Southern Cascadia episodic slow earthquakes

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Received 6 April 2004; revised 19 July 2004; accepted 27 July 2004; published 24 August 2004.

[1] Continuous GPS and seismic data from northern California show that slow earthquakes periodically rupture the Gorda-North America plate interface within southern Cascadia. On average, these creep events have occurred every 10.9 ± 1.2 months since at least 1998. Appearing as week-long GPS extensional transients that reverse secular forearc contraction, the data show a recurrence interval 22% shorter than slow events recognized to the north. Seismic tremor here accompanies the GPS reversals, correlated across as many as 5 northern California seismometers. Tremor occurs sporadically throughout the year, but increases in duration and intensity by a factor of about 10 simultaneous with the GPS reversals. Beneath west-central Oregon, three reversals are also apparent, but more stations are needed to confirm sporadic slip on the plate interface here. Together, these measurements suggest that slow earthquakes likely occur throughout the Cascadia subduction zone and add further evidence for the role of fault-fluid migration in controlling transient slow-slip events here. *INDEX TERMS:* 1206 Geodesy and Gravity: Crustal movements—interplate (8155); 1243 Geodesy and Gravity: Space geodetic surveys; 7230 Seismology: Seismicity and seismotectonics. **Citation:** Szeliga, W., T. I. Melbourne, M. M. Miller, and V. M. Santillan (2004), Southern Cascadia episodic slow earthquakes, *Geophys. Res. Lett.*, *31*, L16602, doi:10.1029/2004GL020824.

1. Introduction

[2] Slow faulting events recently recognized along convergent margins globally are now understood to constitute a fundamental mode of moment release that both trigger and are triggered by regular earthquakes [*Dragert et al.*, 2001; *Heki et al.*, 1997; *Hirose et al.*, 1999; *Kawasaki et al.*, 1995; *Kostoglodov et al.*, 2003; *Larson et al.*, 2004; *Linde and Silver*, 1989; *Lowry et al.*, 2001; *Miller et al.*, 2002; *Obara*, 2002; *Ozawa et al.*, 2002; *Rogers and Dragert*, 2003; *Sagiya and Ozawa*, 2002]. In the Pacific Northwest, continuous GPS has detected nine slow earthquakes occurring at 13.9 ± 0.9 month intervals within the northern Cascadia plate interface [*Dragert et al.*, 2001; *Miller et al.*, 2002] accompanied by harmonic tremor largely absent when slow earthquakes are not occurring [*Obara*, 2002; *Rogers and Dragert*, 2003]. To date, no observations of Cascadia transients, also called slow earthquakes, silent earthquakes, or episodic tremor and slip events, have been made outside of the northern Puget basin, suggesting either that the unique bend in the Juan de Fuca plate here is somehow conducive to slow slip or that instrument density is insufficient outside this region for confident detection. Since slow earthquakes may modulate seismogenic rupture either by reducing the size of a future earthquake, delaying its recurrence, or acting as a trigger, along-strike variability in the existence of slow faulting yields important clues about partitioning, particularly seismogenic segmentation, of the Cascadia subduction zone. In this report we present continuous GPS and seismic data from northern California and Oregon that indicates periodic slow earthquakes occur throughout Cascadia, and with quite variable recurrence rates.

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2. GPS Data

[3] Continuous GPS data from the Pacific Northwest Geodetic Array and the Bay Area Regional Deformation Array [*Miller et al.*, 2001; *Murray et al.*, 1998] were processed with the Gipsy-Oasis II software [*Lichten and Border*, 1987] (Figure 1). Precise point positioning and precise orbits and clocks were used to analyze the phase data with ambiguity resolution applied [*Heflin et al.*, 1992; *Zumberge et al.*, 1997]. Daily solutions for station positions and corresponding matrices of the covariance among the three position components were determined within the International Terrestrial Reference Frame (ITRF 2000) [*Altamimi et al.*, 2002] using daily frame data products provided by the International Geodynamics Service [*Zumberge et al.*, 1997]. A regional stabilization was subsequently applied to each daily position, using a reference set of 42 stations from the North America plate region; 23 of these are concentrated in the Pacific Northwest, the remainder are distributed on the stable plate interior or in other regional networks in western North America. Of the 42 stations, 33 have published positions and velocities in ITRF 2000. This stabilization transformation minimizes network-wide position discrepancies, or common-mode errors. Final time series were simultaneously detrended and corrected for known artifacts that include offsets due to hardware upgrades, earthquakes, and annual and semi-annual sinusoidal signals introduced by mismodeled tropospheric delays and other seasonal effects [*Blewitt and Lavallée*, 2002; *Nikolaidis*, 2002].

