ENERGY BUDGET ANALYSIS OF SLOW-SLIP TREMOR EVENTS
ALONG THE CASCADIA SUBDUCTION ZONE USING
CONTINUOUS GPS ARRAY DATA

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by
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ABSTRACT

ENERGY BUDGET ANALYSIS OF SLOW-SLIP TREMOR EVENTS ALONG THE CASCADIA SUBDUCTION ZONE USING CONTINUOUS GPS ARRAY DATA

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Seismic hazards poised to cities by subduction zones are strongly controlled by fault slip along the deeper extent of the two plate interaction closest to population densities. In Cascadia, where $M_w = 9$ size events are known to occur from a variety of sources, modeling of leveling data has suggested that the region of maximum slip lies well offshore and diminishes rapidly inland. However, over two dozen slow slip distributions have been imaged using Global Positioning System (GPS) along the lower reaches of the northern Cascadia locked zone between 30 and 40 km in depth. Averaged over many episodic tremor and slip events, the upper limit of transient slip in the vicinity of Seattle, Washington and Vancouver, British Columbia comes close to the heavily urbanized regions. Moreover, these events appear to dissipate approximately half of the total tectonic convergence energy in the region, implying that approximately half of the energy will be available in the next megathrust earthquake. This inference is supported by agreement with observed interseismic deformation patterns, which is
consistent with significant plate coupling extending closer to urbanized areas than has been previously thought. The hazard potential incurred by this scenario necessitates a sober mitigation readjustment given that the stress is likely accumulating much closer to the population centers of the Pacific Northwest than previously supposed.
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CHAPTER I

INTRODUCTION

Within the past several years the Global Positioning System (GPS) array network from the Pacific Northwest Geodetic Array (PANGA), positioned along the Cascadia subduction front, has been substantially enriched with hundreds of new stations. These new stations are adding a sharper precision to the array’s usefulness as a geophysical research tool. In addition to the PANGA network, other array networks including Earth Scope’s Plate Boundary Observatory (PBO), the Western Canada Deformation Array (WCDA), and the United States Geological Survey have added, and will continue to add, to the density of coverage. This analytical precision has been needed due to an established direct correlation between the density of a GPS array network and the network’s effectiveness as a spatial-temporal faulting location tool (Rogers and Dragert, 2003; Szeliga et al., 2004). As a result, it has become possible to begin an accounting of the energy disbursement emanating from the Cascadia rupture front and establish a rate of energy accumulation versus energy release analysis.

Along the Cascadia margin (Figure 1) great earthquake events, moment magnitude $M_w > 8.0$ (Dragert and Hyndman, 1995; Atwater and Hemphill-Haley, 1997; Miller et al., 2002), have occurred over its length from central British Columbia to northern California as the Juan de Fuca plate subducts the North American plate. Over the past ~20 years new tools in paleoseismology have revealed a robust Holocene seismic record, including evidence for recurring catastrophic coastal subsidence, tsunamis, and ground shaking (Atwater, 1987).
Figure 1. Area map identifying the Juan de Fuca plate. Subduction occurs off the coast of the Pacific Northwest of the U.S. and British Columbia. The secular east-northeast velocity of the Juan de Fuca plate relative to the North American plate approximates 38 mm/yr (Dragert et al., 2001; Rogers and Dragert, 2003). The inland directed bight off the coast of northern Washington and Vancouver Island identifies where the subduction angle decreases and therefore where the seismogenic zone increases the width of its surficial expression. Modified from Hyndman and Wang (1995).

Taken with evidence for a circum-Pacific tsunami (Satake et al., 1996), these data point to a major earthquake on the Cascadia subduction front at 1700 A.D. On the basis of the size of the tsunami in Japan, the entire length of the Cascadia margin likely broke in a single event of $M_w \approx 9.0$ (Satake et al., 1996). The simultaneity of turbidite triggering in submarine canyons and channels along the length of the Cascadia margin during the Holocene supports the case for large earthquakes that erupt the length of the
convergent margin rather than smaller events along a seismically segmented subduction zone (Adams, 1990). The turbidite data further suggests recurrence for such events on average ~600 years (Atwater and Hemphill-Haley, 1997; Miller et al., 2001). However, in recent years a much slower form of strain release has been detected in many subduction zones, including the Cascadia margin, which adds a further degree of complexity to the system. Until the recent past, these slow-slip events (SSE) went undetected in large part due to the dearth of data from the geophysical networks needed to resolve the subtle signatures of the SSE.

GPS has been used for continuous long-term tectonic deformation and static displacements from large earthquake events (Dragert and Hyndman, 1995; Bock et al., 2004). The GPS component is necessary for a complete long term budget analysis because it is still unknown whether plate creep occurs in the absence of detectable seismicity, simultaneous with seismicity, or not at all. The mechanism by which measured surface deformation represents the integrated deformation from a large number of discrete seismic slip events has not yet been determined (Szeliga et al., 2007). A better accounting of the mechanism and its relationship to mega-thrusts will increase the possibility of improved seismic hazard prediction.

The behavior of accumulated strain and release down-dip of subduction seismogenic plate contacts has not been well understood due to the absence of quality dense GPS array analysis. However, quantitative data models of most subduction zones have indicated that the plate interface moves with increasing depth from purely stick-slip seismogenic behavior to stable slipping. Between these two zones lies a transition
zone; a regime where released ruptured faults cannot self-nucleate (Melbourne et al., 2002). The locked and transition zones have been defined by dislocation modeling of geodetic data and using thermal constraints (Hyndman and Wang, 1995; Wang et al., 2003) (Figure 2). The result of my model differs slightly, but with significant implications, from previous models. The seismogenic zone (maximum rupture area for a great earthquake) extends much further inland from the coast. If all the seismogenic area ruptures in a single event, triggered by the accumulated energy from SSE, the magnitude would be $\sim M_w = 9.0$ (Dragert et al., 1994; Hyndman and Wang, 1995).

