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UNIVERSITY OF CALIFORNIA, SAN DIEGO

**Using GPS to Rapidly Detect and Model Earthquakes and Transient Deformation
Events**

A dissertation submitted in partial satisfaction of the
requirements for the degree
Doctor of Philosophy

in

Earth Sciences

by

Brendan W. Crowell

Committee in charge:

Yehuda Bock, Chair
C. David Chadwell
Yuri Fialko
Jose I. Restrepo
David T. Sandwell
Peter M. Shearer

2013

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The dissertation of Brendan W. Crowell is approved, and it is acceptable in quality and form for publication on microfilm and electronically:

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University of California, San Diego

2013

EPIGRAPH

The good thing about science is that it's true whether or not you believe in it.

—Neil deGrasse Tyson

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Mw 9.0 Tohoku-oki earthquake with seismogeodesy, *Geophys. Res. Lett.*, in revision.

Crowell, B.W., Y. Bock, and D. Melgar (2012), Real-time inversion of GPS data for finite fault modeling and rapid hazard assessment, *Geophys. Res. Lett.*, 39, L09305.

Melgar, D., Y. Bock, and **B.W. Crowell** (2012), Real-time centroid moment tensor determination for large earthquakes from local and regional displacement records, *Geophys. J. Int.*, 188, 703-718.

Bock, Y., D. Melgar, and **B.W. Crowell** (2011), Real-time strong-motion broadband displacements from collocated GPS and accelerometers, *Bull. Seism. Soc. Am.*, 101, 2904-2925.

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Crowell, B.W., and Y. Bock (2013), Automated detection of transient deformation in continuous GPS time series: A case study of the Cascadia subduction zone, *J. Geophys. Res.*, in prep.

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Geng, J., Y. Bock, D. Melgar, **B. W. Crowell**, and J. S. Haase (2013), A new seismogeodetic approach applied to GPS and accelerometer observations of the 2012 Brawley seismic swarm: Implications for earthquake early warning, *Geochem. Geophys. Geosyst.*, in review.

Crowell, B.W., Y. Bock, D.T. Sandwell, and Y. Fialko (2013), Geodetic investigation of the deformational makeup of the Salton trough, *J. Geophys. Res.*, in review.

Crowell, B.W., D. Melgar, Y. Bock, and J.S. Haase (2013), Rapid modeling of the 2011 Mw 9.0 Tohoku-oki earthquake with seismogeodesy, *Geophys. Res. Lett.*, in revision.

Melgar, D., Y. Bock, D. Sanchez, and **B.W. Crowell** (2013), On robust and reliable automated baseline corrections for strong motion seismology, *J. Geophys. Res.*, 118.

Crowell, B.W., Y. Bock, and D. Melgar (2012), Real-time inversion of GPS data for finite fault modeling and rapid hazard assessment, *Geophys. Res. Lett.*, 39, L09305.

Melgar, D., Y. Bock, and **B.W. Crowell** (2012), Real-time centroid moment tensor determination for large earthquakes from local and regional displacement records, *Geophys. J. Int.*, 188, 703-718

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ABSTRACT OF THE DISSERTATION

**Using GPS to Rapidly Detect and Model Earthquakes and Transient Deformation
Events**

by

Brendan W. Crowell

Doctor of Philosophy in Earth Sciences

University of California, San Diego, 2013

Yehuda Bock, Chair

The rapid modeling and detection of earthquakes and transient deformation is a problem of extreme societal importance for earthquake early warning and rapid hazard response. To date, GPS data is not used in earthquake early warning or rapid source modeling even in Japan or California where the most extensive geophysical networks exist. This dissertation focuses on creating algorithms for automated modeling of earthquakes and transient slip events using GPS data in the western United States and Japan. First, I focus on the creation and use of high-rate GPS and combined seismogeodetic data for applications in earthquake early warning and rapid slip inversions. Leveraging

data from earthquakes in Japan and southern California, I demonstrate that an accurate magnitude estimate can be made within seconds using P wave displacement scaling, and that a heterogeneous static slip model can be generated within 2-3 minutes. The preliminary source characterization is sufficiently robust to independently confirm the extent of fault slip used for rapid assessment of strong ground motions and improved tsunami warning in subduction zone environments. Secondly, I investigate the automated detection of transient slow slip events in Cascadia using daily positional estimates from GPS. Proper geodetic characterization of transient deformation is necessary for studies of regional interseismic, coseismic and postseismic tectonics, and miscalculations can affect our understanding of the regional stress field. I utilize the relative strength index (RSI) from financial forecasting to create a complete record of slow slip from continuous GPS stations in the Cascadia subduction zone between 1996 and 2012. I create a complete history of slow slip across the Cascadia subduction zone, fully characterizing the timing, progression, and magnitude of events. Finally, using a combination of continuous and campaign GPS measurements, I characterize the amount of extension, shear and subsidence in the Salton Trough, one of the most complex zone of active faulting and seismicity in California. I show the implications that faulting in the Salton Trough has for the evolution of the Brawley Seismic Zone, and more importantly, the southern San Andreas fault.

Chapter 1

Introduction

1.1 Thirty Years of the Global Positioning System

Continuous Global Positioning System (CGPS) networks for geophysical applications began in the early 1990's on a global scale with the International GPS Service for Geodynamics (IGS) to provide a global terrestrial reference frame, and on a regional scale in the western United States to monitor plate boundary deformation and the hazards that they pose (e.g., *Bock et al.* [1997]; *King et al.* [1995]). Campaign mode GPS measurements during the 1980's proved valuable for studies of seismic cycle deformations in southern California [*Dong and Bock*, 1989; *Feigl et al.*, 1990], Alaska [*Beutler et al.*, 1987], Iceland [*Foulger et al.*, 1987] and South America [*Kellogg and Dixon*, 1990]; however, the limited temporal scale of the measurements motivated the transition to continuous operation. The limited CGPS network available in the early 1990's in southern California provided the first measurements of coseismic deformation and geodetic estimates of earthquake magnitude during the 1992 Mw 7.3 Landers earthquake [*Blewitt et al.*, 1993; *Bock et al.*, 1993]. The 1994 Mw 6.7 Northridge earthquake and resulting losses in an urban environment provided the rationale for a significant increase in the number of CGPS in southern California culminating in the 250-station Southern California Integrated GPS Network (SCIGN) project under the auspices of the Southern California Earthquake Center (SCEC). Other networks were established

throughout the 1990's across the active margins of the western United States including the Bay Area Regional Deformation Network (BARD), the Basin and Range Geodetic Network (BARGEN), the Pacific Northwest Geodetic Array (PANGA), the Alaska Deformation Array (AKDA), and the Eastern Basin-Range and Yellowstone Hotspot GPS Network (EBRY) for earthquake and volcano monitoring. EarthScope's Plate Boundary Observatory (PBO) project merged significant portions of the existing CGPS infrastructure into the PBO network with 1100 CGPS stations [Jackson, 2003]. Uniform GPS station design including highly stable anchored monuments, antenna mounts and radomes, and station enclosures were adopted and refined by PBO using the SCIGN project as a model, resulting in a robust crustal motion monitoring infrastructure. As such, the PBO and its predecessors evolved into an indispensable tool for geophysical monitoring alongside traditional seismic networks.

The dissemination of vast numbers of CGPS stations around the world since the early 1990's has led to an increased need for automated methods in time series analysis. In recent years, the discovery of slow slip events along the Cascadia margin [Dragert *et al.*, 2001; Rogers and Dragert, 2003] has further motivated the desire to investigate transient signals in GPS time series. Transient deformation is broadly defined as deformation that is not associated with traditional earthquakes, and generally has a rupture speed much smaller than the shear wave velocity [Ohtani *et al.*, 2010]. Many times, this deformation is associated with non-traditional earthquakes such as tremor or is completely aseismic. Moreover, there is usually a rather large discrepancy between the observed geodetic and seismic moment (e.g., Lohman and McGuire [2007]), which can have profound impacts on our understanding of long-term seismic hazards. The geodetic community has recognized the problem as extremely important such that it has become the topic of a recent series of workshops at the Southern California Earthquake Center [Murray-Moraleda and Lohman, 2010]. The goal of the workshops has been to develop automated criteria for identifying and modeling transient deformation to aid in regional earthquake hazards studies. Proposed methods of automated transient detection have included, but are not limited to, the network inversion filter [Segall and Matthews, 1997] and the subsequent network strain filter [Ohtani *et al.*, 2010], principal compo-

ment analysis [Dong *et al.*, 2006], covariance descriptor analysis [Kedar *et al.*, 2010], and Gaussian wavelet transforms [Melbourne *et al.*, 2005].

Since their inception, the primary operating mode for CGPS networks has been daily download of data sampled at 15-30 seconds, and production of daily position time series and custom position solutions after seismic events using many hours of data to estimate interseismic, coseismic and postseismic deformation. Since then, many of the CGPS stations in the western United States have been upgraded to high-rate (1 Hz or higher - HRGPS). Using the full spectrum of HRGPS measurements is generally termed "GPS Seismology" (e.g., Larson [2009]) as opposed to only using the derived coseismic displacements ("Earthquake Geodesy"). In the following section I provide a comprehensive overview of numerous science and engineering advances made with HRGPS.

1.1.1 High-Rate GPS Science and Engineering Advances

Earthquake Source Modeling

Nikolaidis *et al.* [2001] provided the first estimates of seismic waves measured by HRGPS during the 1999 Hector Mine earthquake. Since then, many advancements and improvements of our understanding of the earthquake source have arisen from HRGPS. Miyazaki *et al.* [2004] looked at HRGPS during the 2003 Tokachi-oki earthquake (M8.3) and saw good agreement with integrated accelerometer records except the HRGPS measurements perform better at low frequencies. Other studies have found comparable results for the 2003 San Simeon (M6.5) [Ji *et al.*, 2004; Wang *et al.*, 2007], the 2004 Parkfield (M6.0) [Langbein *et al.*, 2005], the 2008 Iwate-Miyagi Nairiku (M6.9) in Japan [Yokota *et al.*, 2009], the 2008 Wenchuan (M7.9) in China [Shi *et al.*, 2010], and the 2010 El Mayor-Cucapah earthquakes (M7.2) [Zheng *et al.*, 2012]. Larson *et al.* [2003] and Bock *et al.* [2004] were able to measure teleseismic waves from the 2002 Denali earthquake in Alaska (M7.9) throughout North America and in southern California respectively. Since, Ohta *et al.* [2006] was able to measure surface wave passage as well as directivity from the 2004 Sumatra-Andaman earthquake (M9.2) at regional and teleseismic distances. More recently, Davis and Smalley [2009] used HRGPS mea-

measurements in North America from the same earthquake to determine the Love wave phase velocity dispersion curve in that area. *Kobayashi et al.* [2006] performed strong-motion, HRGPS, and joint inversions for the 2005 west off Fukuoka Prefecture earthquake (M7.0) and found the HRGPS measurements provided the best constraints on a slip asperity directly underneath one of the most heavily damaged zones. *Heki* [2011] commented on the uniqueness of HRGPS for source modeling in that it is able to reconstruct complex ruptures, such as the 2010 Maule, Chile earthquake (M8.8) that changed directions numerous times, and the 2011 Tohoku-oki earthquake (M9.0). *Avallone et al.* [2011] showed from analysis of the 2009 L'Aquila earthquake (M6.3) that very-high-rate (10-20 Hz) GPS can be very useful as strong-motion stations for earthquake source studies. *Avallone et al.* [2012] followed up with a study of the 2012 Emilia-Romagna seismic sequence that both the strong-motion and HRGPS stations were dominated by low frequency energy, an area in which HRGPS excels. A number of studies have looked at HRGPS inversions during the 2011 Tohoku-oki earthquake and have concluded similar or improved performance over traditional strong-motion studies [*Fukahata et al.*, 2012; *Wei et al.*, 2012; *Yue and Lay*, 2011]. Using this data, *Fukahata et al.* [2012] was able to image the rupture process of the event up-dip, eventually breaking the trench.

Tsunami Modeling

Blewitt et al. [2006] first showed the important role HRGPS has to play for tsunami early warning, postulating that if near-real-time data connections existed during the 2004 Sumatra-Andaman earthquake, a reasonable model could have been ascertained within 10-15 minutes. The framework for this operational system was provided by *Blewitt et al.* [2009]. This concept was later expounded upon by *Sobolev et al.* [2007], who showed that using a shield-array of HRGPS, one can provide a 10 minute warning time for harmful tsunamis. *Song* [2007] used HRGPS to characterize tsunami-genic earthquakes over traditional earthquakes by using near-field stations to measure continental slope displacements and tilts. *Ohta et al.* [2012] showed using pre-computed tsunami simulations and very simple HRGPS slip-inversions, they are able to provide a

reasonable model for the 2011 Tohoku-oki earthquake and tsunami. *Hill et al.* [2012] showed from HRGPS measurements during the 2010 Mentawai earthquake (M7.8) that the rupture needed to be very shallow, at a low dip angle, and very close to the trench, a result that matches observed run up heights.

Volcano Monitoring

Active volcanoes generally exhibit very long deformational cycles that are ideal for HRGPS measurements [*Larson et al.*, 2010]. *Mattia et al.* [2004] showed using rapidly deployed HRGPS stations (some of which were destroyed later) at the Stromboli Volcano in 2003, they were able to image for the first time the shallow magma chambers that give rise to the characteristic Strombolian explosive activity. The shallow plumbing system was also seen during the 2007 eruption at Stromboli, and measurements agreed well with tiltmeters nearby [*Bonaccorso et al.*, 2008]. Also during the 2007 event, *Patanè et al.* [2007] were able to visualize the deep origin of magma prior to eruption. *Cervelli et al.* [2006] and *Mattia et al.* [2008] were both able to characterize the chronology of a discrete deformational event during the 2005-2006 Mt. Augustine eruption in Alaska.

Load Responses of Large Buildings and Bridges

Celebi and Sanli [2002] first showed the utility of HRGPS to monitoring long-period structures. In this case, HRGPS was seen to have a distinct advantage over conventional monitoring using accelerometers due to the ability to measure displacement across the top of the building in real time. *Kogan et al.* [2008] showed a similar advantage on the Verrazano-Narrows bridge during the New York City marathon. In that case, the dichotomy of the two measurement types was stark; the accelerometers picked up the initial vibrations of the runners whereas the GPS was able to recover the full deflection of the bridge, both of which are useful to understanding the the long-term health of such structures. *Yi et al.* [2010] and *Yi et al.* [2012] showed in bridges and tall buildings respectively that HRGPS has a significant role to play in the long-term

monitoring of such structures and provides important information of the thermal, wind and seismic response. In *Bock et al.* [2011], they were able to show the precision of combined HRGPS and accelerometer data was able to exactly mimic the input response for a series of shake table tests on top of a seven story building, further emphasizing the need for multiple geophysical measurements to appropriately characterize building response.

1.1.2 A Transition to Real-Time and Beyond

The achievements of HRGPS motivated the creation of real-time GPS networks (RTGPS), defined herein as high-rate (1-5 Hz or greater) with relatively low latency (less than a few seconds). One of the first RTGPS networks created was the California Real Time Network (CRTN) [*Aydin et al.*, 2007; *Genrich and Bock*, 2006] where, starting in 2002 and in collaboration with the surveying community [*Andrew*, 2003], more than 150 stations from SCIGN (USGS and SOPAC) and PBO have been upgraded to a continuous stream of 1 Hz with a latency of about 0.5 s in southern California and in the Parkfield area. The NASA supported READI project (Real-time Earthquake Analysis and Disaster Mitigation Network) has been responsible for the upgrade of more than 550 stations along the west coast of the US (<http://sopac.ucsd.edu/projects/realtime/READI/>). The Japanese network GEONET [*Miyazaki et al.*, 1998] operated by the Geospatial Information Authority of Japan (GSI) is the most ambitious foray into RTGPS with over 1200 stations at an optimal 20-40 km spacing. RTGPS has since been deployed along most of the world's active plate boundaries.

Bock et al. [2000] introduced the method of instantaneous positioning, which was used by *Nikolaidis et al.* [2001] to produce the first GPS total displacement waveforms. In this method, dual-frequency integer-cycle phase ambiguities and positions are resolved independently at each epoch for a network of GPS receivers. Positions are computed relative to one fixed station in the network. *Langbein and Bock* [2004] reported 6 mm horizontal precision at the 99% confidence level for a few seconds to a few hours of 1 Hz instantaneous positions, with an order of magnitude less precision in

the vertical. *Genrich and Bock* [2006] demonstrated that instantaneous positioning at 10-50 Hz using modern, geodetic-grade GPS receivers, is comparable in precision with lower frequency (1 Hz) sampling. They concluded that very-high frequency response is flat ("white noise") above 0.05-0.5 Hz (< 2 -20 s). Instantaneous positioning has distinct advantages over precise point positioning (PPP) [*Zumberge et al.*, 1997] due to the canceling out of common terms (such as satellite orbits, clocks, and troposphere delays). PPP requires very accurate estimates of satellite orbits, clocks, and other higher-order terms in order estimate positions properly. However, since instantaneous positioning is performed in a relative network setting, seismic motions throughout the network will cause possible biases in the estimated station positions due to the lack of a stable reference frame, whereas stations processed with PPP are independently processed with respect to a global reference frame. The development of PPP with ambiguity resolution (PPP-AR) takes the strengths of both these systems to produce a highly accurate position estimate without dealing with local reference frame issues [*Ge et al.*, 2008; *Geng et al.*, 2012, 2013; *Laurichesse et al.*, 2009]. In short, a larger regional reference network is created and processed in a relative sense to estimate individual station phase ambiguities, which are then used on the local scale with ultra-rapid orbits and processed with PPP. This leads to a much simpler computational problem and allows for full positioning at the station, ideal for an earthquake early warning system.

While the dissemination of RTGPS over the past decade has been significant, there are still very few examples of effectively using the data in real time. *Crowell et al.* [2009] demonstrated that RTGPS could be used to detect the onset of large earthquakes, locate their hypocenters, and roughly model the event on the order of a few minutes after the initial earthquake rupture. *Melgar et al.* [2012] used the static offsets recorded after an earthquake to compute a point-source centroid moment tensor solution within minutes of a large earthquake. This methodology was extended to a line-source by *Crowell et al.* [2013a] for the 2011 Tohoku-oki earthquake (M9.0) to provide a more robust moment tensor estimate with significant variance reduction. *Crowell et al.* [2012] showed that reliable, heterogeneous finite-fault slip inversions could be obtained within a few minutes of an event, important for emergency first responders and tsunami inundation

modeling. *Allen and Ziv* [2011] determined how to rapidly ascertain static offsets using the short-term-average over long-term-average method with RTGPS, and concluded that RTGPS provides important constraints for earthquake early warning.

The importance of RTGPS data to the earthquake early warning community has been explicitly stated as necessary for large events with very long ruptures since traditional seismic early warning methodologies break down for great earthquakes [*Böse and Heaton*, 2010], however, no efforts have been made to include RTGPS into earthquake early warning, even in Japan with the most complete RTGPS network to date. *Blewitt et al.* [2006] showed how powerful GPS is for rapidly ascertaining a true moment magnitude for the 2004 Mw 9.2 Sumatra-Andaman earthquake as opposed to seismic methods which may take several hours. In addition to seismic hazard monitoring, RTGPS is a valuable tool for volcano monitoring, tsunami modeling, water-cycle observations, troposphere modeling, and space weather problems [*Hammond et al.*, 2011]. The German Indonesian Tsunami Early Warning System (GITEWS) is under development to combine real-time positioned GPS buoys and ocean bottom pressure sensors in order to provide near-field measurements of the tsunami wave before coastal inundation occurs [*Schöne et al.*, 2011]. This will provide ample warning time to locals, assist emergency first responders, and aid in updating models of slip in real time.

Although GPS excels in providing critical estimates of static offsets, GPS derived dynamic motions by themselves are not accurate enough to identify mm-level or even smaller amplitude P waves. Furthermore, P wave arrivals have most of their energy in the vertical direction, making it more difficult for GPS because of the significantly less precise vertical component (e.g., *Bock et al.* [2000]). This has motivated the creation of seismogeodetic data through the optimal combination of RTGPS and high-rate accelerometer data. RTGPS data are noisier than accelerometer data and usually are collected at a lower sampling rate (e.g., 1-5 Hz vs. 100-200 Hz). However, GPS performs much better in the lower frequency end of the displacement spectrum [*Emore et al.*, 2007; *Genrich and Bock*, 2006] and has the advantage of capturing the coseismic displacement. Combining accelerometer and GPS displacement estimates on the fly using a multi-rate Kalman filter [*Bock et al.*, 2011; *Smyth and Wu*, 2006] takes advantage of

the strengths of the individual measurements while minimizing their weaknesses. *Bock et al.* [2011] demonstrated that one can obtain seismogeodetic broadband displacement waveforms at the sampling rate of the accelerometers with sub-mm level displacement uncertainties and velocity uncertainties of < 0.1 mm/s. Seismogeodetic waveforms can be observed close to the source without clipping as with GPS-only waveforms. However, the precision is such that P waves are visible in the both the displacement and velocity records in all three components (north, east and vertical), which is not achievable by GPS alone. Typical earthquake early warning systems (e.g., *Allen et al.* [2009]; *Gasparini et al.* [2007]) depend on conventional seismic instruments, and employ P wave detection to predict the arrival and intensity of destructive S and surface waves [*Allen and Kanamori*, 2003; *Heaton*, 1985; *Nakamura*, 1988]. *Crowell et al.* [2013b] showed possible deterministic earthquake rupture using P wave displacements from seismogeodetic waveforms for 7 earthquakes. *Geng et al.* [2013] showed that using the seismogeodetic approach for events during the 2012 Brawley Seismic Swarm, one can obtain P and S wave arrivals out to regional distances for magnitudes down to 4.6.

Since their inception, GPS and seismic networks have been developed independently, leading to little overlap between station locations. One approach to achieving a seismogeodetic capability is to upgrade the existing GPS stations with low-cost MEMS accelerometers, which have been shown by extensive large outdoor shake table tests to be adequate for this purpose (e.g., *Bock et al.* [2011]). Prototypes are currently being tested in southern California. Many of the ideas presented herein will hopefully motivate the seismic and geodetic communities to integrate seismogeodesy into earthquake early warning and rapid response.

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Chapter 2

High-Rate Global Positioning System Measurements for Earthquake Early Warning and Rapid Hazard Assessment

2.1 Abstract

Responses to recent great earthquakes and ensuing tsunamis in Sumatra, Chile, and Japan, with the resulting loss of life and damage to infrastructure demonstrate that our ability to ascertain the magnitude, source, and full extent of slip of catastrophic earthquakes and their tsunamigenic potential in the first seconds to minutes after the initiation of rupture, even with modern seismic instrumentation, is problematic. Earthquake early warning and rapid characterization of large earthquake are important for mitigation of seismic hazards. Saturation of near field seismometers and problems with the double integration of accelerometer data into displacements in real time make it difficult to achieve timely and accurate magnitude estimates and fault slip models. Regional GPS networks such as those in western North America and Japan are complementary

to seismic networks by being able to directly measure displacements close to the source during large earthquakes in real time. I highlight rapid detection, source characterization, and P wave detection for real-time GPS and seismogeodetic data during large earthquakes in Japan and southern California. The preliminary source characterization is sufficiently robust to independently confirm the extent of fault slip used for rapid assessment of strong ground motions and improved tsunami warning in subduction zone environments. The methods reduce the time to identify and characterize large earthquakes correctly by an order of magnitude over traditional seismic methods currently in place.

2.2 Introduction

Rapid detection and characterization of medium-to-large earthquakes is important for earthquake early warning and rapid hazard assessment to mitigate risk and assist emergency first responders. Earthquake early warning systems (EEWS) currently exist in Japan [Nakamura, 1989], Mexico [Espinosa-Aranda *et al.*, 1995], Taiwan [Wu and Kanamori, 2005], Turkey [Erdik *et al.*, 2003], and Romania [Böse *et al.*, 2007; Wenzel *et al.*, 1999], and are under development in California [Allen and Kanamori, 2003; Heaton, 1985] and the western United States. However, EEWS have relied solely upon traditional seismic instrumentation. Seismic early warning of events requires continuous monitoring of a suite of stations looking for changes in key parameters, most notably the P wave displacement (P_d), the predominant period of the P wave (τ_c), or the maximum predominant period of the P wave (τ_p^{max}). With seismic data, displacement has to be obtained by an integration of the broadband sensors, or a double integration of the strong motion sensors. Due to the bandwidth and the dynamic range limits of seismometers the accuracy of the derived displacements is poor, and generally require high-pass filtering. Seismic instruments are also affected by sensor tilts, long-term drifts, and DC offsets. Experience with the great earthquakes and ensuing tsunamis of the last decade has shown that traditional seismic monitoring is lacking in its ability to rapidly esti-

mate accurate earthquake magnitude and fault slip parameters in the near field because broadband seismometers clip and accelerometer data cannot be objectively integrated in real time [Boore and Bommer, 2005]. These problems are not present in displacements measured by continuous GPS instruments. While the seismic measurements provide a powerful constraint on the much noisier GPS measurements, unlike the seismometer, the GPS receiver measures displacement directly and never clips making it particularly sensitive to large earthquakes, which is the main focus of EEW. Continuous GPS instruments can be used to directly measure dynamic displacements ("GPS Seismology" - Nikolaidis *et al.* [2001]; Larson *et al.* [2003]) as well as static (coseismic) displacements ("Earthquake Geodesy" - Bock *et al.* [1993]; Blewitt *et al.* [1993]), from which earthquake magnitude and its source can be inverted.

Modeling of the spatial extent and variability of fault slip following large earthquakes is a very common practice in geodesy and seismology as part of studies of earthquake physics and crustal rheology. Methods tend to focus on post processing after assembling all relevant observations followed by forward and/or inverse modeling to best fit the data and underlying physical assumptions. Currently, the USGS's ShakeMap and PAGER alerts system utilizes simple slip inversions from seismic data sources for earthquake early warning and to assist emergency first responders, researchers, and the general public. GPS networks have the advantage of capturing motions throughout the entire earthquake cycle whether or not these are accompanied by seismic shaking. Blewitt *et al.* [2006] postulated that if near-real-time connections to GPS data had existed during the 2004 Mw 9.2 Sumatra-Andaman earthquake, even with the low density network at the time, a proper seismic moment and tsunami warning could have been made within 15 minutes, potentially saving thousands of lives. Crowell *et al.* [2009] demonstrated that real-time GPS (RTGPS) could be used to detect the onset of large earthquakes, locate their hypocenters, and roughly model the event within a few minutes after the initial earthquake rupture. Contrast this with W phase inversion from the recent 2011 Mw 9.0 Tohoku-oki earthquake that took 20 minutes to obtain a reasonable magnitude estimate because of the reliance on teleseismic data [Duputel *et al.*, 2011]. Melgar *et al.* [2012] showed that centroid moment tensors (CMT) can be computed us-

ing the static offsets in RTGPS time series, with similar accuracy to traditional CMT methodologies in a fraction of the time. *Crowell et al.* [2012] extended CMT solutions as one way to define the fault plane for finite-fault slip inversions. Furthermore, *Bock et al.* [2011] showed that RTGPS and accelerometer data can be combined in real time to estimate very high-rate broadband strong motion displacements (otherwise known as seismogeodetic displacements) with sufficient accuracy (mm-level) in all three components to be able to detect near-source P waves for Mw 6.0+ earthquakes.

In this chapter, I highlight the advances made in RTGPS and seismogeodesy in order to better characterize the earthquake source for earthquake early warning and rapid hazard assessment.

2.3 Data and Network Adjustment

2.3.1 GPS Data Processing

We focus our efforts on 1 Hz GPS data collected in real time during the 2003 Mw 8.3 Tokachi-oki, the 2010 Mw 7.2 El Mayor-Cucapah, and the 2011 Mw 9.0 Tohoku-oki earthquakes, as well as data from four events during the 2012 Brawley Seismic Swarm (M4.59 to 5.44). The data from the largest three events were processed using the method of instantaneous positioning within subnetworks of stations that utilizes dual-frequency, double-differencing of the GPS phase information as well as estimating dual-frequency integer-cycle phase ambiguities on an epoch-by-epoch basis [*Bock et al.*, 2000]. The subnetworks are created through a Delaunay triangulation and processed while keeping one station fixed and allowing the other two to freely move. For the 2003 Tokachi-oki earthquake, we used 10 minutes of data from 356 stations within the GEONET network [*Miyazaki et al.*, 1998], most located on Hokkaido Island, with some on Honshu Island. For the 2010 El Mayor-Cucapah earthquake, we utilized 95 stations within CRTN [*Genrich and Bock*, 2006]. For the Tohoku-oki earthquake, we processed 785 stations throughout Honshu and Hokkaido Islands. It is of note that for these three earthquakes, the GPS networks are not ideally located around the event.

The Tohoku-oki and Tokachi-oki earthquakes both occurred 100 km offshore and the El Mayor-Cucapah earthquake occurred in northern Baja California, Mexico, where no RTGPS stations exist currently, and about 75 km southeast of the closest RTGPS station in southern California.

For the Brawley Seismic Swarm, we processed data from four earthquakes (Mw 4.59, 5.32, 5.44 and 4.61) on 26 August 2012 at three stations (P506, P494, and BOMG). The stations and events are shown in Figure 2.1. To process the data, we used the method of precise point positioning with ambiguity resolution (PPP-AR) [Ge *et al.*, 2008; Geng *et al.*, 2012, 2013; Laurichesse *et al.*, 2009]. The basic theory of this processing method holds true to the underlying tenets of traditional PPP [Zumberge *et al.*, 1997] (i.e. estimate precise satellite orbits and clocks), but leverages the strengths of relative positioning in determining non-integer phase ambiguities that arise mostly from atmospheric delays. In short, a larger regional reference network is created and processed in a relative sense to estimate individual station phase ambiguities, which are then used on the local scale with ultra-rapid orbits and processed with PPP. A detailed overview of the method as used in the Brawley Seismic Swarm is provided in Geng *et al.* [2013]. We choose the PPP-AR method for these events due to the small displacements incurred by these earthquakes and future plans to transition processing to the station rather than a central processing facility as is required with instantaneous positioning.

2.3.2 Network Adjustment

For network positioning, stations are processed in a subnetwork, with positions computed with respect to a single station within the subnetwork. This method allows for the estimation of integer-cycle phase ambiguities and zenith tropospheric delays [Blewitt, 1989; Dong and Bock, 1989]. Analyzing the data as a network results in the cancellation of GPS receiver clock and satellite clock errors that are common across the network of stations. Instantaneous positioning is a subset of network positioning; the underlying difference between the two methods lies with the resolution of single-epoch integer-cycle phase ambiguities in instantaneous positioning [Bock *et al.*, 2000]. How-

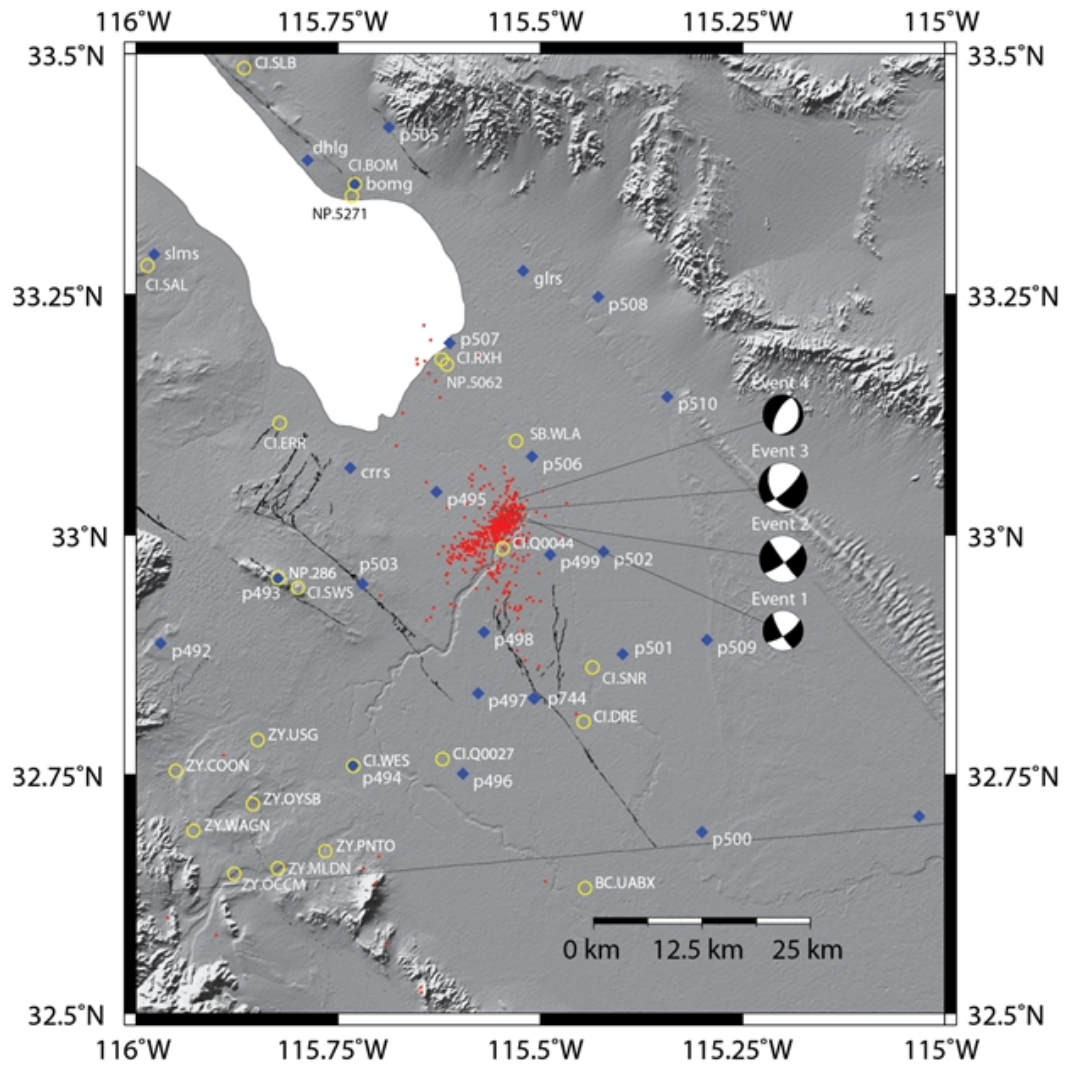


Figure 2.1: Brawley seismic swarm (red dots) surrounded by real-time continuous GPS stations (blue diamonds) and available continuous strong motion stations (open yellow circles) for the period of the swarm. The focal mechanisms are those computed by the Southern California Earthquake Center for the 4 events considered in this study (Mw 4.59, 5.32, 5.44, and 4.61 for Events 1-4 respectively). From *Geng et al.* [2013].

ever, processing the data in a relative sense is not intuitive for understanding earthquake cycle deformations due to large internal deformations required within the subnetwork in order to confidently characterize the deformation. Because of this, we employ a least squares network adjustment that takes the baseline measurements within each sub-network and corrects them to a single, far-away station in the network. The network adjustment solution is simply $Ax = b$, where A is an $n + 1$ by m design matrix, x is a vector of positions of length m , and b is a vector of baselines of length $n + 1$. The plus 1 in the b vector corresponds to the fixed station in the subnetwork, where the true-of-date ITRF position for the station is input into the baseline vector. A simple example with five stations would be

$$\begin{bmatrix} 1 & 0 & 0 & 0 & 0 \\ 1 & -1 & 0 & 0 & 0 \\ 0 & 1 & -1 & 0 & 0 \\ 1 & 0 & 0 & -1 & 0 \\ 0 & 0 & 0 & 1 & -1 \end{bmatrix} \begin{bmatrix} x_1 \\ x_2 \\ x_3 \\ x_4 \\ x_5 \end{bmatrix} = \begin{bmatrix} x_1 \\ x_1 - x_2 \\ x_2 - x_3 \\ x_1 - x_4 \\ x_4 - x_5 \end{bmatrix} \quad (2.1)$$

During a large earthquake, the selection of the reference station is fairly complex, but as long as the station is sufficiently far away (several 100s of km), has good station history (i.e. stability of monument, no latency and data completeness issues), and is contained within a subnetwork that has relatively short baselines, it will serve to track the strongest ground motions closest to the hypocenter. For real time integration, we would have a list of good candidate reference stations, from which a reference station would be selected based on its distance to the hypocenter.

