An earthquake-like magnitude-frequency distribution of slow slip in northern Cascadia

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[1] Major episodic tremor and slip (ETS) events with M_w 6.4 to 6.7 repeat every 15 ± 2 months within the Cascadia subduction zone under the Olympic Peninsula. Although these major ETS events are observed to release strain, smaller "tremor swarms" without detectable geodetic deformation are more frequent. An automatic search from 2006–2009 reveals 20,000 five-minute windows containing tremor which cluster in space and time into 96 tremor swarms. The 93 inter-ETS tremor swarms account for 45% of the total duration of tremor detection during the last three ETS cycles. The number of tremor swarms, N, exceeding duration τ follow a power-law distribution $N \propto \tau^{-0.66}$. If duration is proportional to moment release, the slip inferred from these swarms follows a standard Gutenberg-Richter logarithmic frequency-magnitude relation, with the major ETS events and smaller inter-ETS swarms lying on the same trend. This relationship implies that 1) inter-ETS slip is fundamentally similar to the major events, just smaller and more frequent; and 2) despite fundamental differences in moment-duration scaling, the slow slip magnitude-frequency distribution is the same as normal earthquakes with a b-value of 1. **Citation:** Wech, A. G., K. C. Creager, H. Houston, and J. E. Vidale (2010), An earthquake-like magnitude-frequency distribution of slow slip in northern Cascadia, *Geophys. Res. Lett.*, *37*, L22310, doi:10.1029/2010GL044881.

1. Introduction

[2] The recent discovery of episodic tremor and slip (ETS) and several other slow slip phenomena challenges our understanding of how earthquakes behave. For equivalent fault slip, these slow events take much longer to rupture and suggest a fundamental difference in the relationship between earthquake size and duration from normal earthquakes. With regular earthquakes, the seismic moment M_o, which is proportional to fault area times average fault displacement, scales with the slip duration τ as

$$M_o \sim \tau^3$$

However, for this new branch of slow earthquakes including ETS, the seismic moment is observed to be directly proportional to duration [*Ide et al.*, 2007b]

$$M_o \sim \tau$$

For magnitudes bigger than about 2, the duration is much larger than an ordinary earthquake with a similar moment [*Ide et al.*, 2007b].

[3] Comprising this new category of slow earthquakes are different types of events that span the duration spectrum from discrete, 1 s low-frequency earthquakes (LFE) [*Shelly et al.*, 2006] and 20–200 s very- and ultra-low-frequency (VLF and ULF) seismic events [*Ide et al.*, 2008; *Ito and Obara*, 2006] to geodetically-observed slow slip events (SSE) that can last anywhere from days to years [*Schwartz and Rokosky*, 2007]. ETS in northern Cascadia falls near the larger, longer end of observed slow phenomena, with seismically-observed non-volcanic tremor lasting hours to weeks, the longest of which coincide with geodetically-observed slow slip lasting days to weeks [*Rogers and Dragert*, 2003].

[4] In this paper we define a new class of events, "tremor swarms" by searching for tremor that clusters in both space and time. The largest tremor swarms are the well-studied ETS events that are observed both from tremor and slow slip. Some intermediate-sized ETS events have been detected by measuring tilt in Japan [*Obara et al.*, 2004] and strain in Cascadia [*Dragert et al.*, 2001], but here we detect even smaller tremor swarms. These smaller swarms do not yet correspond to geodetically observed slip, but are interpreted to fall below current geodetic resolution. Using each swarm's duration as a measure of seismic moment, we find that the slow slip inferred from this new class of tremor swarms has a frequency-magnitude distribution similar to that of regular earthquakes.