It has been demonstrated that a velocity weakening interface will not creep yet a velocity strengthening interface will accumulate shear stress across the interface until an

Figure 2. Cascadia subduction interface. Oregon is at the lower left-hand corner of the graphic and Vancouver Island at the upper right-hand corner. The relatively placed locked and transition zones are identified and the decrease in slope angle paralleling the coastline bight, which can also be seen in Figure 1, is represented here as well. Modified from Flück et al. (1997).
equilibrium creep rate equal to the plate convergence rate is established (Wang et al., 1995; Szeliga et al., 2004). Neither state, as reported, yields a long term steady creep rate approaching that of the convergence rate. Since the metastable region is often inferred to be equal to, or of greater total area than, the seismogenic region, it represents a potentially important contribution to the energy budgets of subduction zones in Cascadia as elsewhere (Melbourne et al., 2002). The documentation by Szeliga et al. (2007) of 35 SSEs along the Cascadia megathrust, of which over a dozen have occurred along the northern Cascadia interface, suggests that these events might be of use in quantifying the amount of potential seismogenic rupture that could take place along the deeper reaches of the locked portion closest to major metropolitan regions. While previous modeling of interseismic deformation vectors have focused on fitting long-term interseismic vectors, the increasing spatial resolution of GPS-based slip inversions allows a more detailed look at strain accumulation beneath regions of the megathrust that have had long-term monitoring; the Olympic Peninsula is one of the best candidates for such monitoring (Hyndman and Wang, 1993; Flück et al., 1997; Miller et al., 2001; Mazzotti et al., 2003; Wang et al., 2003).

The goals for this research were two-fold. First, an organizational effort for the GPS stations’ time series in the array blanketing the area dominated by western continental North America was planned. The GPS monuments’ time series exist as a list of clusters of stations which are queued relative to geographic and physiographic similarity, as well as spatial proximity. This undertaking is expressed in the form of an internet Web page which exists as an appendage to the PANGA home Web site.
Second, a model of the strength of the margin’s effective transition zone (ETZ) was fine tuned as a means to apply careful differentiation of the interseismic strain that results from subduction zone coupling and secular deformation on crustal faults. This effort is necessary to illuminate the processes which drive deformation within plate margins and constrains seismic risk. The time series from a subset of the GPS array’s stations in western Washington and Vancouver Island were used to demonstrate that the plate coupling from the Cascadia convergent thrust retains ~50% of its strength to as far east as the Puget Lowlands. This is significant because the implications from such a model would extend the seismogenic zone width, and therefore the potential hazard from a mega-event, closer to the major population centers of the region.
CHAPTER II
DATA CAPTURE AND ORGANIZATION

In the Pacific Northwest typical lengths of earthquake recurrence intervals are an order of magnitude or longer than the decades-long history of geophysical monitoring (Malone and Bor, 1979; Darienzo and Peterson, 1995; Miller et al., 2001). As mentioned, there is a direct positive correlation between the density of a GPS array and its usefulness as a geophysical analytical tool. This study demonstrates the effectiveness of geodesy for rapid determination of the pattern and magnitude of regional deformation. With the build-up of the GPS array also comes the multitudinous volume of constant and continuously streamed time-series data. The initial organizational challenge began by retrieving the GPS time series from the various regional array data sets where they are hosted and then manipulating the sets into a structured fashion which maximizes usage.

Array Networks

The GPS time series assembled for this study exists additionally as part of separate smaller networks managed locally by autonomous groups and agencies. These local networks cover a more or less continuous stretch from northern and interior Alaska southward to the United States-Mexico border. The density of the GPS station monuments varies from local network as well as within the local network coverage area itself (Figures 3 and 4). For this study, the assembled network consists of stations largely from the smaller PANGA, PBO, and WCDA networks.
Figure 3. GPS station array in the northwestern United States circa 2002. The dearth of stations at this time created large area expanses where data had to be greatly interpolated rendering GPS a significantly blunt analytical tool for surficial deformation due to subduction compression. Modified from Google Earth (2008).
Figure 4. GPS station array in the northwestern United States circa 2007. The station array has been substantially enlarged since 2002 lending itself as a more precise Geophysical tool, as there is less area between stations to be interpolated. Modified from Google Earth (2008).

**PANGA Stations Network**

PANGA, a network of GPS receivers optimally recording in perpetuity, is maintained by several governmental and academic organizations. In Washington and Oregon the collaborating institutions include Central Washington University, the University of Washington, Oregon State University, and the Cascade Volcano Observatory. The instruments that they operate, together with receivers operated by the United States Coast Guard and the Geological Survey of Canada, constitute the PANGA array (Miller et al., 1998). The geodetic network operated by the PANGA consortium monitors crustal deformation from accumulating tectonic stress with mm-level precision. Currently, the time series for ~350 continuously operating, high-precision GPS receivers
located throughout the Pacific Northwest are operating under the monitoring efforts of
the PANGA consortium (Miller et al., 2008).

**PBO Stations Network**

The PBO network, to date, includes 875 GPS instruments which are dispersed
across the western United States, including Alaska (Blackman, 2008). PBO is a construct
of the University (Navigation Signal Timing and Ranging [commonly known as
NAVSTAR]) Consortium (commonly known as UNAVCO) which consists of over 90
international universities and organizations (Anderson, 2004; Blackman, 2008). PBO, as
a product of the Earth Scope project, has a mandate to support and promote Earth science
by advancing high-precision techniques for the measurement and understanding of
Earth’s crustal deformation (Blackman, 2008). The GPS array was designed to study the
three-dimensional strain field resulting from deformation across the active boundary zone
between the Pacific and North American plates in the western United States (Anderson,
2004). The observatory consists of arrays of GPS receivers and strainmeters, which
eventually, will be used to deduce the strain field on timescales of days to decades. The
observatory data are also used for geologic and paleoseismic investigations to examine
the strain field over longer time scales (Anderson, 2004).

**WCDA Stations Network**

The WCDA is a permanent GPS network established by the Geological Survey of
Canada as part of the Canadian National Earthquake Hazards Program (Schmidt et al.,
2000). Specifically, the array is used to investigate crustal deformation in southwestern
British Columbia as part of a comprehensive multi-disciplinary study of seismic hazard in
the densely populated area of the lower mainland and Vancouver Island (Geological
Survey of Canada, 2008). As part of the Canadian National Earthquake Hazards Program, the Geological Survey of Canada embarked on a program of crustal deformation measurements in 1981 in order to monitor present horizontal strain at the northern end of the Cascadia subduction zone (CSZ) (Schmidt et al., 2000; Geological Survey of Canada, 2008). The network spans the most seismically active and most densely populated region in western Canada. The first site to be installed was DRAO (at Penticton, British Columbia) in 1991 which has been used as the baseline terminus on the stable North American plate when compared to the other stations in the WCDA (Schmidt et al., 2000).