This methodology has been successfully implemented in real time for CRTN and also applied to three earthquakes: the 2003 Mw 8.3 Tokachi-oki, the 2010 Mw 7.2 El Mayor-Cucapah, and the 2011 Mw 9.0 Tohoku-oki. For the Tokachi-oki earthquake, 1-Hz data was utilized from Japan's GEONET throughout Hokkaido and Honshu islands. The network was subdivided into an ideal network through Delaunay triangulation, and positions were computed using the instantaneous positioning approach described earlier. Even though many stations in the near-field lost connectivity after a few minutes, the net-

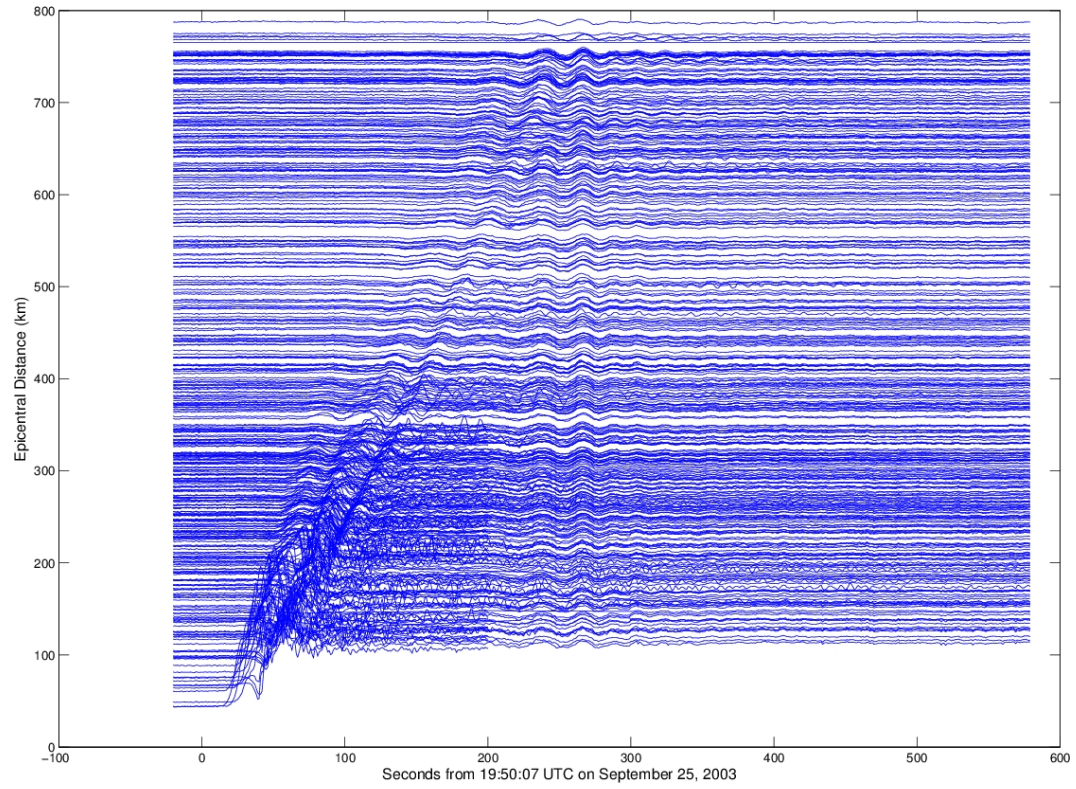


Figure 2.2: East displacement for all GEONET stations during the 2003 Tokachi-oki earthquake ordered as a function of epicentral distance.

work adjustment encountered no problems throughout the entirety of seismic shaking, as evidenced in Figure 2.2.

Figure 2.3 shows all of the stations of CRTN during the 2010 Mw 7.2 El Mayor-Cucapah earthquake in northern Baja California. As can be seen, there are small scale errors that get propagated throughout the entire network due to large signals in a few baseline measurements. However, these signals are small and can be removed through smoothing or combining accelerometer data through a multi-rate Kalman filter.

The third example shows truly how powerful RTGPS can be. Figure 2.4 is the displacements at station CVHS in southern California after the 2011 Tohoku-oki earthquake, clearly showing the S wave and Love wave arrivals at 1400 and 2000 seconds. Surface waves have been detected before at teleseismic distances using post-processed high-rate GPS (e.g. *Larson et al.* [2003] and *Bock et al.* [2004] for the 2002 Mw 7.9 Denali earthquake), but never in a simulated real-time environment. However, these measurements are still too noisy for P wave detection, but this can be mitigated by using a seismogeodetic approach of combining GPS and accelerometer data.

2.4 Strain Analysis in Subnetworks

Because of the small magnitudes of displacement during the initiation of a large earthquake, it is advantageous to use strain for anomaly detection. Strain is a measure of the internal deformation of an object and it is independent of reference frame. It also has the added benefit of only requiring relative length changes, which is what is computed using instantaneous relative positioning. After computing baseline length changes within the Delaunay triangulated subnetworks, we follow the methods of *Feigl et al.* [1990] to compute the two principle components of the strain rate within the triangles. We first start by defining the deformation of the triangle at each vertex. We define the length of a side $\Delta x_0^{(m)}$ for m between 1 and 2. Here, $x_0^{(0)}$ is the initial distance from the center of the triangle for the reference vertex, which can be decomposed to north

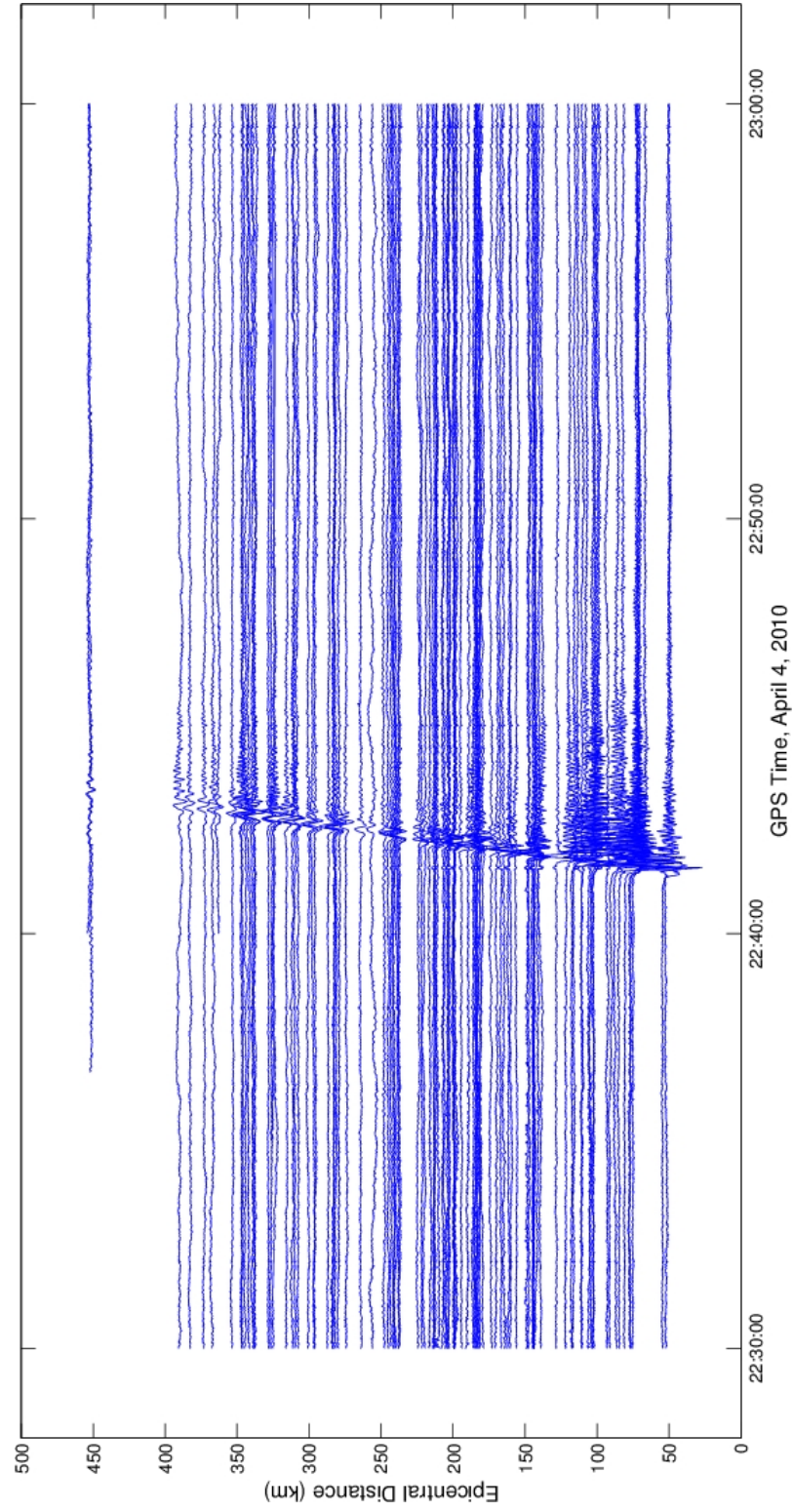


Figure 2.3: North displacement for all CRTN stations during the 2010 El Mayor-Cucapah earthquake ordered as a function of epicentral distance.

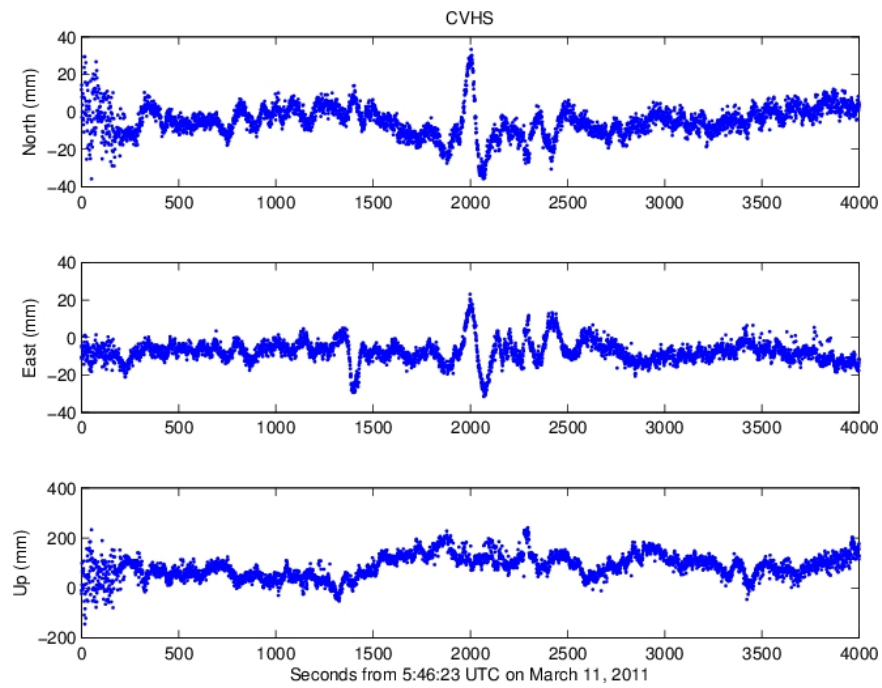


Figure 2.4: The north, east and up displacements at station CVHS in southern California during the 2011 Tohoku-oki earthquake.

and east components. The value $\Delta x_0^{(m)}$ is the length of side m at time zero.

$$\Delta x_0^{(m)} = x_0^{(m)} - x_0^{(0)} \quad (2.2)$$

We then relate the velocity to the velocity gradient tensor, L , by

$$u^{(m)} = x^{(0)} + L(x_0^{(m)} - x_0^{(0)}) \quad (2.3)$$

where $u^{(m)}$ is the difference between the length of the side m at the current time minus the length of the side m at time zero such that

$$u^{(m)} = \Delta x_1^{(m)} - \Delta x_0^{(m)} \quad (2.4)$$

The velocity gradient tensor can be split into two parts: a strain rate tensor and spin tensor

$$L = \dot{E} + \dot{W} \quad (2.5)$$

For the two dimensional case, the velocity gradient tensor is expressed as

$$L = \begin{bmatrix} \dot{E}_{ee} & \dot{E}_{en} \\ \dot{E}_{en} & \dot{E}_{nn} \end{bmatrix} + \begin{bmatrix} 0 & \dot{w} \\ -\dot{w} & 0 \end{bmatrix} \quad (2.6)$$

We then directly solve for the velocity gradient tensor in equation 2.3 by creating a velocity and displacement tensor in 2-D such that $U = LX$

$$U = \begin{bmatrix} \Delta u_e^{(1)} & \Delta u_n^{(1)} \\ \Delta u_e^{(2)} & \Delta u_n^{(2)} \end{bmatrix} \quad (2.7)$$

$$X = \begin{bmatrix} \Delta x_e^{(1)} & \Delta x_n^{(1)} \\ \Delta x_e^{(2)} & \Delta x_n^{(2)} \end{bmatrix} \quad (2.8)$$

We then plug equations 2.7 and 2.8 into equation 2.3 to obtain the velocity gradient tensor

$$L = \frac{1}{\det(X)} \begin{bmatrix} \Delta x_n^{(2)} \Delta u_e^{(1)} - \Delta x_n^{(1)} \Delta u_e^{(2)} & \Delta x_n^{(2)} \Delta u_n^{(1)} - \Delta x_n^{(1)} \Delta u_n^{(2)} \\ \Delta x_e^{(1)} \Delta u_e^{(2)} - \Delta x_e^{(2)} \Delta u_e^{(1)} & \Delta x_e^{(1)} \Delta u_n^{(2)} - \Delta x_e^{(2)} \Delta u_n^{(1)} \end{bmatrix} \quad (2.9)$$

The normal and shear strains are

$$\dot{E}_{ee} = \frac{\Delta x_n^{(2)} \Delta u_e^{(1)} - \Delta x_n^{(1)} \Delta u_e^{(2)}}{\det(X)} \quad (2.10)$$

$$\dot{E}_{nn} = \frac{\Delta x_e^{(1)} \Delta u_n^{(2)} - \Delta x_e^{(2)} \Delta u_n^{(1)}}{\det(X)} \quad (2.11)$$

$$\dot{E}_{en} = \frac{\Delta x_e^{(1)} \Delta u_e^{(2)} - \Delta x_e^{(2)} \Delta u_e^{(1)}}{\det(X)} + \dot{w} \quad (2.12)$$

The rotation rate is given as

$$\dot{w} = \frac{(\Delta x_n^{(2)} \Delta u_n^{(1)} - \Delta x_n^{(1)} \Delta u_n^{(2)}) - (\Delta x_e^{(1)} \Delta u_e^{(2)} - \Delta x_e^{(2)} \Delta u_e^{(1)})}{2\det(X)} \quad (2.13)$$

The two eigenvectors are computed with respect to the azimuth to obtain the two principle components of strain rate. The azimuth is determined from the Mohr's Circle representation.

$$\hat{e}_1 = \begin{bmatrix} \cos\theta \\ -\sin\theta \end{bmatrix} \quad (2.14)$$

$$\hat{e}_2 = \begin{bmatrix} \sin\theta \\ \cos\theta \end{bmatrix} \quad (2.15)$$

$$\dot{\epsilon}_i = \hat{e}_i^T \dot{E} \hat{e}_i \quad (2.16)$$

From equations 2.14-2.16 and 2.6, we obtain the two principle components of strain

$$\dot{\epsilon}_1 = \dot{E}_{ee} \cos^2\theta + \dot{E}_{nn} \sin^2\theta - 2\dot{E}_{en} \sin\theta \cos\theta \quad (2.17)$$

$$\dot{\epsilon}_2 = \dot{E}_{ee} \sin^2\theta + \dot{E}_{nn} \cos^2\theta + 2\dot{E}_{en} \sin\theta \cos\theta \quad (2.18)$$

The azimuth, given by θ , is defined as the angle from north to the ϵ_2 axis

$$\theta = \frac{1}{2} \tan^{-1} \left(\frac{\dot{E}_{nn} - \dot{E}_{ee}}{2\dot{E}_{en}} \right) \quad (2.19)$$

After computing the principal components of strain within each triangle, we set a threshold of 40 μ strain/s and issue alarms every time a threshold exceedance is obtained. Figure 2.5 shows the principal components of strain during the 2003 Mw 8.3 Tokachi-oki earthquake, 60 seconds after earthquake initiation. Strain threshold exceedance occurs well before the strongest shaking occurs, making it a perfect first-order warning system. This point becomes clearer in Figure 2.6 when looking at strain threshold exceedance

8 seconds after a hypothetical Mw 7.8 earthquake along the southern San Andreas fault for the 2008 ShakeOut experiment [Jones *et al.*, 2008]. Strong shaking during the event is not expected in the Los Angeles basin for almost two minutes while strain is exceeded within seconds of the event. This is the general basis behind EEWS: characterize large events before strong shaking occurs.



Figure 2.5: Delaunay triangulation of GEONET stations on Hokkaido Island and the two principal components of strain in each triangle one minute after the 2003 Mw 8.3 Tokachi-oki earthquake. From Crowell *et al.* [2009].

2.5 Hypocenter Determination

The determination of hypocenter using GPS was first outlined in Crowell *et al.* [2009] for the 2003 Mw 8.3 Tokachi-oki earthquake offshore Hokkaido Island in Japan. The method used was rather straightforward and uses two separate grid searches, first to obtain the proper time offset and then secondly to obtain the appropriate location, which is similar to that used by Shearer [2009].

The method starts when four or more GPS stations exceed preset criteria of total

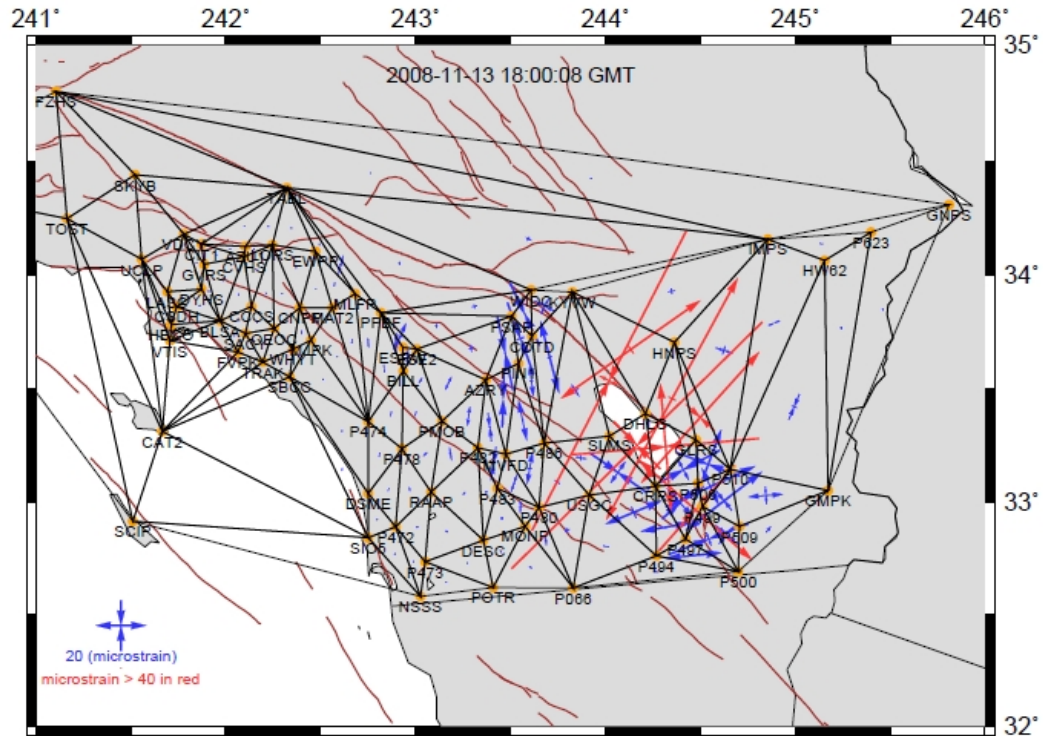


Figure 2.6: Delaunay triangulation of the two principal components of strain rate at 18:00:08 GMT during the 2008 ShakeOut simulated earthquake, using total displacement waveforms computed at each CRTN station. The earthquake initiates at 18:00:00 GMT near station DHLG and stops near station TABL. The red lines represent the triangles in which the strain rate has exceeded the threshold of 40 μ strain and the blue lines are under the threshold. From *Crowell et al.* [2009].

displacement. For the threshold, 0.1 meters offers a good compromise because it is 5-10 times greater than the precision of single station displacement estimates [Genrich and Bock, 2006] but still small enough that the timing will not be too late after the P wave. The first station to exceed the displacement limit is set to $t = 0$. Prior to hypocenter determination, a preset grid is created for the area which encompasses the entire geographic area down to 100 km depth in the global ITRF reference frame with 5 km grid spacing. Once four stations have exceeded the displacement limit, I computed the distance from each GPS station to every grid point in the local grid, defined as d . Then the L_1 norm of the predicted time difference of arrival between the four GPS stations is minimized, which is the distance difference divided by the average wave speed, v , minus the actual time difference of arrivals, t , between the four GPS stations such that

$$r_{min} = \min \left(\sum_{i=1}^3 \sum_{j=2}^4 \left| \frac{d_j - d_i}{v} - (t_j - t_i) \right| \right) \quad (2.20)$$

The value of r_{min} is associated with a distance from the earthquake source to the first GPS station, which is then divided by v to obtain the time offset, t_o , of earthquake rupture. Then the L_2 norm of the actual arrival time minus the predicted arrival time from each local grid point is minimized.

$$s_{min} = \min \left(\sqrt{\sum_{i=1}^4 \left| t_i - t_o - \frac{d_i}{v} \right|^2} \right) \quad (2.21)$$

The value of s_{min} is associated with an ITRF position in the grid, which is the hypocenter estimate. The hypocenter estimate for the 2003 Tokachi-Oki earthquake is shown in Figure 2.7. The epicenter estimate from the GPS measurements is roughly 23 km and 43 km away from the JMA and USGS epicenter estimates respectively. The depth of the earthquake was computed to be 5.9 km, which is much shallower than the 42 km and 27 km estimates from the JMA and USGS respectively.

There are a few caveats with this method using 1-Hz GPS measurements. The first is the low sampling rate effectively rounds up the arrival time at each GPS station, which can easily cause errors in the hypocenter measurement on the order of the wave

speed. The second major cause for concern is the displacement threshold of 0.1 m. While this seems to be a good determinant from visual inspection of records that a large earthquake has started to affect the GPS station, this may not be representative of the actual arrival time of the earthquake, further making the estimate error prone. This problem can be alleviated by using combined GPS and accelerometer data [Bock *et al.*, 2011] to pick out the actual P wave arrivals. The third major source of error is the selection of grid spacing of 5 km. A smaller grid would provide more accuracy, but this increases the computation time. For example, when running this grid search algorithm in MATLAB for the 2003 Tokachi-Oki earthquake, a grid spacing of 10 km takes roughly 0.5 seconds, but a grid spacing of 5 km takes about 3 seconds.

2.6 Rapid Slip Inversions

2.6.1 Methodology

We investigate two different methods of rapidly inverting for earthquake slip using real-time GPS data collected in the near field of large earthquakes. One method utilizes pre-defined fault planes from a fault catalog while the second starts with the fault planes from a separate point source moment tensor inversion as an initial approximation. Otherwise, both methods are identical and obtain their Green's functions from *Okada* [1985] to solve for slip on a finite fault in a homogeneous elastic half-space.

The first approach is a simple finite-fault slip inversion using a predefined fault plane and a generalized regularization equation, which we refer to as the Inverse Method (IM). For the Tokachi-oki earthquake, we use the Slab1.0 [Hayes *et al.*, 2012] plate interface of the Kuril-Japan Trench around Hokkaido Island down to a depth of 200 km and subdivided into 30 along-strike and 20 along-dip segments. We constrain movement on the sides and bottom of the fault to zero to ensure a realistic slip distribution. For the El Mayor-Cucapah earthquake, we define a 200 km long vertical fault centered along the Laguna Salada fault that is 30 km deep.

The inversion utilizes a first-order Tikhonov regularization which aims to reduce

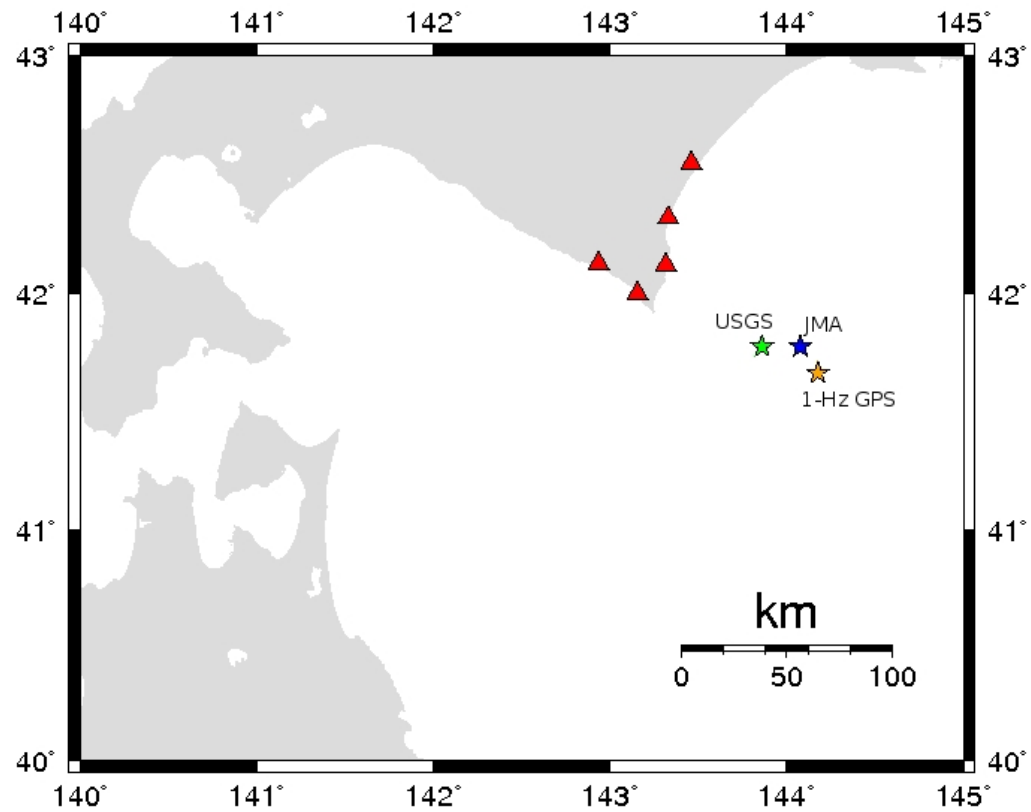


Figure 2.7: Hypocenter determination for the 2003 Mw 8.3 Tokachi-oki earthquake off the coast of Hokkaido Island, Japan. The triangles represent the first five stations to have more than 0.1 m of total displacement during the earthquake sequence, and thus used in determining the hypocenter of the earthquake. The stars represent the USGS and JMA epicenters determined using seismic data and out estimated hypocenter using 1-Hz GPS data. From *Crowell et al.* [2009].

the first derivative in slip between adjoining segments such that

$$\begin{bmatrix} WG \\ \lambda T \end{bmatrix} S = \begin{bmatrix} Wu \\ 0 \end{bmatrix} \quad (2.22)$$

where G are the Green's functions from *Okada* [1985], T is the smoothness matrix, S is the slip on the fault patch (we solve for both strike-slip and dip-slip), W is the weighting matrix, λ is the smoothness constant, and u is the time-averaged GPS data. We generalize the smoothness constant and define it as

$$\lambda = \frac{1}{mn_t^2 |WG|} \quad (2.23)$$

where n_t is the number of rows (or nearest neighbors on the grid) in the smoothness matrix and m is a constant determined from thousands of synthetic iterations to be between 4 and 20. We choose a value of 4 for m , which errs on the side of a smoother model. Equation 2.23 is important because it allows for variable smoothness as the station configuration and fault geometry changes, which is essential for real-time integration.

The weight matrix incorporates weighting the displacements by their standard deviation and inverse distance weighting. We assume that the north and east components have a standard deviation of 1 while the vertical component is equal to 5, essentially to give less credence to the vertical component. *Langbein and Bock* [2004] reported that for 1-Hz real-time GPS data, the rms scatter of the vertical component is about 5 times larger than the horizontal components, so this is a reasonable assumption. The distance weighting algorithm is similar in nature to the one used in time domain waveform moment tensor inversions [*Dreger, 2003*] except we weight by $1/r^2$ to conform more to the natural decay of permanent deformation [*Aki and Richards, 2002*].

For data input, we employ a moving average of 50 seconds over each component of motion for every station and set the three components of motion to zero if the horizontal motion at a station is less than 15 mm, which is about 3 times the expected one-sigma precision for single-epoch instantaneous GPS positions in the horizontal components [*Genrich and Bock, 2006; Langbein and Bock, 2004*]. This is important because a small spurious motion at a station far from the fault will lead to a large estimate of slip during

the inversion. There is also a tradeoff between the amount of averaging, the accuracy of the final solution, and the time taken to obtain the final magnitude. Figure 2.8 shows the tradeoff between different moving averages, the time taken to reach the final solution, the magnitude overrun, and the final root-mean-square (rms) of the data fit for both earthquakes. From Figure 2.8, we determine that the moving average should be between 30 and 60 seconds for optimal performance. For this exercise, we choose a moving average time of 50 seconds to coincide with the rms minimum for the Tokachi-oki earthquake.

In our second approach, we solve for fault slip with no a priori information on the fault geometry, a method we call the CMT Method (CM). We start by computing a CMT to find a point source model that solves for the hypocenter, strike, dip, rake and magnitude of the event using a moving average of the GPS displacements, as outlined in *Melgar et al.* [2012]. The only difference is that we used a shorter moving average window than the more conservative 120 s used by *Melgar et al.* [2012]. While we use the CMT solution computed from our GPS displacements, this is not necessary. We could just as easily implement the CM using a rapid CMT solution from other sources, for example from accelerometers, if it was available. All that is necessary is a reasonable estimate of source parameters from which fault planes are created.

We create two fault planes (main and auxiliary) utilizing the strike, dip and hypocenter information; the one that minimizes the L_2 -norm of the model fit for each epoch is chosen. To create the fault planes, we use the scaling relationships from *Dreger and Kaverina* [2000] to determine the fault plane size based on the CMT solution. While the point source approximation for the CMT begins to fail in the near-field for large earthquakes, we only require a rough estimate of the fault zone geometry; our method then computes a more complex heterogeneous finite slip distribution. We find that the scaling of *Dreger and Kaverina* [2000] is sufficient for the along-strike dimension, but the along-dip dimension for the Tokachi-oki earthquake is inadequate due to the edges of the fault affecting the slip distribution. Because of this, we use 75% of the along-strike dimension for the along-dip dimension; however, a larger along-dip dimension is acceptable. For the El Mayor-Cucapah earthquake, we use the scaling from *Dreger and*

Kaverina [2000] directly. The corresponding fault is then subdivided along-strike and along-dip to roughly match the fault patch sizes from the IM. The rest of the CM process is exactly the same as the IM.

To determine when the final stable solution is obtained for the two methods, we looked at the trailing horizontal variance of the station with the largest motion using the previous 50 seconds, computed every second. During the period of large ground shaking, the trailing variance will grow until it reaches a maximum value. After the intense shaking, the trailing variance slowly drops. When it drops down to 10% of the maximum value, we call the solution final. A similar empirical approach is used by *Melgar et al.* [2012] to ascertain the final CMT solution.

2.6.2 The 2003 Tokachi-oki Earthquake

The 2003 Tokachi-oki earthquake occurred on the Japan-Kuril Trench on 25 September 2003 at 19:50:07 UTC [*Yamanaka and Kikuchi*, 2003], causing significant ground motions of over a meter, permanent displacements up to 0.5 m on land [*Crowell et al.*, 2009], and a local tsunami with run-up heights as large as 4 m [*Tanioka et al.*, 2004]. Results from previous studies are rather varied, with seismic moments between $M_0 = 1.0 \times 10^{21}$ Nm and 3.0×10^{21} Nm (M_w 8.0 to 8.3), maximum slip between 4 and 7 m, and rakes between 90° and 130° [*Honda et al.*, 2011; *Miura et al.*, 2004; *Tanioka et al.*, 2004; *Yagi*, 2004; *Yamanaka and Kikuchi*, 2003]. For the purposes of this study, we assume a shear modulus of 30 GPa for computing seismic moment. The final results for the two schemes, and the evolution of the rms of the residual and moment magnitude are shown in Figure 2.9. The peak trailing variance for station 0521 (42.9389°N , 143.1706°E) is reached in 83 seconds and is reduced by 90% at 116 seconds, at which point we call our solution final.