2. Data Analysis

[5] Using waveform envelope correlation and clustering (WECC) analysis [*Wech and Creager*, 2008], we automatically searched for non-volcanic tremor in northern Cascadia during the continuous four-year interval 2006–2009. WECC attempts to locate tremor in every 50%-overlapping 5-minute time window. Vertical-component seismograms are band-passed filtered at 1.5–5 Hz, rectified by calculating envelope functions, then low pass filtered at 0.1 Hz. We determine the source location that maximizes a weighted sum of cross-correlations over all station pairs evaluated at the lag times of predicted differential S-wave travel times. Resulting epicenters are kept if boot-strap error estimates are less than 5 km and there is another epicenter with 0.1 deg during that day. Of the resulting 22,000 locations, we keep the 20,000 that have at least 7 other epicenters within 30 km and 1.5 days. The 10% of epicenters that are discarded by this procedure removes 80% of geographic outliers. The remaining 20,000 epicenters naturally cluster into 96 tremor swarms that are separated temporally or latitudinally by gaps

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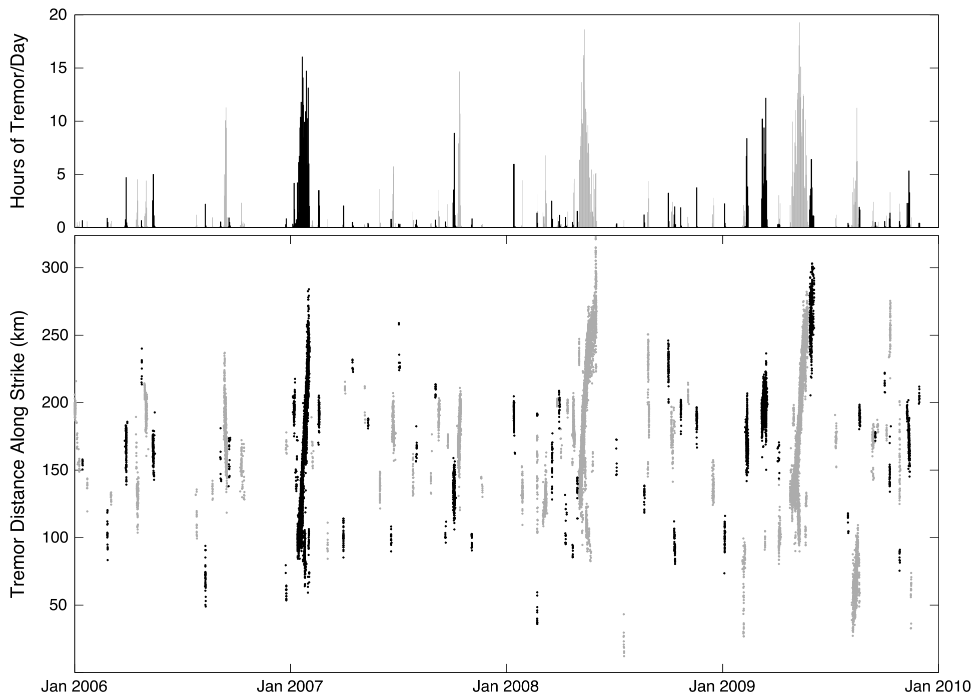

Figure 1. (top) Hours of tremor per day and (bottom) date versus tremor distance along strike during the four-year period 2006 through 2009. The 20,000 tremor epicenters cluster in time and space into 96 tremor swarms ranging in duration from less than one hour to the three ETS events containing more than 200 hours each. Tremor swarms are displayed alternately as gray and black to elucidate the results of our automatic clustering algorithm.

exceeding 1.5 days or 30 km respectively. Using these parameters, WECC identifies tremor swarms that span durations from 0.7 to nearly 300 hours, detecting both ETS and inter-ETS activity. This approach is affected by cultural noise that decreases detection during daytime hours such that the 12-hour period centered on local midnight provides 2/3 of the detections. However, this day/night distribution is nearly identical for the largest (ETS) swarms, intermediate-size swarms and the smaller (<20 hour) swarms so it should not significantly bias the statistics presented in this paper.

[6] Although we have started to detect tremor Cascadia-wide [*Wech*, 2010], we limit our tremor catalog here to northern Washington because it is where we have the most homogeneous tremor record, which is critical for frequency magnitude distribution studies. In this region, ETS events repeat every 15 ± 2 months [*Miller et al.*, 2002; *Rogers and Dragert*, 2003]—except for May 2009—and are remarkably similar in duration, area, and moment. For each of these events, the tremor pattern is in strong agreement with geodetic slip patterns [*Wech et al.*, 2009].