Grouped Time Series

*Clustering Criteria*

The GPS stations are grouped together in 34 clusters based on a subset of criteria. First, the physiographic region where the station is located was considered. Next, if the physiographic descriptor spanned state or provincial boundaries then the cluster would usually be split along the territorial boundary (i.e., the Cascade mountain range in Washington, Oregon, or California). Lastly, a station’s proximity to other stations in a cluster was also given weight for inclusion. It should be noted that to insure seamless coverage many stations, especially near a cluster boundary, exist in multiple clusters. As a result, a repeated station time series will exist for each cluster that contains that station. This duplication allows for direct comparison of time series from different spatial perspectives.

Due in part to the geographical reality of the CSZ, the focus of the analysis was necessarily limited to the western-most flank of North America. Therefore, with few exceptions the territorial boundaries of Alaska, British Columbia, Washington, Oregon,
and California were used as the eastern demarcation for station inclusion consideration. The near border exceptions were made in favor of station inclusion rather than station exclusion. The inclusion exceptions were made either because the paucity of local stations near the eastern territorial border would be substantially enriched by the station inclusion or because the station was part of a group of stations whose arrangement was due to physiographic interests (Figure 5).

![Figure 5](image.png)

Figure 5. Eastern territorial boundary example. Figure is displaying a portion of the California Basin and Range cluster at the California-Nevada state boundary. Site p627, in the background, is included in the cluster because it is part of the local physiographic identity, although it is clearly in Nevada. Modified from Google Earth (2008).

*Web Page*

The 34 station clusters are maintained, displayed, and updated daily on the Web page wusdaily (www.geodesy.org/wusdaily). The wusdaily page is an appendage to the PANGA home page (www.geodesy.org). Each cluster of stations includes an area map created from kml image files displayed in Google Earth format and arranged with the
most northerly clusters displayed at the top of the page while progressing southerly towards the page bottom. In addition to the 34 standard clusters, a section of custom composed station clusters is appended to the bottom of the Web page. These clusters were compiled after specific tectonic events and consist of the stations that most accentuated surficial deformation from that particular event.

The time series for each station in the cluster is included with the cluster map image (see Appendix for details regarding manipulating the wusdaily Web page and station clusters). Each time series is separated into the two horizontal measurement components (longitude and latitude) and the radial component. Each station in the cluster is, in turn, displayed from the northern most at the top while progressing southerly towards the bottom of the page. A thin horizontal line is used to separate the stations which share a special proximity from the other stations in the cluster. In order to keep the time series distinct and separate from each other, a medium of spatial economy was settled on with 10 stations per sheet. In the case of station clusters consisting of more than 10 stations, multiple sheets are generated as needed. The time series are displayed on a Cartesian axis with time (years) being the independent variable and displacement (mm) the dependent. The displacement scale is displayed in increments of 10 with “0” in the center, “150” at the top, and “−150” at the bottom (for the horizontal time series). Because the accuracy for the radial displacement is significantly less than the accuracy for the horizontal components, the radial displacement scale ranges from “300” to “−300.” In effect, the radial displacement is exaggerated by a factor of 2 in order to see the same level of detail as in the horizontal time-series dynamics. The time scale on the x-axis progresses from 4.5 years at the axis origin towards the current date. The time scale
is essentially a sliding window with the time differential fixed at 4.5 years and the current
day’s date dictating the terminus for the most recent displacement data (Figures 6-9).

Before the GPS time-series data was used for daily uploads it first went through
systematic processing. Continuous GPS data from the PANGA array were processed with
the Gipsy-Oasis II software (Miller et al., 2001). Precision point positioning, orbits and
clocks were used to analyze the phase data with ambiguity resolution applied (Szeliga,
2005). A particular reference frame localized to North America, the Stable North
American Reference Frame (SNARF), is used to determine the daily GPS solutions for
station positions and corresponding matrices of the covariance among the position
components (Altamimi et al., 2005; Szeliga, 2005). A regional stabilization is latterly
applied to the daily GPS position which effectively minimizes network-wide position
discrepancies or common-mode errors (Szeliga, 2005). Tropospheric delays and other
seasonal effects which compound known time-series offset error artifacts (hardware
upgrades, earthquakes, and annual and semi-annual sinusoidal signals) were corrected for
when the final time series were detrended (Szeliga, 2005).

Summary

GPS acquires its prowess as an analytic tool from station longevity and array
density. Since the mid 1990s there has been a concerted effort from the various interested
groups to enrich the array from a few hundred stations to the current ~1500 stations. This
effort has favored the opportunity to parse out deformation details within the active
margin by providing information where previously there was only intra-station
smoothing. In concert, the stations’ longevity provides an advanced level of deformation
trend clarity. Although the local arrays which compose the encompassing array network
Figure 6. Area map of a custom station cluster. The May 2008 event is a representative example. Unless stations are added to, or deleted from, the cluster, the map graphic need only be generated once. Modified from Google Earth (2008).
Figure 7. May 2008 custom stations cluster longitude series. As the secular displacement and the transient reaction is predominantly in an east-west direction in the northern Cascadia subduction thrust, the longitude time series is usually the most useful for picking SSE. The May 2008 event is highlighted by the rectangular box. The time series is generated daily from a cron script indicated on the Web page by a date and time stamp located on each time series sheet.
Figure 8. May 2008 custom stations cluster latitude series. As the north-south component of horizontal displacement has a lower magnitude than the east-west component in the northern Cascadia subduction thrust the SSE appearance is much more subtle in the time series. In fact, some stations do not record any discernable north-south displacement. The rectangular box is highlighting the same area as in Figure 7.
Figure 9. May 2008 custom stations cluster vertical series. The y-axis is exaggerated in order to display the displacement with a relative magnitude from the two horizontal displacement time-series graphics (Figures 7 and 8). The rectangular box is highlighting the same time period as in Figures 7 and 8; however, with the radial time series, the event is not as discernible.