The final moment magnitudes for the two scenarios are 8.23 for the IM and 8.28 for the CM, and both methods exceed M_w 8.0 at about 40 s from the earthquake onset. The maximum slip obtained is 4.0 m and 2.2 m for the IM, and CM scenarios respectively. The distribution of slip tends to be more diffuse and smaller in magnitude than

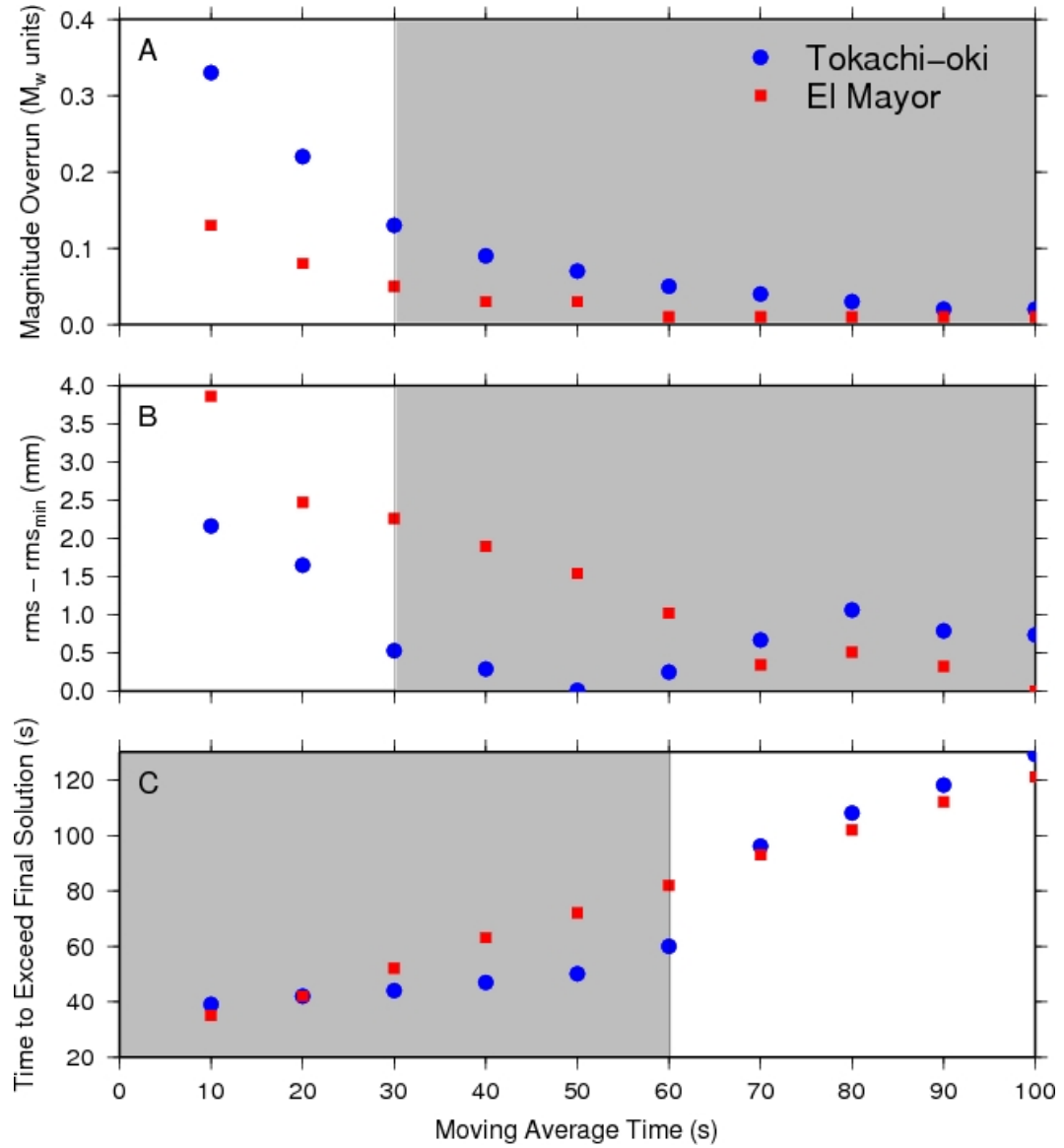


Figure 2.8: (A) Moment magnitude overrun over final magnitudes, (B) rms minus minimum rms, and (C) time to exceed the final magnitude vs. moving average time for the 2003 Tokachi-oki earthquake and 2010 El Mayor-Cucapah earthquake. The grey area on each plot indicates the ideal range for each of the three parameters. From *Crowell et al.* [2012].

in previous studies [*Crowell et al.*, 2009; *Miura et al.*, 2004; *Yagi*, 2004; *Yamanaka and Kikuchi*, 2003]. However, our results agree with the areas of maximum slip, have similar average rake vectors (103° for IM and 88° for CM), and the slip extent is on the same order of magnitude as previous studies (~ 100 km by 100 km). The GPS-derived CMT (gpsCMT) and the Global CMT (GCMT) are located very near to each other, indicating that the centroid of slip is well located within both methods. There are some issues with the CM method, mainly due to using a fault plane with a single dip angle. This has a tendency to allow greater slip at shallow depths and also results in a much larger slip patch than the IM. Also of note is the fact that we use a generalized regularization equation instead of finding the ideal regularization for each inversion. Our regularization tends to err on the side of more smoothness to prevent an overestimate of magnitude. This will cause the slip distribution to become more diffuse and have a lower maximum slip.

2.6.3 The 2010 El Mayor-Cucapah Earthquake

The Mw 7.2 El Mayor-Cucapah occurred on 4 April 2010 at 22:40:42 UTC, starting as a distinct normal faulting event and then bilaterally rupturing a series of northwest trending faults, most notably the Borrego and Pescadores faults [*Hauksson et al.*, 2011; *Wei et al.*, 2011]. Slip is thought to be shallow (less than 10 km), very localized with discrete slip patches, and reaching up to 6 m [*Wei et al.*, 2011]. The tectonic setting consist of a series of right-lateral and normal faults that accommodate the transition from the spreading centers within the Gulf of California to the transform boundary of the San Andreas fault system. Because of the nature of faulting in the area, selection of the appropriate fault plane is difficult, but not impossible, and may not be critical to the final solution.

As explained in the methodology section, for the IM, we used a generalized representation of the Laguna Salada fault, a fault which is parallel to the Borrego and Pescadores faults, but did not experience any slip during the earthquake [*Wei et al.*, 2011]. The Borrego and Pescadores faults were thought not to be active, so assuming the slip occurred on the Laguna Salada fault is more indicative of the IM method in

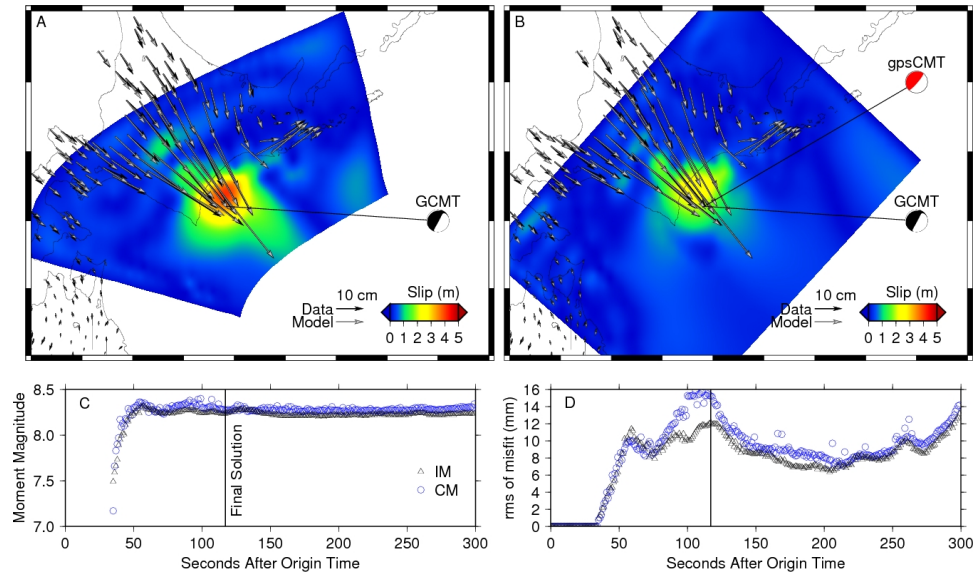


Figure 2.9: Results for the 2003 Tokachi-oki earthquake. The final slip at 116 seconds after 19:50:07 UTC on 25 September 2003 is shown for the IM (A) and CM (B). The colored slip profiles correspond to the surface fault lines on each plot. The location of the CMT solution computed using GPS to determine the fault plane in the CM is shown in (B) as gpsCMT. The Global CMT is indicated by GCMT. The moment magnitudes are 8.23 for the IM and 8.28 for the CM. The evolution of moment magnitude is in (C) and rms of the misfit between the data and model is in (D) as a function of time from the earthquake onset. From *Crowell et al.* [2012].

real-time given our knowledge at the time of the earthquake. The slip results for both methods and the evolution of moment magnitude and rms are shown in Figure 2.10. The peak trailing variance for station P494 (32.7597°N , 115.7321°W) is reduced by 90% in 113 seconds at which point our result is final.

The final moment magnitudes for the two scenarios are 7.16 for the IM and 7.22 for the CM. The maximum slip obtained is 1.4 m and 2.0 m for the IM and CM scenarios, respectively, which are considerably smaller than from Wei *et al.* [2011]. It should be noted that our solutions are very smoothed out over most of the fault, although our slip distributions are confined to the upper 10 km of fault and located near the GCMT epicenter location. Also, all of our stations are located north of the United States-Mexico border whereas the Wei *et al.* [2011] result uses SAR and SPOT imagery directly over the fault to constrain the fault slip at the surface. Some notable differences between the IM and CM come from the different fault planes. There is more deep slip in the CM, mainly due to the westward shifting of the fault compared to the fault used in the IM (note location of gpsCMT versus GCMT on Figure 2.10). Also, the magnitude of slip in the CM is greater than the IM because the along-strike dimension of the CM is 70 km shorter than the IM, which amplifies the magnitude of slip. The higher rms of the CM versus the IM indicates that the fault dimensions and strike for the CM is slightly less preferred than the IM, although both rms values are fairly low considering the magnitude of the GPS vectors (~ 8 mm over 95 stations). However, as Figure 2.10 indicates, the gpsCMT is better at locating the centroid of slip compared to the Global CMT, making the CM solution more centered along the fault. The rake angles for the two methods are 163° and 173° for the IM and CM respectively, falling in line with a mostly right-lateral earthquake with some reverse faulting.

2.6.4 Discussion

Both methodologies are able to rapidly model both earthquakes in under two minutes while keeping the solution stable and accurate. There are some concerns that need to be accounted for before full implementation can be achieved. The most serious

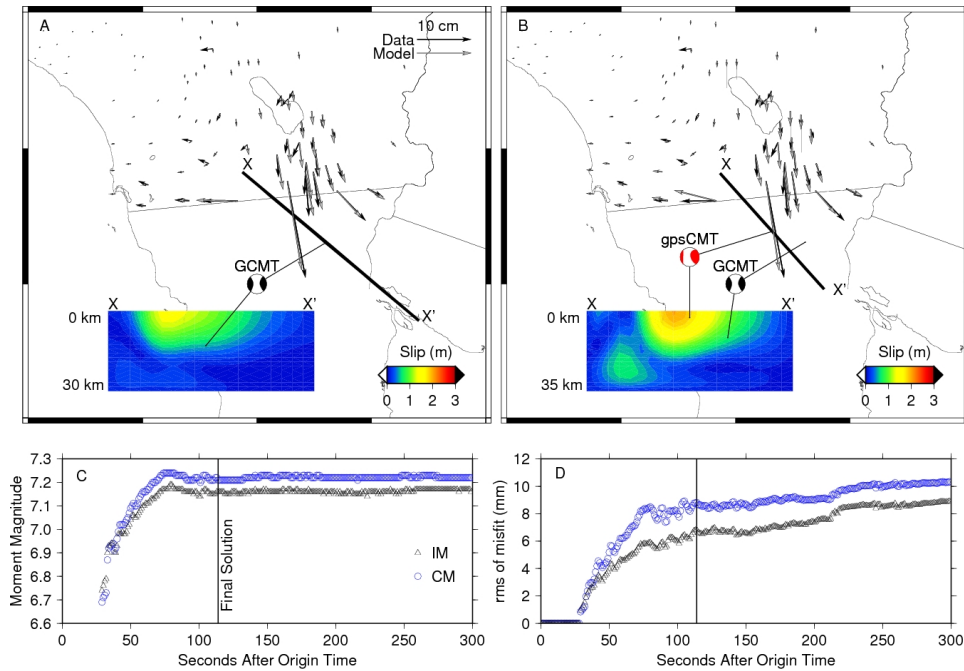


Figure 2.10: Results for the 2010 El Mayor-Cucapah earthquake. The final slip at 113 seconds after 22:40:42 UTC on 4 April 2010 is shown for the IM (A) and CM (B). The colored slip profiles correspond to the surface fault lines on each plot. The location of the CMT solution computed using GPS to determine the fault plane in the CM is shown in (B) as gpsCMT. The Global CMT is indicated by GCMT. The moment magnitudes are 7.16 for the IM and 7.22 for the CM. The evolution of moment magnitude is in (C) and rms of the misfit between the data and model is in (D) as a function of time from the earthquake onset. From *Crowell et al.* [2012].

concern for the IM is the determination of a good approximation of the correct fault plane. This is not a trivial problem in that this requires a system with pre-defined scenarios based on the environment the GPS network encompasses. In areas of complex faulting, there must be an inherent error that the user is willing to accept if the earthquake is located between two major faults or includes motion that is not common to that region. In megathrust environments, this is much less of a concern, although large events in the overriding plate and the outer-rise can occur, but it is unknown what the error would be if we simply placed these events on the main plate interface. This may be a more significant problem from a tsunami or landslide warning perspective than a rapid earthquake response perspective. Moreover, determining the proper fault dimensions for the CM in the along-dip direction needs to be investigated further since for megathrust earthquakes the spatial extent of slip along-dip will be significant and the point source assumption may be significantly violated. We have in fact observed that the point source CMT solution from *Melgar et al.* [2012] is problematic for the very large near-field displacements experienced during the 2011 Mw 9.0 Tohoku-oki earthquake. This has been addressed by further expansions of the CMT algorithm with line sources to account for finite extent sources [*Crowell et al.*, 2013a]. In modeling that event, we would just need to substitute the higher-order CMT solution into our CM method.

Computation time is not an issue for both approaches, running a single epoch in ~ 1 s on a single core PC workstation in MATLAB. Data transmission and processing latencies are currently on the order of ~ 1 s for CRTN, so a reasonable operational latency when these methods are integrated is ~ 2 -5 s. Furthermore, computation of the CMT as in *Melgar et al.* [2012] is as fast as the inversion for slip described in this paper so the fact that the CM method requires two steps adds very little delay, compared to the IM method.

Using a moving average on the GPS time series is a good method for removing any dynamic signal, but does delay the final solution for 20 to 30 seconds. As the catalog of large earthquakes observed with near-field GPS networks grows, we can refine scaling relationships between the peak ground displacements and final coseismic displacements to ascertain the final displacements very rapidly. *Crowell et al.* [2009]

demonstrated that the peak ground displacement for the 2003 Tokachi-oki earthquake was attained 10 to 15 seconds after the first arrival at coastal stations, which could speed up the final slip inversion by up to a minute. We can improve on this further by using the combined accelerometer/GPS data [Bock *et al.*, 2011] to look at traditional EEW parameters, although this requires many more seismic/GPS collocations than currently exist. A higher level product would be to use the full waveform for a kinematic inversion and thus obtain the source-time function of the earthquake. There are some difficulties in implementation as an inverse problem (mainly due to varying source-receiver time delays), but an iterative forward method could be implemented. Using these other methods, we would only be limited by the network geometry around an event and the rupture times. Rupture time can be improved upon by using a probabilistic approach; however, there are still difficulties in regards to large earthquakes [Cua and Heaton, 2007]. Furthermore, the sensitivity of RTGPS is such that the minimum earthquake size that can be reasonably estimated is $M_w \sim 6$, although with gains in precision made through the combination of accelerometer and RTGPS data [Bock *et al.*, 2011], this limit has been decreased down to $M 4.6$ [Geng *et al.*, 2013].

The ability to rapidly characterize an earthquake using near-source observations represents a powerful contribution to tsunami warning for those living in nearby coastal regions. Considering that for the Tokachi-oki earthquake we are able to compute a final stable solution within two minutes, a simple forward tsunami propagation model could have been computed to determine wave amplitudes in the near-field well before the first arrival, using available detailed bathymetry. From a hazards perspective, it is extremely important to differentiate between small and large tsunamis to aid in evacuation efforts and curb complacency within the general public.

2.6.5 Conclusions

Near-source real-time GPS measurements fill an important gap for early earthquake detection, characterization, and rapid response for medium and greater earthquakes where significant fault rupture occurs and tsunamigenic potential exists. The

Inverse Method performs better for the Tokachi-oki earthquake due to the varying dip angles with depth; however, the CMT Method performs well for both earthquakes and has the added benefit of requiring no a priori information on fault geometry, making it the preferred method in complex tectonic environments such as southern California. We obtained reasonable models for both earthquakes in terms of slip, magnitude, and rake estimates in about 2 minutes using both methods, an order of magnitude improvement compared to existing seismic methods for monitoring large earthquakes in the near field, thereby allowing for more effective earthquake response and tsunami warning.

2.7 The Seismogeodetic Solution: Example of the 2011 Tohoku-oki Earthquake

The need for more accurate and rapid estimates of the characteristics of large earthquakes has been poignantly illustrated by actual responses to recent great tsunami-genic earthquakes in Sumatra, Chile, and Japan, which to date have relied on traditional seismic methods [Cardenas-Jiron, 2012; Lay and Kanamori, 2011; Lay *et al.*, 2005, e.g.]. Broadband seismometers saturate in the near field and so magnitude estimation relies on teleseismic waves recorded much later at distant seismic stations. To overcome this limitation, seismic stations are augmented with strong-motion instruments (accelerometers). However, a double integration is required to convert accelerations into displacements, which is unreliable at low frequencies because tilts of the instruments are indistinguishable from translations [Boore and Bommer, 2005] and can be amplified in the integration. Subjective correction algorithms are available [Boore and Bommer, 2005] but the permanent (coseismic) deformation that accompanies large earthquakes may be filtered out or not estimated correctly in the process. These problems affect EEWS during large events [Allen *et al.*, 2009]; EEWS are based on detecting the arrival of primary (P) seismic waves and using that information at one or more stations to predict the arrival and intensity of destructive secondary (S) waves and surface waves [Allen and Kanamori, 2003; Allen *et al.*, 2009; Gasparini *et al.*, 2007; Heaton, 1985].

The current state-of-the-art in seismic monitoring is best illustrated in Japan, which has the most advanced earthquake and tsunami early warning system in the world. During the 2011 Mw 9.0 Tohoku-oki earthquake, the seismic EEWS operated as designed [Hoshiba *et al.*, 2011], and alerted residents up to a minute in advance that strong ground shaking was imminent. However, early estimates of earthquake magnitude from the Japan Meteorological Agency (JMA) were too low by 1-2 orders of magnitude; an estimate of M 7.2 was determined after 30 s and revised to M 8.0 by 107 s [Hoshiba *et al.*, 2011]. It took 20 minutes for a reliable estimate of magnitude to be made teleseismically through the W phase method [Duputel *et al.*, 2011], which cannot be applied in the near field of great earthquakes due to violation of the point source assumption. The first tsunami waves arrived near Sendai within 30 minutes of the mainshock, with the early estimates of wave heights underestimated by several meters [Ozaki, 2011]. Many of the coastal towns possess seawalls that can protect the residents from moderate tsunamis, but they could not protect against such large waves. A more accurate estimate of wave heights could have resulted from a better estimate of the size and geometry of the earthquake source, emphasizing why rapid estimation and characterization of large earthquakes is necessary [Lay and Kanamori, 2011].

Japan also possesses the largest real-time GPS network; the GEONET operated by the Geospatial Information Authority of Japan (GSI) [Miyazaki *et al.*, 1998], which contains over 1200 stations streaming 1 Hz data in real time. However, the national seismic and EEWS in Japan have not incorporated these data although high-rate GPS networks have been shown to provide estimates of permanent displacements not available from seismic monitoring, as well as dynamic motions without saturation [Bock *et al.*, 2004; Larson *et al.*, 2003]. Furthermore, it has been recently demonstrated that real-time GPS can reduce the delay time for obtaining source information for large earthquakes. By using the static displacements estimated as the earthquake rupture progresses, rapid estimates of centroid moment tensor (CMT) solutions [Melgar *et al.*, 2012; O'Toole *et al.*, 2012] and finite fault slip models [Allen and Ziv, 2011; Crowell *et al.*, 2009, 2012; Ohta *et al.*, 2012; Wright *et al.*, 2012] are feasible using one of several real-time GPS analysis approaches. It would seem that GPS has an important role to play in EEWS for

large ($>M7$) earthquakes. Nevertheless, operational early warning systems such as the one in Japan have not yet incorporated GPS data into rapid determination of earthquake magnitude. Real-time finite fault rupture models are still limited to seismic methods, although there are now thousands of real-time GPS stations in seismically active zones.

One factor precluding the operational use of real-time GPS is that displacement waveforms are much less precise than seismic measurements. The precision (and accuracy) of real-time GPS is about 1 cm in the horizontal components and 5-10 cm in the vertical whether using relative positioning or precise point positioning methods [Bock *et al.*, 2011; Genrich and Bock, 2006; Ohta *et al.*, 2012; Wright *et al.*, 2012]. GPS displacement waveforms are not sensitive enough to detect the arrival of P waves even in the near field of the largest earthquakes. A practical outcome is that today there is a general lack of overlap between seismic and geodetic networks.

GPS data are noisier than accelerometer data and usually are collected at a lower sampling rate (e.g., 1-5 Hz vs. 100-200 Hz). However, GPS performs much better in the lower frequency end of the displacement spectrum [Emore *et al.*, 2007; Genrich and Bock, 2006] and has the advantage of capturing the coseismic displacement. Combining accelerometer and GPS displacement estimates on the fly using a Kalman filter takes advantage of the strengths of the individual measurements while minimizing their weaknesses. Bock *et al.* [2011] demonstrated that one can obtain seismogeodetic broadband displacement waveforms at the sampling rate of the accelerometers with sub-mm level displacement uncertainties and velocity uncertainties of <0.1 mm/s. Seismogeodetic waveforms can be observed close to the source without clipping as with GPS-only waveforms. However, the precision is such that P waves are visible in the both the displacement and velocity records in all three components (north, east and vertical), which is not achievable by GPS alone (Figure 2.11). The resulting seismogeodetic displacements from 138 GPS/accelerometer pairs contain an unprecedented view of the progression of both dynamic and static ground displacements with sub-mm precision, which has not previously been available for earthquake studies. This can be seen clearly in the waveforms (Figure 2.11) from a collocated GPS instrument (GEONET 0914) and strong-motion accelerometer (K-NET MYG003) about 205km northwest of the 2011

Mw 9.0 Tohoku-oki epicenter (Figure 2.12).

Using regional seismogeodetic data for the 2011 Mw 9.0 Tohoku-oki earthquake we demonstrate the ability to rapidly estimate the earthquake hypocenter and magnitude from the detected P wave arrivals and to model total slip on the fault as the rupture and permanent deformation progresses. These abilities provide distinct advantages for mitigating earthquake and tsunami hazards.

2.7.1 Seismogeodetic Data Creation: The Kalman Filter Solution

To create seismogeodetic data, we follow the example of *Bock et al.* [2011] in combining high-rate GPS data with accelerometer data using a multi-rate Kalman filter, first proposed for structural engineering by *Smyth and Wu* [2006] and applied to bridge monitoring by *Kogan et al.* [2008]. The GPS processing methodology is irrelevant in this solution; all that is required is an estimate of the noise characteristics and high-rate measurements (1-Hz or greater). For further details on the approach as it applies to GPS/accelerometer combination, see *Bock et al.* [2011]. In short, this methodology uses the constraint of the GPS positions in the double integration of the accelerometer data.

For this earthquake, we use the GPS positions from the 785 stations throughout Honshu and Hokkaido Islands processed using instantaneous positioning. We divide the GPS network into subnetworks using a Delaunay triangulation scheme, and individual triangles are processed independently for relative station positions. The triangles are combined through a real-time network adjustment [*Crowell et al.*, 2009], and station positions are referenced to GEONET station 0848 on the northern tip of Hokkaido Island, 900 km northwest of the hypocenter. We then combine the GPS displacements with data from 138 accelerometers in the K-NET and KiK-net networks operated by Japan's National Research Institute for Earth Science and Disaster Prevention (NIED, *Aoi et al.* [2004]) that are within 1.5 km of a GPS station using the multi-rate Kalman filter.

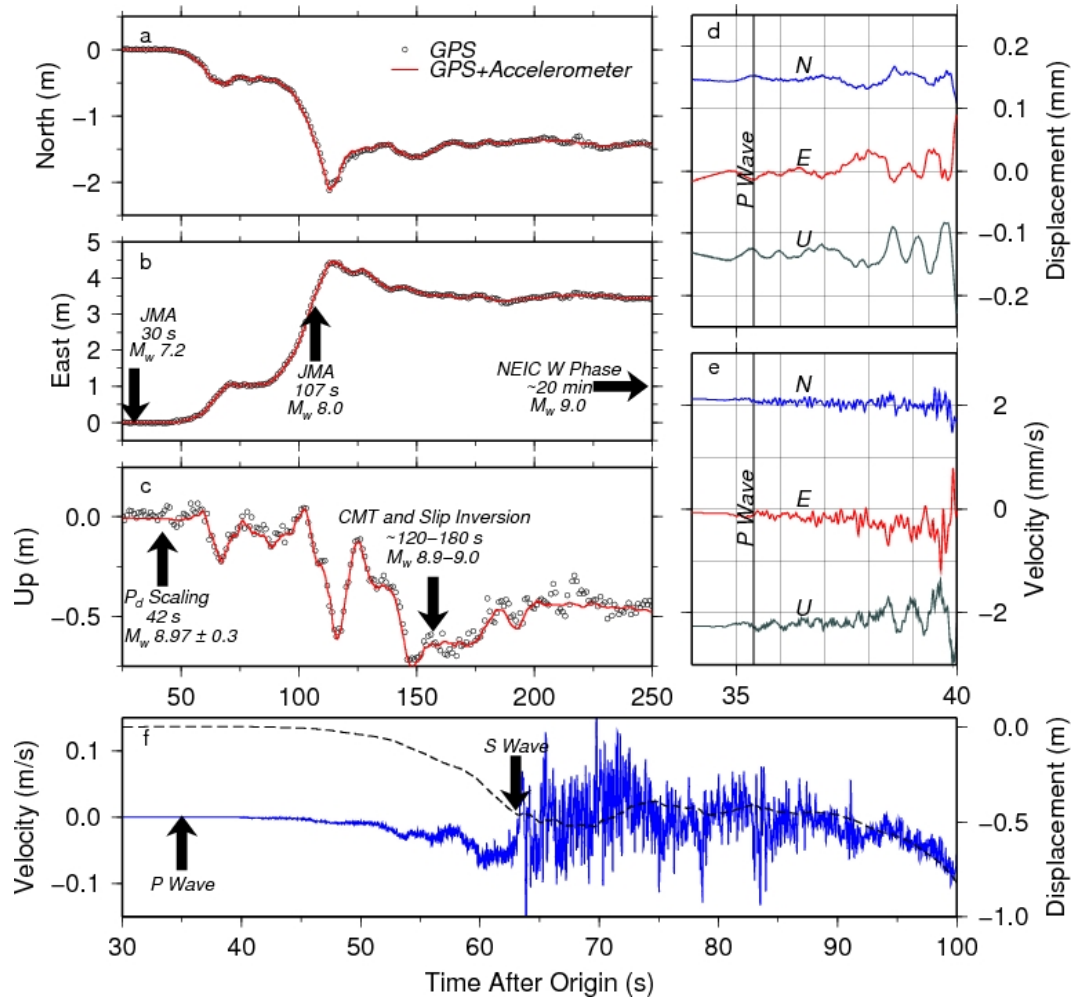


Figure 2.11: The north (A), east (B) and up (C) broadband displacements for GEONET GPS station 0914 (black circles) and K-NET accelerometer MYG003, 205 km from the hypocenter (location shown in Figure 2.12). The Kalman-filtered 100 Hz GPS and accelerometer solution is shown by the solid lines; the 1 Hz GPS-only solution is shown by circles. Magnitude estimates from the NEIC and JMA are shown in (B) and from the seismogeodetic methods in (C). (D) The combined displacements for the three components zoomed into the first five seconds after the P wave arrival at 35.4 s. We remove a cubic fit from each of the components to show details of the P wave arrival more clearly. (E) The combined velocities for the three components for the first five seconds after the P wave arrival. (F) The north velocity (solid) and displacement (dashed) records up to 100 seconds after the earthquake start showing the locations of the P and S wave arrivals and the progression of dynamic and static slip. From *Crowell et al. [2013b]*.

2.7.2 Rapid Source Modeling

Rapid earthquake response predicates itself on discerning the geometry and location of the seismic source. As demonstrated by the Mw 8.6 strike-slip event off Sumatra, Indonesia, on 11 April 2012 [Satriano *et al.*, 2012], one cannot simply assume all large earthquakes in a subduction setting rupture the mega-thrust. To this end, the first step in higher order modeling is to determine the CMT solution for the earthquake source, which indicates the style of faulting (e.g., strike, dip and rake) and magnitude. Near real-time CMT algorithms are usually based on regional broadband recordings [Dreger, 2003] that will clip in the near field of high magnitude events. To take advantage of the early availability of observations in the near field, we employ the fastCMT technique [Melgar *et al.*, 2012], which relies on rapid estimation of coseismic offsets at a stable time determined from a trailing variance technique. Static offsets are computed at 157 s after origin time using the seismogeodetic data from 138 collocated GPS and accelerometer stations from Japan’s K-NET and KiK-net [Aoi *et al.*, 2004] and ingested into the inversion algorithm. We find that the initial fastCMT yields a solution with a high magnitude and poorly located centroid because the large spatial extent of slip is not well approximated by a point source. Because static deformation modeling has no temporal dependence, we extend the fastCMT algorithm to include a linear superposition of point sources providing a line source estimate. We then invert for the moment tensors at these predefined points along the line source, and grid search for the proper geometry. We pre-compute Green’s functions at inversion nodes on a 0.1° by 0.1° by 2 km in depth grid to cover all of Japan with the same strategy as originally described in Melgar *et al.* [2012]. The line source length is 7° with 30 point sources. These numbers are of course arbitrary and must be tailored according to the observational goals of a particular network. We choose a line source long enough to fully encompass the largest expected events plus 2° .

We perform an L_1 minimizing inversion with first order Tikhonov smoothing for the five deviatoric moment tensor components. This first derivative regularization is along the line direction. Thus sharp variations in a particular component of the moment

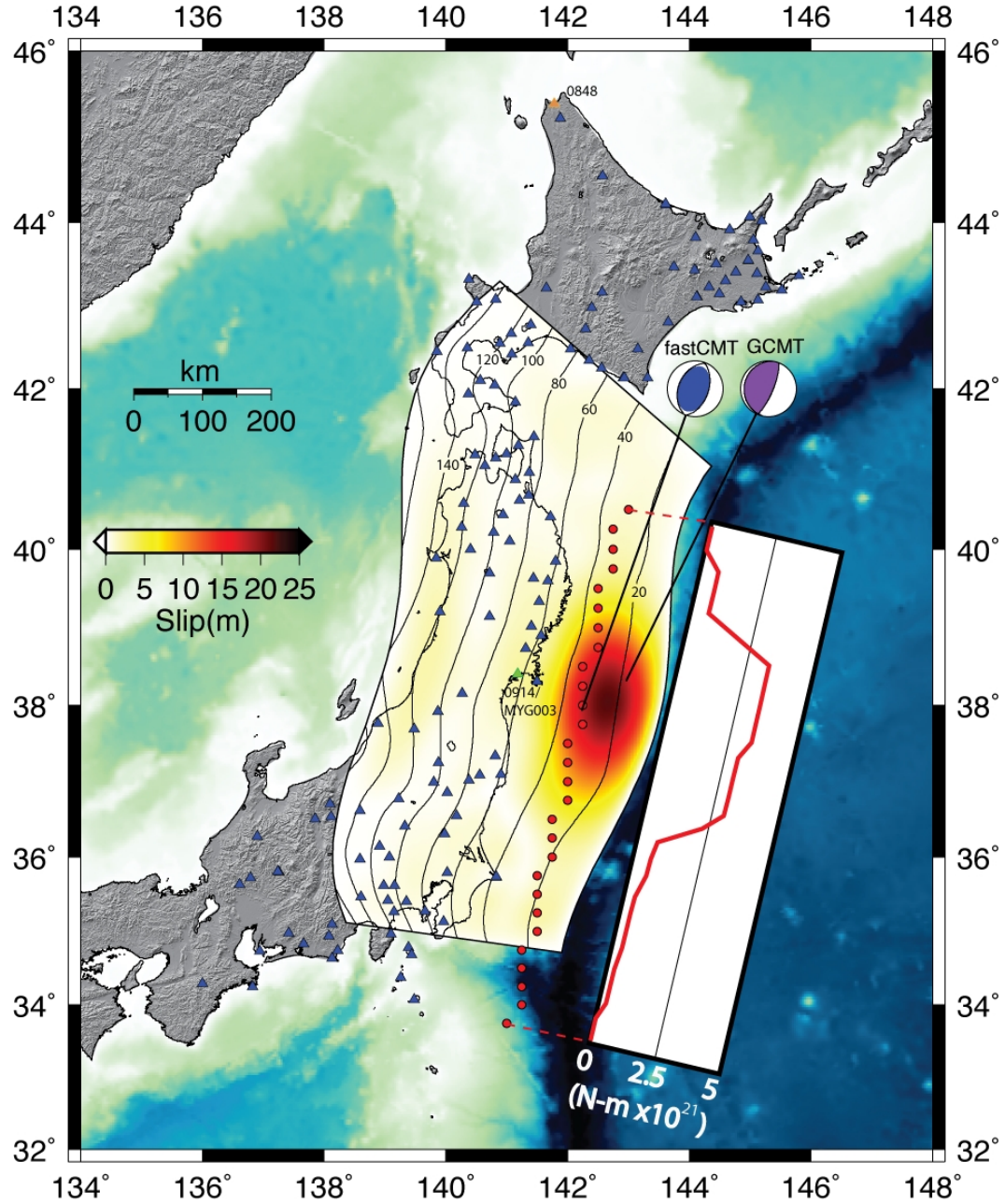


Figure 2.12: Red circles show the extent of the line source of CMT solutions, the final averaged solution shown as fastCMT, and the Global CMT solution shown for comparison. The inset shows the moment release from the line source of CMTs as a function of distance along fault. Shown along the fault interface with 20 km depth contours from the Slab 1.0 model [Hayes et al., 2012] is the result of the slip inversion. The triangles indicate the locations of all the GPS stations used for computing the slip inversion and CMT solution. The two colored triangles represent the fixed GPS station (0848) and the GPS/accelerometer pair (0914/MYG003) in Figure 2.11. From *Crowell et al.* [2013a].

tensor between neighboring point sources are penalized. We also experimented with L_2 inversions but found that they yielded results that were too smooth and did not bracket the main asperity in a fashion as sharply as the L_1 inversions.

