduction zone using two different methods, Eos Trans. AGU, 87, *Fall Meet. Suppl*. Abstract T41A-1533.

Ide, S., D. R. Shelly, and G. C. Beroza (2007a), Mechanism of deep low frequency earthquakes: Further evidence that deep non-volcanic tremor is generated by shear slip on the plate interface, *Geophys. Res. Lett.*, 34, L03308, doi:10.1029/2006GL028890.

Ide, S., G. C. Beroza, D. R. Shelly, and T. Uchide (2007b), A new scaling law for slow earthquakes, *Nature*, 447, 76–79.

Ide, S. (2008), A Brownian walk model for slow earthquakes, *Geophys. Res. Lett.*, 35, doi:10.1029/2008GL034821.

Ide S., Imanishi K, Yoshida Y., Beroza G.C., and Shelly D.R. (2008), Bridging the gap between seismically and geodetically detected slow earthquakes, *Geophys. Res. Lett.*, 35, L10305, doi:10.1029/2008GL034014.

Iglesias, A., S.K. Singh, A.R. Lowry, M. Santoyo, V. Kostoglodov, K.M. Larson, S.I. Fracno-Sanchez (2004), The silent earthquake of 2002 in the Guerrero seismic gap, Mexico (Mw=7.6): Inversion of slip on the plate interface and some implications, *Geofisica Int.*, 43, 309–317.

Ishii, M., P. M. Shearer, H. Houston, and J. E. Vidale (2005), Extent, duration and speed of the 2004 Sumatra-Andaman earthquake imaged by the Hi-Net array, *Nature*, 435, doi: 10.1038/nature03675.

Ito Y., K. Obara, K. Shiomi, S. Sekine, and H. Hirose (2007), Slow earthquakes coincident with episodic tremors and slow slip events, *Science*, 315, 503–506, doi:0.1126/science.1134454.

Johnston, M.J.S., R.D. Borcherdt, A.T. Linde, and M.T. Gladwin (2006), Continuous borehole strain and pore pressure in the near field of the 28 September 2004 M 6.0 Parkfield, California, earthquake: Implications for nucleation, fault response, earthquake prediction, and tremor, *Bull. Seismol. Soc. Am.*, 96, S56–S72.

Julian, B.R. (2000), Period doubling and other nonlinear phenomena in volcanic earthquakes and tremor, *J. Volcanol. Geothermal Res.*, 101, 19–26.

Julian, B. R. (2002), Seismological Detectic of Slab Metamorphism, *Science*, 296, 1625–1626.

Kao, H. and S-J. Shan (2004), The source-scanning algorithm: Mapping the distribution of seismic sources in time and space. *Geophys. J. Intl*. 157, 589–594.

Kao, H., S. Shan, H. Dragert, G. Rogers, J. F. Cassidy, and K. Ramachandran (2005), A wide depth distribution of seismic tremors along the northern Cascadia margin, *Nature*, 436, 841–844.

Kao, H. S-JShan, H. Dragert, G. Rogers, J.F. Cassidy, K. Wang, T.S. James, and K. Ramachandran (2006), Spatial-temporal patterns of seismic tremors in northern Cascadia. *J. Geophys. Res.*, 111, doi:10.1029/2005JB003727.

Kao, H., P. J. Thompson, G. Rogers, H. Dragert, and G. Spence (2007a), Automatic detection and characterization of seismic tremors in northern Cascadia, *Geophys. Res. Lett.*, 34, L16313, doi:10.1029/2007GL030822.

Kao, H., S.-J. Shan, G. Rogers, and H. Dragert (2007b), Migration characteristics of seismic tremors in the northern Cascadia margin, *Geophys. Res. Lett.*, 34, L03304, doi:10.1029/2006GL028430.

Katsumata, A., and N. Kamaya (2003), Low-frequency continuous tremor around the Moho discontinuity away from volcanoes in the southwest Japan, *Geophys. Res. Lett.* 30, doi:10.1029/2002GL015981.

Kawasaki I., Asai Y., Tamura Y., Sagiya T., Mikami N., Okada Y., Sakata M., and Kasahara M., (1995), The 1992 Sanriku-Oki, Japan, ultra-slow earthquake, *J. Phys. Earth*, 43, 105–116.

