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Pole-Tide Modulation of Slow Slip Events at Circum-Pacific Subduction Zones

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Abstract Episodic slow slip (ESS) events have been detected at several circum-Pacific subduction zones, such as Cascadia, Japan, and Mexico. Notably, at least eight ESS events along the northern Cascadia subduction zone recurred with a period of 13–16 months. We study the relationship between pole-tide (associated with Chandler wobble with a period of ~14 months)-induced stress and the occurrence of the ESS events. Our quantitative analysis shows that 14 of the 20 documented ESS events occurred during the ascension phase, prior to the maximum, of a pole-tide-induced Coulomb failure stress change, and three events occurred at the stress-change peak. The pole tides modulate the stress field at the downdip edge of the transition zone along the plate interface and may trigger ESS events when conditions are favorable. The phase advance of the triggered events with respect to the induced Coulomb failure stress change may reflect that the fault slip is dictated by a rate- and state-dependent friction law inferred from laboratory experiments.

Introduction

Slow slip events have been observed around the circum-Pacific subduction zones, and some are found to recur almost periodically. The most remarkable example of such repeating ESS events is found at the northern Cascadia subduction zone (Dragert *et al.*, 2001; Miller *et al.*, 2002; Rogers and Dragert, 2003). Since 1993, eight ESS events have been detected by continuous Global Positioning System (GPS) measurements in the vicinity of Vancouver Island, British Columbia, where stations were found to reverse their motion directions with respect to the interior of the North American plate for 1–3 weeks every 13 to 16 months (Dragert *et al.*, 2001) (Fig. 1). Emissions of tremor-like seismic signals were subsequently discovered accompanying such slip events (Rogers and Dragert, 2003). Recently, slow slip and tremor events have also been reported at the southern Cascadia subduction zone. Eight ESS events were detected by a continuous GPS station located near the California/Oregon border 1997–2004, with seismic tremors observed accompanying the last five events (Szeliga *et al.*, 2004). At the northwestern tip of Vancouver Island seismic tremor events were observed, with a 6-month phase lag behind the south island events (Dragert *et al.*, 2004). Modeling of GPS-observed episodic displacements has placed the source of such ESS events near the downdip edge of the coupled plate interface, at 25–45 km depth (Dragert *et al.*, 2001), downdip of the seismogenic zone of the megathrust, which is shallow and mostly offshore (Wang *et al.*, 2003). The total slip for each ESS event is about 2–4 cm, with the seismic moment release equivalent to an $M \sim 6.5$ earthquake (Dragert *et al.*, 2004).

More ESS events have been discovered at other circum-Pacific subduction zones such as along the Japanese islands and at Guerrero, Mexico. Two transient slip events were captured by a continuous GPS network off Bungo Channel, Japan, in 1996–1997 and 2003, respectively (Hirose *et al.*, 1999; Miyazaki *et al.*, 2003; Ozawa *et al.*, 2004). The 1996–1997 event lasted for about a year, from November 1996 to November 1997. The event had a gentle start, and accelerated later from July to November 1997. The 2003 event took place from July 2003 to February 2004, with most of the deformation accumulated between August and November 2003 (Ozawa *et al.*, 2004). Low-frequency tremors were also recorded during this time period, mostly between 1 September and 1 December 2003, coinciding with the time period when most of the transient slip was taking place (Ozawa *et al.*, 2004). Modeling of the GPS displacement data placed the sources of the two slip events at the same segment of the slab interface between the Eurasian plate and the subducting Philippine Sea plate (Ozawa *et al.*, 2004). The tremors were found to occur at the downdip edge of the slip area, and likely spurred subsequent, more rapid, updip slip (Ozawa *et al.*, 2004). Obara *et al.*, (2004) reported more tremor events in the region between January 2001 and December 2003, with a recurrence period of around 6 months. They also observed tilt changes concurrently with these events. Most of these events, however, lasted only a few days and would result in surface displacement no more than 2 mm (Obara *et al.*, 2004). Consequently, no slow slip was observed by GPS concurrently with these events, except for the one that oc-