are still largely managed and maintained at that level, the cooperation and data availability outside of the local system is more fluid due to organizational oversight. Due to the sheer volume of continuous data stream, the time-series organization has become a pressing need; therefore, an internet Web page consisting of the stations’ time series was
created with physiographic and spatial proximity as a structural guide. As structural
guides, and not dominant constraints, these parameters were trumped by a north-south or
east-west progressing station order organizing decisions.
CHAPTER III
GEODETIC CONSTRAINTS ON GPS RECEIVERS

Vertical Accuracy

The vertical component of GPS measurement is the algebraic result of the geoid separation differential from the ellipsoid height: the difference in geoid and ellipsoid height from the Earth’s centroid of mass at a given surface point (Lambeck, 1988; Strang and Borre, 1997). GPS computes height from the smooth World Geodetic System 1984 (commonly referred to as WGS84) reference frame directly and GPS receivers can only compute latitude, longitude, and ellipsoidal height. Because orthometric (mean sea level) heights are what are desired, GPS receivers incorporate a table of geoid height values. Unfortunately, because of storage constraints, these values are not sufficiently accurate. Most GPS receivers use a 10 × 10 geoid model which stores one geoid height value per 10° × 10° defined area. However, geoid values can vary greatly within a 10° × 10° area. The GPS receiver computes the height:

\[ h = H + N \]

where \( h \) is the ellipsoid height, \( H \) is the orthometric height and \( N \) is the geoid height (Figure 10). The distance from the Earth’s centroid of mass to the surface of the ellipsoid is determined by the functions of trilateration and free-space ranging (Blewitt, 2003).

Three-dimensional positions are calculated using the pseudo-range between satellite and receiver at a given epoch to determine a position (Blewitt, 2003; Altamimi et al., 2005). If more than four satellites are in view, then most receivers overdetermine the position using all combinations of four satellites from those in view to establish
Figured 10. Vertical reference frames. Most GPS receivers use a $10 \times 10$ geoid model that stores one geoid height value per $10^\circ \times 10^\circ$ area (Blewitt, 2003). Geoid heights can vary a great deal within a $10^\circ \times 10^\circ$ area.

positions, then perform a least squares solution to attempt to determine which satellite combination is best (Lambeck, 1988). This combination is tested periodically, but usually not for every solution. To get a good three-dimensional position, a satellite is needed directly overhead and three others maximally spaced below the horizon in a constellation. Since GPS signals do not traverse Earth material very well, this is impractical for surface based receivers; as a result, vertical accuracy suffers. However, using a strong receiver network for adjustment, long period (multiple hours) data acquisition, data decimation to remove autocorrelation effects, and careful post processing to achieve a good solution to submit to least squares adjustment a vertical accuracy of a factor of $\sim 3$ relative to the horizontal accuracy can be achieved (Lambeck, 1988; Strang and Borre, 1997).
Stable North American Reference Frame

For time-series analysis solutions the PANGA laboratory utilizes SNARF as a particular reference frame for continental North America (Altamimi et al., 2005). Specifically, GPS is used to measure the movements of \( \sim 1000 \) points spanning the North American-Pacific plate boundary and these motions must be defined relative to a terrestrial reference frame (Blewitt, 2008). Such a frame requires the definition of its Cartesian coordinate axes and the evolution of these axes in time, as well as precise models of the Earth. The method used by PANGA to establish SNARF relies on the inclusion of sufficient stations from the stable continent to solve for an Euler pole for the internally consistent data set (Miller et al., 2001).

Fundamentally, a reference frame is required because GPS alone does not provide unambiguous coordinates. GPS data are relatively insensitive to global rotations of the entire system (Blewitt, 2003; Calais, 2006). Fixing the rotation according to a well documented scientific scheme can facilitate geophysical interpretation. GPS stations are used to measure the movements of points spanning the plate boundaries. These motions must be defined relative to SNARF. SNARF requires the definition of its Cartesian coordinate axes and the evolution of these axes in time, as well as precise models of the Earth (Blewitt, 2003). The motions of the Earth’s surface due to tectonic processes are most naturally expressed with respect to the stable interiors of crustal plates. SNARF, as a standard reference frame, therefore allows for a more straightforward interpretation of the geodetic data in terms of where the total budget of relative plate motion is being
accommodated, and how deeply the plate boundary dynamics penetrate into the plate interior (Lambeck, 1988; Blewitt, 2003).

SNARF provides a common framework for comparison of geodetic data and geophysical models. Defining a stable frame at the sub millimeter level requires adequate characterization of Earth deformation processes across the stable plate interior (Blewitt, 2003); a region that by definition is relatively unaffected by plate boundary processes (Lambeck, 1988). The plate interior, therefore, provides a stable platform from which to view plate boundary deformation. The stable plate interior actually deforms very slowly in a complex way due to phenomena such as glacial isostatic adjustment and other mantle-scale processes coupled to a heterogeneous lithosphere which is occasionally host to large intra-plate earthquakes (Lambeck, 1988).

Summary

GPS geodesy, as it is used in this research, is constrained three-dimensionally by the accuracy of its vertical component. Although vertical accuracy is constantly being improved it will likely not ever be as dependable as the horizontal components because of the physical realities of gleaning the vertical component on spheroidal surfaces with a satellite constellation. As it stands, the horizontal component accuracy is ~3 times that of the vertical.

Any reference frame that is used for high precision crustal deformation will have constraints inherent to the system. The motions of the Earth’s surface due to tectonic processes of the region spanning the boundary between the North American and Juan de
Fuca plates are most naturally expressed with respect to the stable interiors of the North American plate. A standard reference frame therefore makes it easier to interpret the geodetic data in terms of where the total budget of relative plate motion is being accommodated.
CHAPTER IV

SUBDUCTION THRUST, ENERGY DISBURSEMENT

Slow-Slip Tremor Events

The CSZ harbored directly off the coast of the northwestern United States and southwestern Canada is an ~1100-km thrust fault that stretches from Cape Mendocino in northern California to northern Vancouver Island, British Columbia. The CSZ is truncated to the south at the Mendocino Fracture Zone created by the triple plate interaction of the Pacific, North American, and Gorda tectonic plates. Similarly, the CSZ is truncated to the north at the Queen Charlotte Fault created by the triple plate interaction of the Juan de Fuca, North American, and Explorer tectonic plates. The Juan de Fuca plate subducts the North America plate at 3 to 4 cm/yr (Savage, 1983; DeMets and Dixon, 1999; Miller et al., 2001) along an interface known from several lines of evidence to be seismogenic and rupture margin wide in relatively infrequent $M_w = 9$ events averaging every 500-600 years (Satake et al., 1996; Atwater and Hemphill-Haley, 1997).