3. Seismic Tremor Data

[4] Continuous horizontal component 100-hz seismic data from Guralp 40T's and 20-hz seismic data from a combination of both STS-1's and STS-2's spanning four years from 2000 through 2003 were downloaded from the Northern California Seismic Network (Figure 1). Eight stations are available with minor outages that together span northernmost California, with average spacing of 136 km. Four stations lie within 150 km of the trench, two of which (YBH and WDC) are among the quietest of stations in northern California based on our examination of four years

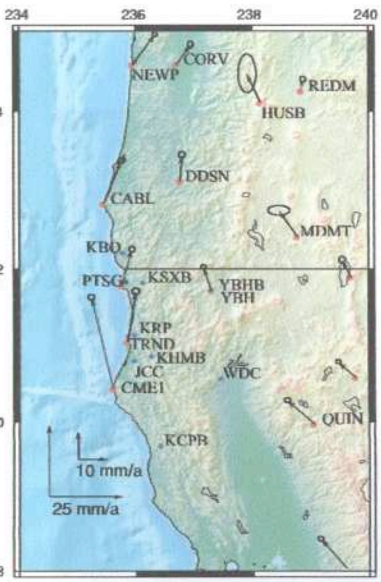

Figure 1. Topographic/bathymetric map of Northern California. Red circles represent continuous GPS stations, blue diamonds represent seismic stations used in this study. Note the sparse distribution of continuous GPS stations in Northern California and Coastal Oregon. Vectors represent motion of each station relative to stable North America. Note the northwestward movement of station YBHB. This is due to a summation of east- west oriented compression from subduction, westerly compression from Basin and Range expansion, and northwesterly translation of the Sierra Nevada microplate. During slow earthquakes, fault fluid migration along the plate interface allows the upper plate (North America) to move west. This is demonstrated in Figure 3a where westerly resets are observed at station YBHB.

of continuous seismic data from all available stations. Four additional stations lie sufficiently west and south of where tremor is expected to be visible and can be used to assess background noise when picking tremor. Due to the distances between instruments, signals correlated across stations must have their genesis in deep-earth processes and cannot be attributed to anthropogenic, meteorologic or other local noise sources. Tremor signals are readily correlated by eye (Figure 2), and their spectra show predominant frequencies in the 1–5 Hz band, similar to that reported in Japan and northern Cascadia [*Obara*, 2002; *Rogers and Dragert*, 2003]. All seismic data in this study were band-passed between 1 and 5 Hz frequencies and gain-normalized to enhance tremor identification. The data record a multitude of signals that include local non-tectonic noise, teleseismic and local earthquakes; tremor signals are distinguished by waveform and coda correlation across adjacent stations. However, due to the emergent nature of the signal [*Rogers and Dragert*, 2003], and the lack of accurately identifiable phases, constraining event onset time with the precision required to determine source depth and location becomes highly assumption-dependent and was not performed in this study.

[5] Identification of tremor entailed plotting all gain-removed, horizontal seismic traces in spatial and temporal proximity, similar to historical drum recordings. Tremor was then identified as signals correlated both temporally and spatially across at least three stations. Periods during which no correlated tremor is evident have background seismic velocities typically less than 0.07 micrometers per second. We therefore summed the rate of visibly correlated tremor whose maximum velocities exceed 0.5 micrometers per second, or roughly 10-times background noise. Figure 2 shows a typical example, approximately 21 minutes of tremor recorded on 5 seismic stations. This window was taken from a much longer burst recorded on 12/10/2002, two days after the onset of transient westward movement of the GPS station YBHB that began on 12/08/2002. During the time of this GPS reversal, correlated tremor activity increased to approximately 90 hours per week.