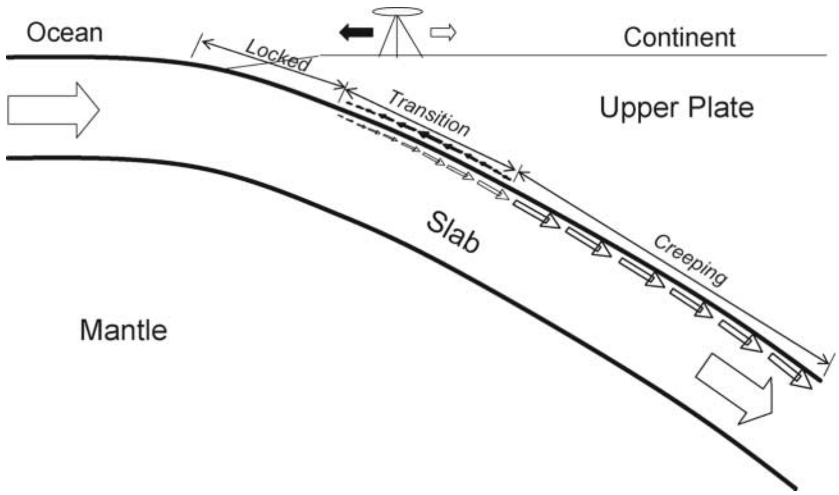

Figure 1. Schematic view of tectonic setting of subduction zone and occurrence of ESS events. Large open arrows mark the secular motion of the oceanic plate relative to the continental plate, with the relative motion across the plate interface shown by open arrows. The small open arrow next to the GPS antenna points to its secular motion direction with regard to the interior of the continental plate. Solid arrows on the plate interface show the episodic slow slip events that occurred in the locked-to-creeping transition zone, with their associated surface motion direction indicated by a solid arrow next to the GPS antenna.

curred late 2003, and we consider these events to be minor, with mechanisms that differ from what we discuss here.

About 700 km east-northeast from the Bungo Channel, two transient events took place around the Boso Peninsula in 1996 and 2002 and were recorded by the GEONET GPS network (Ozawa *et al.*, 2002, 2003). The 1996 event occurred for perhaps more than a year, but most of the transient slip took place from 8 April to 10 June 1996 (Ozawa *et al.*, 2002), or 16 May to 23 May 1996 (Sagiya, 2004). The 2002 event lasted for about 50 days from 4 October to 2 December (Ozawa *et al.*, 2003). The two events show similar deformation patterns: centimeter-level surface displacement in the southeast direction with respect to the interior of the Japanese islands (Ozawa *et al.*, 2002, 2003), except that a recent study showed that surface motion of the 1996 event was more southward, rather than parallel to the relative plate motion direction (Sagiya, 2004). Modeling of the displacement field placed the source of deformation at the upper surface of the subducting Philippine Sea plate off the Boso Peninsula (Ozawa *et al.*, 2002). Low-frequency seismic activity was also observed during these two time periods (Ozawa *et al.*, 2003). Another slow slip event was recorded in the Tokai area, starting as early as July 2000 through at least June 2002. This event was possibly triggered by a swarm of seismo-volcanic activities around the Izu Islands from July to September 2000 (Ozawa *et al.*, 2002). Because of its unique triggering mechanism and long duration, it is not included in this analysis; however, the initiation of the event apparently occurred during ascending pole-tide stress.

Two slow slip events have been reported at the Guerrero subduction zone, Mexico, between the Cocos and North America plate (Lowry *et al.*, 2001; Kostoglodov *et al.*, 2003). The first event occurred between December 1997 and June 1998, with ~3-cm reversed displacement accumulated at a continuous GPS station. The second event took place from October 2001 to May 2002, with up to ~7 cm of anomalous displacement accumulated at some GPS sites during this time period. Recently, Lowry *et al.* (2004) reported detection of two more slow slip events that occurred since 2002; one occurred around the beginning of 2003, and the other started in January 2004. Additional slow slip events may have occurred in 1999, 2000, and 2001, some of which might be associated with preceding seismic events (DeMets *et al.*, 2004). Similar to the Cascadia and Japanese island events, these Guerrero events are believed to have occurred at the transition zone of the subducting plate interface, about 60–180 km landward from the trench. Equivalent seismic moments released for the 1998, 2002, and 2003 events are on the order of M_w 7.2, M_w 7.4, and M_w 7.2 earthquakes, respectively, based on geodetic deformation modeling (Lowry *et al.*, 2004). In this article, we examine only the events of 1998 and 2002, since we do not know precise timings and durations of the other events yet.