There is geological and historical evidence that suggests that the last event occurred in 1700 A.D. as a, at least, $M_w = 9$ earthquake and caused widespread tsunami damage in Japan (Atwater, 1987; Satake et al., 1996; Satake et al., 2003). The secular deformation field from ongoing interseismic convergence manifests itself as NE-directed compression along the arc that drops off markedly with distance from the arc. SSEs are known to occur along the CSZ length and manifest themselves as spatially correlated transient reversals of the GPS velocities that last from 1 to 2 weeks at any given station, and are observed to propagate across the geodetic array (Dragert et al., 2001). These
events are interpreted as slow faulting along the deeper CSZ by aseismically slipping, down-dip of the locked zone, at around 25 km to 45 km depth (Figure 11). Similar events

![Image of SSE events](image)

Figure 11. SSE lasting 5.5 weeks in early 2003. Event is propagating through the CSZ. Plate interface depth contours are shown (white lines) running parallel to the subduction strike. The event can be seen as a slip coloration moving through the system between the 25 and 45 km depth contours.

have been found with recurrence of ~14 months in northern Cascadia (Miller et al., 2002) and 18 and 11 months in central Oregon and northern California, respectively (Szeliga et al., 2007). To date, over three dozen events have been identified along the Cascadia arc since 1997, including eight in the northern Cascadia region.

Temperature differential is an important constraint for the dip profile (Hyndman and Wang, 1995) and depth on the subducting slab is the dominant controlling factor on temperature variation. However, the angle of plate dip also has a noticeable influence and
therefore must be considered. Since the down thrusting plate, landward of the
deformation front, provides less of a heat sink the heat flow decreases landward. Further,
the increasing precision with which these downdip limits can be constrained is advanced
by recent GPS-inferred moment estimates (Aguiar, 2007; Szeliga et al., 2007) that have
helped define the downdip extent of the seismogenic zone when compared with seismic
tremor data.

The current downdip extent of SSE rupture in northern Cascadia was first
estimated from several avenues of geodetic measurement techniques (Dragert and
Hyndman, 1995; Hyndman and Wang, 1995). Repeated high-precision leveling has been
used to establish vertical strain data and, additionally, long-term trends in tide gauge data
and repeated very accurate gravity measurements have also contributed to the strain data
(Hyndman and Wang, 1995). As repeated triangulation measurements (GPS and laser
ranging trilateration) have helped to establish horizontal motion, this early effort has been
used to ensconce a linear downdip SSE rupture rate and extent. Deformation data have
constrained the dislocation model with linearly defined widths to represent 90 km for the
locked zone with a 90 km transition zone in northern Cascadia, where the subduction dip
angle is much shallower.

Wang et al. (2003) employed the effective transition zone (ETZ) to allow a
downdip decrease in slip deficit along the fault and to account for the effect of
viscoelastic relaxation of the mantle wedge in an elastic model. Further, Wang et al.
(2003) uses an exponential function algorithm to allow for slip deficit to employ a faster
decrease rate in the seaward component of the transition zone and a slower rate in the
landward component, which effectively allows the downdip extent of rupture to extend
landward to fit the data. With this more flexible model, the downdip limit of the ETZ is not defined by fault properties and the width should increase with interseismic time differentials. McCaffrey et al. (2007) modified the depth distribution of locking from Wang et al. (2003) by constraining the instantaneous creep fraction to decrease with depth. This group also forced the slope to increase or remain approximately constant with depth, which effectively constrains the seismogenic zone to a narrow width (McCaffrey et al., 2007).

The Tokai SSE in southwest Japan is thought to be interesting, in part, because the slip region is adjacent to the expected source area of the Tokai earthquake and the SSE might promote the occurrence of the mega-earthquake (Hirose and Obara, 2006). Hirose and Obara (2006) report that episodes of SSE and tremors have repeatedly occurred in the Tokai region and further suggest that these ETS episodes are a characteristic behavior of the deep region of the subducting plate boundary where nonvolcanic tremor occur along the Nankai trough. The annual displacement released by means of SSEs along the Nankai trough is comparable to the relative plate convergence rate; this indicates that the SSEs may be the process of strain release which is accumulated by the relative plate motion (Hirose and Obara, 2006). Further, this may also support that the SSEs occur on the subduction plate interface, which is estimated by the SSE fault inversion.

The recurrence features of the SSEs at the Nankai trough suggest that the short-term SSE may be the process of strain release which is accumulated by the relative plate motion. The frequency of SSE episodes then may be at least partly controlled by a plate convergence rate (Hirose and Obara, 2006). Tremor similarities and tectonic environment
between the Cascadia margin and southwest Japan, at their respective source depth (between 25 and 45 km), has been observed by Obara et al. (2004). This group reports observed data that can be explained by slow slips with dislocations corresponding to an equivalent $M_w = 6.0$ on faults which are located just above the dipping seismic zone in the subducting slab (Obara et al., 2004). The updip limit of the slip is reported to correspond to the source area of tremors and the deeper part of the rupture area to megathrust earthquakes (Obara et al., 2004; Hirose and Obara, 2006).

The SSE exertion in northern Cascadia drifts along strike of the subduction zone coincident with the deep slip events at rates varying from ~5 to ~15 km/day (Rogers and Dragert, 2003). The activity through the affected region deviates from gradual displacement to a marked transformation from one region of the subduction fault to another so that the maximum extreme of the amplitude spectrum can be detected at least as far as 300 km from the source of nucleation (Melbourne et al., 2005). GPS analyses, used to detect the surface displacement patterns, have been implemented and have efficaciously modeled the deformation (Dragert et al., 2001) using first order dislocations of 2 to 4 cm on the plate interface, bounded by the 25 km and 45 km depth constraints, which strongly suggests a spatial correlation with the area of tremor nucleation (Rogers and Dragert, 2003).