4. Northern California and Central Oregon Transients

[6] Purely from the standpoint of deformation, westerly resets at YBHB are expected for slow earthquakes along the deeper Gorda-North American plate interface. Surface deformation from such events results from a sum of contraction from shallow plate locking and extension from the slow faulting itself. Since secular deformation in

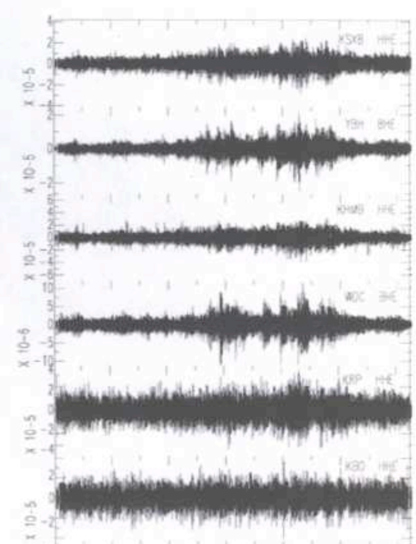

Figure 2. Approximately 20 minutes of tremor recorded on stations from the Northern California seismic network. Vertical axis is in cm/s and horizontal axis is in hundreds of seconds after 09:00 UTM on 12/10/2002. Note the overall waveform correlation between the top 5 seismic stations. The bottom 3 seismic stations are located on the coast and do not show evidence of tremor. Obvious digitization errors in the form of step functions were manually station KHMB were manually removed.

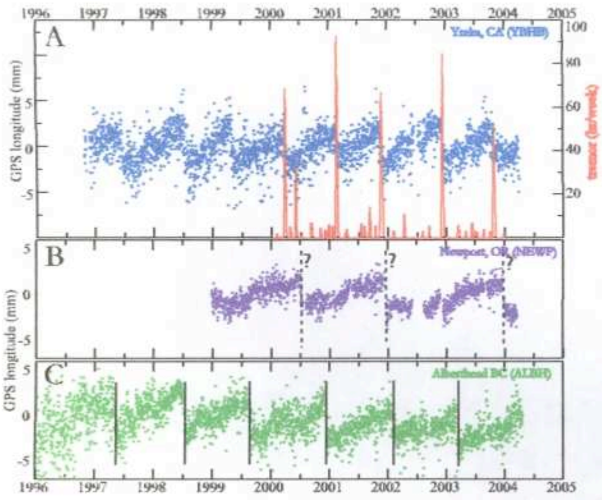

Figure 3. GPS eastings from Yreka, CA, Newport, OR and Alberthead, BC, and seismic tremor histogram from Yreka. a.) Blue points are daily GPS station positions in mm of the longitudinal component of station YBHB. Solid red line is a plot of the hours of tremor per week at seismic station YBH. Note the similarity of shape displayed by ALBH (Figure 3c) and YBHB. The correlation between GPS offsets and increased tremor activity indicates that slow faulting occurs beneath Northern California. b.) Purple points represent daily solutions of station position for the longitudinal component of GPS station NEWP from Newport, Oregon. Note the similarity of NEWP offsets (dashed black lines) to those at ALBH. The lack of seismic and continuous GPS stations near NEWP precludes the definitive identification of slow earthquakes here at the present time. c.) Green points represent daily position solutions of the longitudinal component of ALBH. Note the characteristic sawtooth reset shape of the timeseries due to slow faulting events. For correlation between increased tremor and GPS offsets at station ALBH, see *Rogers and Dragert* [2003]. Solid black lines denote times of known slow earthquakes at ALBH.

southern Cascadia is influenced by roughly east- northeast directed contraction, westerly oriented Basin and Range extension, and northwesterly translation due to Sierra Nevada block and Pacific plate entrainment, slow transient thrust faulting should appear as westerly jumps seen predominantly in the longitude, as is the case. Since it is thought that slow earthquakes result from fault fluid migration along the subduction interface, this lubrication acts to relieve the east- northeast directed contraction caused by subduction of the Gorda Plate, thus the resets seen at YBHB should be and are opposite to the direction of subduction. Figure 3a shows GPS residuals at station YBHB demonstrating periodic resets. For comparison, longitude resets from Alberthead, British Columbia (ALBH), the time series from which episodic slow Cascadia earthquakes were first identified, are shown in Figure 3c. Residuals from YBHB in northern California show similar characteristics as ALBH, particularly westerly jumps of up to 4 mm occurring at 1997.46, 1998.52, 1999.30, 2000.24, 2001.12, 2001.90, 2002.93 and 2003.81. The amplitudes are similar to those at ALBH, but the "interseismic" interval is significantly shorter: 10.9 ± 1.2 months as opposed to 13.9 ± 0.9 month. By contrast, time series from nearby stations TRND, CME1, PTSG and MDMT (Figure 1) show no such resets, indicating the observed resets are not reference-frame artifacts.