Although these ESS events have been documented and their mechanical processes have been hypothesized, little is known about what triggers them. Miller *et al.* (2002) pointed out a remarkable correlation between the periods of the Cascadia ESS recurrences and Chandler wobble (the latter refers to wobbling of the Earth's rotation pole around its inertia figure, caused by mass movement of the Earth) (Melchior, 1978): both are about 14.5 months. However, they dismissed the latter as a plausible triggering source on the grounds of the small stress perturbation it may generate. But how much and of what phase is the stress change induced by the Chandler wobble, and can that, after all, trigger the ESS events?

The Model

To answer these questions we calculate the stress change produced by the pole tide at the locked-to-creeping transition zone interface at the subduction zones mentioned above, which are subsequently transformed to a coordinate system referenced to the plate interface. Gravitational potential induced by variation of Earth rotation can be expressed as (Wahr, 1985):

$$V(\theta,\lambda) = -\frac{1}{2}\Omega_0^2 r^2 [\sin 2\theta(m_x \cos\lambda + m_y \sin\lambda) + 2m_z \sin^2\theta]$$

where θ and λ are colatitude and longitude respectively; Ω_0 is the Earth's rotation rate, r the position radius, m_x and m_y are the vector components of the rotation pole change and can be obtained from the International Earth Rotation and Reference System server (IERS, www.iers.org/iers/products/eop), and m_z is the change of rotation rate, which is an order of magnitude smaller than m_x and m_y. Pole-tide-induced displacement at the Earth's surface, given by Wahr (1985) and Gipson and Ma (1998) is:

$$U_r = \frac{h}{g} V(\theta,\lambda,m) = -h\frac{\Omega_0^2 r^2}{2g}[\sin 2\theta(m_x \cos\lambda + m_y \sin\lambda) + 2m_z \sin^2\theta]$$

$$U_\lambda = \frac{l}{g}\frac{1}{\sin\theta}\frac{\partial}{\partial\lambda}V(\theta,\lambda,m) = l\frac{\Omega_0^2 r^2}{g}\cos\theta(m_x \sin\lambda - m_y \cos\lambda)$$

$$U_\theta = \frac{l}{g}\frac{\partial}{\partial\theta}V(\theta,\lambda,m) = -l\frac{\Omega_0^2 r^2}{g}[\cos 2\theta(m_x \cos\lambda + m_y \sin\lambda) + m_z \sin 2\theta]$$

where l and h are Love numbers, given as $l \approx 0.085$ and $h \approx 0.600$; g is the gravitational constant.

We derive pole-tide-induced strains of three components from above equations:

$$\varepsilon_{\theta\theta} = \frac{1}{r}\frac{\partial U_\theta}{\partial\theta} + \frac{U_r}{r} = 2l\frac{\Omega_0^2 r}{g}[\sin 2\theta(m_x \cos\lambda + m_y \sin\lambda) - m_z \cos 2\theta] - h\frac{\Omega_0^2 r}{2g}[\sin 2\theta(m_x \cos\lambda + m_y \sin\lambda) + 2m_z \sin^2\theta]$$

$$\varepsilon_{\lambda\lambda} = \frac{1}{r\sin\theta}\frac{\partial U_\lambda}{\partial\lambda} + \frac{U_\theta}{r} ctg\theta + \frac{U_r}{r} = l\frac{\Omega_0^2 r}{g} ctg\theta[(1 - \cos 2\theta)(m_x \cos\lambda + m_y \sin\lambda) - m_z \sin 2\theta] - h\frac{\Omega_0^2 r}{2g}[\sin 2\theta(m_x \cos\lambda + m_y \sin\lambda) + 2m_z \sin^2\theta]$$

$$\varepsilon_{\theta\lambda} = \frac{1}{2}\left(\frac{1}{r\sin\theta}\frac{\partial U_\theta}{\partial\lambda} + \frac{1}{r}\frac{\partial U_\lambda}{\partial\theta} - \frac{U_\lambda}{r} ctg\theta\right) = -l\frac{\Omega_0^2\, r}{g}\sin\theta(m_x \sin\lambda - m_y \cos\lambda)$$

