Extent and duration of the 2003 Cascadia slow earthquake

Timothy I. Melbourne, Walter M. Szeliga, M. Meghan Miller, and V. Marcelo Santillan

Geodesy Laboratory and PANGA Data Analysis Facility, Department of Geological Sciences, Central Washington University, Ellensburg, Washington, USA

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*,

0094-8276/05/2004GL021790$05.00

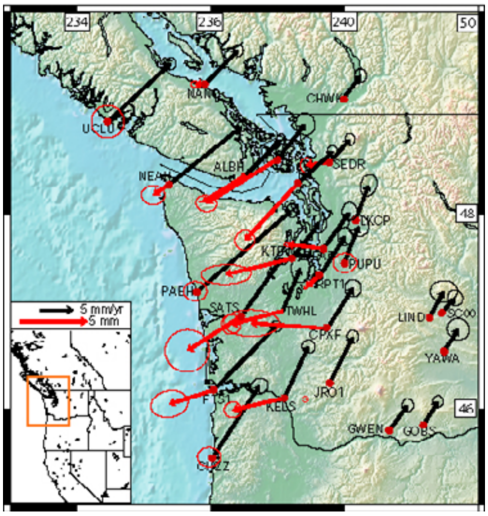

Figure 1. Slow earthquake displacements (red arrows) during the 2003 Cascadia event and interseismic deformation vectors (black) from continuous PANGA and WDCA GPS networks. 12 stations record discernible, transient reversals from NE-directed contraction to WSW-directed extension. The transient event 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).

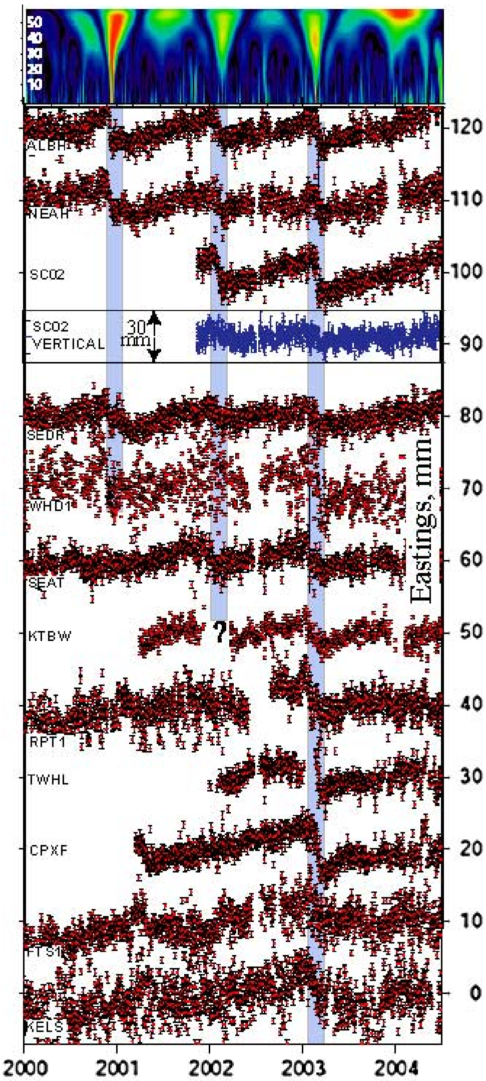

Figure 2. (Bottom) Daily longitude positions record the last three episodic slow earthquakes in Cascadia (thick blue lines). Unlike many previous events, the 2003 event ruptured as far south as the Oregon border. TWHL has irrecoverable data outages at the onset of the event and cannot be used to constrain 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).

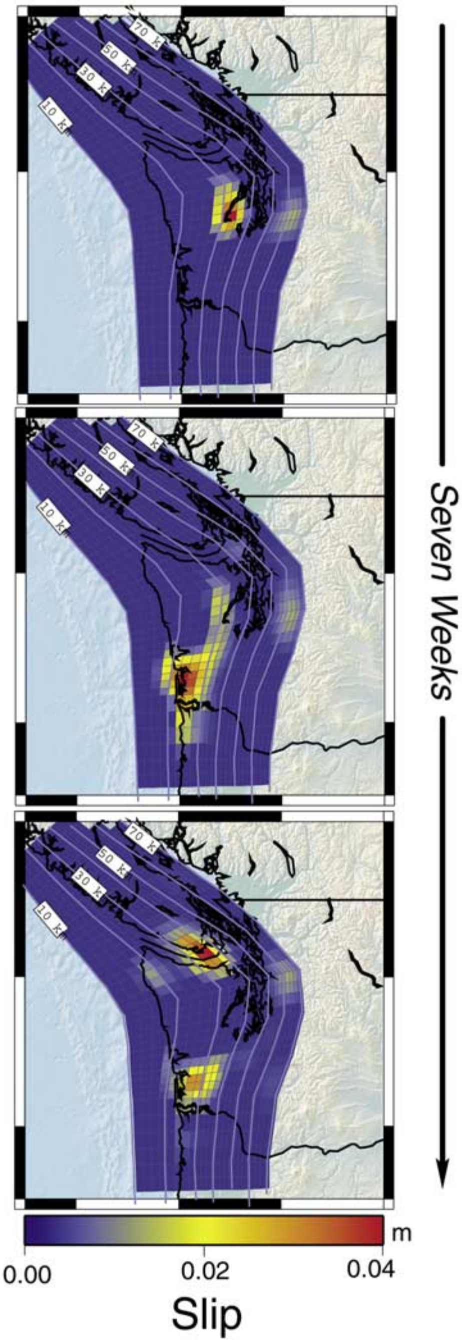

Figure 3. Early 2003 transient creep propagation along the Juan de Fuca- North America plate interface, inverted from geodetic GPS data using non-negative least squares. The event nucleated beneath the southern Puget basin and propagated bi-lateral of ~300 km over six weeks. Slip propagation is estimated 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.