To determine the line location and azimuth we invert for multiple combinations of line geometries, that is, we vary the origin of the line source, its azimuth and depth. The line source geometry that minimizes misfit is the best solution. Because the inversion is computationally simple we can perform thousands of such inversions without a large computational overhead.

The smoothing parameter (λ) controls the complexity of the inversion results, thus it is critical to have an objective way of determining it, such that no user interaction is required. For such real time determination of λ we compute the tradeoff curves of data misfit versus model norm (L curves) for all values of λ and their optimal corners for all possible source geometries. We then compute the non-parametric kernel smoothing density estimate of the probability density function (pdf) of all the L-curve corner λ 's. We select the mode of the resulting pdf as the preferred smoothing parameter λ^* . Then, we compare the misfit of all inversions at the λ^* smoothing level and, as in the point source fastCMT method, select the one with the smallest misfit. The final inversion yields a variance reduction of 87% (one minus variance of misfit divided by variance of data) with the GPS data alone and a slight improvement to 89% with the Kalman filtered seismogeodetic displacement data.

After finding the best fitting line source, we compute the weighted average of all moment tensors over that line source based on their moment release and place the centroid at the location of mean moment release to obtain a single CMT estimate, shown in Figure 2.12. The implementation of fastCMT is unique in that no a priori information is required on the fault geometry. The process is computationally simple, automatable, and results in a magnitude of 9.0, average strike, dip, and rake of 204° , 30° , and 95° respectively, with a source extent of 340 km along strike. It is interesting to point out that while the extended source CMT found the correct fault orientation without any a priori information, the initial W-phase CMT inversion found an incorrect strike of 230° , which delayed the rapid finite fault inversion and ultimately required re-alignment to the

Slab1.0 geometry [Hayes *et al.*, 2012]. The resulting source extent of 340 km is sufficiently close to the final extent that we infer it would have value for guiding the initial release of hazard maps, such as the USGS's ShakeMap and PAGER products [Hayes *et al.*, 2011].

After receiving information on the style of faulting and the hypocenter from the fastCMT and determining that it is a thrust event close to the slab, we invert for static slip using the same offsets used in the fastCMT using the method of Crowell *et al.* [2012] (Figure 2.12). For the static slip inversion, we first locate the closest fault segment from the Slab1.0 model to the line source [Hayes *et al.*, 2012]. The fault is subdivided into 40 along-strike and 20 along-dip segments. Green's Functions are computed for a homogeneous half-space [Okada, 1985]. We use a Laplacian smoothness constraint that is weighted by a factor that depends inversely on the data uncertainties, and the mean value of the Green's functions and which has been calibrated through synthetic tests [Crowell *et al.*, 2012]. Thus the factor is predetermined for the particular station configuration and fault geometry and it is not necessary to construct L-curves for a range of models, which would require a prohibitive amount of time for the real-time implementation. The data is smoothed using a 50 s moving average on all three components of motion to get a more stable estimate of the static offset as soon as possible after the dynamic motion ends. We constrain the edges of the fault to zero to ensure that non-physical motions do not exist in the model (i.e., step-function motion at the edges). Motion at the trench is constrained to zero due to lack of resolution because the network geometry is located exclusively to the west of the Japan trench. The inversion is solved using an L_2 -norm minimizing solution.

Our model indicates Mw 8.9, a slip patch that is roughly 340 km along-strike and 155 km along-dip, and maximum slip of 21.6 m located near the center of the fastCMT location at 28 km depth. Using seafloor geodesy, surveyed up to a month after the earthquake, static slip inversions produced estimates of maximum slip closer to 50 m [Sato *et al.*, 2011]; however, seafloor geodetic measurements are currently not available in real time and would include some sizable amount of postseismic deformation, so our model is indicative of a realistic real-time scenario using just land-based seismogeodetic

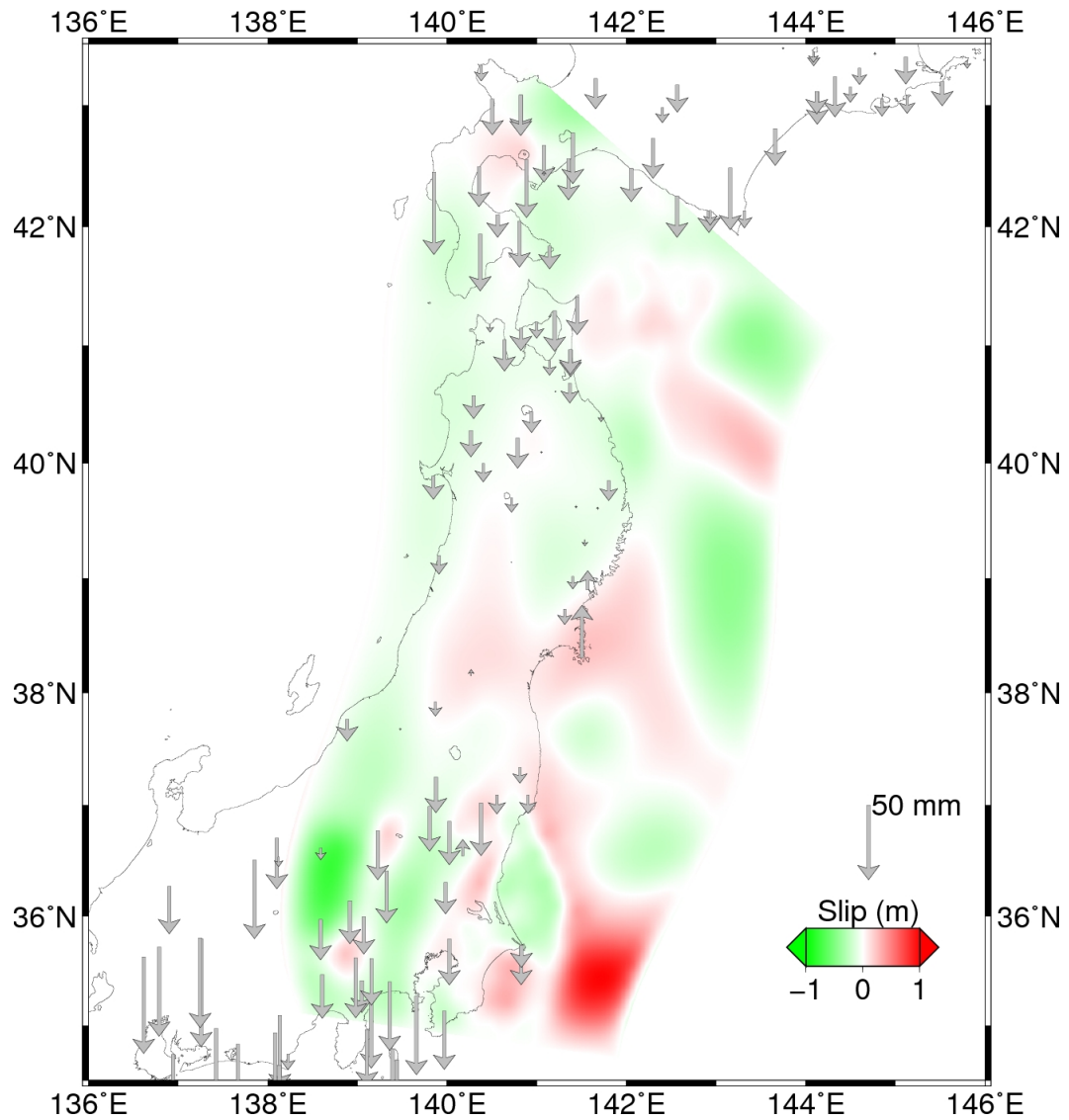


Figure 2.13: Total slip inversion difference between seismogeodetic and GPS-only solutions. The colors show the total slip difference along the fault plane between the seismogeodetic data and the GPS-only data, with positive values indicating more slip in the seismogeodetic inversion and negative values indicating more slip in the GPS-only inversion. The vectors show the difference in vertical deformation between the two solutions. From *Crowell et al. [2013a]*.

data. A joint inversion using seafloor pressure gauge data and GPS offsets was able to produce a maximum slip of 50 m [Simons *et al.*, 2011] and could possibly be available in real-time in the near future. The moment release from the line source fastCMT overlain upon the slip distribution in Figure 2.12 shows the slip distribution is bounded by the moment release along strike, indicating that both methods independently agree on the spatial extent of slip.

We also perform the fastCMT and slip inversion using just the GPS data for the same stations to investigate improvements with seismogeodetic data. We find that vertical bias (median difference of 17 mm) in the GPS-only solution leads to far more deep slip than the seismogeodetic solution (Figures 2.13 and 2.14) and a 5° difference in the average rake (83° for combined data and 88° for GPS-only data). There is little difference in the line source fastCMT solution between the GPS-only and seismogeodetic solution, due in part to the one dimensional nature of the line source (Figure 2.15). We conclude that the seismogeodetic dataset leads to a more robust real-time solution for rapid hazard modeling.

2.7.3 Applicability to Earthquake Early Warning: P_d Scaling

Real-time seismogeodetic displacement and velocity data can also be useful for EEW methods that are based on the amplitude (P_d) or predominant period (τ_c or τ_p^{max}) of the initial P wave [Nakamura, 1988; Wu and Kanamori, 2005]. The empirical relationships between magnitude and P_d and τ_p^{max} have been determined for several compilations of earthquakes in California [Tsang *et al.*, 2007], Japan [Brown *et al.*, 2011], and Taiwan [Wu and Kanamori, 2005; Wu *et al.*, 2006]. The use of P_d to predict magnitude has been particularly troublesome because the relationship appears to saturate at high magnitude, that is, for a small number of events with $M > 7$, in particular the 1999 Mw 7.6 Chi-Chi and 1999 Mw 7.1 Hector Mine earthquakes, P_d does not continue to increase as a function of magnitude [Brown *et al.*, 2011; Wu and Zhao, 2006; Wu *et al.*, 2006]. For Japan, the relationship appears to break down for $M > 8$ [Brown *et al.*, 2011]. This makes the use of P_d for seismic early warning unreliable without the use of predominant period,

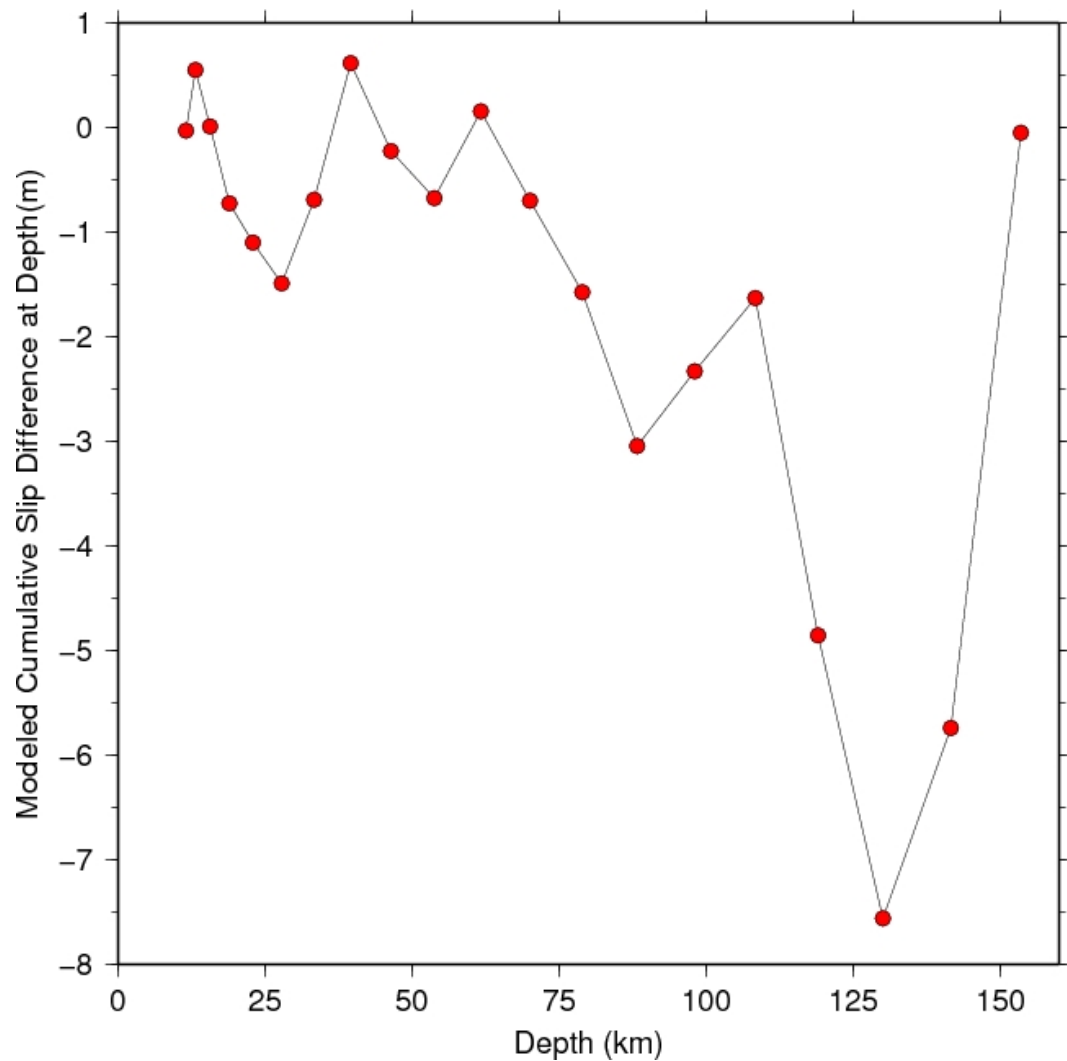


Figure 2.14: Slip inversion difference between seismogeodetic and GPS-only solutions at depth. Each point is the sum of all slip differences at each depth interval. Positive differences indicate more total slip in the seismogeodetic solution where negative values indicate more total slip in the GPS-only solution. From *Crowell et al.* [2013a].

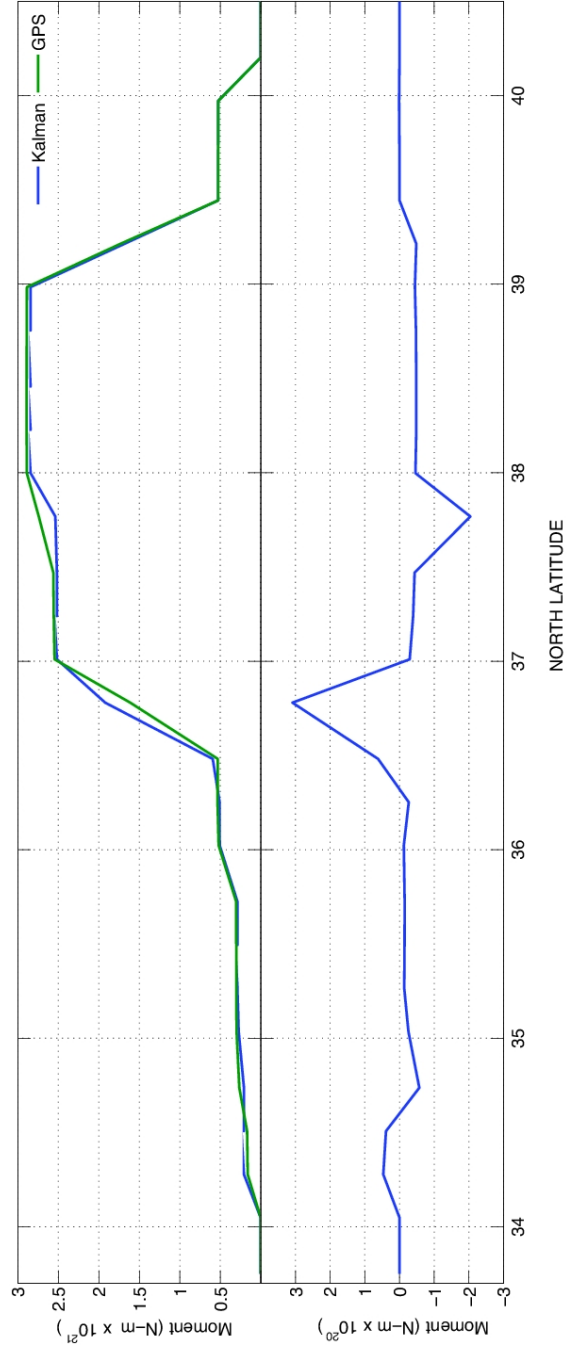


Figure 2.15: Top: Comparison of the moment release as a function of latitude for the GPS-only and seismogeodetic fastCMT inversion results. **Bottom:** Seismogeodetic minus GPS only moment release highlighting the small variations between the two results. From *Crowell et al.* [2013].

which is more robust and appears not to saturate for large events [*Brown et al.*, 2011; *Wu and Kanamori*, 2005].

The seismogeodetic displacements may provide a significant improvement in the scaling relationship with P_d because they contain the low frequency component of shaking for large amplitude signals, and do not require the typical high pass filtering at 0.075 Hz (13 s) that is applied to seismic records to assure stable integration to displacement. For large events at nearby stations, the amplitude of the P wave displacement monotonically increases with time within the few seconds, making it a robust estimate of earthquake size independently of the predominant frequency. First, we test P_d scaling using the first 3 seconds after the P wave arrival in the vertical direction as is done with traditional seismic P_d scaling (e.g., *Wu and Zhao* [2006]). We use 7 earthquakes: the 2011 Tohoku-oki (M9.0), the 2003 Tokachi-oki (M8.3), the 2010 El Mayor-Cucapah (M7.2), and 4 events from the 2012 Brawley Seismic Swarm (M4.59 to 5.44). P wave picks are made manually on the resultant vertical velocity waveforms from the Kalman filter combination of accelerometer and GPS data. Figure 2.16 shows the results of our scaling relationship with that derived by *Wu and Zhao* [2006]. Of note is that both regressions are fairly similar, with the chief difference being the attenuation of our regression being less steep. Also, for the larger events, there is much more scatter in the displacements which leads to less confidence in properly characterizing the magnitude of larger earthquakes. Because of these issues, we look at the first 3 to 5 seconds of displacement in the horizontal direction. For stations with S wave arrivals less than 5 seconds after the P wave arrivals, we use 3 second P_d scaling, otherwise 5 second scaling is utilized. The resultant P_d relationship appears to hold for large earthquakes (Figure 2.17) and the scatter is much improved. For the Tohoku-oki earthquake, we estimate $M 9.1 \pm 0.3$ within 39 s of origin time using a single station, or $M 9.0 \pm 0.3$ in 42 s with three stations. Our scaling relationship indicates that earthquake rupture may be deterministic [*Olson and Allen*, 2005]; however, we need to investigate the scaling of more large earthquakes to back up this claim. The lack of saturation indicates that the sensitivity to low frequencies of the seismogeodetic data provides important source physics information that cannot be obtained with seismic data alone. One possible ex-

planation for earthquake determinism is the probability of future rupture is proportional to the initial stress drop (i.e. large earthquakes have a larger initial energy release) [Zollo *et al.*, 2006].

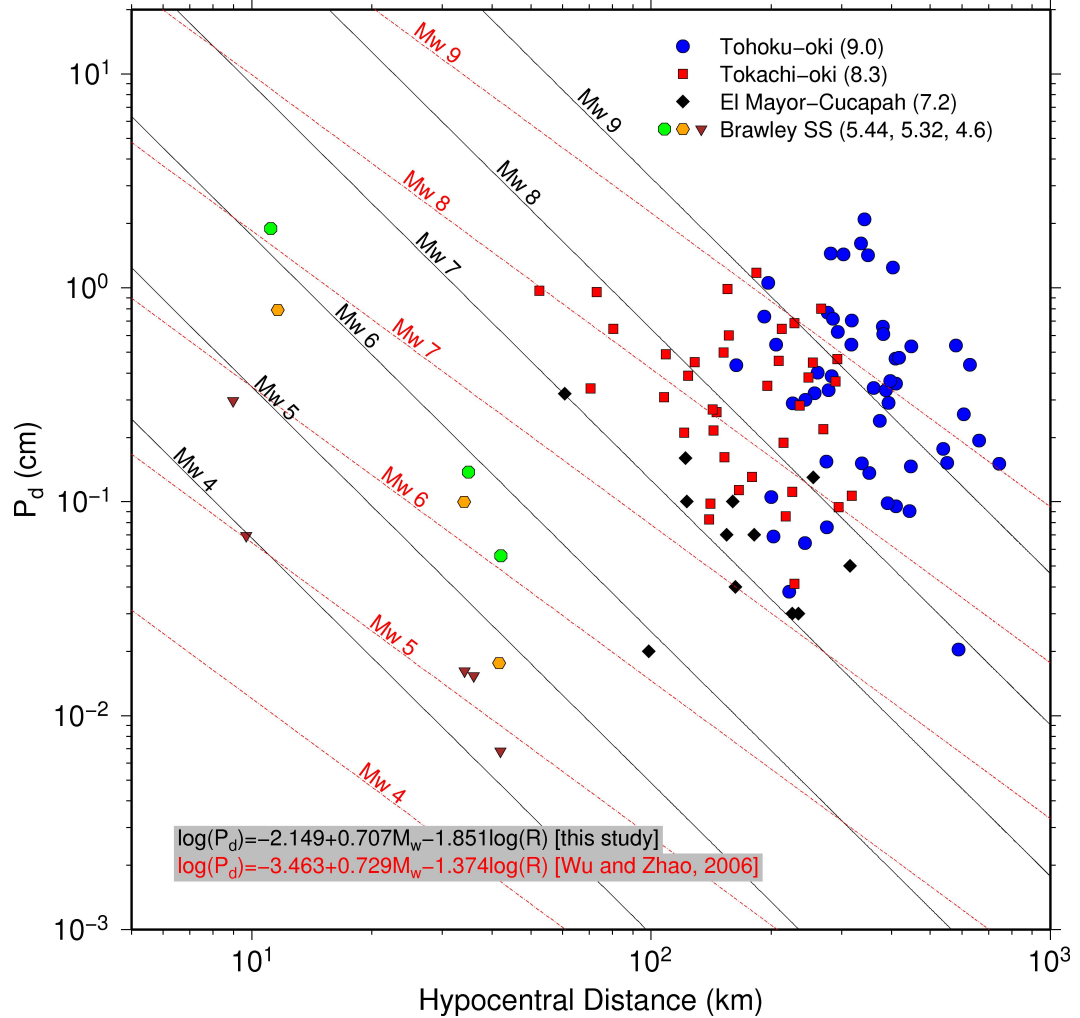


Figure 2.16: The vertical P wave displacement as a function of epicentral distance after three seconds for the 2011 Mw 9.0 Tohoku-oki, 2003 Mw 8.3 Tokachi-oki and 2010 Mw 7.2 El Mayor-Cucapah earthquakes, as well as four events from the August 2012 Brawley Seismic Swarm (Mw 5.44, 5.32, 4.61 and 4.59) using collocated GPS and accelerometer stations. The lines show predicted scaling for magnitudes 9, 8, 7, 6, 5 and 4 for our study (solid black lines) and from *Wu and Zhao* [2006] (red dashed lines). The result of the regression is shown on the figure where R is the hypocentral distance.

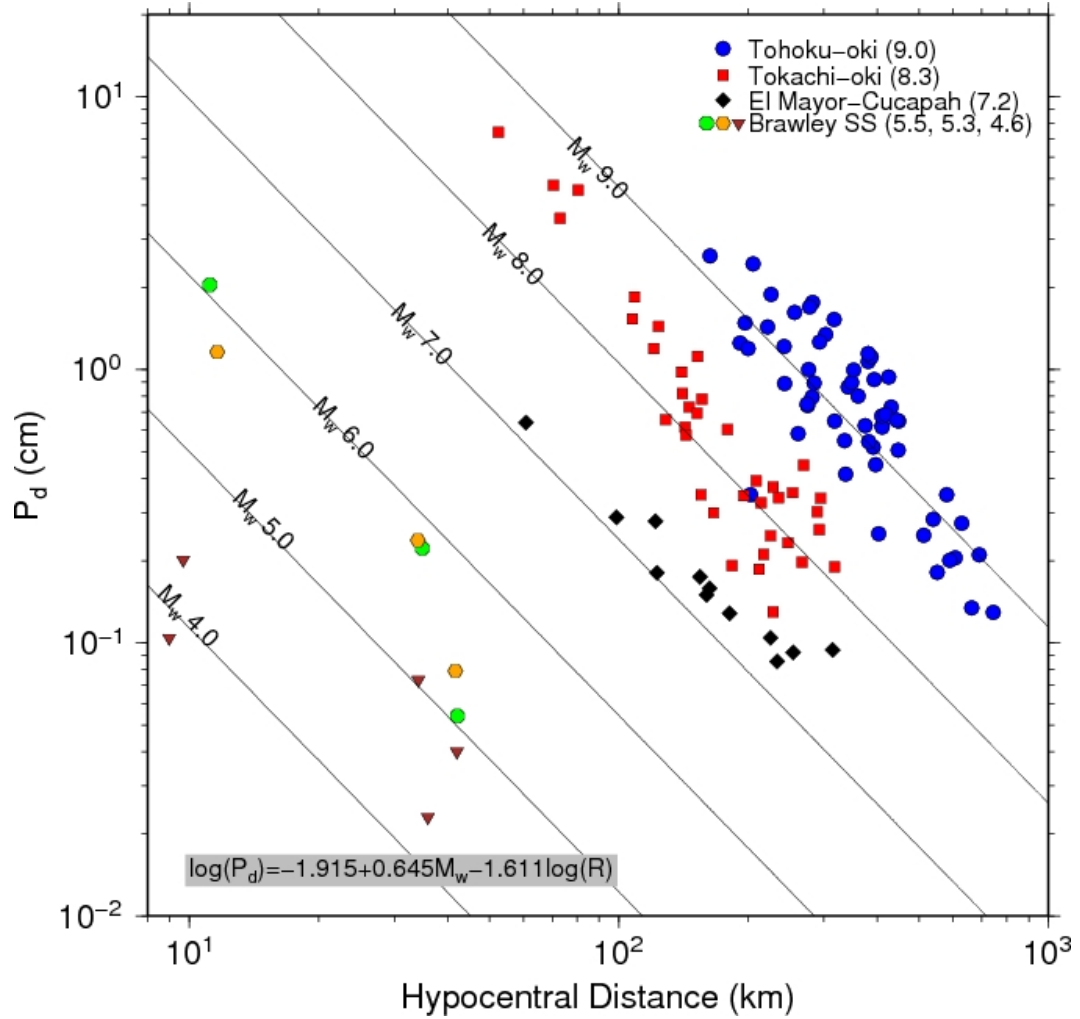


Figure 2.17: The P wave displacement as a function of epicentral distance after five seconds for the 2011 Mw 9.0 Tohoku-oki, 2003 Mw 8.3 Tokachi-oki and 2010 Mw 7.2 El Mayor-Cucapah earthquakes, as well as four events from the August 2012 Brawley Seismic Swarm (Mw 5.44, 5.32, 4.61 and 4.59) using collocated GPS and accelerometer stations. The lines show predicted scaling for magnitudes 9, 8, 7, 6, 5 and 4. The result of the regression is shown on the figure where R is the hypocentral distance. From *Crowell et al.* [2013b].

2.7.4 Timeline of Event Analysis from Seismogeodetic Data

Based upon the aforementioned approach, we propose the following timeline of effective warnings for an augmented seismogeodetic earthquake early warning and monitoring system. The origin time and hypocenter are calculated from the first detections of P waves at a subset of seismic or seismogeodetic instruments using existing methodologies. After communicating the timing and hypocenter estimate back to an individual site, P_d scaling could provide a first estimate of moment magnitude, from first arrivals picked directly off of the seismogeodetic waveforms. Alternatively, τ_c or τ_p^{max} could also be calculated on site, providing magnitude using a completely autonomous approach. Line-source CMT solutions, computed from the coseismic offsets, determine a more robust magnitude, location and style of faulting, as well as provide rough constraints on the lateral extent of moment release within two to three minutes. From the line source CMT solutions, we are able to determine an appropriate section of fault to perform a heterogeneous static slip inversion. The slip inversion can run immediately after the CMT computation, so a full heterogeneous slip distribution is determined within two to three minutes. After the slip inversion, obvious next steps would include ingestion into ShakeMap, PAGER maps, tsunami forward modeling programs to ascertain near-field inundation models [Ohta *et al.*, 2012], kinematic slip inversions, and further analysis to capture aftershocks and postseismic deformation. Of note is that all methods proposed are computationally efficient and are able to be implemented in real time.

2.7.5 Conclusions

Considering the review of the sequence of events at the USGS National Earthquake Information Center (NEIC) [Hayes *et al.*, 2011], there are several stages where the independent information provided by the seismogeodetic time series could have been helpful for a great event such as Tohoku-oki. Both the NEIC and the Pacific Tsunami Warning Center obtained preliminary teleseismic P wave locations and mechanisms within 5 minutes of the origin time, but the first public release was delayed until 9.7 minutes. Independent confirmation of the source size would provide confidence for re-

leasing information earlier. The first two versions of ShakeMap and PAGER alerts were released with the affected areas based on point sources. The ShakeMap and Pager alerts were updated to a finite fault source, significantly extending the length of affected coastline after 2 hrs and 42 minutes. The initial result of the line source fastCMT and slip inversion solutions reproduced the main features of the rupture, approximately 340 km long with average slip of 15 m over the main source region. Early access to this information, which in addition had the correct fault orientation of 204° as opposed to 230° for the rapid finite fault inversion based on the W-phase fault orientation, could provide confidence for earlier release of the finite fault Global ShakeMap. Full integration and combination of GPS into seismic monitoring is essential for a hazard system that is more robust than either traditional seismic or geodetic monitoring alone. The ability to produce seismogeodetic displacement and velocity time series in real time implies that the delay for calculating near-real-time kinematic rupture models can also be significantly reduced in the future.

2.8 Conclusions

In this chapter, I have presented methods to create a fully autonomous earthquake early warning and rapid modeling system with real-time GPS in Japan and California. I first described a novel network adjustment algorithm to create absolute displacement waveforms from baseline measurements. I next showed how to use triangulated strain as an initial detection criterion for large earthquakes. From here, hypocenter determination is possible from GPS using a threshold on the total displacements. Next, using the coseismic displacements obtained rapidly after a large earthquake, I showed how to compute a heterogeneous finite-fault slip inversion in real time using a novel generalized regularization scheme. I demonstrated the improvements made using combined seismogeodetic data for centroid moment tensor and slip inversion results. Finally, using the seismogeodetic displacements, I showed the possible deterministic nature of earthquakes from P wave displacements and how this data will improve the future of

earthquake early warning.

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Chapter 3

Automated Detection of Transient Deformation in Continuous GPS Time Series: A Case Study of the Cascadia Subduction Zone

3.1 Abstract

The discovery of transient slow-slip events over the past decade has changed our understanding of tectonic hazards and the earthquake cycle. Proper geodetic characterization of slow-slip events is necessary for studies of regional interseismic, coseismic and postseismic tectonics, and miscalculations can affect our understanding of the regional stress field. Using the relative strength index (RSI), a financial momentum oscillator, we create a complete record of transient deformation in the Cascadia subduction zone from 2003 to 2013. We also characterize the size, duration, and propagation of each event. We identify all previously studied events as well as some unknown events. We perform synthetic tests to characterize the signal detection limits and the existence of false positive detections. Finally, we model the time series using the ascertained detections and

show that significant variance reduction is achieved in the resultant residual time series.

3.2 Introduction

Since the early 1990s, thousands of continuous Global Positioning System (GPS) instruments have been installed along much of the world's active margins, especially in Japan and the western United States. For most stations, modeling the time series is a rather straight forward process consisting of computing through least squares the bias, secular velocity, seasonal terms, and any jumps, whether related to equipment changes or coseismic deformation [Nikolaidis, 2002]. However, many stations exhibit extremely non-linear behaviors that are not easily picked out or are previously unknown, and leads to poor modeling. This can have serious consequences on studies pertaining to inter-seismic strain accumulation, where knowing minute changes in the secular velocity is important to ascertain the amount of loading on a fault. Generally, many transient deformation signals are accompanied by little to no seismicity to the extent that seismic moment magnitude calculations are significantly underestimated. Transient deformation, or slow slip, is the result of slip on a fault with rupture velocities much less than the shear wave velocity [Ohtani *et al.*, 2010].

One of the first major transient deformational features noted in GPS time series is episodic tremor and slip (ETS) in the Cascadia subduction zone [Dragert *et al.*, 2001; Rogers and Dragert, 2003], where slow-slip is accompanied by non-volcanic tremor instead of traditional earthquakes. In the decade since the discovery of ETS, the phenomena has proven to be rather cyclical, with average recurrence periods between 14 and 18 months depending on location, has also been located within other subduction zones [Beavan *et al.*, 2007; Lowry *et al.*, 2001; Obara *et al.*, 2004; Schwartz and Rokosky, 2007; Shelley *et al.*, 2006] and even in strike-slip environments [Guilhem and Nadeau, 2012; Wech *et al.*, 2012]. ETS has direct implications on earthquake hazards due to the changing stresses along the subduction interface.