[7] Transient slow faulting in northern Cascadia was recognized primarily from deformation reversals correlated across nearby continuous GPS stations, but the GPS instrument density in northern California is currently insufficient for any similar correlation. The nearest continuous GPS station to YBHB lies on the coast (PTSG) at a distance of 120 km; by comparison, there are seven stations within 60 km of each other in the northern Puget basin. Nonetheless, Figure 3a shows the longitude component of YBHB overlying a histogram of hours of correlated tremor from a nearby seismic station (YBH). The remarkable correlation between tremor rate and GPS deformation reversals is readily apparent and confirms that slow earthquakes occur beneath northern California. Although background tremor here is detected during many weeks of the year when no GPS reversals are evident, the rate of tremor increases by an order of magnitude during GPS reversals.

[8] Coastal Oregon also shows preliminary evidence of westerly resets at station NEWP, located in Newport, Oregon. These reversals have similar amplitudes to those at YBHB and northern Puget stations, but do not yet show periodic behavior. NEWP shows three resets in longitude, at 2000.52, 2001.98 and 2003.99. These offsets are not observed at the GPS station CORV located 60 km inland in Corvallis, Oregon. The absence of offsets at station CORV is consistent with relatively narrow, offshore locked and transition zones at this latitude, also suggested from vertical deformation rates [*Mitchell et al.*, 1994]. Thus, CORV may lie well east of the down-dip edge of the transition zone where slow earthquakes occur. At the present time, however, the dearth of GPS or seismic data close to NEWP precludes determination of spatially coherent events.

5. Discussion

[9] The northern California data demonstrate that slow Cascadia earthquakes are not confined to the structural bight in the Juan de Fuca plate beneath the northern Puget Basin, and argue that they occur throughout Cascadia and many other subduction zones. More importantly, these results follow *Obara* [2002] and *Rogers and Dragert* [2003] in linking seismic tremor and slow faulting to one underlying cause, most likely fault fluid transport [*Melbourne and Webb*, 2003]. Analysis of tremor alone for source processes that might constrain such transport is complex, since the lack of discernible phases prohibits discrimination between path and source contributions to tremor coda. For example, delta-function sources, propagated through complex crustal media, have been shown to cause harmonic volcano tremor originally attributed to resonance at the source [*Chouet et al.*, 1987; *Kedar et al.*, 1998; *Koyanagi et al.*, 1987]. Moreover, if Cascadia tremor does indeed result from a harmonic source at depth, a host of distinct driving mechanisms could produce source resonance and identical surface observations, again obfuscating the underlying physics [*Chouet et al.*, 1987; *Koyanagi et al.*, 1987].

If both tremor and slow slip are manifestations of hydraulic transport resonating and unclamping fault walls that sandwich pore fluids, an important next step will be to implement experiments that can constrain near-field (static), non-double couple components of moment release. These, in turn, will likely be of great use in constraining slow earthquake physics at a resolution higher than that afforded by either GPS or tremor.

[10] **Acknowledgments.** This research was supported by National Science Foundation Grant EAR-0208214, US Geological Survey NEHERP award 04HQGR0005, the National Aeronautics and Space Administration grant SENH-0000-0264, and Central Washington University. GPS data collection supported was by the National Science Foundation under Grants No. EAR-0318549 to UNAVCO, Inc., EAR-0002066 and EAR-9616540, to CWU; by the National Aeronautics and Space Administration NASA Contracts No. NAG5-13728 and NAG5-7672; and by the U.S. Geological Survey NEHRP Award 04HQAG0007 and Central Washington University. We thank Craig Scrivner (CWU) for discussion and help with software needed for automated downloading and viewing of NCSN seismic data in record section format. We thank the Pacific Northwest Geodetic Array, the Western Canadian Deformation Array operated by the Pacific Geoscience centre for the Geological Survey of Canada, the Scripps Institute of Oceanography, the Northern California Earthquake Data Center (NCEDC), the Northern California Seismic Network, U.S. Geological Survey, Menlo Park and the Berkeley Seismological Laboratory, University of California, Berkeley for GPS and seismic time series.