GPS resolution has been a blunt tool to infer details of slip, due mostly to historic scarcity of stations in the array. However, since the 2001 Nisqually earthquake the area encapsulating the Puget Lowlands has been blanketed with dozens of stations. Because of the array growth, especially in this region, Szeliga et al. (2007) were able to demonstrate that slip inversions based on the Green’s smoothing function can reliably estimate total
equivalent magnitude. Since this is a reliable predictive method for earthquake response it is therefore prudent to then systematically invert the largest creep events recorded for slip on many stations. While previous modeling of interseismic deformation vectors have focused on fitting long-term interseismic vectors, the increasing spatial resolution of GPS-based slip inversions allows a more detailed look at strain accumulation beneath regions of the megathrust that have had long-term monitoring and the Olympic Peninsula is one of the best candidates for such monitoring (Flück et al., 1997; Miller et al., 2001; Mazzotti et al., 2003; Wang et al., 2003).

GPS Data Analysis, Inversion for Slip

Raw GPS-phase observables from the combined networks of the PANGA and WCDA arrays were processed with the GIPSY (Zumberge et al., 1997) software package as described in Szeliga et al. (2007). The resultant time series of Cascadia GPS positions relative to cratonic North America were then decomposed into a set of basis functions that include linear trends, annual and semiannual sinusoids, and a summation of step functions introduced at times of known earthquakes, slow earthquakes, or GPS instrumentation upgrades. This approach of simultaneous decomposition yields the full covariances of all estimated parameters, and the transient deformation due to the slow slip events discussed here is shown in Figure 12.

To invert for slip the plate boundary is specified by linearly interpolating between depth contours supplied by Flück et al. (1997). This surface is then divided into variable sized subfaults whose typical dimensions are around 25 km along strike and 15 km downdip. Positivity (thrust-only slip) is enforced in the inversion by employing the non-
Figure 12. Nine years of GPS longitude measurements. This series from the CSZ shows evidence of over 30 slow slip events. Vertical tick marks are 10 mm. Blue lines indicate slip events either well-recorded with GPS or corroborated by observations of subduction-zone tremor. Red lines indicate spatially coherent transient GPS deformation recorded on less than 4 stations and uncorroborated by tremor. Maximum geodetic offsets are 6 mm and correspond to the spatially largest event in early 2003. The February 2001 $M_w$ 6.7 Nisqually earthquake appears on two stations.
negative least squares algorithm of Lawson and Hanson (1995). To avoid highly oscillatory and nonunique slip distributions, smoothing is enforced by row-augmenting the matrix of Green’s functions with a finite-difference approximation to the Laplacian operator and augmenting the corresponding rows of the data vector with an equal number of zeros. This requires finding an optimum weighting factor to control the degree of smoothing, which is performed by solving a data-reduced vector and constructing a bootstrap estimate of the remaining data to predict the missing data subsets (Efron and Tibshirani, 1994). Although smoothing trades off with maximum slip, the resultant moment inverted from the transient data is largely invariant with respect to smoothing, and changes inverted moment by less than one percent over four orders of magnitude change in the smoothing parameter. Details of the parameter estimation, inversion, and signal filtering can be found in Szeliga et al. (2007).

Results for the largest events over the period from April 1997 to April 2007 lay between moment magnitudes ranging between 6.3 and 6.8. Further criteria dictated that GPS-inferable moment magnitudes were not below $M_w = 6.3$ because events lower in magnitude are not resolvable with GPS and the maximum inverted slip is 3.6 cm for all events (Szeliga et al., 2007) (Figure 13). In order to obtain a first order estimation of the plate coupling stress release and retention, the nine GPS stations in the region with a sufficiently long data collecting history were selected and coalesced as a summation of total transient slip recorded over the time differential. The transient slip rate was first adjusted by subtracting the 4 cm/yr secular convergence, based on the NUVEL1 plate motion model, and then the slip was normalized so that the areas with no slip are 100% locked. Then, the difference between the total accrued slip over the approximate 9-year
Figure 13. Slip distributions for Cascadia SSE. The largest 12 events from the last decade are shown. GPS-inferable moment magnitudes range from 6.7 down to 6.3, below which they are not resolvable. Maximum inverted slip is 3 cm for all events, but this is a function of smoothing coefficient b, shown beneath each slip amount. Red vectors show misfit between data and modeled vectors.
period and the accrued slip at a particular location in our study area gives a value relative to total slip (Figure 14).

Figure 14. Convergence accrued along the Olympic Peninsula. This section of the Cascadia megathrust is from 1997-2005, assuming 34 mm/yr convergence rate, minus the summation of the slip distributions for the 12 largest slow slip events. Red regions indicate where no known slip has occurred; blue regions indicate where a significant amount of the total accrued convergence has been dissipated by SSE that occurred between 1997 and 2006. The integration of the slip distributions show that up to ~50% of the total Cascadia convergence rate may be accumulating along the deeper edge of the Cascadia locked zone, and in close proximity to the major metropolitan regions of the Pacific Northwest. We test this by modeling long-term PANGA time series with a coupling profile. Note that regions to the north and south of this area have not had sufficient GPS monitoring density or duration to perform this type of analysis outside of the Olympic region.

Interseismic Modeling

The GPS station density was sufficiently robust at the time of my investigation for a first order approximation to estimate the slip distribution from GPS deformation for the transients. The plate boundary surface was conditioned by linearly interpolating between depth contours, as provisioned by Flück et al. (1997). The surface is then partitioned into subfaults of varying dimensions; the along-strike lineation is ~25 km, whereas the
downdip lineation is normally ~15 km. However, the precise dimensions of a particular subfault fluctuate with the geometry. In northern Cascadia the surficial geometry is quite complex, exaggerated by the deflection of the oblique angle of dip which differs greatly along strike. By following the shoreline north along the seaward side of Vancouver Island the inland bight betrays a simulacrum of the underlying subduction obliqueness. The plate interface here displays the conspicuous three-dimensional bend north of 47°N, changing strike from northeast to northwest (Flück et al., 1997; Miller et al., 2001; Szelig et al., 2007). This equates to flexible subfault strikes ranging over 40° over less than 1° of latitude differential and requires that each subfault be autonomously specified with a unique strike, dip, and pitch. The subfaults on the down-thrusting converging plate display 60 rows of cells along strike with 30 cell columns normal to dip. This 30 × 60 area cell matrix, or 1800 area cells, terminates to the north between the latitudes of ~49°N and ~51°N and the southern terminus is at ~42°N latitude. The western extent approximates 232° longitude and terminates to the east at ~238° longitude.