The geodetic community has recognized the problem as extremely important

such that it has become the topic of a recent series of workshops at the Southern California Earthquake Center [Murray-Moraleda and Lohman, 2010]. The goal of the workshops has been to develop automated criteria for identifying and modeling transient deformation to aid in regional earthquake hazards studies. Proposed methods of automated transient detection have included, but are not limited to, the network inversion filter [Segall and Matthews, 1997] and the subsequent network strain filter [Ohtani *et al.*, 2010], principal component analysis [Dong *et al.*, 2006], covariance descriptor analysis [Kedar *et al.*, 2010], and Gaussian wavelet transforms [Melbourne *et al.*, 2005]. Kimura *et al.* [2011] extend the network inversion filter formulation to tiltmeters for automated detection of slow slip in the Nankai subduction zone with the assumption that all slow slip events are the result of a single representative model and the source parameters are determined through a grid-search algorithm. This method however requires coherent signals across many stations, which may not be evident given event and station geometries. All of these methods have their individual strengths and weaknesses and aim to perform similar tasks, but the mark of a good automated transient algorithm must be robust (detect all signals above certain level and minimize false detections), self-contained, run in a recursive fashion, require minimal additional modeling, and is able to fully characterize events (size, duration, and timing).

The Relative Strength Index (RSI) has proved to be a valuable technical indicator on the stock market for the last forty years [Wong *et al.*, 2003]. The main purpose of the indicator is to determine strong changes in stock price over short periods of time to declare a stock "overbought" or "oversold", indicative of bullish or bearish behavior. Chong and Ng [2008] find that use of the RSI generates higher returns than simple buy and hold patterns for determining entry and exit points for a specific security. The RSI is a momentum oscillator that indicates rapid divergence from normal behavior above and beyond normal variability, which lends itself well to geophysical applications, especially towards automated detection of events. To this end, we apply the RSI method to 82 continuous GPS time series throughout the Cascadia subduction zone from 2003 to 2013 to determine changes in momentum of the individual stations associated with transient deformation. Using this method, we are able to track migration patterns, the

temporal scale, and the size of slow slip events. We also apply this method to synthetic events to determine the displacement threshold and the occurrence of false positives within the time series. Finally, we use the detections to model offsets in the individual time series to test improvements in the residual variance to time series modeling not using the observed offsets.

3.3 Data

For this study, we utilize 82 continuous GPS stations (Figure 3.1, grouped by regions) currently processed and combined by the Scripps Orbit and Permanent Array Center (SOPAC) and the Jet Propulsion Laboratory (JPL) with respect to the International Terrestrial Reference Frame (ITRF) 2008 [Altamimi *et al.*, 2011] as part of the NASA MEaSUREs project. The SOPAC positions are processed using GAMIT and JPL positions with GIPSY, which are then combined using the Quasi-Observation Combination Analysis (QOCA) software package (i.e., Dong *et al.* [1998]). After combining the positions, we perform regional filtering with JPL’s `st_filter` package [Dong *et al.*, 2006], which performs a Principal Component Analysis to remove higher-order errors to reduce uncertainties. From the filtered, combined time series, we model velocity, seasonal terms, coseismic and postseismic terms using JPL’s `analyze_tseri` software package [Dong *et al.*, 2002]. The time series are then detrended using the modeled velocities. Detrending the time series is beneficial to the RSI, but not required at the small time windows being looked at (21 days), which are much shorter than the wavelength of interseismic deformation. All of the time series are available on the GPS Explorer data portal at SOPAC as well as packaged in a tar file (<http://geoapp.ucsd.edu/>; <http://geoftp01.ucsd.edu/pub/timeseries/measures/ats/WesternNorthAmerica/>).

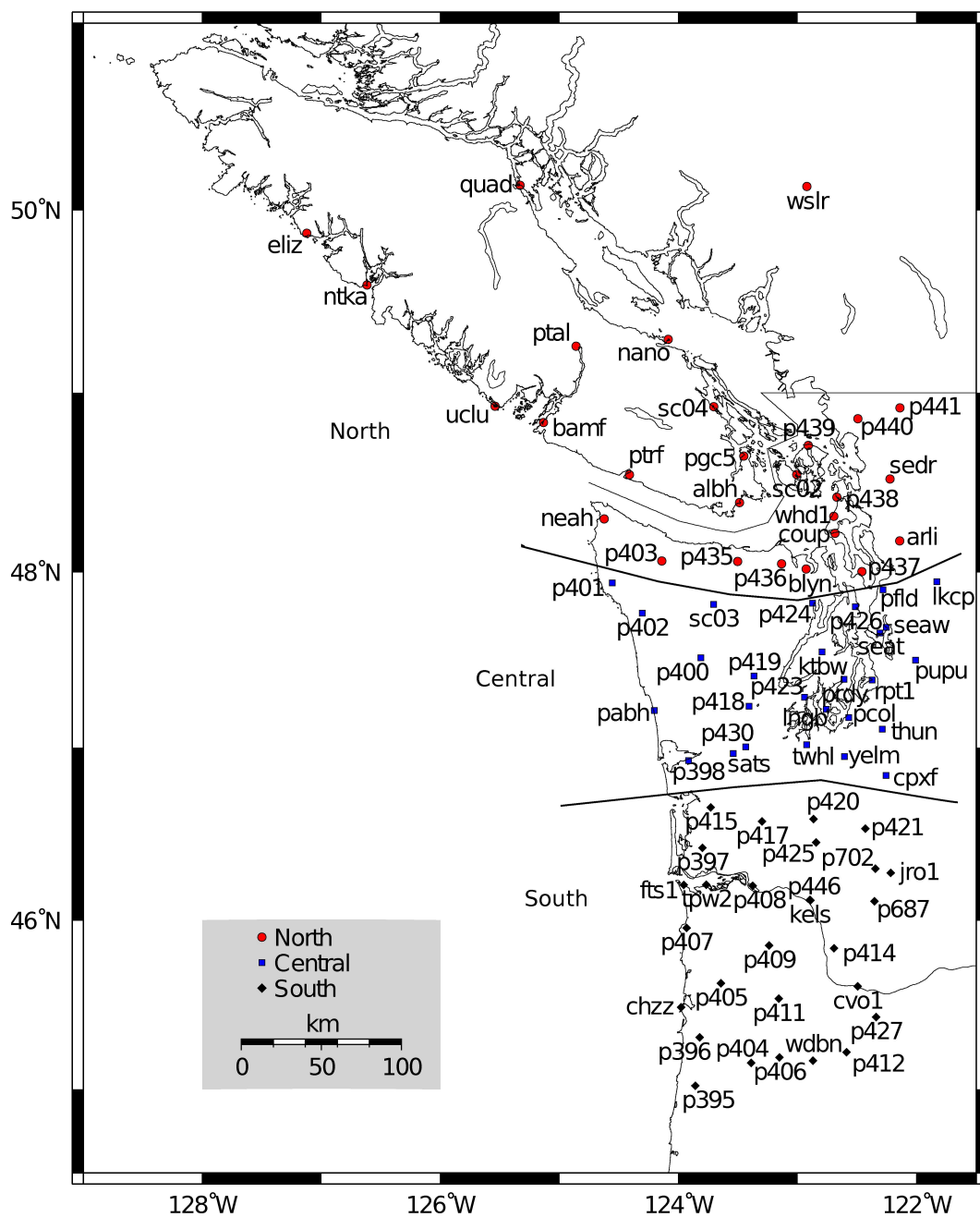


Figure 3.1: Continuous GPS stations in Cascadia analyzed for this study. The different symbols represent how each station is regionally grouped, into north, central and south.

3.4 Methods

The RSI was first presented by *Wilder* [1978] to help explain changes in a stock that are significant above normal volatility or variance. First two functions need to be defined, U and D , for up and down respectively. During an up period for a given displacement time series d_i , the function values are as follows

$$U(t_i) = d_i - d_{i-1} \quad (3.1)$$

$$D(t_i) = 0 \quad (3.2)$$

During a down period, the functions are

$$U(t_i) = 0 \quad (3.3)$$

$$D(t_i) = d_{i-1} - d_i \quad (3.4)$$

Next, simple moving averages are computed for U and D for a given time period n (t_{i-n} to t_i). For this study, we choose $n = 21$. Subsequent iterations of the RSI use exponential moving averages instead of simple moving averages, but we choose to use the original formalism of *Wilder* [1978] since we find the results using exponential moving averages are roughly comparable to simple moving averages. The ratio of the two moving averages is called the relative strength, RS ,

$$RS(t_i) = \frac{\sum_{k=i-n}^i U(t_k)}{\sum_{k=i-n}^i D(t_k)} \quad (3.5)$$

The RS is converted to the RSI, a value that varies from 0 to 100, by

$$RSI(t_i) = 100 - \frac{100}{1 + RS(t_i)} \quad (3.6)$$

It is easy to see that during periods where ups and downs offset each other, the relative strength will be ~ 1 , giving the RSI a value of 50. For this study, we perform an additional step, a simple central moving average of period 21 days to help smooth out small variations and more accurately determine the timing of events.

The general rule of thumb with the threshold values are 70/30 for a 14-day RSI

and 60/40 for the 20-day RSI to determine overbought and oversold conditions [Wong *et al.*, 2003]. However, in the case of geodetic data, the noise characteristics vary temporally and spatially such that we need to treat the threshold RSI values on a station by station basis. To obtain the thresholds, we first take the time series for each station and high-pass filter them to a period of 5 days. We choose this value because most slow-slip events have signals that persist much longer than 5 days. From the high-pass filtered data, we then compute the RSI and perform a simple 21-day central moving average on the filtered RSI. We then define the threshold values to be the minimum and maximum RSI values. Figure 3.2 shows the threshold values for each Cascadia station in all three components. The mean threshold values are 52.82/47.05, 53.86/45.88, and 52.87/47.08 for the north, east and up components respectively.

After computing threshold values for each station, we compute the threshold exceedance and ignore values that lie between the thresholds. We then remove any events that do not persist for at least five days at a given station since we high-pass filter to that period. Using these RSI detections in all three components, we solve for displacements using the `analyze_tseri` software package developed by JPL [Dong *et al.*, 2002]. The modeling is performed as described before in the data section, with the only difference being the offsets are defined as the start times of the RSI detections.

3.5 Results

The westward RSI exceedance of all stations grouped by region is shown in Figure 3.3 from 2003 to 2013. The northern region is 27 stations above 48°N , the central region is 27 stations between 46.8°N and 48°N , and the southern region is 28 stations between 45°N and 46.8°N . Plots of individual years are included in Figures 3.4 to 3.24. From Figure 3.3, it becomes apparent that events are coherent across many stations and some events have much greater relative strength than others. Table 3.1 is an overview of all the events seen in Figure 3.3, including start and stop times, maximum RSI exceedance, the regions affected, and the number of stations. We removed detections

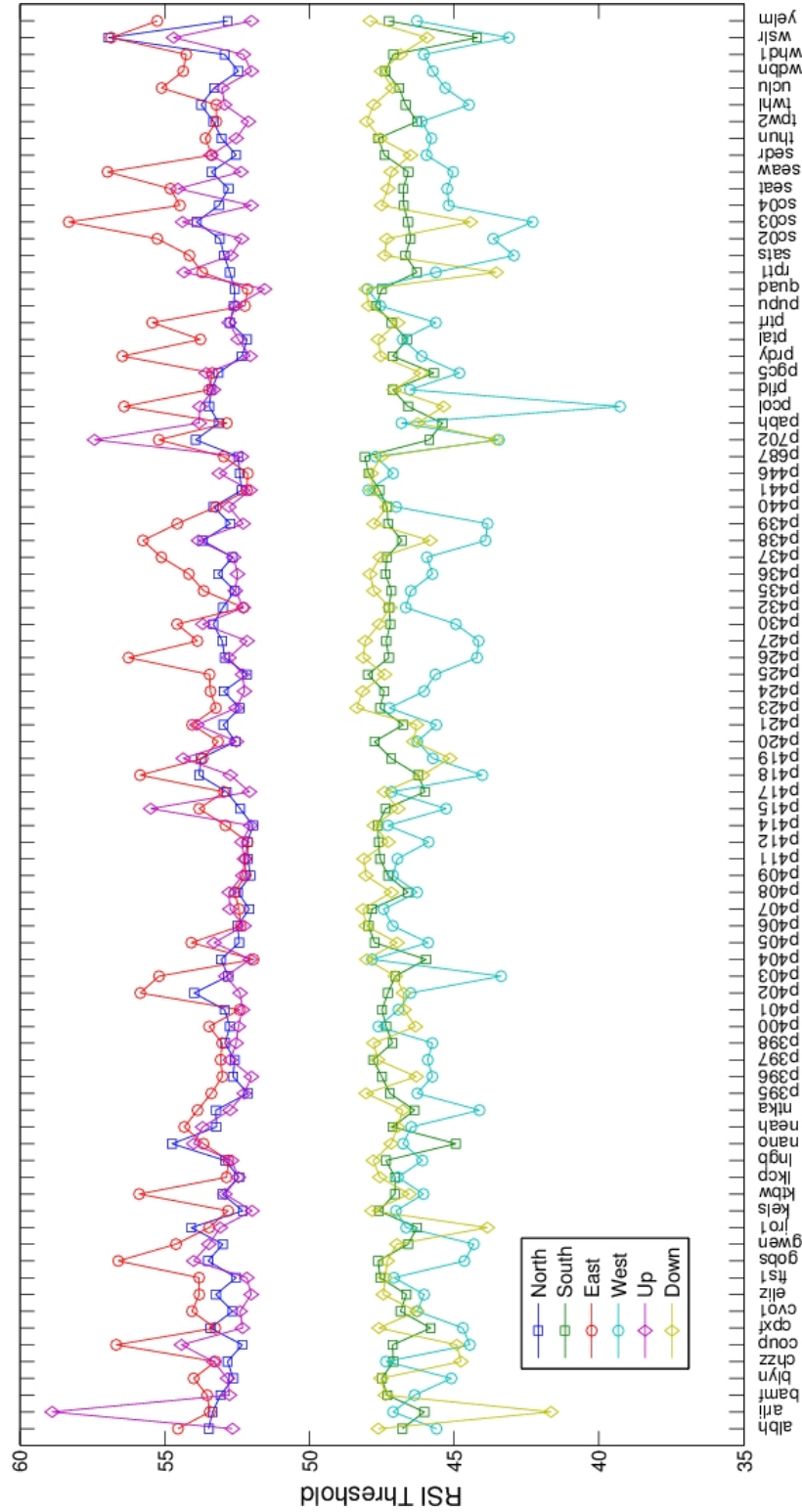


Figure 3.2: The threshold RSI values in each direction for all sites. For north, east and up, all RSI values greater than the threshold are considered events. For south, west, and down, all RSI values less than the threshold are considered events.

recorded at only one station for brevity.

3.5.1 Events

Slip during the February-March, 2003 (Figures 3.4 and 3.5) slow slip event initiated in the first two weeks of February in west-central Washington [*Melbourne et al.*, 2005]. Slip proceeded southwestward during the next two weeks, and then migrated to southern Vancouver Island in the mid-March. Tremor is recorded on the Pacific Northwest Seismic Network (PNSN) between 26 February 2003 and 19 March 2003. *Szeliga et al.* [2008] identified 23 stations with discernible slow slip signals. Out of these 23 stations, we do not analyze 4 of them in this study since they are outside the region of interest (GOBS, GWEN, SC00, and YAWA), and they do not analyze 5 of the stations in this study (BAMF, PGC5, PTAL, SEAW, and TPW2). Out of the 19 overlapping stations, we have positive detections in the RSI at all stations but BLYN, CPXF, and SEDR. BLYN has a data gap at the beginning of 2003 and SEDR has a positive detection for only 2 days so it is eliminated due to not persisting for longer than 5 days. In the case of CPXF, the RSI is just below the threshold due to a very small offset of < 2 mm. Additionally, we have detections at CVO1 and CHZZ, both of which are processed by *Szeliga et al.* [2008], but not assigned an offset. *Holtkamp and Brudzinski* [2010] also do not identify slow slip at CHZZ as well as KTBW, but identify all the other stations as having slow slip evident (with the exception of BAMF, PGC5 and PTAL, which they do not process). The RSI detection times follow a similar pattern to the observed tremor and slip during the event. The first detection is at station RPT1 in late January near southern Puget Sound. The next detections are nearby at KTBW, SEAW, and SEAT. After this, the event moves south and is recorded at many stations in southern Washington and Oregon. Finally, the event switches back north, with detections starting at SC02 and then moving onto ALBH, NEAH, PGC5, PTAL, BAMF and finally UCLU.

The July 2004 event (Figures 3.6 and 3.7) was predominantly centered on northern Washington and Vancouver Island. *Wech et al.* [2009] reported that tremor during this event was predominantly in the northern Olympic Peninsula and along southern

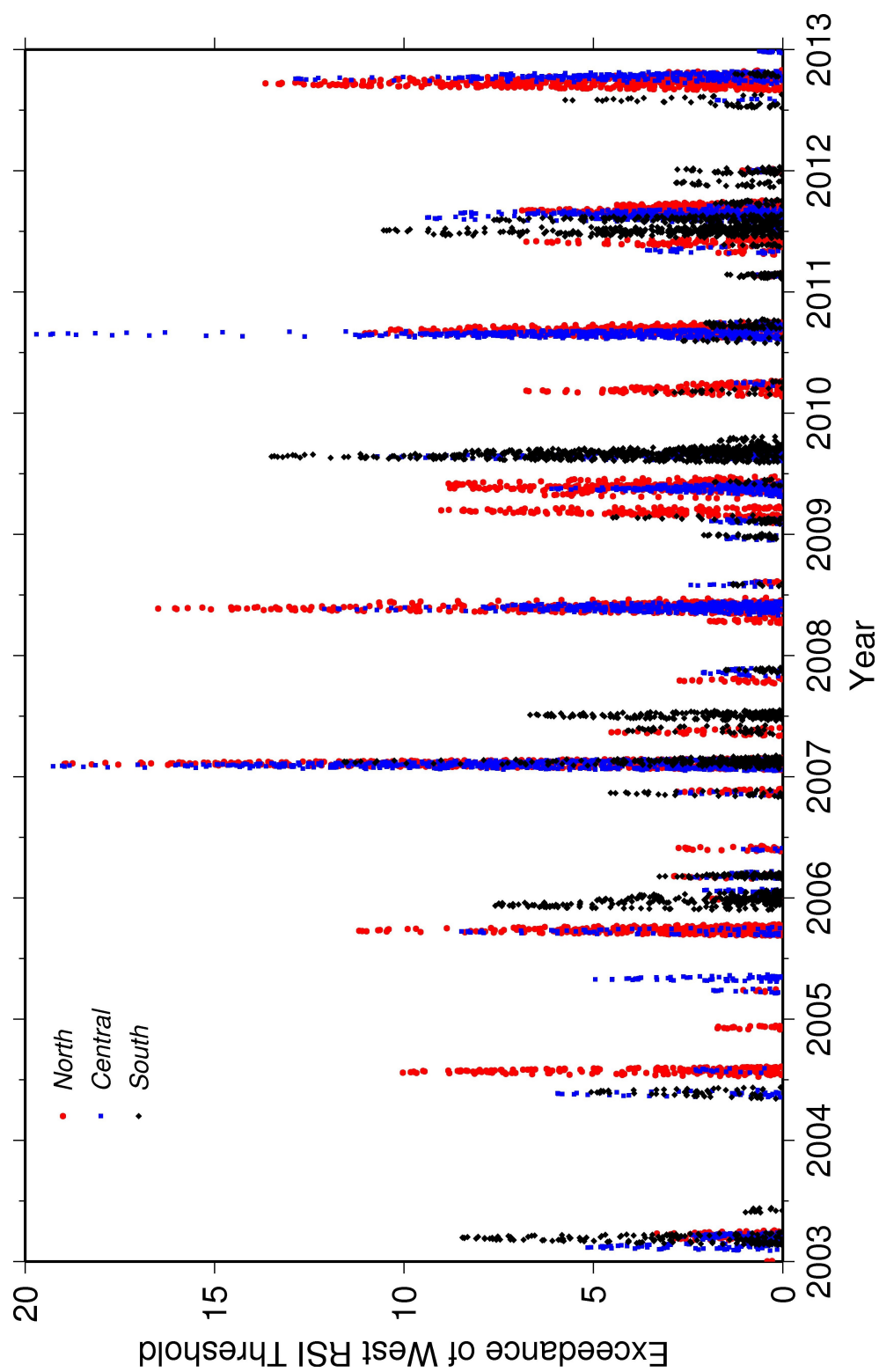


Figure 3.3: Exceedance of RSI in the west direction from 2003 to 2013. The different symbols represent the regional grouping of stations into North, Central, and South.

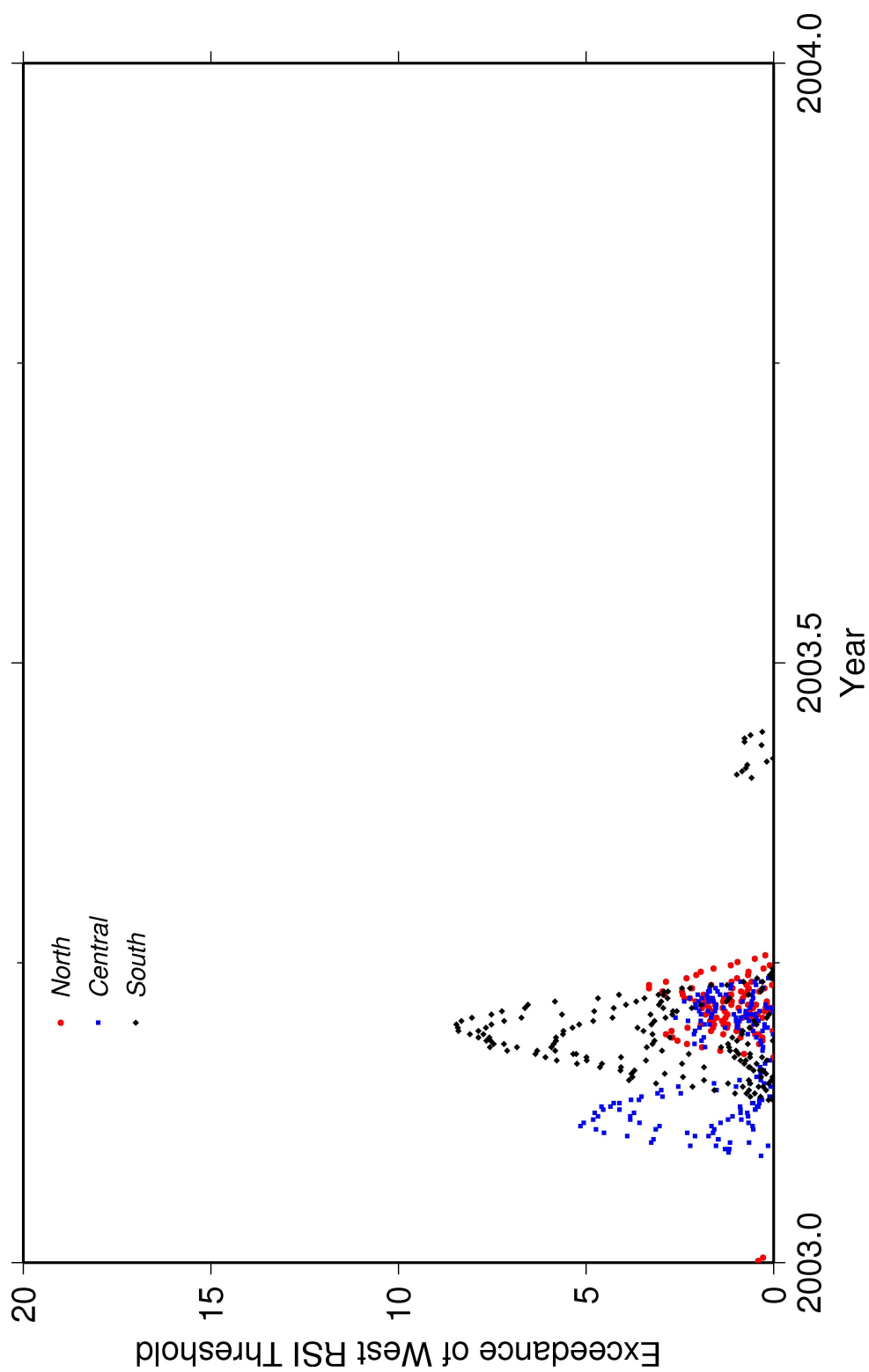


Figure 3.4: Exceedance of RSI in the west direction in 2003. The different symbols represent the regional grouping of stations into North, Central, and South.

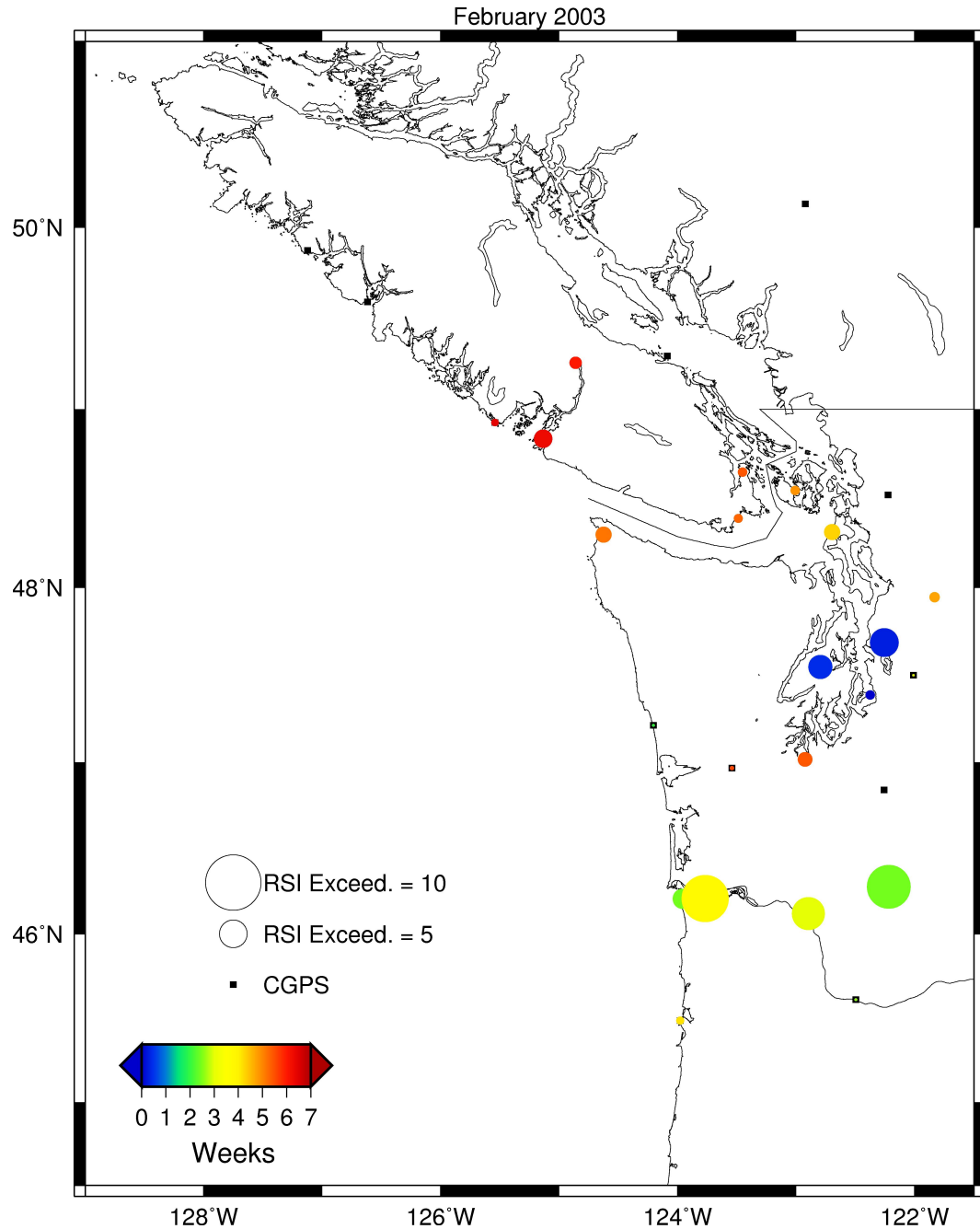


Figure 3.5: RSI detections during the February 2003 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2003.0890.

Vancouver Island. We record this event across 11 stations, of which 8 are recorded by *Szeliga et al.* [2008] and the other 3 they do not process (BAMF, ELIZ, and PGC5). They also have positive recordings at PABH, which we have an east recording for during this event, and SC04, which we have a data gap for. *Holtkamp and Brudzinski* [2010] have a detection at ELIZ, as well as at NTKA and PUPU, in addition to all of the other stations we analyze. NTKA has an RSI exceedance in the east direction and PUPU does not have any discernible displacement during the July 2004 event. With regards to rupture directivity, we have the first detection at ALBH; the transient proceeds north and south with detections at WHD1, SC02, BLYN, and PGC5. Next, rupture migrates eastward towards SEDR and LKCP before finally reaching the stations on northern Vancouver Island (BAMF, NANO, and ELIZ). A similar pattern is seen in the tremor migration from the PNSN [*Wech et al.*, 2009].

There are two significant events in 2005 (Figure 3.8), the first being a northern event in September (Figure 3.9) and the second being a southern event in December (Figure 3.10). The September event was studied in detail by *Szeliga et al.* [2008], and both events were studied by *Holtkamp and Brudzinski* [2010] and *Schmidt and Gao* [2010]. For the September event, we identify events at all of the stations processed by *Szeliga et al.* [2008] (except OTIS which we do not process) in addition to 7 stations they do not process (ARLI, BAMF, COUP, P401, PFLD, PGC5, and PTAL). *Holtkamp and Brudzinski* [2010] process COUP and PFLD but do not record this event. *Wech et al.* [2009] report that the tremor for this event started near the southern tip of Vancouver Island and then proceeded westward before heading northwest up Vancouver Island. This is consistent with our first detection at ALBH, and then progressing northwestward. The largest detections are at ALBH and SC02, consistent with the slip models of *Szeliga et al.* [2008] and *Schmidt and Gao* [2010]. The December event is recorded at 13 stations processed by *Holtkamp and Brudzinski* [2010]. Of those 13 stations, we do not process 5 of them, and the other two (P420 and P421) fall just under the threshold. From the time series of P420 and P421, it appears that the source duration is fairly long, leading the RSI to fall just under the threshold.

Wang et al. [2008] reported on an event under central Vancouver Island in

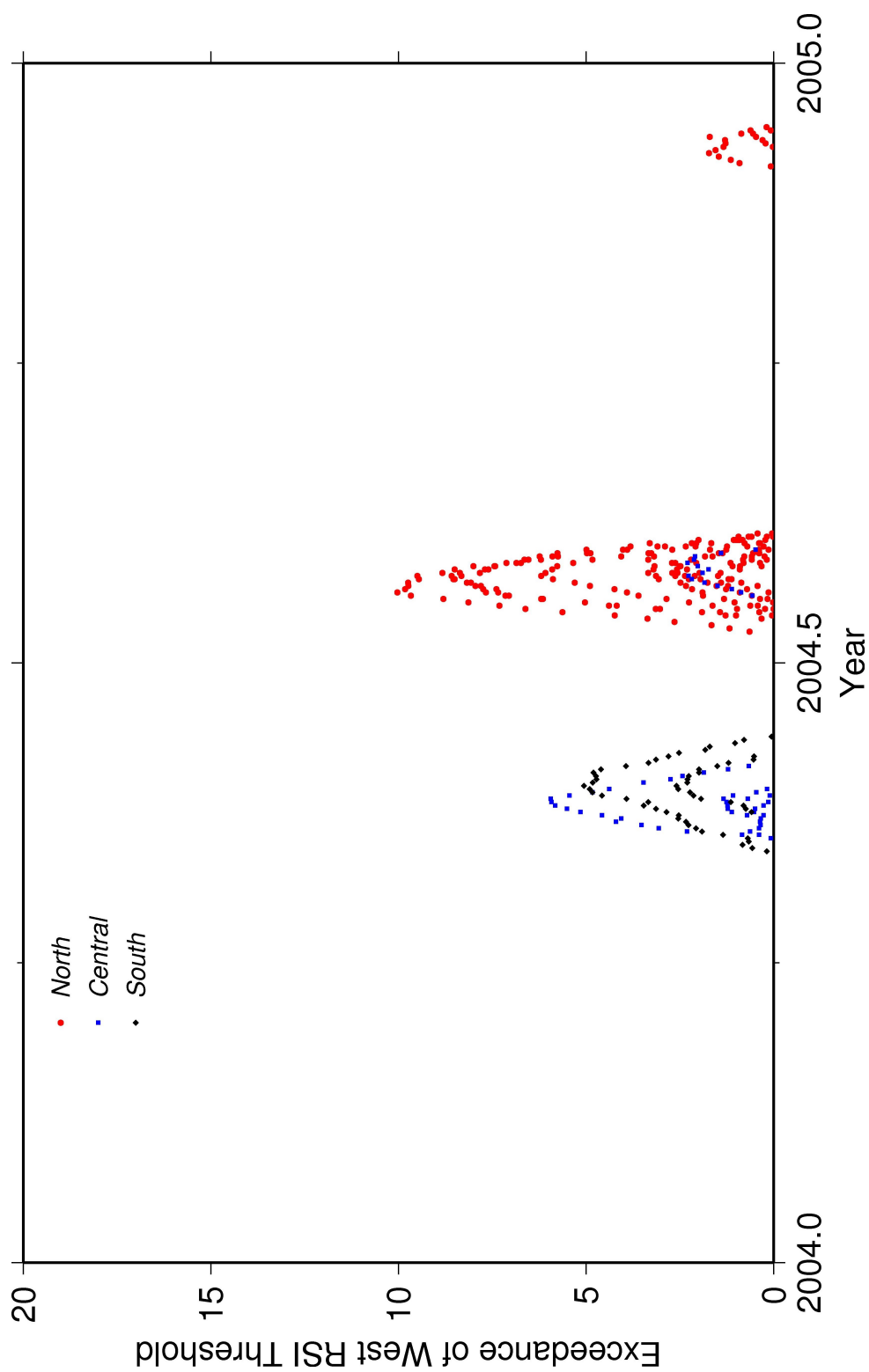


Figure 3.6: Exceedance of RSI in the west direction in 2004. The different symbols represent the regional grouping of stations into North, Central, and South.