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Extent and duration of the 2003 Cascadia slow earthquake

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Received 18 October 2004; revised 4 January 2005; accepted 19 January 2005; published 16 February 2005.

[1] Inversion of continuous GPS measurements from the Pacific Northwest show the 2003 Cascadia slow earthquake to be among the largest of ten transients recognized here. Twelve stations bracketing slow slip indicate transient slip propagated bi-directionally from initiation in the southern Puget basin, reaching 300 km along-strike over a period of seven weeks. This event produced, for the first time, resolvable vertical subsidence, and horizontal displacement reaching six mm in southern Washington State. Inverted for non-negative thrust slip, a maximum of 3.8 cm of slip is inferred, centered at 28 km depth near the sharp arch in the subducting Juan de Fuca plate. Nearly all slip lies shallower than 38 km. Inverted slip shows a total moment release equal to $M_w = 6.6$ and a high degree of spatial localization rather than near-uniform slip. This suggests rupture concentrated along asperities holds for slow earthquakes as well as conventional events. **Citation:** Melbourne, T. I., W. M. Szeliga, M. M. Miller, and V. M. Santillan (2005), Extent and duration of the 2003 Cascadia slow earthquake, *Geophys. Res. Lett.*, *32*, L04301, doi:10.1029/2004GL021790.

1. Introduction

[2] Transient creep events in subduction zones, also known as slow or silent earthquakes, or episodic tremor and slip events, often occur periodically with recurrence intervals that range from months to years [*Beavan et al.*, 1983; *Kawasaki et al.*, 1995; *Larson et al.*, 2004; *Linde et al.*, 1996; *Lowry et al.*, 2001; *Ozawa et al.*, 2001; *Sagiya*, 2004; *Sagiya and Ozawa*, 2002]. In Cascadia, ten slow events have been detected with a 13.9 ± 2 month recurrence near the US-Canadian border and six events with a 10.9 ± 1.2 month periodicity beneath northern California [*Szeliga et al.*, 2004]. They are observed with GPS as spatially coherent reversals from secular forearc contraction to transient extension, and more recently with seismic tremor [*Obara*, 2002; *Rogers and Dragert*, 2003]. However, locating tremor hypocenters remains challenging due both to the lack of pickable phases and because its high frequency content (1–5 hz) renders it sensitive to small-scale crustal structures, thus resulting in hypocenters whose accuracy is difficult to assess. The triggers of transient creep thus remain unknown, but have been hypothesized to stem from pore fluid migration producing conduit resonance simultaneous with reducing fault-normal stress [*Julian*, 2002; *Melbourne and Webb*, 2003].

[3] Besides a remarkable periodicity, Cascadia creep events also show characteristic maximum offsets of typically five to eight mm. Whether this is a result of characteristic slip along specific asperities or is instead a purely elastic masking of adjacent rupture patches in subsequent events is an important mechanical constraint still undetermined. Here we invert GPS measurements that constrain slip during the 2003 Cascadia event. The results suggest that slow earthquakes, like conventional ones, have slip that is coarsely distributed along relatively localized asperities.

2. Data

[4] Continuous GPS data from the Pacific Northwest Geodetic Array [*Miller et al.*, 2001] and Western Canada Deformation Arrays [*Dragert and Hyndman*, 1995] (Figure 1) was processed with the Gipsy-Oasis II [*Lichten and Border*, 1987] software utilizing satellite orbit and clock parameters provided by JPL [*Heflin et al.*, 1992]. Point positioning and precise orbits and clocks were used to analyze the phase data with ambiguity resolution applied [*Zumberge et al.*, 1997]. Daily positions and covariance matrices were determined within the ITRF2000 reference frame [*Altamimi et al.*, 2002] using daily frame products also from JPL. A regional stabilization was applied to each daily position using a reference set of 42 stations from the North America plate region, 23 of which are concentrated in the Pacific Northwest and 33 of which have published positions and velocities in ITRF2000. This stabilization minimizes network-wide position discrepancies and common-mode errors but recovers all differential motion of Cascadia relative to stable North America. Final time series were simultaneously detrended and corrected for hardware upgrades, earthquakes, and annual and semi-annual sinusoidal signals caused by mismodeled seasonal effects [*Blewitt and Lavallée*, 2002; *Nikolaidis*, 2002; *Szeliga et al.*, 2004]. Residuals from this estimation are shown in Figure 2.