Kawasaki, I., Y.Asal, and Y. Tamura, (2001), Space-time distribution of interplate moment release including slow earthquakes and the seismo-geodetic coupling in the Sanriku-oki region along the Japan trench, *Tectonophysics*, 330, 267–283.

Kirby, S., K. Wang, and T. Brocher (2002), A possible deep, long-term source for water in the Northern San Andreas Fualt System: A ghost of Cascadia subduction past? Eos Trans. AGU, 83, *Fall Meet. Suppl.*, Abstract S22B-1038.

Kodaira, S., T. Iidaka, A. Kato, J.-O. Park, T. Iwassaki, and Y. Kaneda (2004), High pore fluid pressure may cause silent slip in the Nankai Trough, *Science*, 304, 1295–1298.

Kostoglodov, V., S.K. Singh, J.A. Santiago, S.I. Franco, K.M. Larson, A.R. Lowry, and R. Bilham (2003), A large silent earthquake in the Guerrero seismic gap, Mexico, *Geophys. Res. Lett.*, 30, doi:10.1029/2003GL017219.

Kuroki, H., H.M. Ito, H. Takayama, and A. Yoshida (2004). 3-D simulation of the occurrence of slow slip events in the Tokai region with a rate- and state-dependent friction law, *Bull. Seismol. Soc. Am.*, 94, 2037–2050.

La Rocca, M., W. McCausland, D. Galluzo, S. Malone, G. Saccorotti, and E. Del Pezzo (2005). Array measurements of deep tremor signals in the Cascadia subduction zone, *Geophys. Res. Lett.*, 32, L21319, doi:10.1029/2005GL023974.

La Rocca, M., D. Galluzzo, S. Malone, W. McCausland, G. Saccorotti, E. Del Pezzo (2008). Testing small-aperture array analysis on well-located earthquakes, and application to the location of deep tremor, *Bull. Seismol. Soc. Am.*, 93, 620–635.

La Rocca, M., K.C. Creager, D. Galluzzo, S. Malone, J.E. Vidale, J.R. Sweet, and A.G. Wech (2009), Cascadia tremor located near plate interface constrained by *S* minus *P* wave times, *Science*, 323, 620–623, doi:10.1126/science.1167112.

Lambert, A., H. Kao, G. Rogers, and N. Courtier (2009), Correlation of tremor activity with tidal stress in the northern Cascadia subduction zone, *J. Geophys. Res.,* 114, B00A08, doi:10.1029/2008JB006038.

Larson, K.M., V. Kostoglodov, S. Miyazaki, and J.A.S. Santiago (2007), The 2006 aseismic slow slip event in Guerrero, Mexico: New results from GPS, *Geophys. Res. Lett.*, 34, L13309, doi:10.1029/2007GL029912.

Liu, Y. and J. R. Rice (2005), Aseismic slip transients emerge spontaneously in three-dimensional rate and state modeling of subduction earthquake sequences, *J. Geophys. Res.*, 110, doi:10.1029/2004JB003424.

Liu, Y. and J. R. Rice (2007), Spontaneous and triggered aseismic deformation transients in a subduction fault model, *J. Geophys. Res.*, 12, B09404, doi:10.1029/2007JB004930.

Liu, Y., J.R. Rice, and K. M. Larson (2007), Seismicity variations associated with aseismic transients in Guerrero, Mexico, 1995–2006, *Earth Planet. Sci. Lett.*, 262, 493–504.

Lowry, A.R. (2006). Resonant slow fault slip in subduction zones forced by climatic load stress, *Nature*, 442, 802–805.

Lowry, A.R., K.M. Larson, V. Kostoglodov, and R. Bilham (2001), Transient fault slip in Guerrero, Southern Mexico, *Geophys. Res. Lett.*, 28, 3753–3756.

Matsubara, M., K. Obara, and K. Kashara (2009), High-V_P/V_S zone accompanying non-volcanic tremors and slow-slip events beneath southwestern Japan, *Tectonophysics,* 472, 6–17, doi:10.1016/j.tecto.2008.06.013.

Mazzoti, S. and J. Adams (2004), Variability of near-term probability for the next great earthquake on the Cascadia subduction zone, *Bull. Seismol. Soc. Am.*, 94, 1954–1959.