The free surface boundary condition provides us with 3 more equations:

$$\tau_{rr} = \tau_{\theta r} = \tau_{\lambda r} = 0$$

Assuming elastic media, we can combine the above equations for the three known stress and three known strain components and their constitutive relationships to derive the other three stress components $\tau_{\theta\theta}$, $\tau_{\lambda\lambda}$, and $\tau_{\theta\lambda}$. To a first-order approximation, we assume that the pole-tide-induced stress does not vary with depth within the shallow depth (tens of kilometers) of the Earth. The above equations then are used to evaluate the Coulomb stress change induced by the pole tide.

Results

We choose the Coulomb stress estimation spot to be located near the downdip end of the transition zone from fully locked to creeping following published estimates. The plate geometry parameters are listed in Table 1. The calculated pole tide induced stresses on the plate interface for all the areas are shown in Figure 2. A remarkable synchronization is evident between the occurrences of the ESS events and phases of pole-tide-induced updip shear stress τ. Most of the recorded ESS events occurred during ascension of τ, and some of them were close to or at the maximum of τ. It is also noted, however, that the pole-tide-induced normal stress σ oscillates with a phase offset of $\sim$180° from τ; that is, the slip events occurred when pole-tide stresses further compressed the fault plane. If we assume a friction coefficient of $\mu = 0.4$, the static Coulomb failure stress change $\Delta\mathrm{CFS} = \tau + \mu\sigma$ is a bit smaller than but still synchronized with τ. For the 20 slow slip events recorded at the five subduction zones described above, 14 occurred during the ascension phase, three at the maximum, and three in the descending phase of the pole-tide-induced Coulomb failure stress change. Among the three events that slipped during the descending phase, two were in southern Cascadia and one was at the Boso Peninsula of southwest Japan.

It should be noted that the elasticity assumption used for stress and deformation estimation above is valid only for the region outside of the fault zone. Within the fault zone, the effect of fault creep downdip from the transition zone should be accounted for. A qualitative model suggests that such an effect may amplify the Coulomb failure stress on the locked plate interface by an order of magnitude. This test is done using a finite-element model, assuming a subducting slab dipping at 30°, with the slab interface completely locked from 0 to 40 km depth, and bearing no shear stress at >40 km depth (for time periods of several months, within which the locked-to-creeping transition zone is assumed with no episodic slipping). The result shows a 5- to 65-fold increase of the induced shear stress changes around the transition zone of the subducting plate at 25–40 km depth. A smoothly tapered fault strength of the transition zone would

Table 1
Parameters of Subduction Zone Plate Setting and Locations of Stress Evaluation

Segment	Strike	Dip	Rake	Latitude	Longitude	Reference
N. Cascadia	N30°W	20°	90°	48.4°N	123.5°W	Wang *et al.*, 2001
S. Cascadia	N5°W	20°	90°	41.8°N	122.7°W	Szeliga *et al.*, 2004
Guerrero	N70°W	20°	80°	17°N	99°W	Kostoglodov *et al.*, 2003
Bungo Channel	N20°E	20°	70°	32°N	132°E	Miyazaki and Heki, 2001
Boso Peninsula	N25°E	20°	80°	35°N	141.5°E	Ozawa *et al.*, 2003

reduce the stress amplification at its lower end, but would increase the amplification at its upper end. Thus it is reasonable to expect the pole-tide-induced Coulomb stress changes at the transition zone to be at the level of 1 kPa or above.