References

Altamimi, Z., P. Sillard, and C. Boucher (2002), ITRF2000: A new release of the International Terrestrial Reference Frame for earth science applications, *J. Geophys. Res.*, *107*(B10), 2214, doi:10.1029/2001JB000561.

Beavan, J., E. Hauksson, S. R. McNutt, R. Bilham, and K. H. Jacob (1983), Tilt and seismicity changes in the Shumagin seismic gap, *Science*, *22*(4621), 322–325.

Blewitt, G., and D. Lavallée (2002), Effect of annual signals on geodetic velocity, *J. Geophys. Res.*, *107*(B7), 2145, doi:10.1029/2001JB000570.

Dragert, H., and R. Hyndman (1995), Continuous GPS monitoring of elastic strain in the northern Cascadia subduction zone, *Geophys. Res. Lett.*, *22*, 755–758.

Fluck, P., R. D. Hyndman, and K. Wang (2000), 3-D dislocation model for great earthquakes of the Cascadia subduction zone, in *Penrose Conference "Great Cascadia Earthquake Tricentennial": Best Western Oceanview Resort, Seaside, Oregon, June 4–8, 2000: Program Summary and Abstracts*, edited by J. J. Clague et al., p. 48, Oregon Dep. of Geol. and Miner. Ind., Portland.

Harris, R. A., and P. Segall (1987), Detection of a locked zone at depth on the Parkfield, California, segment of the San Andreas Fault, *J. Geophys. Res.*, *92*, 7945–7962.

Heflin, M., et al. (1992), Global geodesy using GPS without fiducial sites, *Geophys. Res. Lett.*, *19*, 131–134.

Julian, B. (2002), Seismological detection 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. Int.*, *157*, 589–594.

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

Langbein, J., and H. Johnson (1997), Correlated errors in geodetic time series; implications for time-dependent deformation, *J. Geophys. Res.*, *102*, 591–604.

Larson, K. M., A. R. Lowry, V. Kostoglodov, W. Hutton, O. Sánchez, K. Hudnut, and G. Suárez (2004), Crustal deformation measurements in Guerrero, Mexico, *J. Geophys. Res.*, *109*, B04409, doi:10.1029/2003JB002843.

Lawson, C. L., and R. Hanson (1995), *Classics in Applied Mathematics*, vol. 15, *Solving Least Squares Problems*, Soc. for Ind. and Appl. Math., Philadelphia, Pa.

Lichten, S. M., and J. S. Border (1987), Strategies for high-precision global positioning system orbit determination, *J. Geophys. Res.*, *92*, 12,751–12,762.

Linde, A. T., M. T. Gladwin, M. J. S. Johnston, R. L. Gwyther, and R. G. Bilham (1996), A slow earthquake sequence on the San Andreas Fault, *Nature*, *383*, 65–68.

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.

Melbourne, T. I., and F. H. Webb (2003), Slow but not quite silent, *Science*, *300*, 1886–1887.

Miller, M. M., D. J. Johnson, C. M. Rubin, H. Dragert, K. Wang, A. Qamar, and C. Goldfinger (2001), GPS-determination of along-strike variation in Cascadia margin kinematics: Implications for relative plate motion, subduction zone coupling, and permanent deformation, *Tectonics*, *20*, 161–176.

Nikolaidis, R. M. (2002), Observation of geodetic and seismic deformation with the Global Positioning System, Ph.D. thesis, Univ. of Calif, San Diego.

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

Okada, Y. (1992), Internal deformation due to shear and tensile faults in a half-space, *Bull. Seismol. Soc. Am.*, *82*, 1018–1040.

Ozawa, S., M. Murakami, and T. Tada (2001), Time-dependent inversion study of the slow thrust event in the Nankai Trough subduction zone, southwestern Japan, *J. Geophys. Res.*, *106*, 787–802.

Rogers, G., and H. Dragert (2003), Episodic tremor and slip: The chatter of slow earthquakes, *Science*, *300*, 1942–1944.

Sagiya, T. (2004), Interplate coupling in the Kanto District, central Japan, and the Boso silent earthquake in May 1996, *Pure Appl. Geophys.*, *161*, 2601–2616.

Sagiya, T., and S. Ozawa (2002), Anomalous transient deformation and silent earthquakes along the Nankai Trough subduction zones (abstract), *Seismol. Res. Lett*, *73*, 234–235.

Szeliga, W., T. Melbourne, M. M. Miller, and V. M. Santillan (2004), Southern Cascadia episodic slow earthquakes, *Geophys. Res. Lett.*, *31*, L16602, doi:10.1029/2004GL020824.

Zhu, L., and L. A. Rivera (2002), A note on the dynamic and static displacements from a point source in multilayered media, *Geophys. J. Int.*, *148*, 619–627.

Zumberge, J. F., M. B. Heflin, D. C. Jefferson, M. M. Watkins, and F. H. Webb (1997), Precise point positioning for the efficient and robust analysis of GPS data from large networks, *J. Geophys. Res.*, *102*, 5005–5017.

T. I. Melbourne, M. M. Miller, V. M. Santillan, and W. M. Szeliga, Geodesy Laboratory and PANGA Data Analysis Facility, Department of Geological Sciences, Central Washington University, 400 E. University Way, Ellensburg, WA 98926, USA. (tim@geology.cwu.edu)