McCausland, W., S. Malone and D. Johnson (2005). Temporal and spatial occurrence of deep non-volcanic tremor: From Washington to Northern California, *Geophys. Res. Lett.* 32, doi:10.1029/2005GL024349.

McCaffrey R, Wallace L.M., and Beavan J (2008), Slow slip and frictional transition at low temperature at the Hikurangi subduction zone, *Nat. Geosci.*, 1, 316–320.

McCaffrey, R. (2009), Time-dependent inversion of three-component continuous GPS for steady and transient sources in northern Cascadia, *Geophys. Res. Lett.,* 36, L07304, doi:10.1029/2008GL036784.

McNutt S.R. (2005), Volcanic Seismology, *Ann. Rev. Earth Plaet. Sci.*, 32:461–491, doi:10.1146/annurev.earth.33.092203.122459.

Meade, B.J. and J.P. Loveless (2009), Predicting the geodetic signature of MW>=8 slow slip events, *Geophys. Res. Lett.,* 36, L01306, doi:10.1029/2008GL03634.

Miller, M. M., Melbourne, T., Johnson, D. J. & Sumner, W. Q. (2002) Periodic slow earthquakes from the Cascadia subduction zone. *Science* 295, 2423.

Miyazawa, M. and E. E. Brodsky (2008), Deep low-frequency tremor that correlates with passing surface waves, *J. Geophys. Res.*, 113, B01307, doi:10.1029/2006JB004890.

Miyazawa, M., E.E. Brodsky, and J. Mori (2008), Learning from dynamic triggering of low-frequency tremor in subduction zones, *Earth Planets Space*, 60, e17–e20.

Miyazawa, M. and J. Mori, (2005), Detection of triggered deep low-frequency events from the 20032005 Tokachi-oki earthquake, *Geophys. Res. Lett.*, 32, doi:10.1029/2005GL022539.

Miyazawa, M. and J. Mori (2006), Evidence suggesting fluid flow beneath Japan due to periodic seismic triggering from the 2004 Sumatra-Andaman earthquake, *Geophys. Res. Lett.,* 33, doi:10.1029/2005GL025087.

Montgomery-Brown, E.K., P. Segall, and A. Miklius (2009), Kilauea slow slip events: Identification, source inversions, and relation to seismicity, *J. Geophys. Res.,* 114, B00A03, doi:10.1029/2008JB006074.

Nadeau, R., A. Thomas, and R. Burgmann (2008), Tremor-tide correlations at Parkfield, CA, *Eos Trans. AGU,* 89, Fall Meet Suppl., Abstract U33A-0054.

Nadeau, R.M. and A. Guilhem (2009), Nonvolcanic tremor evolution and the San Simeon and Parkfield, California, Earthquakes, *Science,* 325, 191–193, doi:10.1126/science.1174155.

Nadeau, R.M. and T. V. McEvilly (1999), Fault slip rates at depth from recurrence intervals of repeating microearthquakes, *Science* 285, 718–721, DOI: 10.1126/science.285.5428.718.

Nadeau, R.M., A. Michelini, R.A. Uhrhammer, D. Dolenc, and T.V. McEvilly (2004). Detailed kinematics, structure, and recurrence of micro-seismicity in the SAFOD target region, *Geophys. Res. Lett.*, 31, L12S08, doi:10.1029/2003GL019409.

Nadeau, R. M. & Dolenc, D. (2005) Nonvolcanic tremors deep beneath the San Andreas fault. *Science* 307, 389; published online 9 December 2004 (10.1126/science.1107142).

Nakata, R., N. Suda, and H. Tsuruoka, (2008), Non-volcanic tremor resulting from the combined effect of Earth tides and slow slip events, *Nat. Geosci.*, 1, 676–678, doi:10.1038/ngeo288.

Nakatani, M., and C. H. Scholz (2004), Frictional healing of quartz gouge under hydrothermal conditions: 1. Experimental evidence for solution transfer healing mechanism, *J. Geophys. Res.*, 109 B07201, doi:10,1029/2001JB001522.

Nugraha, A.D. and J. Mori (2006). Three-dimensional velocity structure in the Bungo Channel and the Shikoku area, Japan, and its relationship to low-frequency earthquakes, *Geophys. Res. Lett.*, 33, L24307, doi:10.1029/2006GL028479.

Obara, K. (2002), Nonvolcanic deep tremor associated with subduction in southwest Japan. *Science*, 296, 1679–1681.