[5] Identification of creep onset times with GPS is difficult due to the low signal to noise ratio of the measurements. As an alternate to manual event picking, we use the Gaussian wavelet transform to better identify initiation of rupture. This approach employs the fact that succeeding wavelet basis functions are increasingly sensitive to temporal localization of any given signal, unlike the periodic sinusoids of the Fourier transform. Slow faulting at depth, which effectively produces a Heaviside step at the onset of faulting, appears in the wavelet transform as an amplitude spike that pervades the wavelet power spectrum (Figure 2). Faulting initiation is precisely identified from the temporal location of this spike in amplitudes of wavelets with greatest localization. Besides being repeatable and less prone to human or reference-frame biases, the wavelet transform also allows clear discrimination of slow faulting deformation from other transient, non-solid earth signals such as those that arise from colored noise [*Langbein and Johnson*,

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Figure 1. Slow earthquake displacements (red arrows) during the 2003 Cascadia event and interseismic deformation vectors (black) from continuous PANGA and WDCA GPS networks. 12 stations record discernible, transient reversals from NE-directed contraction to WSW-directed extension. The transient event emerged over seven weeks and spanned nearly 300 km along-strike, from the Oregon to Canadian borders of Washington State. Error ellipses are 1σ and variable size reflects time series scatter.

1997]. Furthermore, times picked from the wavelet transform produce a significant reduction in chi-squared misfits in event-offset parameter estimation, at least for short-duration transients lasting less several weeks. Finally, this technique is appealing in that it forms the basis for automated transient detection in large geodetic networks, such as the US Plate Boundary Observatory, where manual picking of ~4000 data channels will not be feasible.

3. 2003 Cascadia Slow Earthquake

[6] Total offsets for the 2003 event (Figures 1 and 2) are sensible in that they suggest a spatially localized but temporally staggered pattern of simultaneous, N-S bi-directional propagation of reversals from contraction to transient extension throughout, but limited to, the northern Cascadia forearc. The first significant departure from secular contraction is recorded simultaneously beneath the southern Puget Basin in late January 2003 on stations SEAT, KTBW, and RPT1 (Figures 1 and 2). Within a time span of less than one month, transient reversals then appear simultaneously to the north (SEDR and WHD1). By mid-February, 2003, about three weeks after its initiation, creep had spread ~200 km north and south, reaching southwestern British Columbia (SCO2, NEAH, ALBH) and southernmost Washington (CPXF, KELS, FTS1, JR01). By March 2003, six weeks after nucleation, the transient is evident on 12 stations. Although its termination is difficult to precisely identify, the data suggest that by mid-March slow slip had terminated along the entire margin. Six ±1 mm of displacement is recovered in the southern Puget basin (CPXF), resolvable extension reaches as far south as the Oregon border, and vertical subsidence of 5 ± 2 mm is visible in the northern Puget basin (SC02).