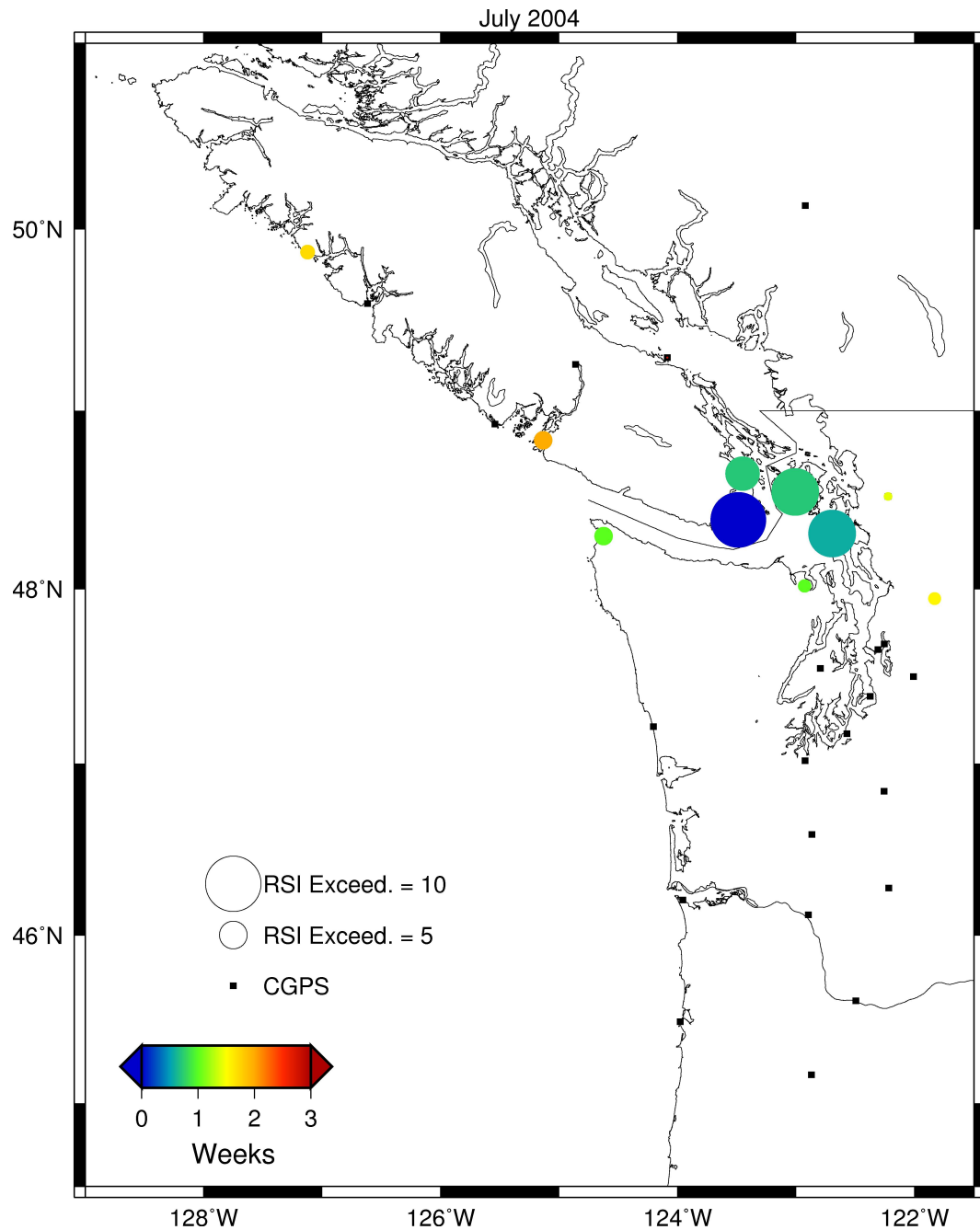


Figure 3.7: RSI detections during the July 2004 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2004.5260.

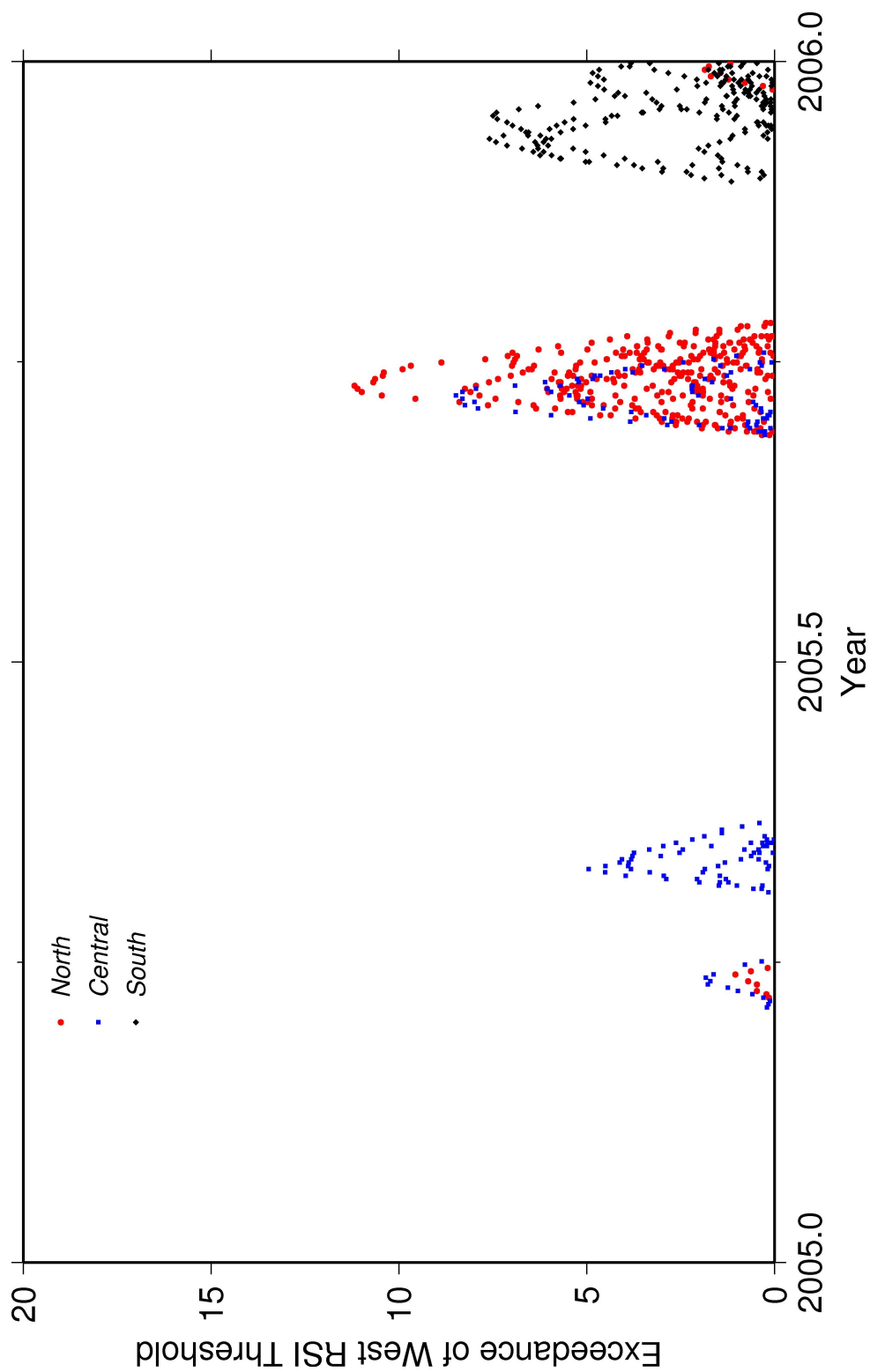


Figure 3.8: Exceedance of RSI in the west direction in 2005. The different symbols represent the regional grouping of stations into North, Central, and South.

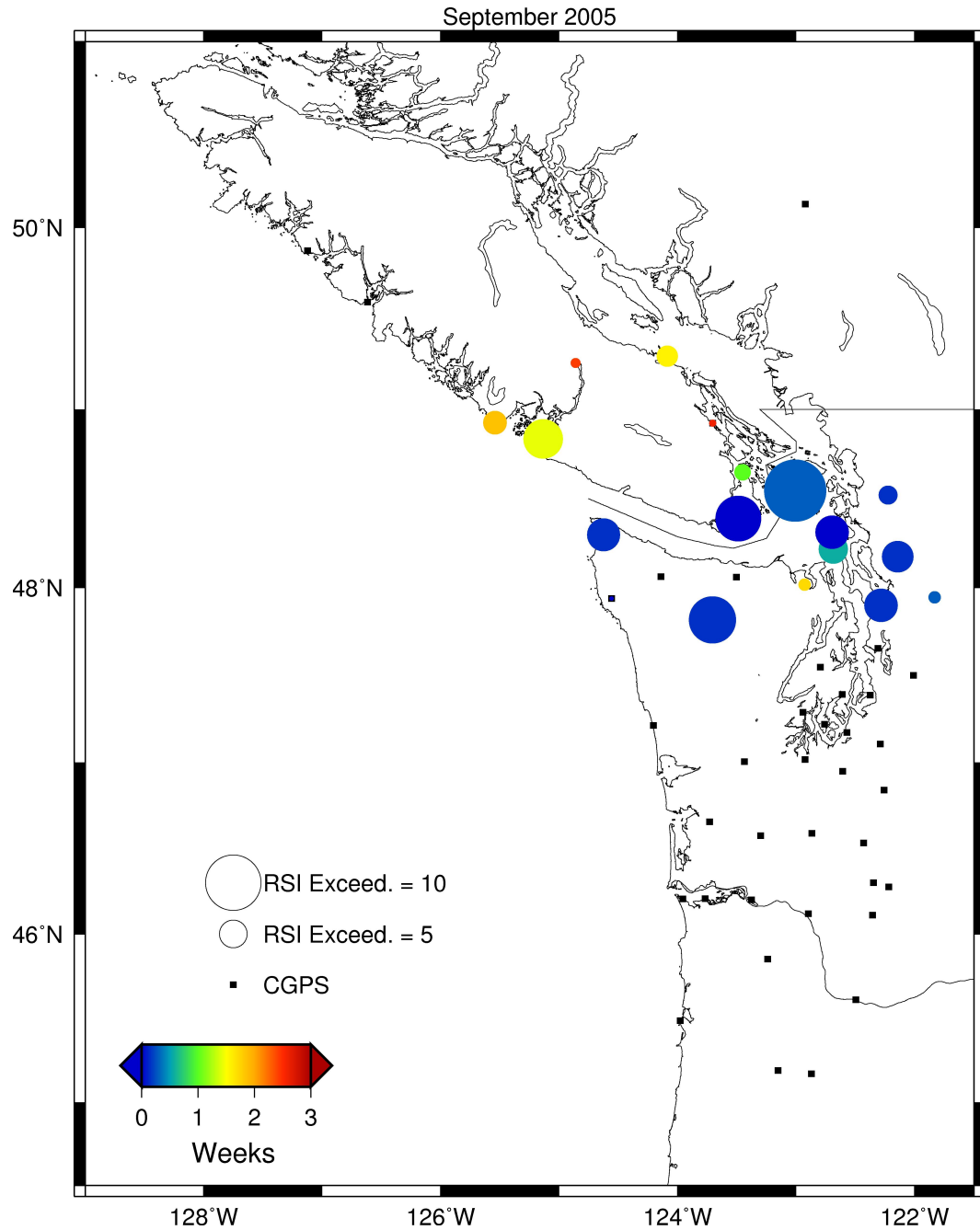


Figure 3.9: RSI detections during the September 2005 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2005.6890.

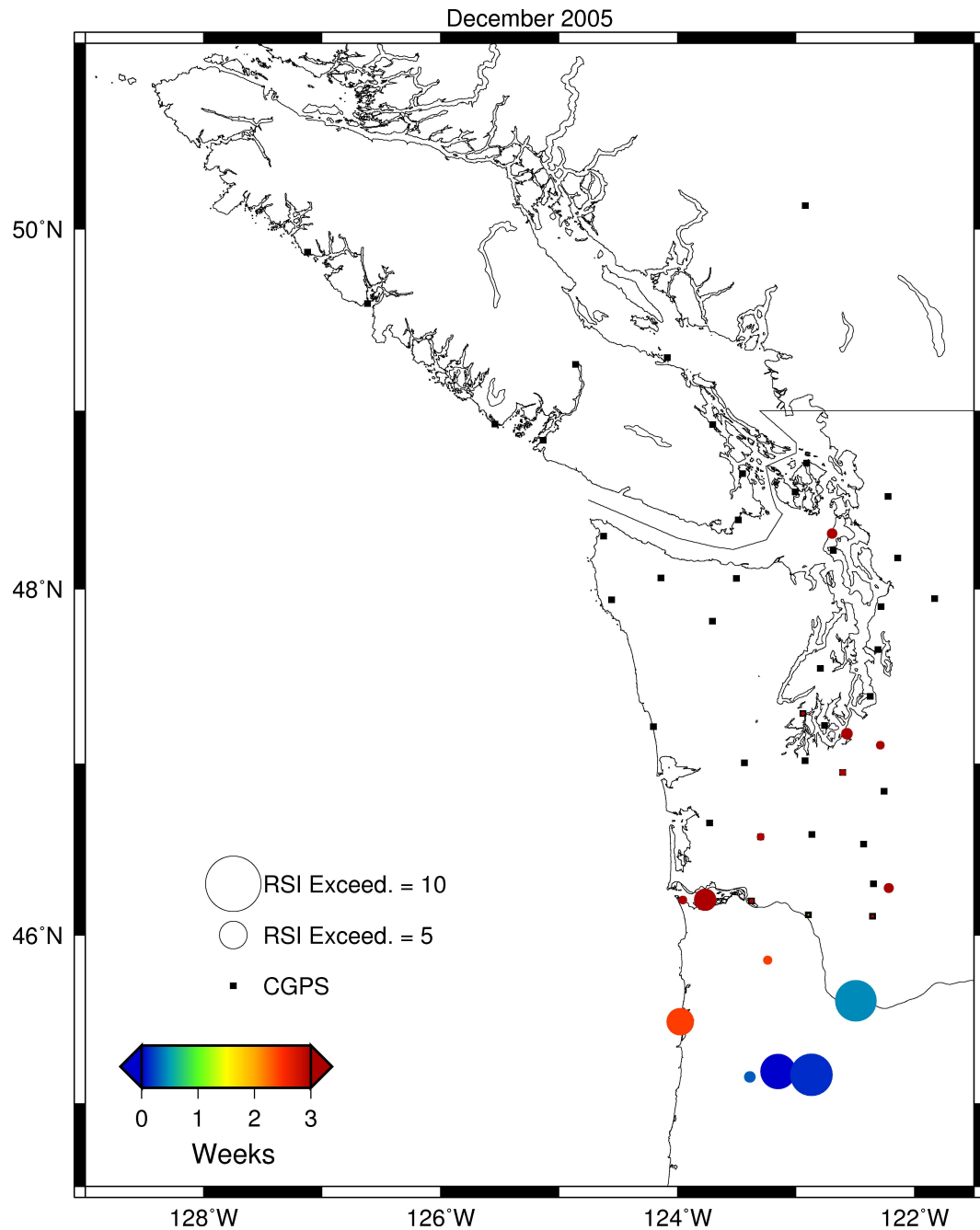


Figure 3.10: RSI detections during the December 2005 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2005.9000.

November 2006 (Figure 3.11) that was smaller in duration and slip. They identified subtle signals at UCLU, BAMF, NANO, and PTAL of less than 2 mm. We have detections at UCLU, BAMF, and PTAL, with the strongest signal at BAMF, which is consistent with their slip model.

The January 2007 (Figures 3.12 and 3.13) slow slip event was the most widespread event covered in this study, affecting the majority of the stations (53), and displaying very sharp signals in the RSI. The event started in southern Puget Sound near SEAR and quickly progressed to all surrounding stations within days. Out of the first 10 stations, 8 of them display RSI threshold exceedances greater than 10. Next, the event ruptures bilaterally, with very strong signals recorded near Vancouver Island (ALBH, P438, and SC02). In total, detections are made as far north as PTAL and as far south as P395, with the strongest signals in the greater Puget Sound area. This is independently confirmed by looking at the modeled slip distributions from *Aguilar et al.* [2009] and *Schmidt and Gao* [2010], and the progression roughly follows the tremor patterns observed [Wech et al., 2009]. *Holtkamp and Brudzinski* [2010] also process this event with continuous GPS, and for the overlapping stations between our studies they do not have positive identifications for 10 stations that the RSI picked up (FTS1, LKCP, NANO, P420, P421, P687, P702, PABH, PUPU, and UCLU). The RSI exceedance values for these stations are all under 3.04 (except PABH at 5.16), although visual inspection of these time series reveals small but discernable signals in the displacement on the order of 1-2 mm.

The May 2008 (Figures 3.14 and 3.15) event started near central Puget Sound and then migrated northward into Vancouver Island before migrating southward into southern Puget Sound [Wech et al., 2009]. Our first detections occur at PFLD, P437, and ARLI, and then progresses westward towards COUP and P436. The final detections are at YELM and P398 in the south and NANO and SC04 in the north. The largest detections are at COUP, P403, P437, and P426, which is consistent with the regions of maximum slip modeled from *Schmidt and Gao* [2010].

In 2009, there were 3 distinct events: February-March, April-June, and August-October (Figure 3.16). The February event is most likely two different events, with a

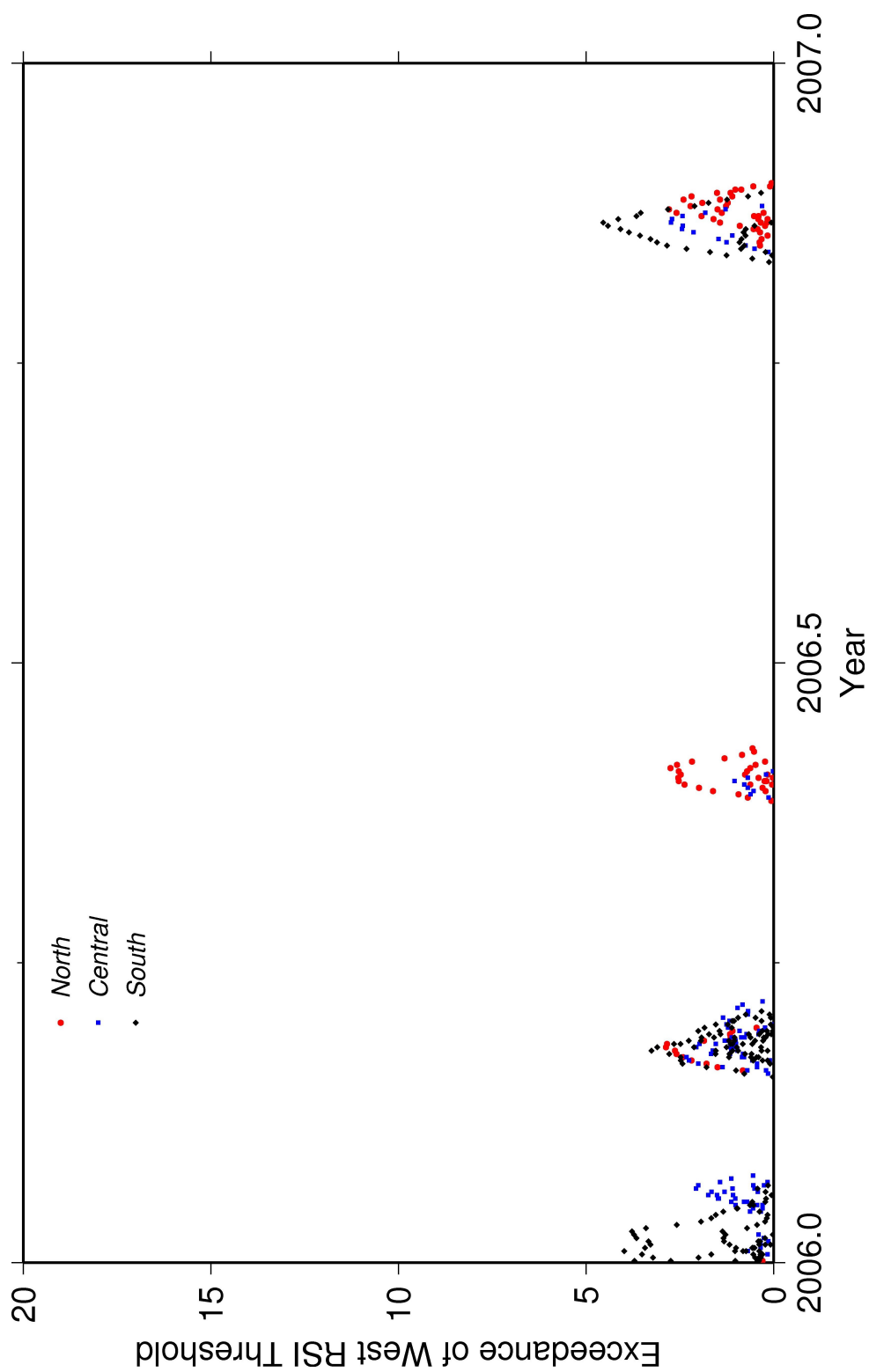


Figure 3.11: Exceedance of RSI in the west direction in 2006. The different symbols represent the regional grouping of stations into North, Central, and South.

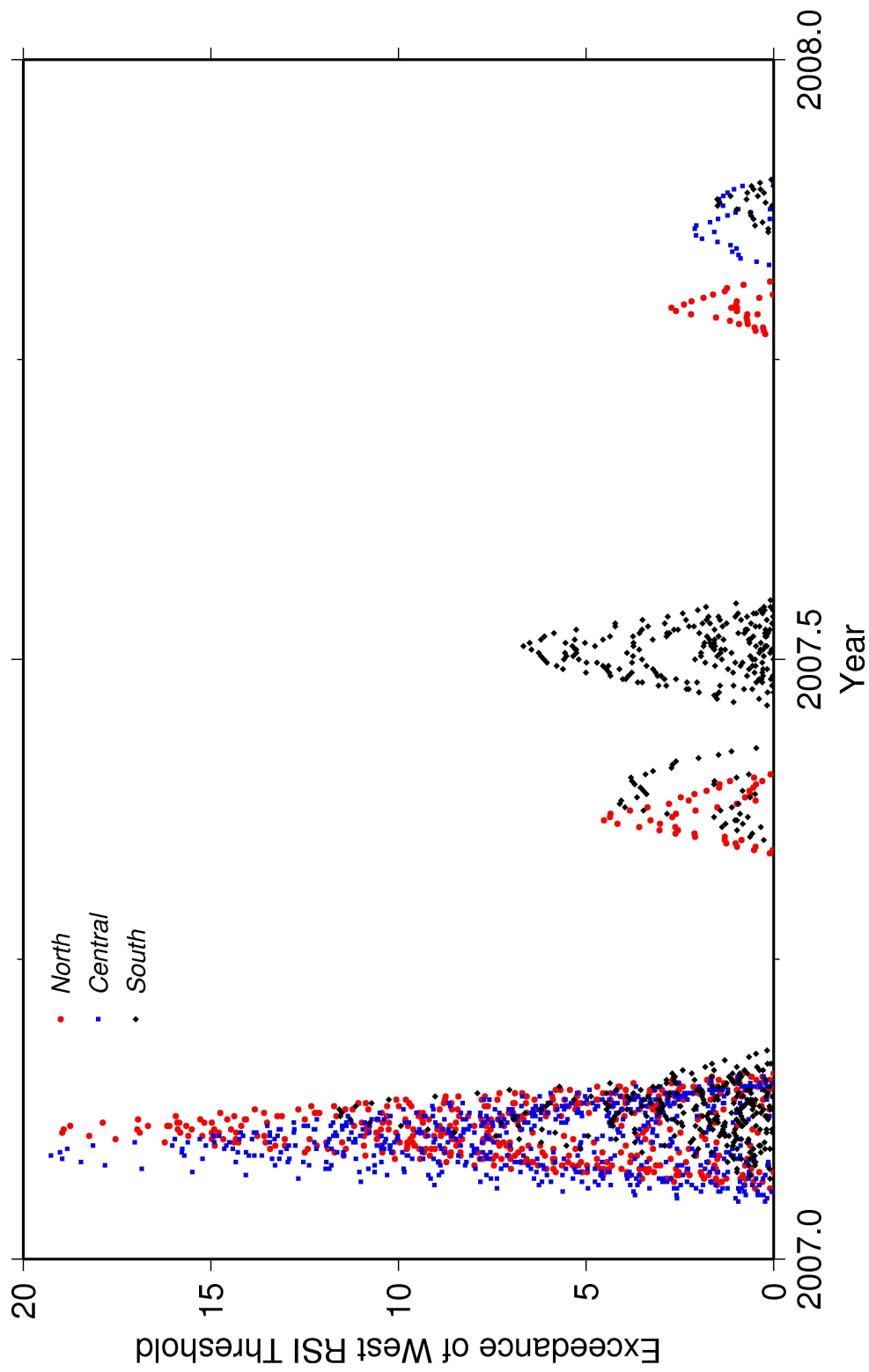


Figure 3.12: Exceedance of RSI in the west direction in 2007. The different symbols represent the regional grouping of stations into North, Central, and South.

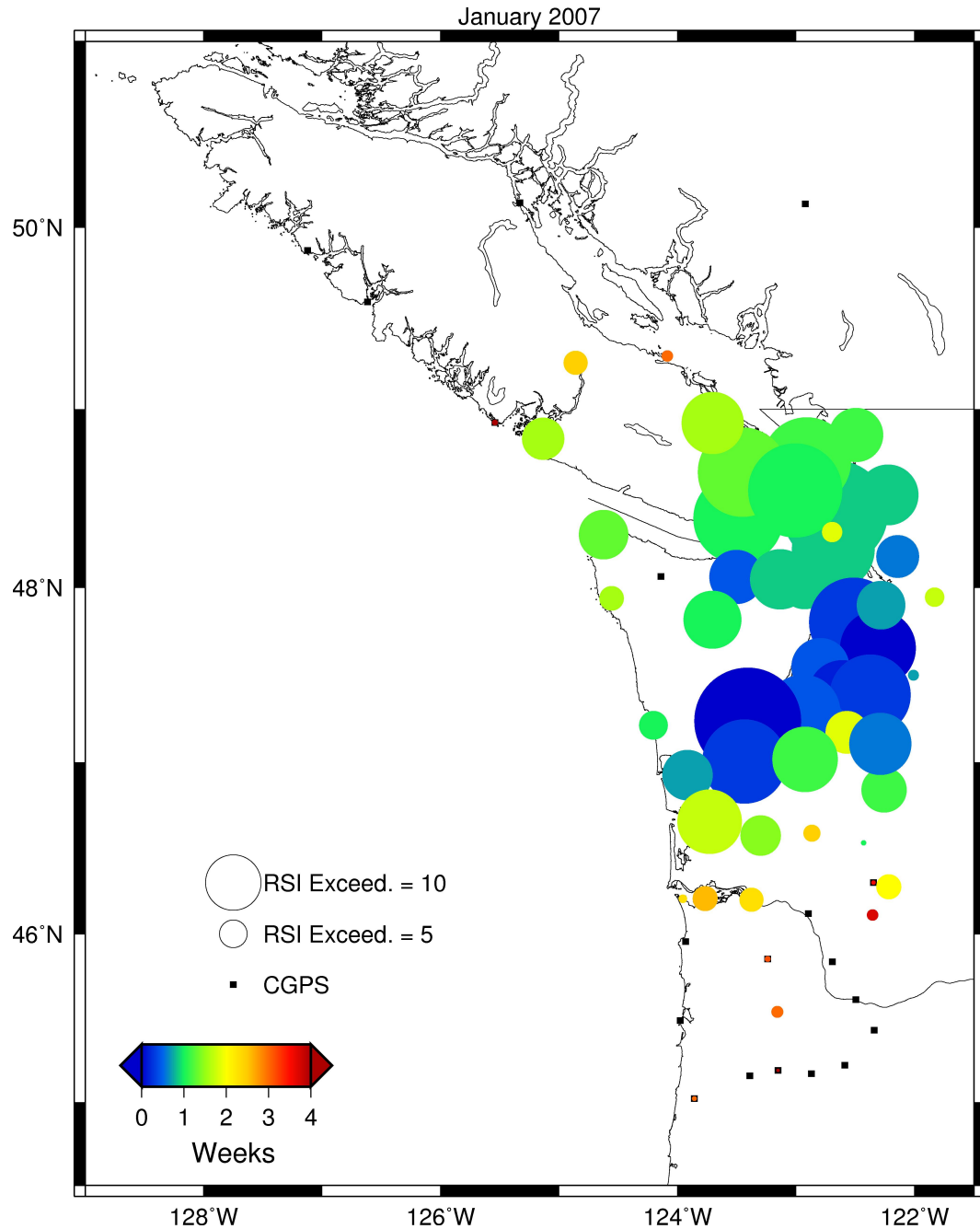


Figure 3.13: RSI detections during the January 2007 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2007.0479.

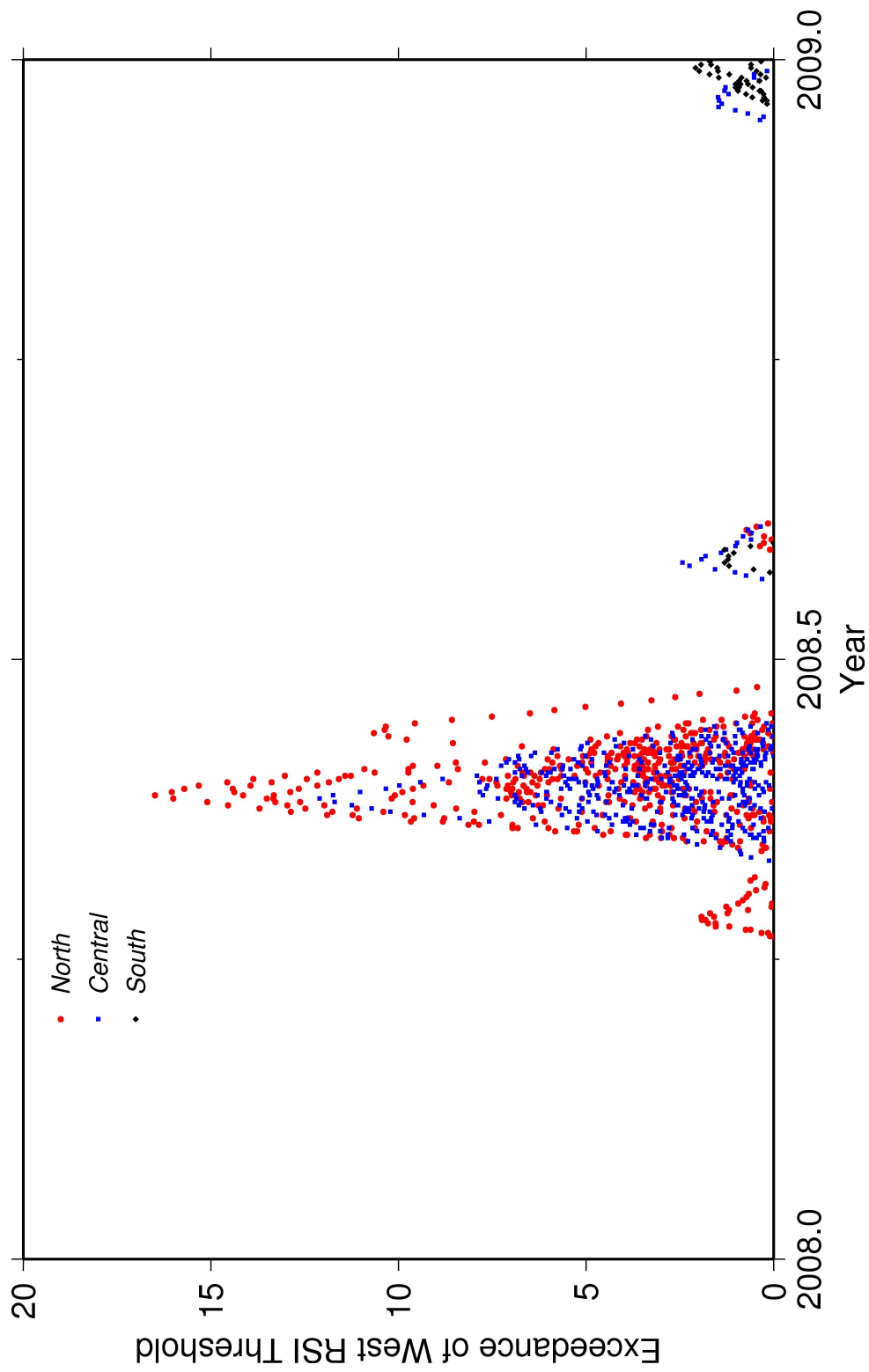


Figure 3.14: Exceedance of RSI in the west direction in 2008. The different symbols represent the regional grouping of stations into North, Central, and South.