Obara, K. (2003), Time sequence of deep low-frequency tremors in the Southwest Japan Subduction Zone: Triggering phenomena and periodic activity, *Chigaku Zasshi (J. Geogr.)*, 112, 837–849 (in Japanese).

Obara, K., H. Hirose, F. Yamamizu, and K. Kasahara (2004), Episodic slow slip events accompanied by non-volcanic tremors in southwest Japan subduction zone. *Geophys. Res. Lett.* 31, doi:10.1029/2004GL020848.

Obara, K., K. Kasahara, S. Hori, and Y. Okada (2005), A densely distributed highsensitivity seismograph network in Japan: Hi-net by National Research Institute for Earth Science and Disaster Prevention. *Rev. Sci. Instrum.* 76, doi:10.1063/1.1854197.

Ohmi, S. and K. Obara (2002), Deep low-frequency earthquakes beneath the focal region of the Mw 6.7 2000 Western Tottori earthquake, *Geophys. Res. Lett.*, 29, doi:10.1029/2001GL014469.

Ohmi, S., I. Hirose, and J. Mori (2004), Deep low-frequency earthquakes near the downward extension of the seismogenic fault of the 2000 Western Tottori earthquake, *Earth Planets Space*, 56, 1185–1189.

Ohta, Y., J. T. Freymueller, S. Hreinsdóttir, and H. Suito, (2006), A large slow slip event and the depth of the seismogenic zone in the south central Alaska subduction zone, *Earth Planet. Sci. Lett.*, 247, 108–116.

Ohta, K., and S. Ide (2008), A precise hypocenter determination method using network correlation coefficients and its application to deep low frequency earthquakes, *Earth Planets. Space*, 60, 877–882.

Ozawa, S., M. Murakami, M. Kaidzu, T Tada, T. Sagiya, Y. Hatanaka, H. Yarai, and T. Nishimura (2002), Detection and monitoring of ongoing aseismic slip in the Tokai region, central Japan, *Science*, 298, 1009–1012.

Ozawa, S., S. Miyazaki, Y. Hatanaka, T. Imakiire, M. Kaidzu, M. Murakami (2003), Characteristic silent earthquakes in the eastern part of the Boso peninsula, Central Japan, *Geophys. Res. Lett.*, 30, doi:10.1029/2002GL016665.

Payero, J.S., V. Kostoglodov, N. Shapiro, T. Mikumo, A. Iglesia, X. Perez-Campos, R.W. Clayton (2008), Nonvolcanic tremor observed in the Mexican subduction zone, *Geophys. Res. Lett.*, 35, L07305, doi:10.1029/2007GL032877.

Peacock, S.M. (2009), Thermal and metamorphic environment of subduction zone episodic tremor and slip, *J. Geophys. Res.,* 114, B00A07, doi:10.1029/2008JB005978.

Peacock, S. M. & Wang, K. (1999) Seismic consequences of warm versus cool subduction metamorphism: Examples from southwest and northeast Japan. *Science* 286, 937–939.

Peng, Z. and K. Chao (2008), Non-volcanic tremor beneath the Central Range in Taiwan triggered by the 2001 MW7.8 Kunlun earthquake, *Geophys. J. Int.,* 175, 825–829, doi:10.1111/j.1365-246X.2008.03886.x.

Peng, Z., J.E. Vidale, K.C. Creager, J.L. Rubinstein, J. Gomberg, and P. Bodin (2008), Strong tremor near Parkfield, CA excited by the 2002 Denali Fault earthquake, *Geophys. Res. Lett.,* 35, L23305, doi:10.1029/2008GL036080.

Peng, Z., J.E. Vidale, A.G. Wech, R.M. Nadeau, and K.C Creager (2009), Remote triggering of tremor along the San Andreas Fault in central California, *J. Geophys. Res.,* 114, B00A06, doi:10.1029/2008JB006049.

Peterson, C.L. and D.H. Christensen (2009). Possible relationship between nonvolcanic tremor and the 1998–2001 slow slip event, south central Alaska, *J. Geophys. Res.,* 114, B06302, doi:10.1029/2008JB006096.