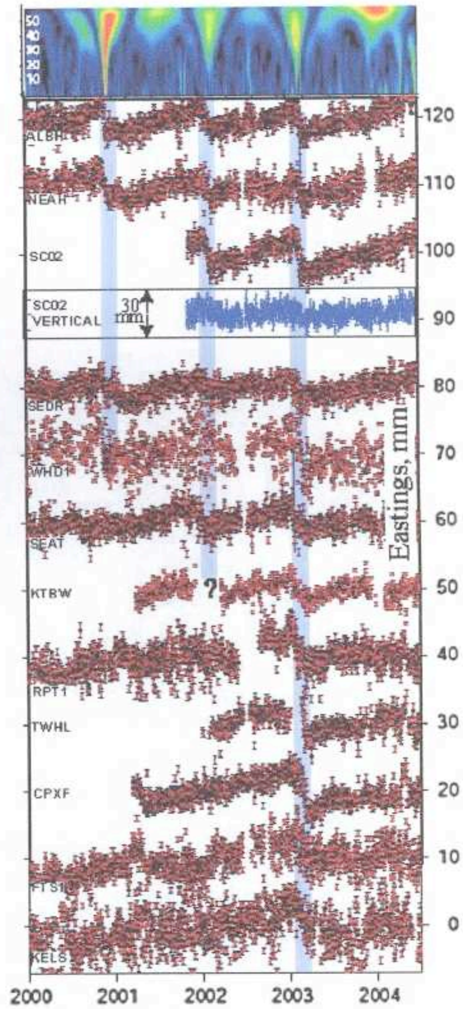

Figure 2. (Bottom) Daily longitude positions record the last three episodic slow earthquakes in Cascadia (thick blue lines). Unlike many previous events, the 2003 event ruptured as far south as the Oregon border. TWHL has irrecoverable data outages at the onset of the event and cannot be used to constrain onset timing at that station. Station SC02 records the first discernible vertical subsidence for slow earthquakes, 5 ± 2 mm during this event (note change of scale). Instruments FTS1, PRT1 and WHD1 are US Coast Guard stations with older antennas mounted on 10-meter towers and have higher intrinsic scatter. (Top) Example of Gaussian-wavelet transform used to pick transient onset times (shown is east component of ALBH, the topmost time series). Y-axis is wavelet scale (temporal extent), X-axis is time, and color denotes relative wavelet coefficient amplitude, with red showing highest amplitudes and blue lowest. Discrete fault slip events produces step-like functions in the geodetic time series that show up equally across all wavelets, providing the basis for automated transient detection and correlation with large geodetic station arrays ($n > 100$).

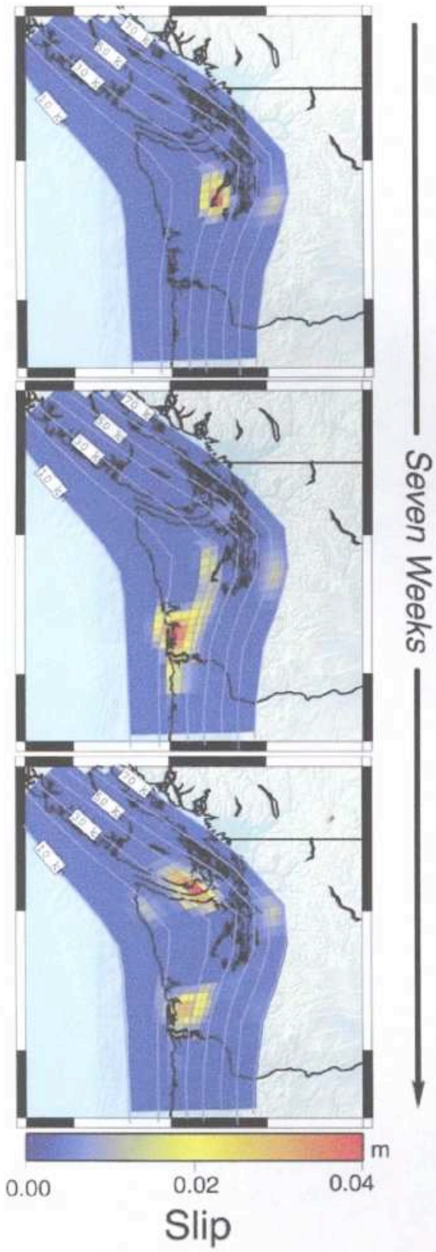

Figure 3. Early 2003 transient creep propagation along the Juan de Fuca- North America plate interface, inverted from geodetic GPS data using non-negative least squares. The event nucleated beneath the southern Puget basin and propagated bi-lateral of ~300 km over six weeks. Slip propagation is estimated at three two-week intervals February 1–14, February 15–28, and March 1–15, 2003. The plate interface is parameterized into roughly 10 × 15 km subfaults, with 10-km depth contours shown with solid lines. Maximum cumulative slip of 3.8 cm occurs at 28 km depth beneath the southern Puget basin, and little slip (less than 15% of maximum) is inferred below 40 km depth. Inversion employs non-negative least squares estimates of thrust slip at the average Juan de Fuca-North America plate convergence direction. Cumulative moment release from beginning to end for this event is $M_w = 6.6$.