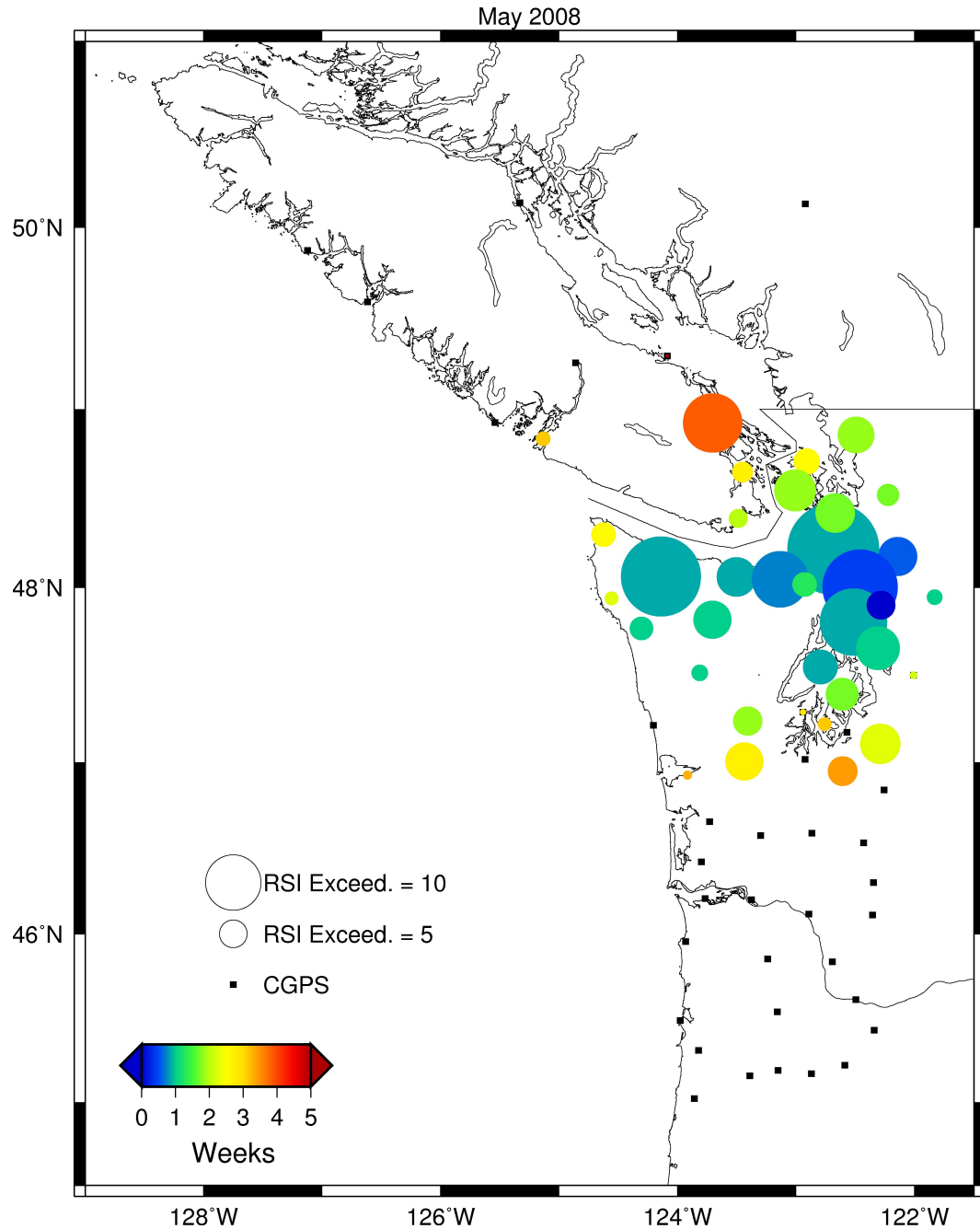


Figure 3.15: RSI detections during the May 2008 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2008.3320.

strong event on northern Vancouver Island and a much smaller event in southern Washington/Oregon. The northern stations during the February event show very strong RSI and displacement signals, starting at PTAL and then moving northward towards QUAD, NTKA, and ELIZ, before affecting NANO and UCLU. The April-June event is a more widespread event with known tremor detections at the PNSN (Figure 3.17). The event starts in the west-central Olympic peninsula near P435, then spreads out to both the east and west. The event then moves south, reaching the Oregon border, and finally goes north onto Vancouver Island, reaching as far as SC04, which is similar to the pattern of observed tremor in the PNSN. The maximum RSI intensity is observed at P437 (8.8) and PTRF (8.8). *Bartlow et al.* [2011] report on the August event and show that slip started near the Washington-Oregon border in early August and then bilaterally ruptured, with most of the energy towards the south, and still rupturing towards mid-late September (Figure 3.18). Our first detections occur under P397, P408, P417, TPW2, and P415. *Bartlow et al.* [2011] also find that the maximum slip is below this region, which matches the high RSI exceedances in the area, most notably at P397 (11.6), P425 (13.5), and P421 (11.1).

In 2010, two distinct events occur, the first being a northern event in February and the second being a bilaterally propagating event in August to September (Figures 3.19 and 3.20). The first event only affects the far north stations on Vancouver Island, although some small detections occur at 2 stations in the south (P404 and P409) and 1 in the central region (PFLD). The August-September event has tremor recorded by the PNSN starting in southern Puget Sound and then bilaterally ruptures. The first RSI detections occur at P419, P423, SEAT, and KTBW. The largest RSI exceedance value in this study is at SC03 during this event (19.7), followed by P419 (11.3), P436 (11.0), and SEAT (10.9).

The mid 2011 slow slip event is complex and can possibly be separated into 3 distinct events, with a separate northern and southern event before July and then a south-to-north progressing event from July to September (Figure 3.21). The earlier northern event consists of only 5 stations on the far northern end of Vancouver Island (ELIZ, QUAD, NTKA, NANO, PTAL, and UCLU). The early southern event (Figure 3.22) af-

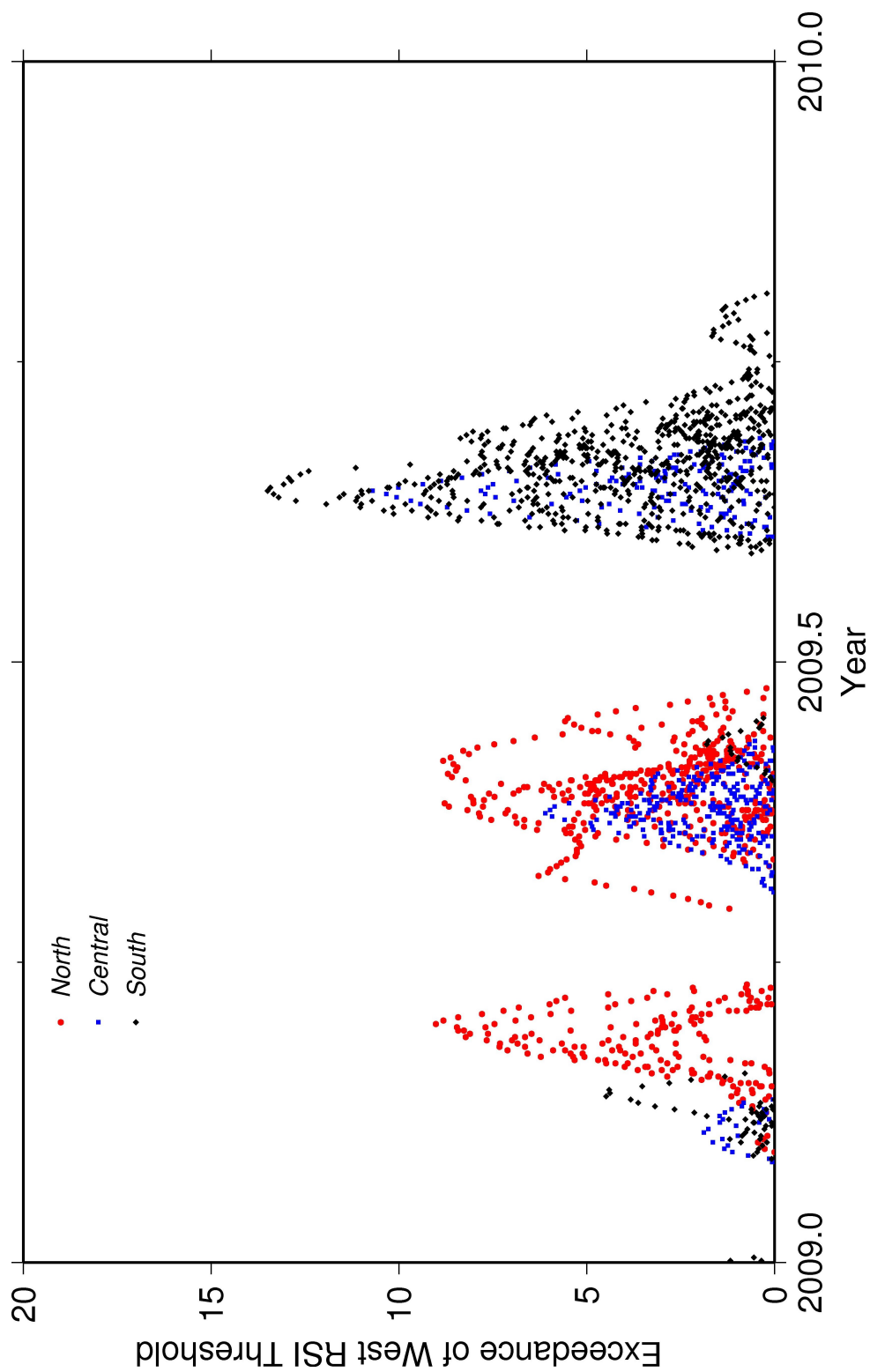


Figure 3.16: Exceedance of RSI in the west direction in 2009. The different symbols represent the regional grouping of stations into North, Central, and South.

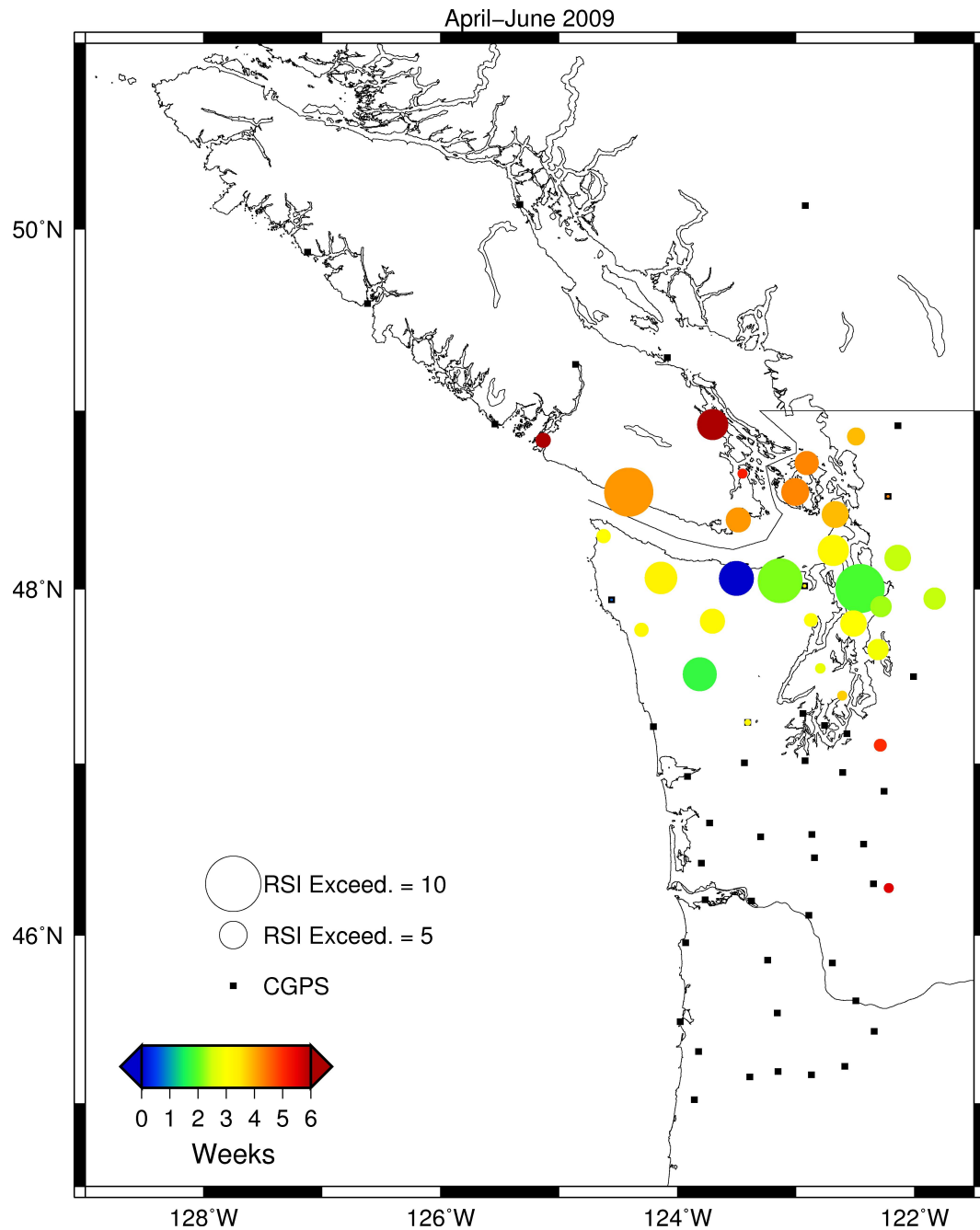


Figure 3.17: RSI detections during the April–June 2009 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2009.2945.

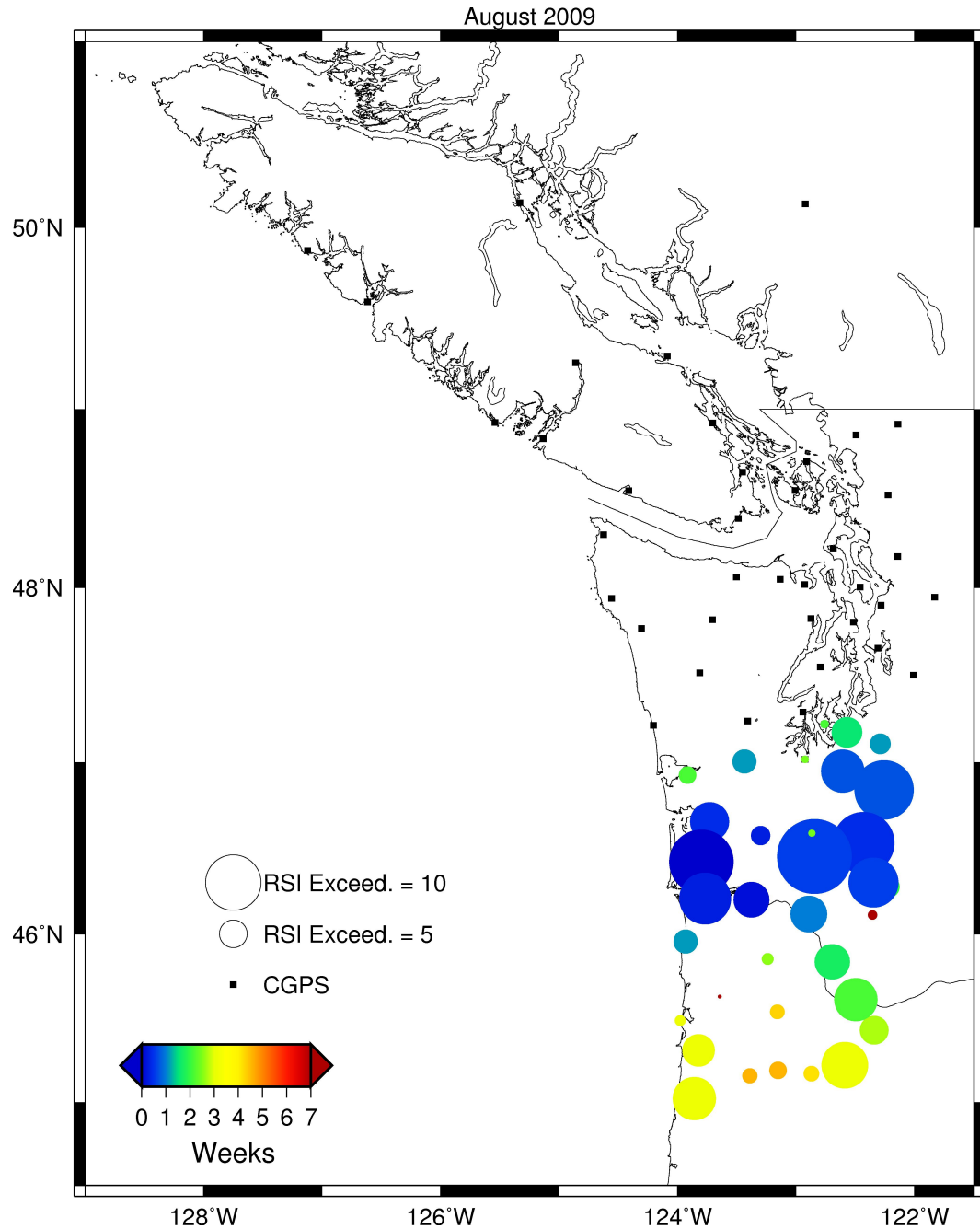


Figure 3.18: RSI detections during the August 2009 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2009.5904.

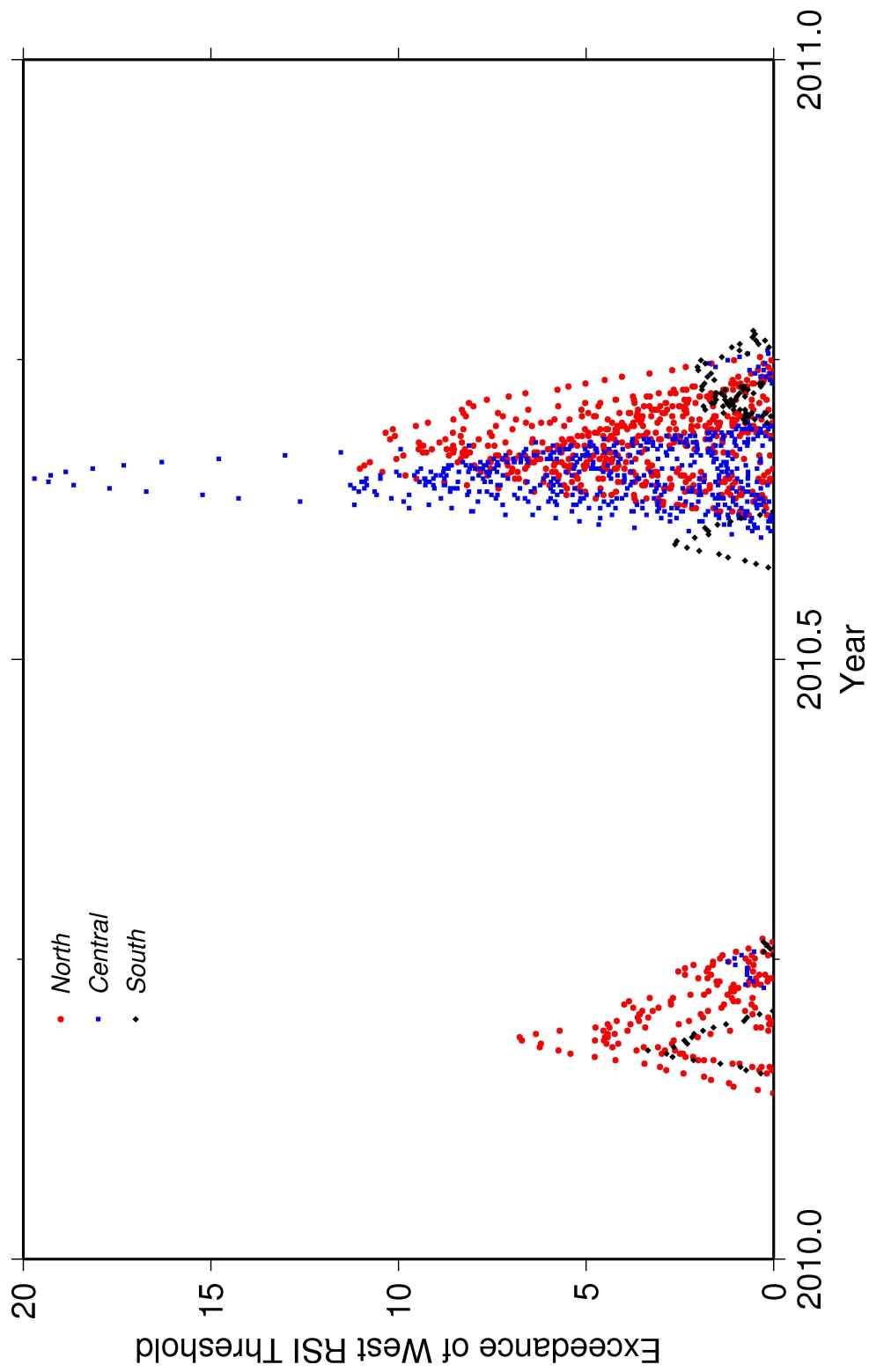


Figure 3.19: Exceedance of RSI in the west direction in 2010. The different symbols represent the regional grouping of stations into North, Central, and South.

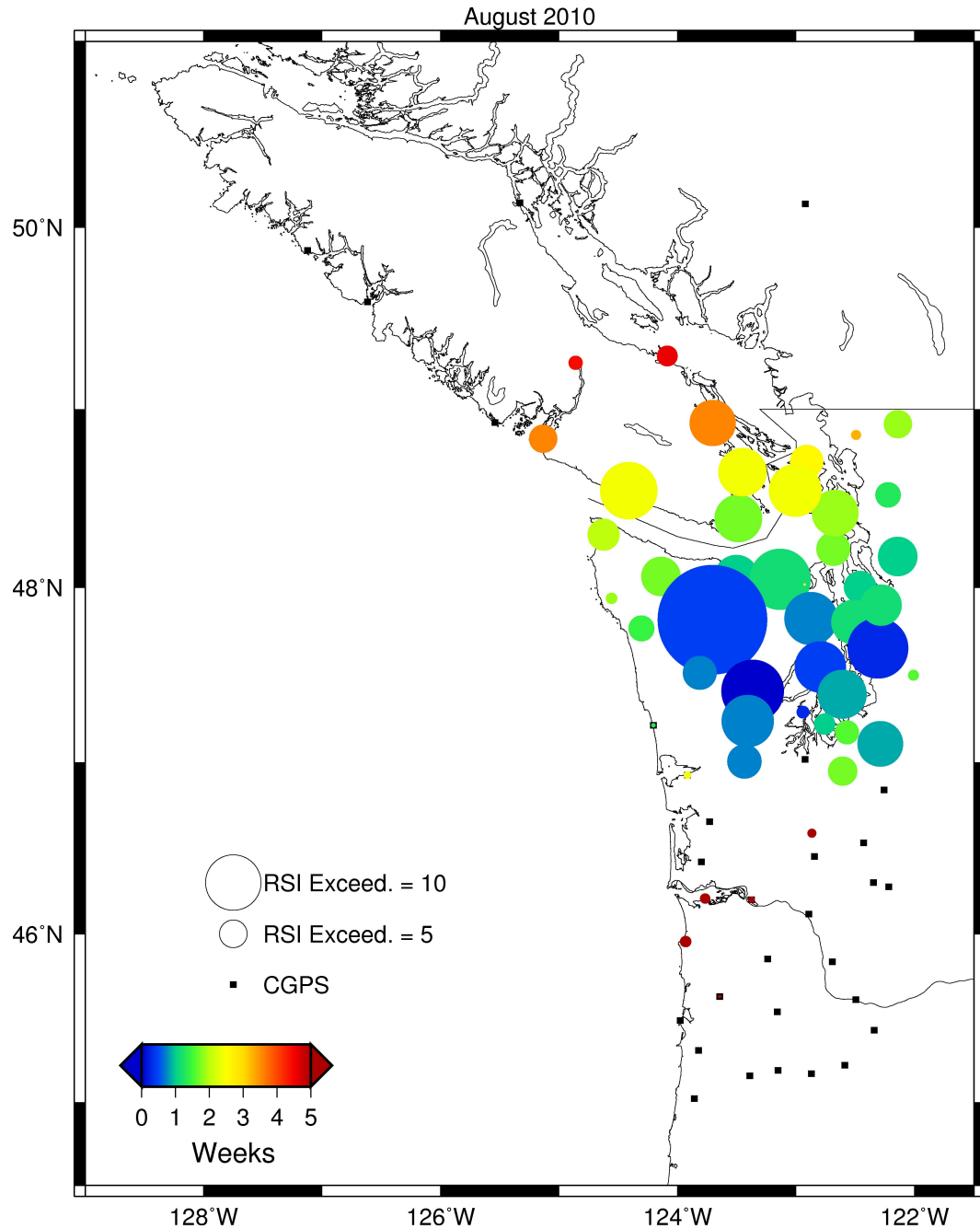


Figure 3.20: RSI detections during the August 2010 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2010.6014.

fects all the stations south of P407, with the maximum RSI occurring at P405 (10.5). Tremor recorded on the PNSN for the July to September event (Figure 3.23) starts near the Oregon border and then moves northward over the next six weeks and ends up on the southern end of Vancouver Island. The first RSI detections occur along the border between the central and southern stations, and then steadily move northward. During the late 2011 event, the southern stations affected by the early 2011 event show very small or no RSI detections, indicating that the earlier event could possibly be part of the later event.

Tremor recorded on the PNSN during the August-October 2012 slow slip event (Figures 3.24 and 3.25) uncharacteristically started in northern Vancouver Island between UCLU and BAMF and progressed southward into south-central Washington as well as progressing northward along the entire extent of Vancouver Island. Our first detection occurs at BAMF, followed by NANO, PTAL, SC04, and QUAD. Within days, all stations on Vancouver Island have a positive detection. By October, the event reaches the Oregon border. The strongest RSI readings are recorded at PTRF (13.7), SC03 (12.9), BAMF (12.4), and NANO (11.8).

3.6 Synthetic Tests

In order to test the detection limit and performance of our method, we create a series of synthetic transients of varying rupture duration and slip magnitude. We first start by creating a noise time series using the individual epoch uncertainties for station ALBH. From the noise time series, we superimpose the transients computed through a dislocation in a homogeneous half-space [Okada, 1985]. The source extent is 300 km long by 15 km wide and stretches from the Washington-Oregon border into Vancouver Island. We use constant strike and dip angles of 345° and 20° respectively. The center of the fault is buried at 40 km depth. For the different synthetic transients, we vary the slip magnitude (pure dip-slip) between 10 mm and 5 m, in 10 mm increments. These slip magnitudes correspond to displacements between 0.16-83.66 mm (east), 0.34-170.2

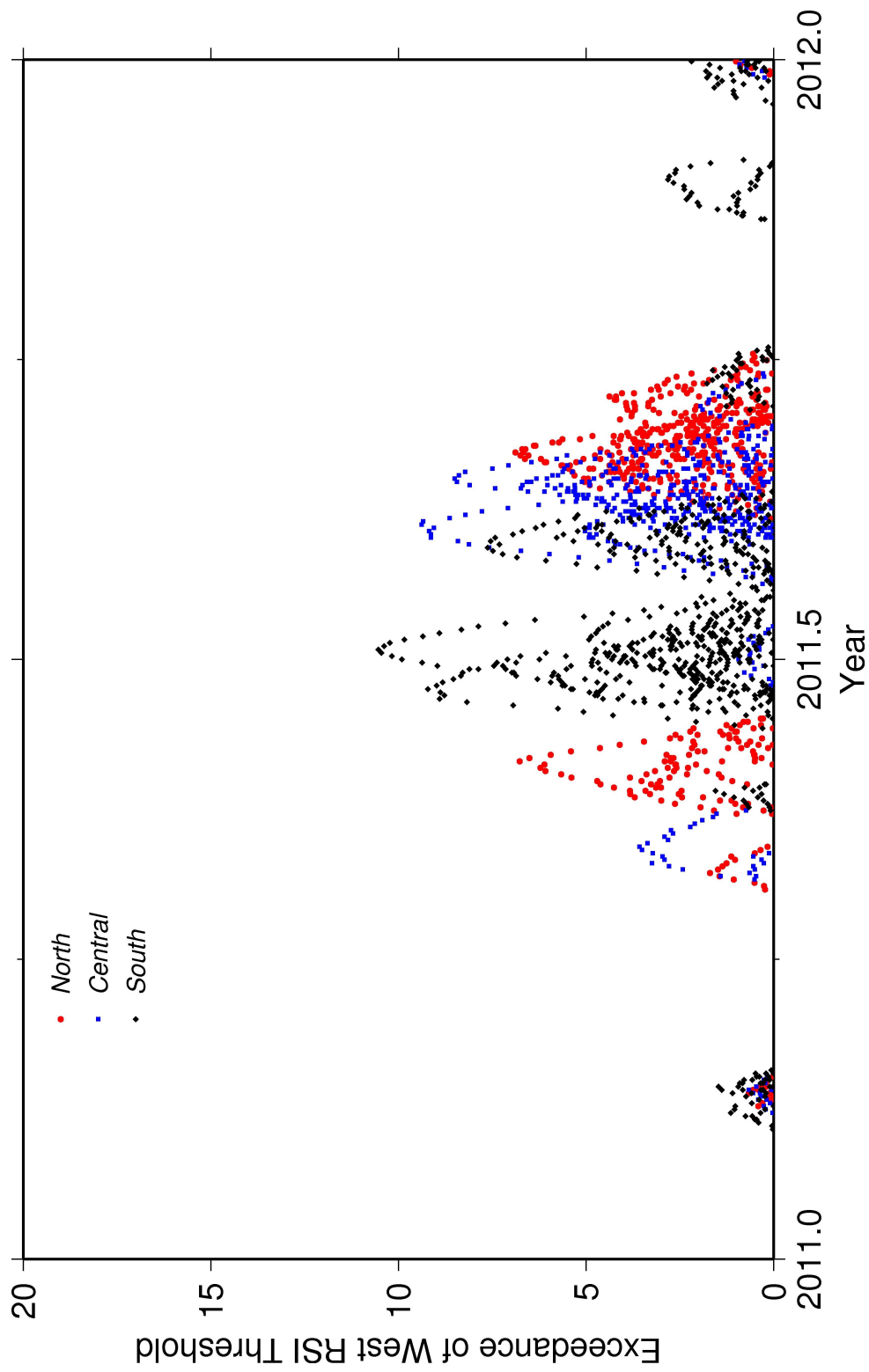


Figure 3.21: Exceedance of RSI in the west direction in 2011. The different symbols represent the regional grouping of stations into North, Central, and South.

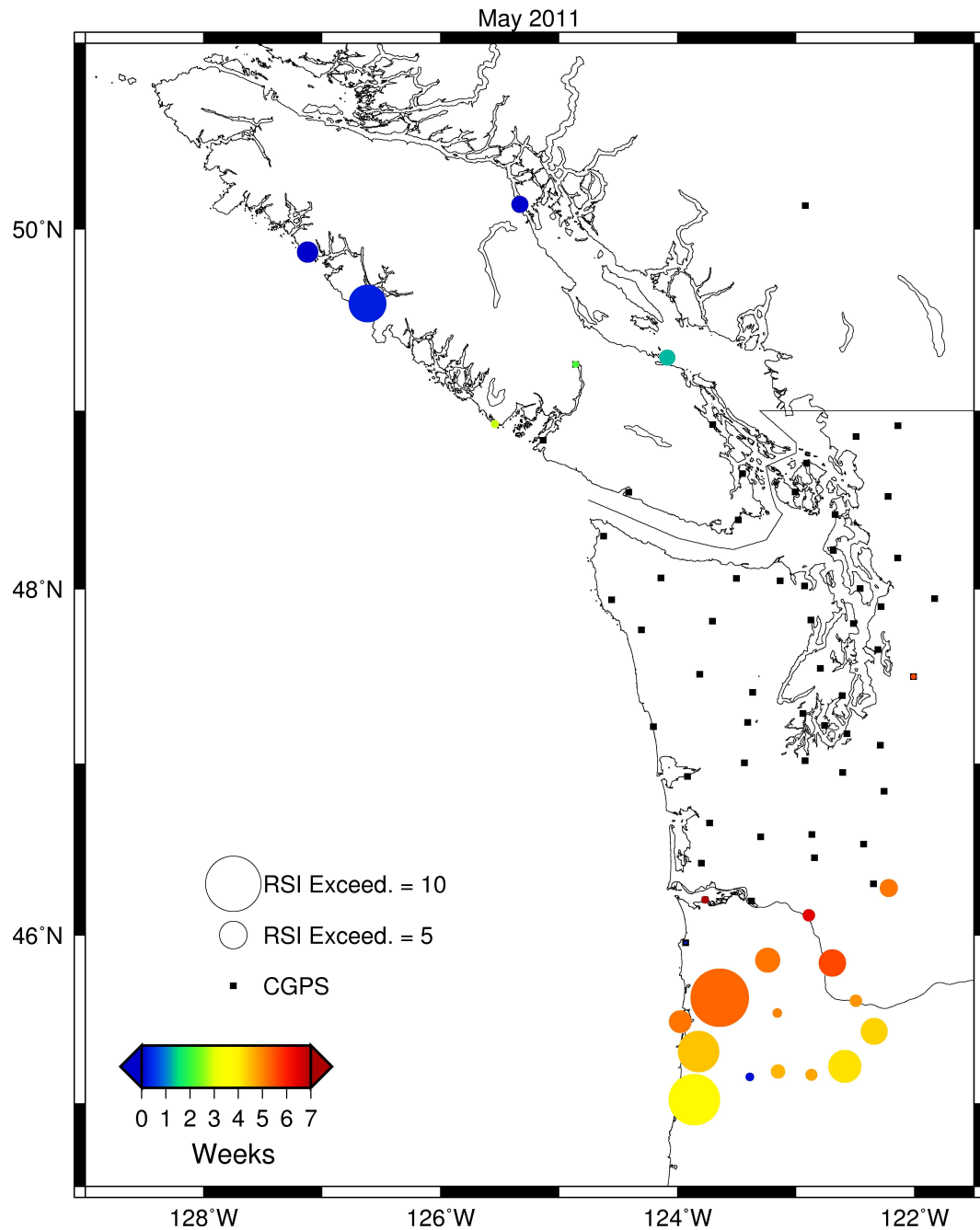


Figure 3.22: RSI detections during the May 2011 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2011.3712.