Poupinet, G., W. L. Ellsworth, and J. Fréchet (1984), Monitoring velocity variations in the crust using earthquake doublets: An application to the Calaveras Fault, California, *J. Geophys. Res.*, 89, 5719–5731.

Reyners, M. and S. Bannister (2007), Earthquakes triggered by slow slip at the plate interface in the Hikurangi subduction zone, New Zealand, *Geophys. Res. Lett.*, 34, L14305, doi:10.1029/2007GL030511.

Rogers, G. and H. Dragert (2003), Episodic tremor and slip on the Cascadia subduction zone: The chatter of silent slip. *Science*, 300, 1942–1943.

Rogers, G. (2007), Episodic Tremor and Slip in Northern Cascadia – Going Back in Time, paper presented at the 2007 Seismol. Soc. Am. Annual Meeting, Waikoloa, Hawaii., 11–13 April.

Royle G.T, Calvert A.J., Kao H (2006), Observations of nonvolcanic tremor during the northern Cascadia slow slip event in February 2002, *Geophys. Res. Lett.*, 33, L18313, doi10.1029/2006GL027316.

Rubin, A.M., D. Gillard, and J.-L. Got (1999), Streaks of microearthquakes along creeping faults. *Nature*, 400, 635–641.

Rubin, A.M. and P. Segall (2007), Episodic slow-slip transients and rate-and-data friction, *Eos Trans. AGU,* 88, Fall Meet. Suppl., Abstract T21A-0374.

Rubin, A.M. (2008), Episodic slow slip events and rate-and-state friction, *J. Geophys. Res.,* 113, B11414, doi:10.1029/2008JB005642.

Rubinstein, J.L., J.E. Vidale, J. Gomberg, P. Bodin, K.C. Creager and S.D. Malone (2007), Non-volcanic tremor driven by large transient shear stresses, *Nature*, 448, doi:10.1038/nature06017, 579–582.

Rubinstein, J.L., M. La Rocca, J.E. Vidale, K.C. Creager, and A.G. Wech (2008), Tidal modulation of non-volcanic tremor, *Science*, 319, 186–189.

Rubinstein, J.L., J. Gomberg, J.E. Vidale, A.G. Wech, H. Kao, K.C. Creager, and G. Rogers (2009), Seismic wave triggering of nonvolcanic tremor, episodic tremor and slip, and earthquakes on Vancouver Island, *J. Geophys. Res.,* 114, B00A01, doi:10.1029/2008JB005875.

Schaff, D. P., G. C. Beroza, and B. E. Shaw (1998), Postseismic response of repeating aftershocks, *Geophys. Res. Lett.*, 25, 4549–4552.

Schaff, D.P., G.H.R. Bokelmann, W.L. Ellsworth, E. Zanzerkia, F. Waldhauser, and G.C. Beroza (2004), Optimizing correlation techniques for improved earthquake location, *Bull. Seism. Soc. Am.,* 94, 705–721, doi:10.1785/0120020238.

Schwartz, S.Y., J.I. Walter, T.H. Dixon, K.C. Psencik, M. Protti, V. Gonzalez, M. Thorwart, and W. Rabbel (2008), Slow slip and tremor detected at the northern Costa Rica seismogenic zone, *Eos Trans. AGU,* 89, Fall. Meet. Suppl., Abstract U31B-06.

Schwartz, S.Y. and J.M. Rokosky (2007), Slow slip events and seismic tremor at circum-pacific subduction zones, *Rev. Geophys.* 45, RG3004, doi:10.1029/2006RG000208.

Segall, P. and A.M. Rubin (2007), Dilatency stabilization of frictional sliding as a mechanism for slow slip events, *Eos Trans. AGU*, 88, Fall Meet. Suppl., Abstract T13F-08.

Segall, P., E.K. Desmarais, D. Shelly, A. Miklius, and P. Cervelli (2006), Earthquakes triggered by silent slip events on Kilauea volcano, Hawaii, *Nature*, 442, 71–74.

Seno, T. and T. Yamasaki (2003), Low-frequency tremors, intraslab and interplate earthquakes in Southwest Japan – from a viewpoint of slab dehydration. *Geophys. Res. Lett.* 30, doi:10.1029/2003GL018349.

Seno, T. (2005), Variation of downdip limit of the seismogenic zone near the Japanese islands, Implications for the serpentinization mechanism of the forearc mantle wedge, *Earth Planet. Sci. Lett.*, 231, 249–262.