Discussion and Conclusions

Can pole-tide-induced stress change be large enough to trigger ESS? This question is not easy to answer, because of our lack of knowledge of most of the relevant factors, such as the tectonic stress/strain rate and state, slip (or slow earthquake) nucleation time, and the appropriate constitutive law at the transition zone of the subducting plate interface. However, recent geophysical studies have provided some clues. One of them is the observation of earthquake triggering under extreme solid Earth tides. The solid Earth tides cause daily stress oscillation in the crust, on the order of several kPa. Previous statistical studies of Vidale *et al.* (1998) found virtually no triggering of microseismicity from solid Earth tides. Their recent study (Cochran *et al.*, 2004), however, showed that in the subset of earthquakes that occurred during the top 1% of absolute Earth-tide-induced Coulomb stress ($\geq \sim$10 kPa), 74% of the events occurred when Coulomb stress was positive, suggesting that the triggering effect is significant (>99% confidence). Such a result suggests that for short-term triggering of earthquakes, Coulomb stress induced by diurnal and semidiurnal tides (on the order of 1 kPa) are normally just below the triggering threshold; they could, however, play a triggering role when they are at their extremes. Such an assessment is corroborated by the study of Tanaka *et al.* (2004), who discovered that tidal triggering may be significant in a small fraction of the subduction zone regions in Japan, where directions of tidal-induced compressional stress at the time of the earthquakes is strongly correlated with the compressional *P* axes of observed earthquakes (Stein, 2004).

If stress induced by the extreme solid Earth tides could barely play a seismic triggering role, what about the pole-tide-induced stress changes, which are estimated to be on the order of 1 kPa, smaller than the interseismic stress build-up during a pole-tide oscillation cycle (Mazzotti and Adams, 2004). The pole tide oscillates at a much lower frequency, and ESS events occur at the plate interface of different depth range and fault rheology than ordinary earthquakes. The faulting process there is determined not only by its static stress state, but also by deformation rate and history (Dieterich, 1981; Ruina, 1983). The triggering effect can depend strongly on the oscillation period of the modulation stress. Laboratory experiments of a fault system loaded by a combination of steady stressing and a low-frequency oscillatory loading indicate that when the period of the sinusoidal stress is long enough (hundreds of seconds, presumably much longer than the nucleation time of cracks), the triggering threshold for stress oscillation is significantly reduced (Beeler and Lockner, 2003). "Earthquakes" could be triggered even if the amplitude of the periodic loading is about one order of magnitude smaller than the accumulated loading stress in an oscillation cycle. Beeler and Lockner (2003) estimated the minimum duration of earthquake nucleation on major fault zones as $\sim$1 year. This may suggest that the triggering threshold for periodic loading stress at the intermediate (months to years) timescale could be much less than that at the short-immediate (hours to days) timescale. That is, pole-tide-induced Coulomb stress may be able to trigger ESS at the intermediate timescale ($\sim 10^2$ days), while the larger diurnal and semidiurnal Earth tides rarely do.

Beeler and Lockner (2003) find in their laboratory experiments that under the combined action of a constant stress loading and a small sinusoidal stress, there is a phase shift between fault slip events and peak periodic loading stress. If the stress oscillation period is longer than the duration of slip nucleation, slip may take place before the system reaches the maximum of the periodic loading stress. This is consistent with our observation that ESS events almost always occur before the maximum pole-tide-induced Coulomb stress change.

If pole-tide-induced Coulomb stress is capable of triggering ESS events, what about other periodic loading forces caused by motions of solid or fluid objects? Murakami and Miyazaki (2001) reported seasonal variation of crustal deformation measured by a continuous GPS network within the Japanese island arc. Heki (2001) linked the annual surface displacements to heavy snow loading in northeast Japan. He estimated that the periodic loading of snow could be equivalent to up to 10 kPa atmospheric pressure change. Interestingly enough, seasonality of earthquake occurrence has also been observed and related to crustal deformation in Japan (Ohtake and Nakahara, 1999). Although the snow loading region is hundreds of kilometers away from the subduction zone, it may still cause a Coulomb stress change at the plate interface on the order of hundreds of Pa (Heki, 2001), similar to the pole-tide-induced Coulomb stress. An ESS event,