[7] We identify 96 distinct tremor swarms from 2006 through 2009, including the three ETS events in January 2007, May 2008 and May 2009 (Figure 1). Inter-ETS tremor was detected in ~9000 windows, some of which overlap by 50%, so inter-ETS tremor was seen 2% of the time. The number of hours of tremor per smaller swarm ranged from about 1 to 68, totaling 746 hours. These smaller tremor swarms generally locate along the downdip side of the major ETS events [*Wech et al.*, 2009], and account for approximately 45% of the time that tremor has been detected during the last three entire ETS cycles [*Wech and Creager*, 2008; *Wech et al.*, 2009]. Also, like the major ETS tremor swarms, many of the smaller swarms are similar in duration, spatial extent and propagation direction (Figure 1).

[8] We interpret the smaller tremor swarms that comprise this distribution to be associated with transient slow slip that falls below geodetic resolution. This interpretation hinges on the relationship between tremor and slip of ETS swarms. Despite being observed separately and on opposite ends of the frequency spectrum, the spatio-temporal coincidence of tremor and slip [*Rogers and Dragert*, 2003; *Wech et al.*, 2009] strongly suggests a causal connection. Evidence from several ETS observations indicates the two phenomena are different observations of the same shear process along the plate interface. Geodetic inversions estimate 2–3 cm of slip on the plate boundary every 15 months [*Dragert*, 2007; *Dragert et al.*, 2001; *Szeliga et al.*, 2004]. Also supporting this model are seismic observations from LFEs comprising tremor in Japan [*Ide et al.*, 2007a; *Shelly et al.*, 2006] and polarization analysis in Cascadia [*Wech and Creager*, 2007] showing a tremor mechanism of reverse-thrusting consistent with slip on the subduction zone plate interface. Tremor-comprising LFEs in both Japan [*Shelly et al.*, 2006] and Cascadia [*Brown et al.*, 2009] as well as S-P wave arrival time differentials in Cascadia [*La Rocca et al.*, 2009] also indicate that tremor occurs on the plate interface, congruent with slow slip. Constraining tremor depths is difficult and perhaps tremor does not always occur on the plate boundary and represents slip or other fluid-related deformation 10 s of km above the plate boundary [*Kao et al.*, 2005; *McCausland et al.*, 2005]. However, mounting evidence suggests that tremor represents slow shear, serving as a good proxy for slow slip. And, consistent with the moment-duration relationship of the other slow shear phenomena [*Ide et al.*, 2007b], its total duration is proportional to the slow slip

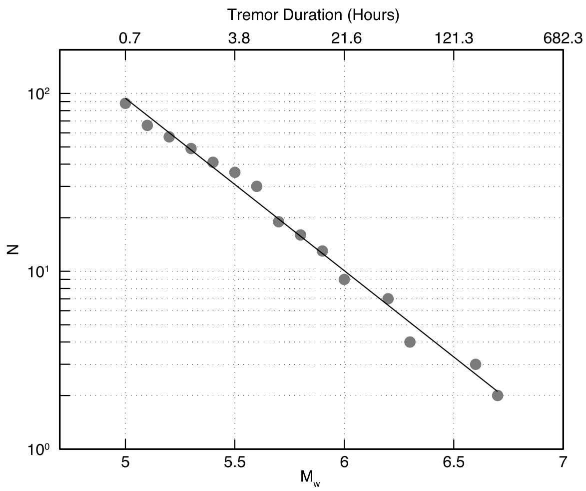

Figure 2. Log of number (N) of tremor swarms exceeding durations given on upper axis can be fit with a straight line indicating that N is proportional to $\tau^{-0.66}$ where τ is the duration of a tremor swarm. We assume that the seismic moment is proportional to tremor duration scaled by M_o (N-m) = $5.2 \times 10^{16}\ \tau$ (hrs) [*Aguiar et al.*, 2009] to equate duration (upper axis) to moment magnitude (lower axis). This allows a standard Gutenburg-Richter style analysis for slow slip events and produces a b-value of 1.0 ($\log N = a - bM$), which is within the range commonly seen for regular earthquakes.

seismic moment [*Aguiar et al.*, 2009]. Geodesy's resolution is limited to surface deformation of 2 mm. Slow slip events at this lower limit equate to $M_w \sim 6.3$ with around 70 hours of associated tremor [*Aguiar et al.*, 2009]. Therefore, although only the largest tremor swarms presented in this paper have corresponding geodetic observations, we interpret each swarm to represent transient shear slip on the plate interface, consider their duration as a measure of the total seismic moment of a slow slip event, and use the terms "swarm" and "slow slip" interchangeably throughout this paper.