[7] The density of stations on which the 2003 slow event was recorded invites a formal inversion of the surface displacement for a variable-slip distribution along the plate interface. We discretized the Juan de Fuca-North American plate interface [*Fluck et al.*, 2000] into 10 × 25-km subfaults along the down-dip and along-strike components, respectively. The plate interface intersects the earth's surface at the geomorphic expression of the offshore deformation front and extends to an absolute depth of 70 km, far below the region of expected faulting. Green's functions for both an elastic half-space and layer-cake were computed using the methodologies of *Okada* [1992] and *Zhu and Rivera* [2002]and were found not to differ significantly for deep (>10 km) sources. A 2nd-order Laplacian smoothing operator is incorporated into the design matrix, following [*Harris and Segall*, 1987], which serves to stabilize the inversion without unduly localizing slip. An optimal smoothing coefficient was derived using a cross-validation method in which single stations were sequentially removed and the remaining data compared with the surface displacements predicted by inversion based on the incomplete data. This procedure is then repeated for each station and for multiple lambda values. The smoothing coefficient which minimized misfit, 3.5e-4, was then adopted for the inversions shown in Figure 3. The design matrix was inverted with QR decomposition constrained to solve for positive thrust slip only [*Lawson and Hanson*, 1995]. Offsets from cleaned time series estimated bi-monthly were inverted for cumulative slip during that time period. Figure 3 shows three bi-weekly time slices starting in early February through mid-March 2003. Transient faulting clearly nucleated below the southern Puget basin, propagated along-strike bi-diagonally from this region, reached maximum slip by mid-March of 2003, and faded in the south prior to the north. A maximum of 3.8 cm of cumulative slip is imaged beneath the southern Puget basin.

4. Discussion

[8] Moment release, which we estimate by summing inverted slip over time, is largely invariant with respect to the details of the slip distribution so long as the inverted slip produces vectors that match the data. The cumulative moment release of this event is equal to $M_w = 6.6$. Among the largest of the Cascadia events (perhaps due to instrumentation), this event is still significantly smaller than other slow events reported elsewhere, for instance in western Mexico ($M_w = 7.5$) [*Lowry et al.*, 2001].

[9] The slip heterogeneity shown in Figure 3 is likely real, in that inversions based on a coarser parameterization of the plate interface fail to fit the data adequately. Moreover, inversion of synthetic time series with the relatively fine subfaults shown in Figure 3 suggests that evenly distributed, wide-spread creep spread over hundreds of km instead of patchy slip localized over several tens of km should be resolved by the 12 stations. Beyond these dimensions the inversions cannot reveal details of slip, which effectively precludes estimating stress drop from surface deformation measurements alone. Tremor studies of slow earthquakes, by contrast, consistently indicate that the deformation observed at the surface likely reflect the summation of elastic strain from a large number of tiny

faulting events clustered in time [*Kao and Shan*, 2004; *Rogers and Dragert*, 2003; *Szeliga et al.*, 2004]. Stress drop of these tiny events therefore will likely constrain their rupture mechanism, and, in their ensemble, of slow earthquakes. As a result, source constraints on these smaller events-deduced from tremor seismicity- will likely prove most fruitful in determining how rupture fronts propagate along the plate interface. Broadband recordings that might document dilatational components of faulting in particular would prove particularly valuable in understanding these new phenomena.

[10] **Acknowledgments.** This research was supported by National Science Foundation Grant EAR-0208214, US Geological Survey NEHERP award 04HQGR0005, the National Aeronautics and Space Administration grant SENH-0000-0264, and Central Washington University. GPS data collection supported was by the National Science Foundation under Grants No. EAR-0318549 to UNAVCO, Inc., EAR-0002066 and EAR-9616540, to CWU; by the National Aeronautics and Space Administration NASA Contracts No. NAG5-13728 and NAG5-7672; and by the U.S. Geological Survey NEHRP Award 04HQAG0007 and Central Washington University. We thank the Pacific Northwest Geodetic Array and the Western Canadian Deformation Array operated by the Pacific Geoscience Centre for the Geological Survey of Canada for use of their data.

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