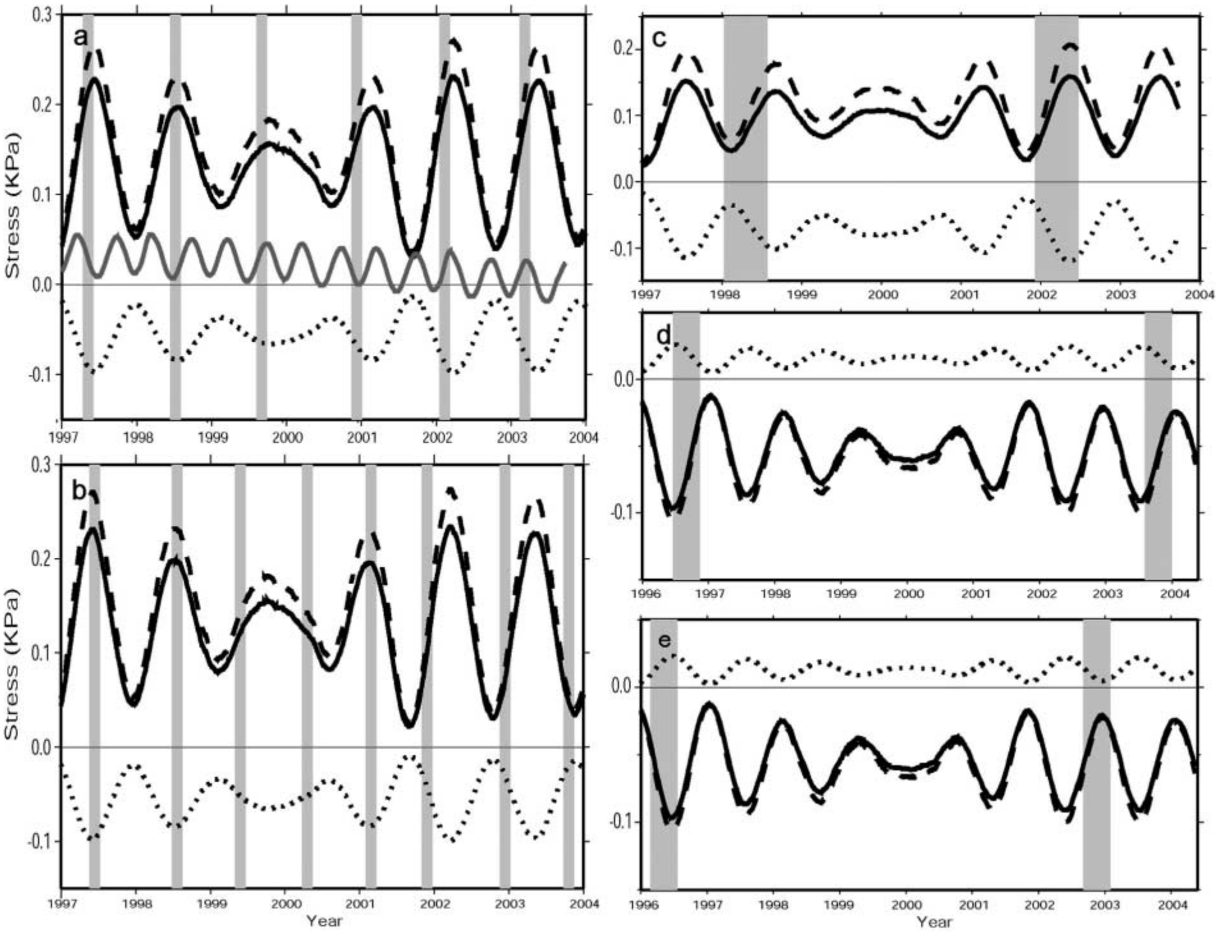

Figure 2. Pole-tide-induced stress-change time series. Solid curve is the Coulomb failure stress change on the slab interface, dashed and dotted curves are its shear and normal components, respectively. Fault normal tension and thrust-sense shear stress are positive. Gray vertical stripes are the epochs of the ESS events. (a) northern Cascadia, (b) southern Cascadia, (c) Guerrero, (d) Bungo Channel, and (e) Boso Peninsula subduction zones. Gray curve in (a) is the Coulomb failure stress change induced by annual and semiannual Earth tides.

which our model cannot fit, at Boso Peninsula from April to June 1996, was in phase with the snow loading-induced Coulomb stress at the interface of the subduction zone (Heki, 2001). This may be more than a coincidence. The high-frequency diurnal and semidiurnal tides, as discussed above, do not appear to play a strong triggering role. Amplitudes of the monthly and semimonthly Earth tides are slightly larger than that of the semiannual Earth tide, but their periods are perhaps still too short to impose a sustained stress for triggering. The annual and semiannual Earth tides, as shown in Figure 2, have much smaller magnitudes than the pole tide (Melchior, 1978), and do not appear to be significant in earthquake or ESS triggering. Annual barometric pressure change would be less than 1 kPa at the depths (25–45 km) we are investigating. Other potential oscillatory loading sources, such as tidal and nontidal ocean loading, contribute most at coastal sites such as those around Vancouver Island, but still less than that of the solid Earth tides, and can be ignored (Scherneck, 1991).