3. Results

[9] We find that 20,000 5-minute windows of tremor during a four year period cluster in space and time into 96 tremor swarms. These swarms follow a power-law relationship such that the number of swarms, N, exceeding duration τ (hr) is given by $N \sim \tau^{-0.66}$. If we assume that seismic moment of slow events is proportional to τ as proposed by *Ide et al.* [2007b] and observed in northern Cascadia by *Aguiar et al.* [2009], we find that these slow slip events follow a standard Gutenberg-Richter (GR) law [*Gutenberg and Richter*, 1954] (Figure 2)

$$\log N = a - bM$$

where N is the number of swarms greater than magnitude M, a is a constant and b, known as the b-value, describes the ratio of large to small events.

[10] Using duration as a measure for seismic moment, scaled by M_o (N-m) = $5.2 \times 10^{16}\ \tau$ (hr) [*Aguiar et al.*, 2009], we obtain magnitudes for each event [*Hanks and Kanamori*, 1979]

$$M_w = \frac{2}{3}[\log M_o - 9.05]$$

and find that the moment-magnitude of slip inferred from these tremor swarms follow a GR frequency-magnitude relation [*Gutenberg and Richter*, 1944], $N \sim 10^{-bM_w}$ with a b-value of 1.0 (Figure 2), which lies in the range for normal earthquake catalogs. A b-value of 1.0 translates to a factor of 10 increase in tremor swarm activity with each incremental decrease in magnitude. Global observations of earthquake magnitudes indicate that $b \sim 1$ [*Frohlich and Davis*, 1993] meaning that, despite releasing seismic moment on different time scales, slow slip events follow the same magnitude-frequency pattern as typical earthquakes. We find that the moment-frequency distribution is best fit by a power law, but this distribution describes the macroscopic occurrence of slow slip and should not be confused with the observed exponential duration-amplitude distribution of individual bursts within the tremor [*Watanabe et al.*, 2007].

[11] Additionally, if we consider the three major ETS events in this study period as large tremor swarms, we find that they fall on the same trend as the inter-ETS swarms (Figure 2), suggesting that the inter-ETS swarms are just smaller, more frequent, down-dip versions of the major 15-month ETS events (A. G. Wech and K. C. Creager, A continuum of stress, strength and slip in the Cascadia transition zone, manuscript in preparation, 2010). Only the largest events coincide with geodetically observed slip, meaning that current geodetic observations may be missing nearly half of the total slip if tremor moment rates are constant across our 96 swarms.

[12] To check the robustness of these procedures, we have applied the clustering algorithm with a variety of clustering parameters. For example, if we require 5, 7, or 9 tremors to be within 1.5 days and 30 km we obtain 114, 96 and 78 tremor swarms respectively and the resulting b-values are 0.99, 0.99 and 0.96. If we require 7 tremors to be within 1.0, 1.5 or 2.0 days as well as 20, 30, or 40 km we obtain 99, 96 and 96 tremor swarms respectively and b-values of 1.07, 0.99 and 0.99 respectively. In each case, the moment-frequency distribution is well fit by a power law and the resulting b-value ranges from 1.0 To 1.1.

4. Interpretation

[13] Standard earthquakes have long been known to follow the GR frequency magnitude distribution with $b \sim 1$ [*Frohlich and Davis*, 1993; *Kagan*, 1999] although the physical basis for it is not entirely understood. However, there have been studies observing differences in b-values with tectonic setting [*Schorlemmer et al.*, 2005; *Wiemer and Wyss*, 2002]. In fact, b-values have even been suggested to vary along different portions of a particular fault [*Ghosh et al.*, 2008], or even change with time within the same region of a fault [*Wiemer and Wyss*, 2002]. Sometimes inferred to relate to the local stress regime [*Schorlemmer et al.*, 2005], b-values therefore may provide a means for comparing the state of stress among different faults as well as serving as an indicator of stress variations in space

and time along a fault. With more extensive monitoring, perhaps spatial and temporal variations in b-values can be observed in tremor swarms leading to further insight into stress variations.