Shearer, P.M. (1999), Introduction to Seismology, Cambridge University Press, Cambridge.

Shelly, D.R., Possible deep fault slip preceding the 2004 Parkfield earthquake, inferred from detailed observations of tectonic tremor, *Geophys. Res. Lett.* (in press).

Shelly, D. R., G. C. Beroza, S. Ide, and S. Nakamula (2006), Low-frequency earthquakes in Shikoku, Japan and their relationship to episodic tremor and slip. *Nature* 442, 188–191.

Shelly, D. R., Beroza, G. C. & Ide, S. (2007a), Non-volcanic tremor and low frequency earthquake swarms. *Nature*, 446, 305–307.

Shelly, D.R., G.C. Beroza, and S. Ide (2007b), Complex evolution of transient slip derived from precise tremor locations in western Shikoku, Japan. *Geochem. Geophys. Geosyst.*, 8, Q10014, doi:10.1029/2007GC001640.

Shelly, D.R., W.L. Ellsworth, T. Ryberg, C. Haberland, G.S. Fuis, J. Murphy, R.M. Nadeau, and R. Burgmann (2009), Precise location of San Andreas Fault tremors near Cholame, California using seismometer clusters: Slip on the deep extension of the fault?, *Geophys. Res. Lett.,* 36, L01303, doi:10.1029/2008GL036367.

Shen, Z-K., Q. Wang, R. Burgmann, Y. Wan, and J. Ning (2005). Pole-tide modulation of slow slip events at circum-Pacific subduction zones. *Bull. Seismol. Soc. Am.*, 95, 2009–2015.

Shibazaki, B., and Y. Iio (2003), On the physical mechanism of silent slip events along the deeper part of the seismogenic zone, *Geophys. Res. Lett.*, 30(9), 1489, doi:10.1029/2003GL017047.

Shibazaki B, Shimamoto T (2007) Modelling of short-interval silent slip events in deeper subduction interfaces considering the frictional properties at the unstable-stable transition regime. *Geophys. J. Intl.* 171, 191–205.

Shimamoto, T., (1986). Transition between frictional slip and ductile flow for Halite shear zones at room temperature, *Science,* 231, 711–714.

Smith, E.F. and J. Gomberg, A search in strainmeter data for slow slip associated with triggered and ambient tremor near Parkfield, California, *J. Geophys. Res.*, in press.

Song, T.-R.A, D.V. Helmberger, M.R. Brudzinski, R.W. Clayton, P. Davis, X. Perez-Campos, S.K. Singh (2009), Subducting slab ultra-slow velocity layer coincident with silent earthquakes in southern Mexico, *Science,* 324, 502–505, doi:10.1126/science.1167595.

Spudich, P. and E. Cranswick (1984), Direct observation of rupture propagation during the 1979 Imperial Valley earthquake using a short baseline accelerometer array, *Bull. Seismol. Soc. Am.*, 74, 2083–2114.

Suda, N.R., R. Nakata, and T. Kusumi, An automatic monitoring system for non-volcanic tremors in southwest Japan, *J. Geophys. Res.,* in press.

Sweet, J., K. Creager, J. Vidale, A. Ghosh, M. Nichols, T. Pratt, and A. Wech (2008), Low Frequency Earthquakes in Cascadia, paper presented at 2008 IRIS Workshop, Stevenson Washington, 9 June 2008.

Szeliga, W., T.I. Melbourne, M.M. Miller, and V.M. Santillan (2004), Southern Cascadia episodic slow earthquakes, *Geophys. Res. Lett.*, L16602, doi:10.1029/2004GL020824.

Szeliga, W., T. Melbourne, M. Santillan, and M. Miller (2008), GPS constraints on 34 slow slip events within the Cascadia subduction zone, 1997–2005, *J. Geophys. Res.*, 113, B04404, doi:10.1029/2007JB004948.

Tanaka, S., M. Ohtake, and H. Sato (2002), Evidence for tidal triggering of earthquakes as revealed from statistical analysis of global data, *J. Geophys. Res.*, 107(B10), 221, doi:10.1029/2001JB001577.