Attempts were made to fit the independently derived tremor data (Aguiar, 2007; Szelig et al., 2007) for a first order approximation from both linear and step digression functions with relative success. However, it became apparent that more flexibility was needed by following Wang et al. (2003) and Miyazaki and Larson (2008) and recognizing that the afterslip updip relative to the ETZ propagates faster and afterslip downdip propagates significantly slower. Therefore, a simple algorithm function \(\frac{1}{ax}\) was manipulated and applied for a relatively rapid dampening effect asymptotically with distance from the deformation front. The proportional distance for each subfault row
through the ETZ is represented by $x$ and from repeated modeling attempts, $a = .1975$ (a constant) (Figure 15).

![Diagram](image.png)

**Figure 15.** Coupling strength normalized from full coupling. Transition is through the locked zone and decreasing in strength moving distally from the deformation front. The 3 trend lines represent the best fitting trends; the overall best fitting trend (a), the best fitting step digression (b), and the best fitting linear digression (c).

For economy of manipulation, each cell is assigned a coupling strength value equivalent to the average value for the row of cells to which it belongs. Because the GPS time-series history becomes more limited the further we move along strike from our area of interest, especially northward, it would be difficult to constrain coupling effect strength further for the more distal cells. For this reason, the area of interest is further constrained to between the latitudes of 47°N and 49°N. An error caveat must be
highlighted here by acknowledging that the coupling strength, as recorded by the stations in the study area that data is being manipulated from, will be greater above the convergence interface where GPS station concentration is the densest. This is manifested by the cells on the perimeter of the study area being skewed more heavily. Even so, this method provided remarkably accurate results for both north-south and east-west horizontal deformation model fits for the nine monument stations used in the study. Of the nine stations (Figure 16) the longitude model trends for the best fitting $\frac{1}{ax}$

Figure 16. Area map showing the nine GPS monuments. These stations have been recording continuously, with few notable exceptions, since 1997. These stations are used to model the interseismic coupling profiles from the best fitting parameter. Modified from Google Earth (2008).

dampening function matched remarkably well (on a relative qualitative scale) for four of the stations (PABH, NEAH, ALBH, BLYN). For three more of the stations the model trends fit reasonably well (SATS, WHD1, SEDR). The model trends for two of the
stations (SEAT, KTBW) applied too much weight to the model strength (Figure 17). The latitude trends fared similarly, if with a slightly different station grouping. Of the nine stations in the study, five stations (NEAH, SATS, BLYN, KTBW, SEDR) displayed

![Figure 17. GPS longitude time series from $1/\alpha_t$ showing horizontal deformation. Time-series data (dots) showing northward and eastward directed deformation of stations shown in Figure 16. Because the GPS data contain both interseismic deformation as well as coseismic responses to known SSE, we include both modeled steady-state interseismic deformation predicted by the coupling profile and 34 mm/yr of convergence. Using Savage (1983) backslip method, the southwest-directed deformation predicted the slow slip distributions (Figure 13) at their time of occurrence. Coupling strength as a function of fault depth and the coseismic slow slip deformation predicted from the slip distributions of Figure 13 fit all deformation profiles within 2 mm/yr, with the majority fitting within 0.5 mm/yr, possibly caused by unknown crustal heterogeneity or faulting. This supports the inference that up to ~50% of the total Juan de Fuca convergence beneath the Olympics is available to drive seismogenic slip during the next megathrust rupture.](image)
remarkable model latitude trends. One station (ALBH) displayed a reasonable fit. Three stations (PABH, WHD1, SEAT) fared more poorly (Figure 18). Of the nine stations, two stations (NEAH and BLYN) displayed remarkably well for both longitude and latitude trend fits. Only one station (SEAT) fared relatively poorly for both horizontal displacement direction trend fits. Model trend differences which are described as reasonably well are still no greater than ~5 mm off the data trend fit over the entire approximate nine-year period (SATS, BLYN, and KTBW have shorter time-series histories). Of the stations which performed poorly, the trends were generally less than ~15% off the data trends (SEAT is a notable exception). The relatively poor showing,
especially at the SEAT station, is likely due to the unusually large amount of sub-faulting in that part of the Puget Lowlands (Atwater and Hemphill-Haley, 1997).

Comparing the data matching results of the $1/ax$ dampening function to the best fitting linear digression results (Figures 19 and 20) and the best fitting step digression results (Figures 21 and 22) we observe a superior match for the more flexible $1/ax$ function results. The best fitting linear application supplies too much weight for the longitude series to the stations near the coast and only has comparable results in the Puget Lowlands region where stress accommodation is likely distorting all approximating models. The linear application fared better for the latitude series for both the SEAT and WHD1 stations but overall is not comparable. Again, these two stations are well within the stress accommodating Puget Lowlands region. The best fitting longitude series for the step function application provides a much better fit for the SEAT station and a slightly better fit for the SATS station but otherwise it is not comparable to the $1/ax$ dampening application. The latitude series for the step function supplies better fits for the KTBW and SEAT stations but otherwise it is also not comparable. Overall, except for the one outlier at the SATS station, the only stations where the $1/ax$ function application did not excel or show comparability are located in the complex stress accommodating Puget Lowlands region.