[14] In addition to the GR law, another important earthquake scaling relation is Bath's law, which concerns aftershocks. Bath's law states that the magnitude of the largest aftershock in a mainshock-aftershock sequence is, on average, about 1.2 magnitude units smaller than the mainshock [*Bath*, 1965]. While the GR character of the tremor swarms is reminiscent of aftershock behavior, the relationship between ETS and tremor swarms differs from that of mainshock-aftershock sequences in at least two respects. First, the size distribution does not follow Bath's law. Instead, the largest non-ETS swarms are only 0.5 magnitude units smaller than the ETS. Bath's law has been explained in terms of the difference between the process of mainshock rupture, which is highly organized and strongly driven in space and time by dynamically-propagating stress waves, and aftershock sequences which span a greater duration, and are not dynamically driven [*Kagan and Houston*, 2005]. Thus, the failure of slow slip processes to follow Bath's law is consistent with the absence of dynamic stressing during the major ETS (i.e., the "mainshocks"), due to their notoriously slow propagation velocities. Second, the swarms concentrate in time before, rather than after, the main ETS, suggesting that they track the buildup of stress over the 15-month cycle. So we might view non-ETS swarms and ETS as a sequence, much like foreshocks, mainshocks and aftershocks, but with a very different evolution and magnitude distribution.

[15] Though the actual scaling factor used [*Aguiar et al.*, 2009] between M_o and τ doesn't affect b, the b-value strongly depends on the relationship between M_o and τ. Considering $M_o \sim \tau^{\gamma}$ (where γ = 3 for normal earthquakes and is postulated to be 1 for slow slip) we can substitute this dependence into the GR relation to find that $N \sim \tau^{-\frac{2}{3}b\gamma}$. Our most direct observations is that $N \sim \tau^{-\mathrm{p}}$ where p = 0.66, so we have the general relation: b = 1.5 p/γ, which for p = 0.66 implies that b is approximately 1.0/γ. If these power-law relations both hold for a wide range of moments, then we can use these to extrapolate how much moment is being released by events that are too small to observe. The integrated moment of all events smaller than M_w is proportional to $10^{\left(\frac{3}{2}-b\right)M_w}$. So for p observed to be 0.66 and γ assumed to be 1, we get a b-value of 1.0 and the total moment released by all events smaller than, say 6.7 (the biggest we observe) is 10 times greater than the total moment released by all events smaller than 4.7 (the smallest we observe). This implies that shorter duration events, though quite numerous, only account for 10% of the total moment, and we are observing the vast majority of it. Of course it is possible that larger, less frequent, slow slip events will occur in the future. If γ is bigger than 1, then b will be smaller than 1 and we would be missing an even smaller percentage of the total moment. For example if γ = 2, b = 0.5, and we would be missing only 1% of the total moment.

5. Conclusions

[16] The observation that tremor swarms follow a GR frequency magnitude distribution has significant implications. First, this discovery provides seismologists a new tool with the potential for investigating stress conditions on faults. Tremor b-values could be used as a metric for comparing varying stress of different tremoring regions as well as comparing both temporal and spatial variations of individual faults. Second, this frequency magnitude distribution indicates that, while slow slip events have different underlying physics, they relieve accumulated stress like normal earthquakes. As seen with typical earthquakes, an incremental decrease in magnitude coincides with a 10-fold increase in tremor swarm activity. Despite being slow, the occurrence and amount of stress released by this new type of earthquake is familiarly predictable. Third, this distribution contributes circumstantial evidence for the $M_o \sim \tau$ scaling relationship of slow events. It is unclear why slow earthquakes behave this way, and we have a limited understanding of why normal earthquakes have a b-value of 1. Nevertheless, if we assume this moment-duration scaling for tremor swarms, we find that slow events have the same magnitude-frequency distribution as regular earthquakes. This observation is consistent with a model in which earthquakes and slow events can be thought of as different manifestations of the same phenomena [*Ide et al.*, 2007b].

[17] Ultimately, the b-value coincidence of normal and slow events is an observational link between the two separate processes. This link may lead to a better grasp of the difference in the underlying processes between slow shear events and typical earthquakes in addition to providing new evidence to guide our understanding of the physics of magnitude-frequency distributions.

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