Uchida, N., T. Matsuzawa, W. L. Ellsworth, K. Imanishi, T. Okada, and A. Hasegawa (2007), Source parameters of a M4.8 and its accompanying repeating earthquakes off Kamaishi, NE Japan - implications for the hierarchical structure of asperities and earthquake cycle, *Geophys. Res. Lett.*, 34, doi:10.1029/2007GL031263.

Vidale, J.E. (1988). Finite-difference travel time calculation, *Bull. Seismol. Soc. Am.*, 78, 2062–2076.

Vidale, J.E., D.C. Agnew, M.J.S. Johnston, and D.H. Oppenheimer (1998). Absence of earthquake correlation with Earth tides: An indication of high preseismic fault stress rate, *J. Geophys. Res.*, 103, 7247–7263.

Voisin, C., J-R. Grasso, E. Larose, and F. Renard (2008), Evolution of seismic signals and slip patterns along subduction zones: Insights from a friction lab scale experiment, *Geophys. Res. Lett.,* 35, L08302, doi:10.1029/2008GL033356.

Waldhauser, F., W. L. Ellsworth, D. P. Schaff, and A. Cole (2004), Streaks, multiplets, and holes: High-resolution spatio-temporal behavior of Parkfield seismicity. *Geophys. Res. Lett.*, 31, doi:10.1029/2004GL02069.

Wang, Z., D. Zhao, O.P. Mishra, and A. Yamada (2006), Structural heterogeneity and its implications for the low frequency tremors in Southwest Japan, *Earth. Planet. Sci. Lett.,* 251, 66–78.

Wang, K., H. Dragert, H. Kao, and E Roeloffs (2008), Characterizing an “uncharacteristic” ETS event in northern Cascadia, *Geophys. Res. Lett.*, 35, L15303, doi:10.1029/2008GL034415.

Watanabe T, Hiramatsu Y, and Obara K (2007) Scaling relationship between the duration and the amplitude of non-volcanic deep low-frequency tremors, *Geophys. Res. Lett.*, 34, L07305, doi:10.1029/2007GL029391.

Wech A. G., K. C. Creager (2007), Cascadia tremor polarization evidence for plate interface slip, *Geophys. Res. Lett.*, 34, L22306, doi:10.1029/2007GL031167.

Wech, A.G. and K.C. Creager (2008), Automated detection and location of Cascadia tremor, *Geophys. Res. Lett.*, 35, L20302, doi:10.1029/2008GL035458.

Wilcock, W.S.D. (2001). Tidal triggering of microearthquakes on the Juan de Fuca Ridge, *Geophys. Res. Lett.*, 28, 3999–4002.

Wolfe, C.J., B.A. Brooks, J.H. Foster, and P.G. Okubo (2007), Microearthquake streaks and seismicity triggered by slow ear-thquakes on the mobile south flank of Kilauea Volcano, Hawai'I, *Geophys. Res. Lett.*, 34, L23306, doi:10.1029/2007GL031625.

Yamasaki T. and T. Seno (2003), Double seismic zone and dehydration embrittlement, *J. Geophys. Res.*, 108, doi:10.1029/2002JB001918.

Yoshida, S. and N. Kato (2003). Episodic aseismic slip in a two-degree-of-freedom block-spring model, *Geophys. Res. Lett.*, 30, doi:10.1029/2003GL017439.

Yoshida, A., K. Hososno, T. Tsukakoshi, A. Kobayashi, H. Takayama, and S. Wiemer (2006), Change in seismic activity in the Tokai region related to weakening and strengthening of the interplate coupling, *Tectonophysics*, 417, 17–31.

Yoshioka, S., T. Mikumo, V. Kostoglodov, K.M. Larson, A.R. Lowry, and S.K. Singh (2004), Interplate coupling and a recent aseismic slow slip event in the Guerrero seismic gap of the Mexican subduction zone, as deduced from GPS data inversion using a Bayesian information criterio, *Phys. Earth Planet. Interior.*, 146, 513–530.

Yoshioka, S., M. Toda, and J. Nakajima (2008), Regionality of deep low-frequency earthquakes associated with subduction of the Philippine Sea plate along the Nankai Trough, southwest Japan, *Earth Planet. Sci. Lett.*, 272, 189–198.

Zhang, H. & Thurber, C. H. (2003) Double-difference tomography: The method and its application to the Hayward fault, California. *Bull. Seismol. Soc. Am.*, 93, 1875–1889.