The occurrence of seismic tremor associated with ESS suggests active involvement of fluids that may be released in slab dehydration reactions (Rogers and Dragert, 2003; Szeliga *et al.*, 2004). Presence of pressurized fluids at the transition zone may lubricate the plate interface and reduce the triggering threshold substantially. Gao *et al.* (2000) discovered that triggered miscoseismicity in hydrothermal or volcanic regions in the western United States was modulated by an annual cycle within 5 years following the 1992 Landers earthquake. They interpreted the periodic seismicity fluctuation as caused by barometric pressure change, which has an annual magnitude of only about 2 kPa. No such annual cycle, however, was found for microearthquakes that occurred in other regions without hydrothermal or volcanic origin, suggesting that presence of a fluid phase in the crust could make a difference in earthquake triggering. Another study by Custodio *et al.* (2003) discovered strong M_2 Earth tide modulation of volcanic tremors at Fogo Volcano, Cape Verde, again suggesting the exceptional sensitivity of triggering in the presence of geophysical fluids. Neuberg (2000) also reported numerous cases of strong correlation between seismo-volcanic activities and periodic loading forces such as the diurnal and semidiurnal tides and barometric pressure change in Italy, New Zealand, and Indonesia. Along the Nankai subduction zones, Kodaira *et al.* (2004) discovered a high Poisson's ratio at the subducting plate interface based on seismic V_p/V_s imaging, and suggested that presence of fluid might lower the triggering threshold substantially for the occurrence of slow slip events there. Obara (2002) and Seno and Yamasaki (2003) specifically hypothesized that some of the tremor events observed at the plate interface were caused by hydro-bursts.

Dragert *et al.* (2004) reported that seismic tremor events recorded at the northwestern tip of Vancouver Island had a 6-month phase offset from those observed south of that region. Modeling the Coulomb stress change at the plate interface in the north is difficult, because very little is known about the tectonic setting and kinematics of the subduction zone north of the Nootka fault. Secular displacements of GPS in the region are due north with respect to the interior of the North America plate as opposed to the east-northeast-directed motions to the south (Mazzotti *et al.*, 2003). Assuming the same subduction plate geometry as in the south, our preliminary model shows that both the along-strike shear and normal components are about 180° phase offset with, but smaller than, the updip shear component in terms of contribution to the Coulomb failure stress change at the plate interface. Because of oblique subduction in the region, along-strike shear stress may play an important role in modulating the Coulomb stress change at the plate interface. However, precise modeling will have to await detailed information about its tectonic setting and kinematics.

ESS events at the southern Cascadia subduction zone during the past 7 years seem to have a period of ~0.9 years, shorter than that of the Chandler wobble (Szeliga *et al.*, 2004). Therefore the events were not always synchronized with the ascension phase of pole-tide-induced Coulomb stress change, and two of the events occurred during the descending phase (Fig. 2). Other physical processes might take place at the plate interface and contribute to the triggering of the ESS events, which we do not fully understand yet. Despite this, our model is able to interpret most of the observations.

In summary the pole-tide-induced Coulomb stress change seems to be a viable source capable of modulating the recurrence of ESS events. After the accumulation of tens or hundreds of such events, one ESS event may eventually trigger a megathrust earthquake that ruptures the entire subduction zone (Ozawa *et al.*, 2002, 2004; Larson *et al.*, 2004; Satake *et al.*, 2003). Modeling studies (e.g., Shibazaki and Iio, 2003; Yoshida and Kato, 2003; Liu and Rice, 2005) of stress transfer at subduction zones, although preliminary, have found ways to "make" slow slip events which lead to the eventual failure and the mega-thrust earthquake. Our work presented here may provide a viable scenario for future theoretical and modeling studies to investigate.

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