The subfaulting then is likely accommodating much of the slip deficit. As can be seen by the coupling strength profile (Figure 23), dictated by the horizontally directed model trend fits, nearly half of the coupling strength is maintained significantly more east of the deformation front than previous models predicted. The summation for the nine
Figure 19. GPS longitude time series from linear model. This is from the best fitting linear model. Digression is showing horizontal deformation. This solution applies too much weight to the near coast receivers and is only comparable with the $1/\alpha \xi$ solution in the poorly constrained Puget Lowlands region.
Figure 20. GPS latitude time series from linear model. This is from the best fitting linear model. Digression is showing horizontal deformation. This solution fared better at the SEAT and WHD1 stations (in the Puget Lowlands region) than the $1/a_x$ solution, but overall it supplied inferior results.
Figure 21. GPS longitude time series from step model. This is from the best fitting step model. Digression is showing horizontal deformation. This solution supplies a slightly better fit at the SATS and SEAT stations but otherwise it is not comparable to the $1/\alpha x$ solution.
Figure 22. GPS latitude time series from step model. This is from the best fitting step model. Digression is showing horizontal deformation. This solution supplies a better fit for the SEAT and KTBW stations (both in the Puget Lowlands region) than the $\frac{1}{ax}$ solution but otherwise it is inferior.
Figure 23. Coupling strength as a function of fault depth. This representation best matches the accrued slip profile of fault depth that compares with the accrued slip profile in Figure 14, but in which the ~50% locking profile under the Olympics is propagated along strike (where little GPS data exists to constrain it). This coupling model is similar to Wang et al. (2003), but has somewhat more accrued slip under the Olympics. The model results for each of the stations are compared here with the data summation over 9 years.
stations involved in the study, when compared with the model trends, displays a compelling case for the accuracy of the study’s base assumptions. The coupling strength that is being maintained further downdip is likely accumulating stress for potential megathrust tremor disturbances further eastward into the heavily populated urban areas.
CHAPTER V
CONCLUSIONS

The coupling strength from the best fit maintains nearly half its energy as far from the deformation front as where the concentration of accrued slip measured over a nearly nine year period is located. Miyazaki and Larson (2008) report that a possible explanation for aftershock from the Tokachi-oki earthquake is that the velocity strengthening zone between two weakening zones is partially locked during the interseismic period. It is further reported that the effect is due to weakening zones on the edges of both plates which are strongly locked preventing the strengthening zone from freely slipping with plate velocity (Miyazaki and Larson, 2008). Further, Miyazaki and Larson (2008) derived that the shear stress in the afterslip regions decreased while the slip rates were nearly constant and a deeper part of the fault eventually slipped. Although the subduction angle at the Kurile Trench convergence is different than that of Cascadia and the secular motion maintains a different rate, these findings have potential significance for our observations in Cascadia. Also, Miyazaki and Larson (2008) report that there is further evidence for slip occurring deeper on the Kurile slip interface.

The SSEs summation from the inversion of slip technique systematically locates the accrued slip focus on the east side of the Olympic Peninsula, considerably further (~100 mi) from the deformation front than previous models have predicted (Wang et al., 2003; McCaffrey et al., 2007). Further, when comparing these results to the independently derived coupling strength profile, the potential is remarkable. The coupling strength derived from this first order fitting algorithm culminated in a profile downdip through the locked zone and the ETZ, which maintains ~50% of its strength through the
strike deflection opposite the Olympics and well into the Puget Lowlands. Not only is the slip potentially occurring deeper on the plate interface, it is consequently widening the seismogenic zone eastward in Cascadia. The significance of this potential should not be underestimated. Most building structures in the Pacific Northwest are vulnerable to a megathrust event of this potential magnitude due to the absence of earthquake coding, especially bridges and reinforced brick buildings. Further, as of the Census 2000 (U.S. Census Bureau, 2003), the Seattle-Tacoma metropolitan area situated in the Puget Lowlands region reported a population of over 3 million people; approximately half of Washington’s total population.

Summarily, if it is assumed that the plate convergence is maintaining ~50% of its energy, then ~3.8 cm/yr of slip for ~500 years approximates a potential $M_w = ~9.2$ (Table 1) earthquake with a hypocenter under the Puget Lowlands and very near the population centers of Portland, Seattle, and Vancouver.

<table>
<thead>
<tr>
<th>Dimension</th>
<th>Magnitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slip length (km)</td>
<td>1100</td>
</tr>
<tr>
<td>Slip depth (km)</td>
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<tr>
<td>Slip amount (m)</td>
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<td>Slip rate (cm/yr)</td>
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<tr>
<td>Duration of slip accumulation (yr)</td>
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</tr>
<tr>
<td>$M_w (^{2/3}<em>{10} \log</em>{10} M_0-10.7)$</td>
<td>9.2</td>
</tr>
</tbody>
</table>
REFERENCES


APPENDIX

The specific steps for manipulating the wusdaily Web-page station clusters:

1. To make or change a GPS cluster copy the desired stations, including latitude and longitude, (the first three columns) from

   /home/chapmanj/list_clusters/master_append to a new file created in
   /home/chapmanj/lists_clusters/station_clusters.

2. Add the new cluster name and information to

   /home/chapman/lists_clusters/station_clusters/custom_abrev or */cluster_abrev
   (depending on whether you are amending the custom clusters or the original clusters.)
   The format of the existing files should be followed.

3. Create the kml files by executing

   /home/chapmanj/list_clusters/station_clusters/kml_files/bin/rewrite_clusterfile
   ..new_cluster1 ../new_cluster2. This process reverses the latitude and longitude order
   which is necessary because GMT demands the order one way and kml demands the
   order in the opposite arrangement. Copy clusters.kml to clusters.kml.bak in the case
   of carelessness.

4. Create the kml files by executing

   /home/chapmanj/lists_clusters/station_clusters/kml_files/bin/lists2kmld*_list.txt >
   clusters.kml.

5. Retrieve the kml files on a computer which is Google Earth enabled. It is best to use a
   secure file transfer client.
6. Make a jpg image of the new cluster in Google Earth. Move, or copy, the image into

/home/chapmanj/array_jpegs on a PANGA UNIX computer.

7. To test the manipulated file, or files, execute

/home/chapmanj/experimental/plot_lists_toweb.pl. Then execute

/chapmanj/lists_clusters/wusdaily/make_wusdaily_web.pl.

8. Finally, refresh the wusdaily Web page for immediate effect; otherwise the daily cron

will generate the update the following day.