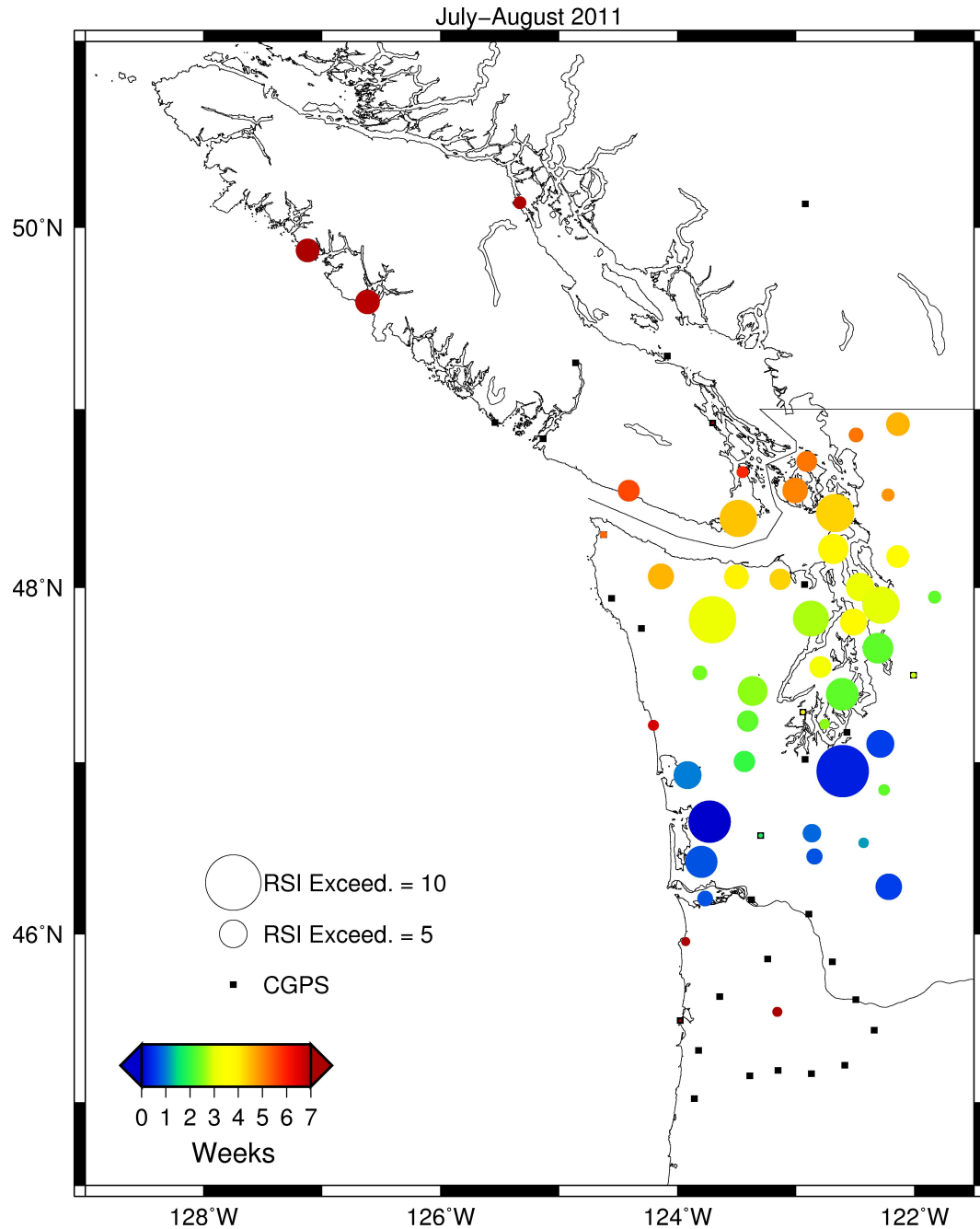


Figure 3.23: RSI detections during the July–August 2011 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2011.5575.

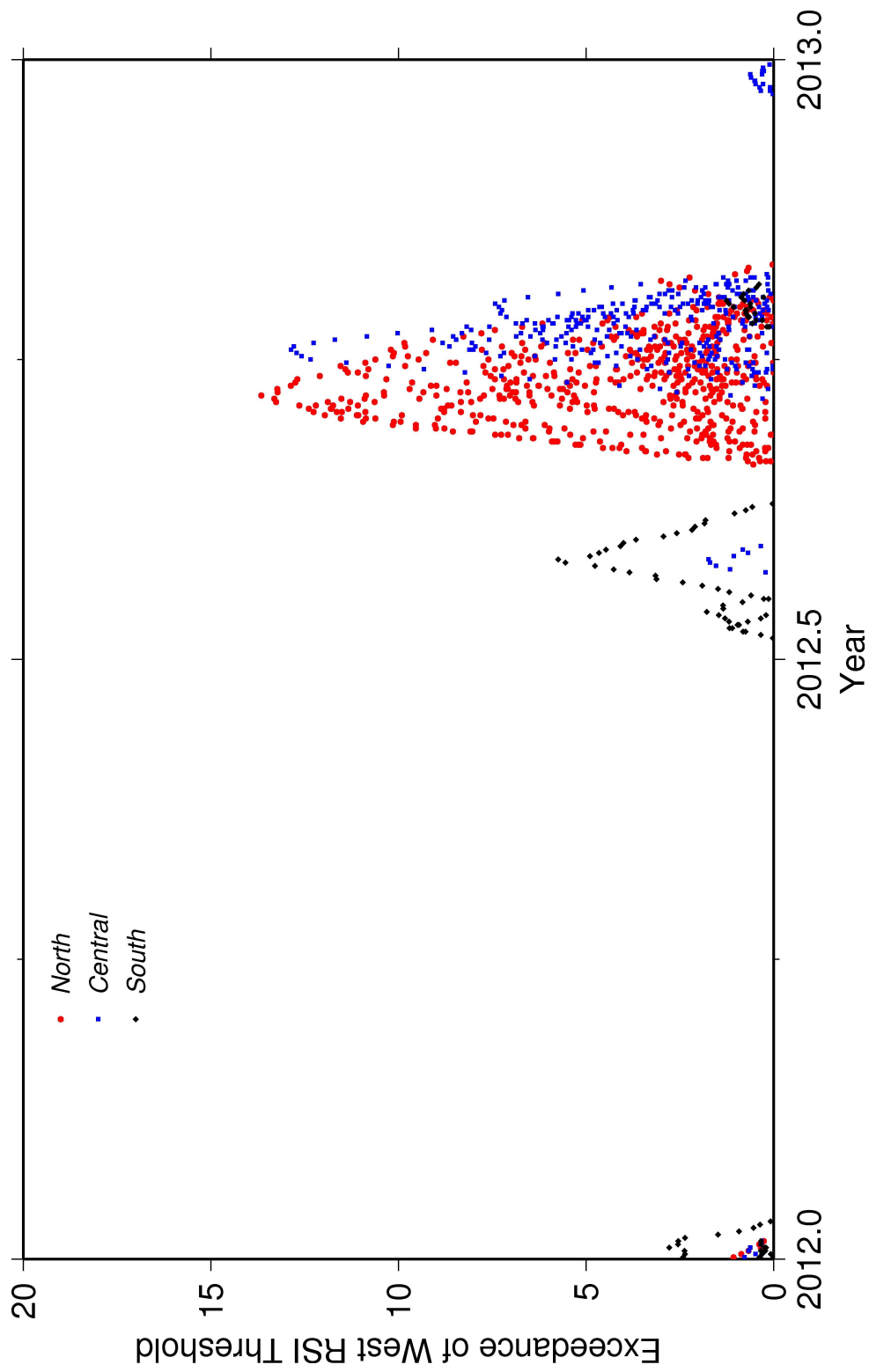


Figure 3.24: Exceedance of RSI in the west direction in 2012. The different symbols represent the regional grouping of stations into North, Central, and South.

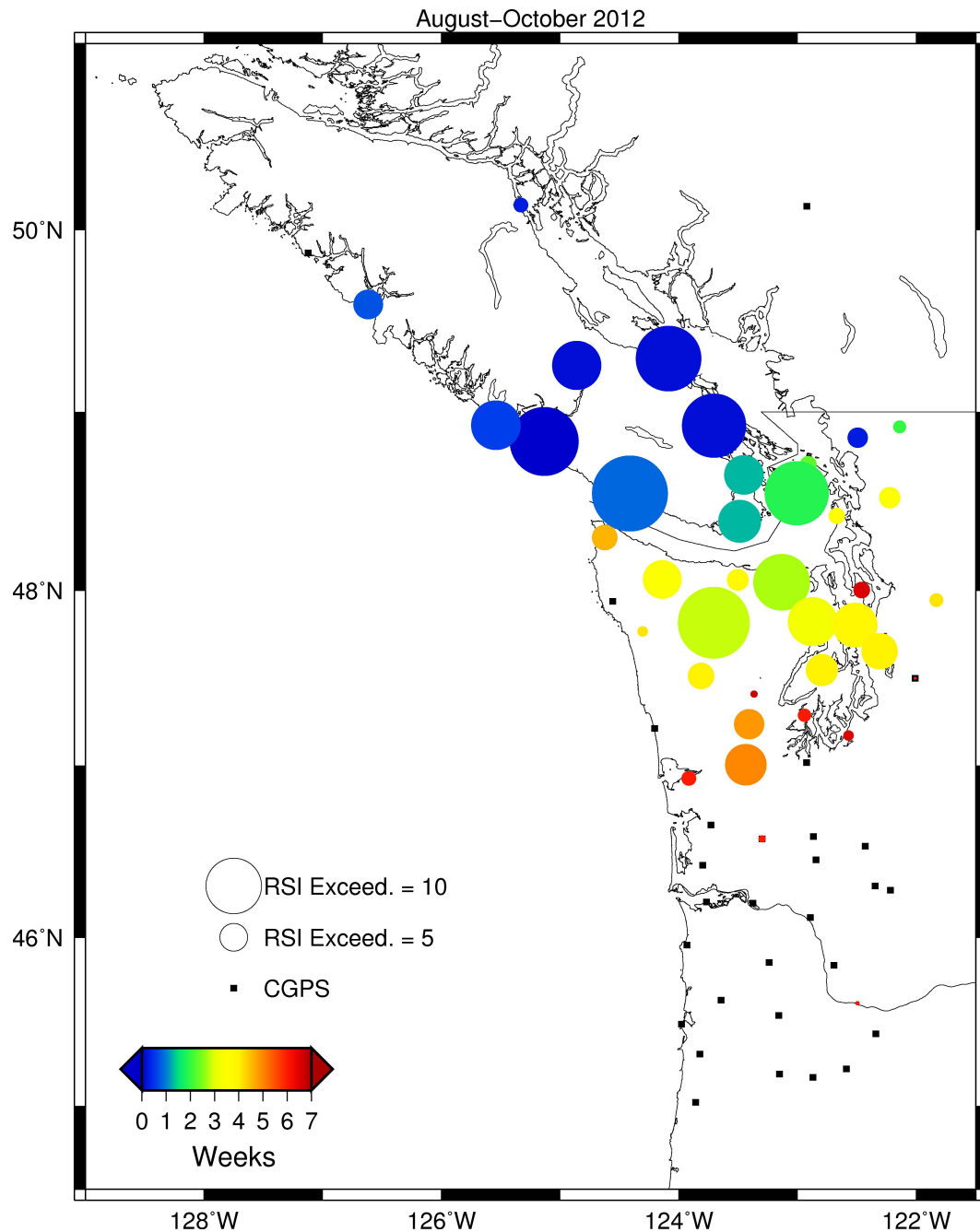


Figure 3.25: RSI detections during the August–October 2012 slow slip event. The circles represent the amount of RSI exceedance and the colors are the arrival times determined from RSI threshold exceedance. The black squares are other CGPS stations available at the time of the event. The time scale is weeks from 2012.6626.

mm (north), and 0.35-173.76 mm (up) at the station. For each slip magnitude, we use 50 different rupture durations between 2 and 100 days, in 2 day increments, for a total of 25000 combinations. We look at different rupture durations to test the ability to detect very long signals that may not have a clear arrival. For the rupture, we slip all of the fault patches together and just vary the amount of slip applied per epoch until the full slip amount is achieved. We choose to do this so we can test how well we are predicting the rupture duration rather than have a propagating rupture that may only affect the station for a few epochs once the rupture is near the station.

Figure 3.26 shows the total number of false positives as a function of displacement. We see that above 10 mm of displacement, there are no false positives in the synthetic tests for all slip durations. The maximum number of false positives is 71, which is about 1% of all epochs. Of note is that as the duration time increases, so does the number of false positives for a similar displacement. Figure 3.27 shows the maximum RSI exceedance value for a false positive as a function of displacement. From Figure 3.27, we see that the maximum RSI exceedance value of a false positive is 1.00. Since this is a low and consistent number, in order to reduce false positive detections, excluding detections not greater than 1.0 might be a worthwhile endeavor without removing meaningful detections.

3.7 Improvements to the Time Series

Using the RSI detections, we model each time series with and without the assumed offsets to quantify improvements in the residuals (real displacements minus modeled displacements). Figure 3.28 shows the variance reduction of the residual time series for all stations, where positive values indicate less variance in the residuals of the time series using the RSI detection offsets such that

$$VR = \left(\frac{\sigma_{withoutoffsets}^2 - \sigma_{withoffsets}^2}{\sigma_{withoutoffsets}^2} \right) \times 100 \quad (3.7)$$

Of note is that variance is reduced for all stations and all components using the RSI detections. The mean and median variance reductions are 29.6/25.9, 46.6/49.3, and

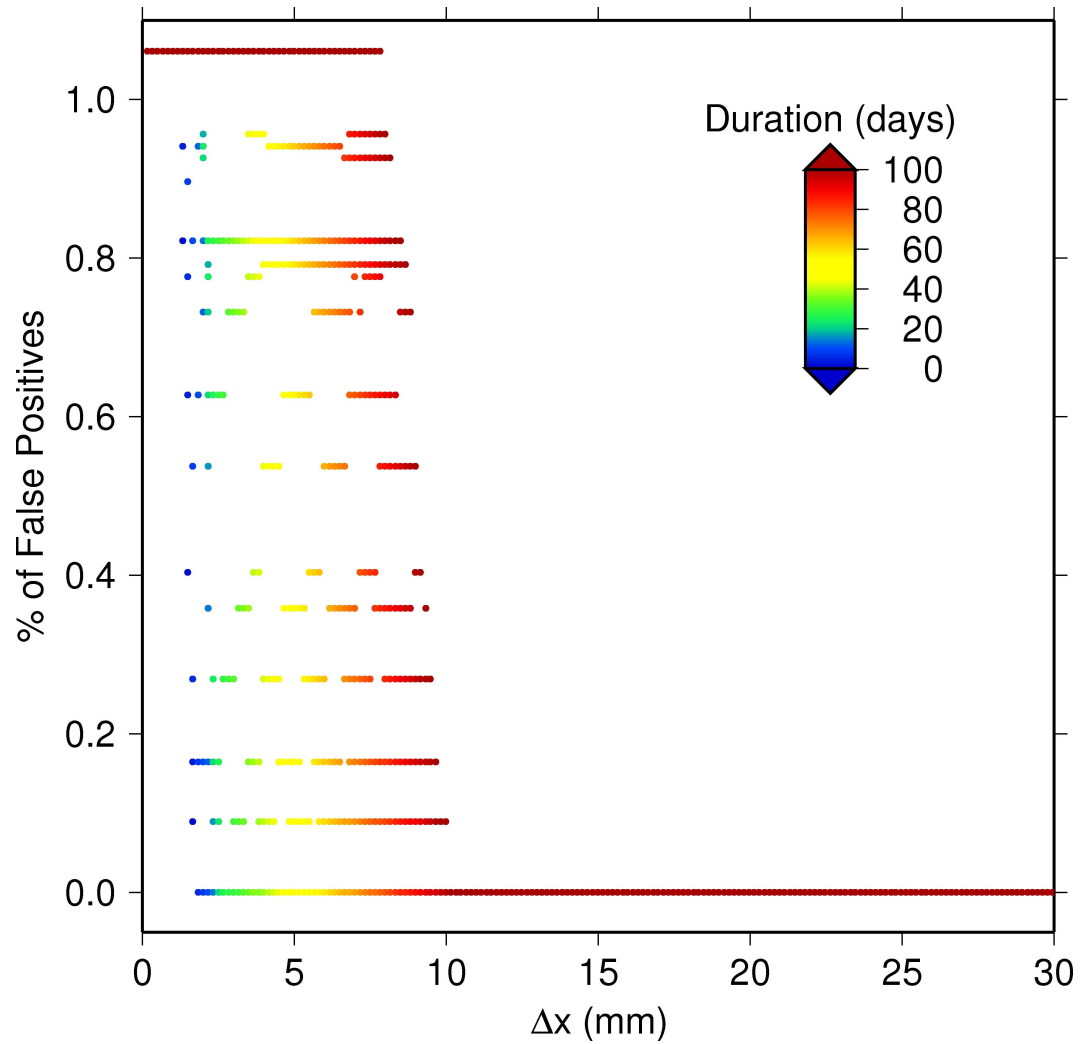


Figure 3.26: Percentage of epochs in the synthetic RSI tests exceeding the threshold values during quiescent periods as a function of displacement. The color bar represents the duration of slip in days.

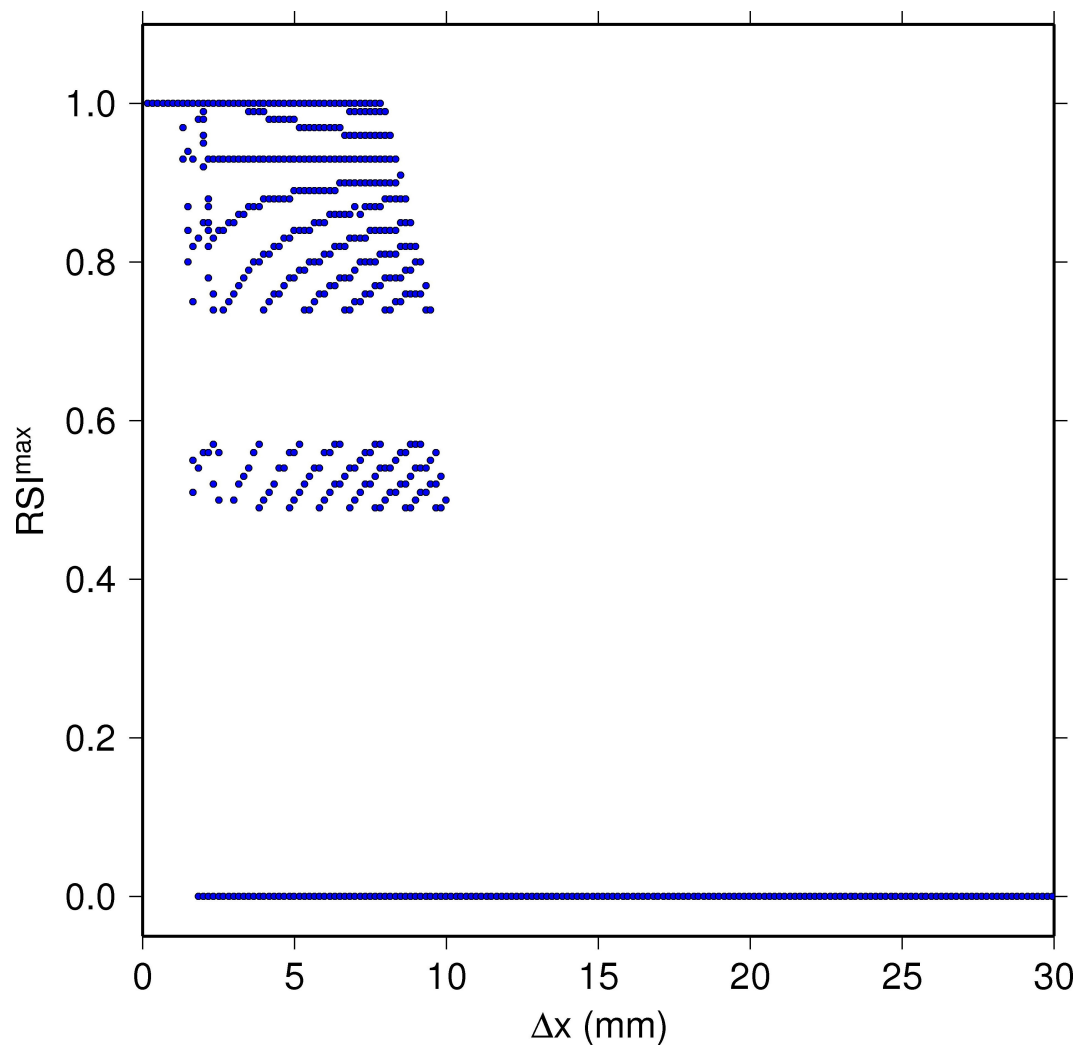


Figure 3.27: Maximum RSI exceedance values for false positive detections in the synthetic tests as a function of displacement.

26.7/23.7 percent for the north, east, and up components respectively. It is of no surprise that the variance reduction is highest in the east component due to the majority of the transient signals being oriented in this direction, although significant gains are made in the other two components. Due to the positive gains made in time series modeling, we can safely conclude that offsets obtained in an automated manner with the RSI are robust enough for ingestion into slip models to monitor geodetic moment release in areas of high seismic hazard.

3.8 Conclusions

We have shown that the RSI is a simple, stable, and objective method for automated detection of anomalous signals in geophysical time series. We determined the characteristics of all ETS events in Cascadia from 2003 to 2013, also detecting events not thought to have significant geodetic signatures. Using this framework, it is easy to apply to all stations in western North America to search for unknown transients. We showed that false positive detections are kept less than 1% of epochs, and considerably less in time series with significant geodetic signals. Finally, we showed that using the detections from the RSI results in significant variance reduction on the modeled residual time series compared to not using the recorded offsets, which lays the groundwork for a fully automated time series modeler.

3.9 Acknowledgements

We acknowledge the Southern California Integrated GPS Network and its sponsors, the W.M. Keck Foundation, NASA, NSF, USGS, SCEC; the Plate Boundary Observatory operated by UNAVCO for EarthScope (<http://www.earthscope.org>) and supported by NSF grant EAR-0323309. This work was funded by NASA grants (SENH) NAG5-13269, (AIST) NNX09AI67G, and (ROSES) NNX12AK24G. This chapter is in preparation for publication as: **Crowell, B.W., and Y. Bock (2013), Automated detec-**

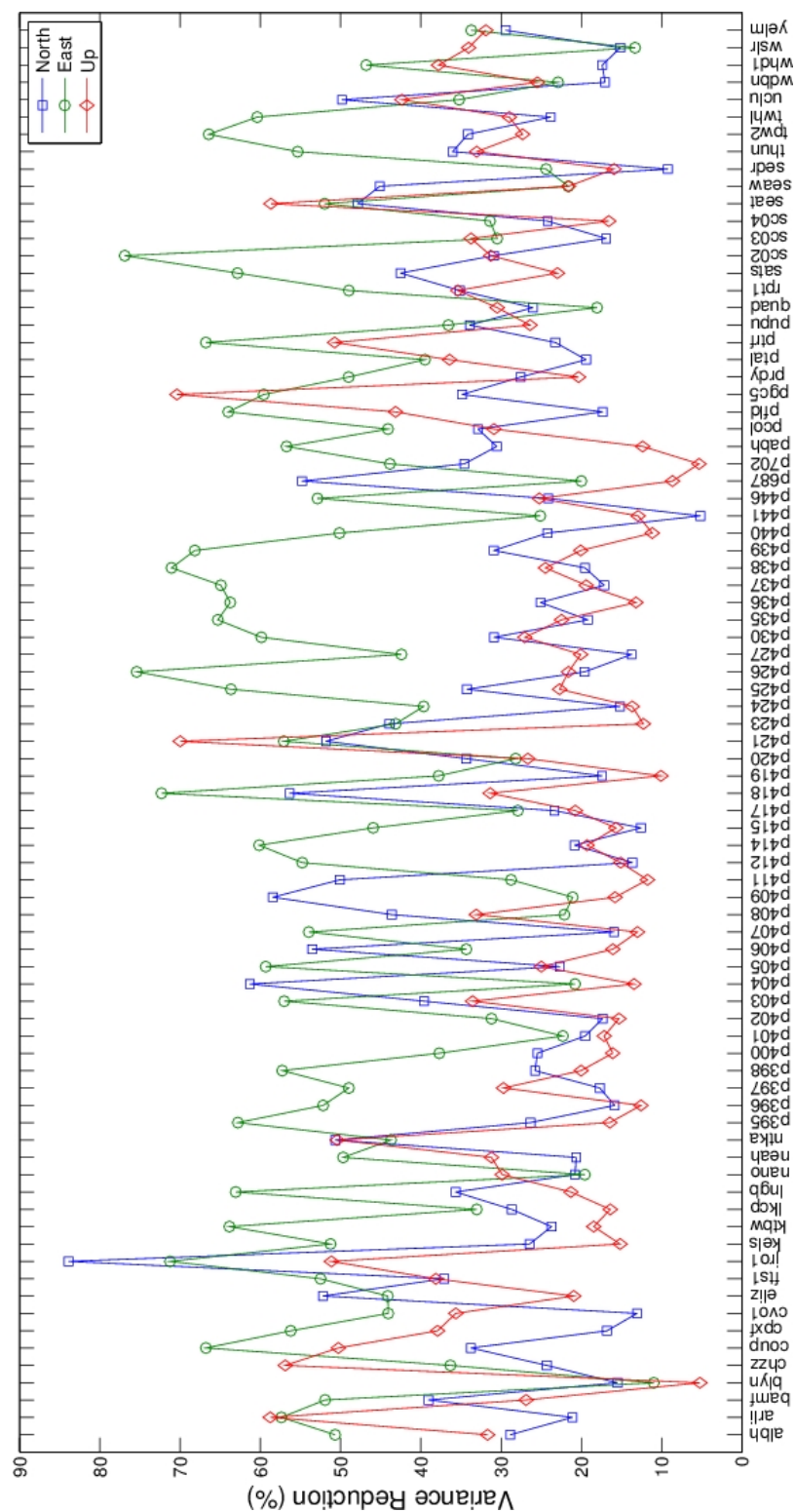


Figure 3.28: Variance reduction from time series modeling using offsets from RSI detections versus no offsets used for all stations.

tion of transient deformation in continuous GPS time series: A case study of the Cascadia subduction zone, *J. Geophys. Res.*, in prep.

Table 3.1: Slow slip events recorded between 2003 and 2013

Event	Start Time	Stop Time	Duration	# of Sites	Max RSI	Region
1	2003.0890	2003.2562	62	23	8.45	N, C, S
2	2003.4041	2003.4425	15	2	0.98	S
3	2004.3429	2004.4385	36	5	5.94	C, S
4	2004.5260	2004.6079	31	11	10.03	N, C
5	2004.9139	2004.9467	13	2	1.72	N
6	2005.2123	2005.2507	15	2	1.83	N, C
7	2005.3082	2005.3658	22	5	4.95	C
8	2005.6890	2005.7822	35	18	11.18	N, C
9	2005.9000	2006.0726	64	18	7.59	N, C, S
10	2006.1548	2006.2178	24	10	3.25	N, C, S
11	2006.3849	2006.4288	17	4	2.75	N, C
12	2006.8342	2006.9000	25	6	4.54	N, C, S
13	2007.0479	2007.1740	47	53	19.26	N, C, S
14	2007.3384	2007.4041	25	2	4.52	N
15	2007.3493	2007.4260	29	2	4.10	S
16	2007.4616	2007.5493	33	11	6.67	S
17	2007.7712	2007.8151	17	3	2.72	N
18	2007.8288	2007.9000	27	4	2.10	C, S
19	2008.2691	2008.3183	19	2	1.92	N
20	2008.3320	2008.4768	54	36	16.49	N, C
21	2008.9495	2009.0041	21	4	2.08	C, S
22	2009.0836	2009.2315	55	14	9.01	N, C, S
23	2009.2945	2009.4781	68	32	8.82	N, C, S
24	2009.5904	2009.8068	80	34	13.49	C, S
25	2010.1384	2010.2671	48	10	6.77	N, C, S
26	2010.5767	2010.7740	73	48	19.70	N, C, S
27	2011.1082	2011.1575	19	9	1.47	N, C, S
28	2011.3082	2011.3740	25	3	3.57	N, C
29	2011.3712	2011.5384	62	24	10.53	N, C, S
30	2011.5575	2011.7603	75	51	9.38	N, C, S
31	2011.8671	2011.9164	19	2	2.81	S
32	2011.9630	2012.0150	20	7	2.78	N, C, S
33	2012.5178	2012.6298	42	4	5.74	C, S
34	2012.6626	2012.8292	62	38	13.65	N, C, S
35	2012.9713	2012.9959	10	2	0.62	C

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Chapter 4

Geodetic Investigation into the Deformational Makeup of the Salton Trough

4.1 Abstract

The Salton Trough represents a complex transition between the spreading center in Baja California and the strike-slip San Andreas fault system, and is one of the most active zones of deformation and seismicity in California. We present a high-resolution interseismic velocity field for the Salton Trough derived from 74 continuous GPS sites and 109 benchmarks surveyed in three GPS campaigns during 2008-2009. We also investigate small-scale deformation by removing the regional velocity field predicted by an elastic block model for southern California from the observed velocities. We find a total extension rate of 11 mm/yr from the Mesquite Basin to the southern edge of the San Andreas fault, coupled with 15 mm/yr of left-lateral shear, the majority of which is concentrated in the southern Salton Sea and Obsidian Buttes. Differential shear strain is exclusively localized in the Brawley Seismic Zone and dilatation rate indicates widespread extension throughout the zone. In addition, we infer counterclockwise rotation of $10^\circ/\text{Ma}$, consistent with northwestward propagation of the Brawley Seismic

Zone over geologic time.

4.2 Introduction

The Salton Trough lies at the transition from the spreading centers of the Gulf of California and East Pacific Rise to the south into the predominately strike-slip San Andreas fault system to the north, and is a classic example of a transtensional basin. The historical and paleoseismic records show that the San Andreas fault in California experiences great earthquakes every 150 years on average, however the southernmost segment of the San Andreas is the only seismogenic part of the fault that has not ruptured in historic times, and is approaching or exceeding the maximum slip deficit known from paleoseismic studies along this part of the fault [Fialko, 2006]. The threat of a large earthquake on the southern terminus of the San Andreas fault has generated increased interest in the seismotectonics of the Salton Trough. *Brothers et al.* [2011] showed that slip along conjugate faults in the Salton Sea causes an appreciable change in Coulomb stress along the southern San Andreas fault, making full characterization of deformation in the Brawley Seismic Zone necessary for hazard assessment. The Imperial fault, at the southern end of the Imperial Valley, is a fast-moving and partially creeping fault that accommodates most of the relative motion between the North American and Pacific plates at that latitude, roughly 40 mm/yr [Genrich *et al.*, 1997; Lyons *et al.*, 2002], and also has been the location for many large earthquakes (e.g. 1940 Mw 7.1 and 1979 Mw 6.4 [Archuleta, 1982]). From the terminus of the Imperial fault and other auxiliary faults such as the Brawley and Superstition Hills faults, there exists a large step over to the San Andreas fault. Many parallel oblique normal N15°E striking faults help bridge the gap between the faults [Brothers *et al.*, 2009]. Large scale subsidence and anomalous heat flow exist in the Imperial Valley near Obsidian Buttes and Mexico's Cerro Prieto volcanic zone caused by both tectonic extension and water extraction for irrigation and geothermal plants [Lachenbruch *et al.*, 1985; Lynch and Hudnut, 2008].

Seismically, the region is one of the most active in California, with many small

swarms of earthquakes such as the 1982 West Moreland, the 2005 Obsidian Buttes, the 2009 Bombay Beach and the 2012 Brawley seismic swarms [Chen and Shearer, 2011; Lohman and McGuire, 2007; Shearer, 2002]. Recent moderate-to-large earthquakes include the 1987 Mw 6.6 Superstition Hills and Mw 6.2 Elmore Ranch earthquakes that occurred 1 day apart [Hanks and Allen, 1989], the 1954 Mw 6.4 San Jacinto earthquake [Sharp, 1967], the 1968 Mw 6.5 Borrego Mountain earthquake [Burford, 1972], and the 2010 Mw 7.2 El Mayor-Cucapah earthquake [Hauksson *et al.*, 2011].

A detailed picture of crustal motion and strain accumulation in this complex tectonic region requires geodetic measurements with high spatial density. The Imperial and Coachella Valleys are both locations of intense agricultural activity resulting in temporal decorrelation in Interferometric Synthetic Aperture Radar (InSAR) [Lohman and McGuire, 2007]. Persistent scattering methods have been partially successful in the areas north of the Salton Sea, but less effective to the south [Lyons and Sandwell, 2003]. A significant number of continuous Global Positioning System (GPS) stations have been installed in this region, but the average spacing between the continuous GPS monuments is ~ 20 km, so local GPS surveying is the only feasible option for resolving small-scale deformation in this area. To increase the spatial density of measurements, we conducted three surveys in the Salton Trough, from the United States-Mexico border into the Coachella Valley to the north. We also used data from previous surveys performed in the early summers of 2000, 2003, 2004 and 2005 to compute station velocities. After creating a velocity field, we computed baseline length changes across regions of high deformation in the Salton Trough and determined the amount of motion through four provinces: the Southern Salton Sea, Obsidian Buttes, the Central Brawley Seismic Zone and the Mesquite Basin. We also removed a component of regional tectonics using a block model [Smith-Konter and Sandwell, 2009] to look at local deformation caused just by auxiliary faulting. To look in more detail at internal deformation, we computed strain rates from the velocity fields throughout the Salton Trough, which helps us characterize the amount of shear, dilatation, and rotation throughout. We also present an analysis of vertical deformation from continuous GPS to complement the extension measurements in the Salton Trough, and argue that the available data indicate

the northwestward propagation of the Brawley Seismic Zone.

4.3 Data

4.3.1 Campaign GPS Surveys and Processing

We performed three campaign GPS surveys in February, 2008, October, 2008, and February, 2009 (Figure 4.1). During each survey, teams autonomously surveyed a subset of sites at least twice during each campaign. Each occupation was between 12 and 20 minutes at a one second sampling rate. Occupations of the same monument were performed at different times during the day, approximately 6 and 18 hours away from the first occupation, to obtain a different satellite set to reduce multipath errors. Using the method of instantaneous positioning [*Bock et al.*, 2000], we were able to carry out the surveys in a rapid and efficient manner so that many monuments could be surveyed with modest manpower and processed on-the-fly with respect to continuous GPS stations in the California Real Time Network (CRTN) (<http://sopac.ucsd.edu/projects/realtime/CRTN/>). Raw receiver data, either from Leica GMX902 or Ashtech Z-X113 GPS receivers, were streamed via Bluetooth protocol to Verizon XV6700 smart phones running Geodetics, Inc. RTD Rover software. The raw data were streamed to a server at the Scripps Orbit and Permanent Array Center (SOPAC), where positions were computed using the International Terrestrial Reference Frame (ITRF) 2005 [*Altamimi et al.*, 2007] coordinates of the CRTN stations valid for that day, ultra-rapid SOPAC orbital parameters (<http://garner.ucsd.edu/pub/products>) and the NOAA Tropospheric Signal Delay Model (NOAATrop) [*Gutman et al.*, 2004]. Positions and statistics were then streamed back to the user where they were stored on the memory card of the smart phone. Raw receiver data were stored on the server as well as the memory card for post-processing.

The February, 2008 survey consisted of 34 geodetic monuments established by Imperial College, London [*Mason*, 1987] and 4 National Geodetic Survey (NGS) sites centered on the Imperial fault. Previous GPS surveys of this dense array of monuments

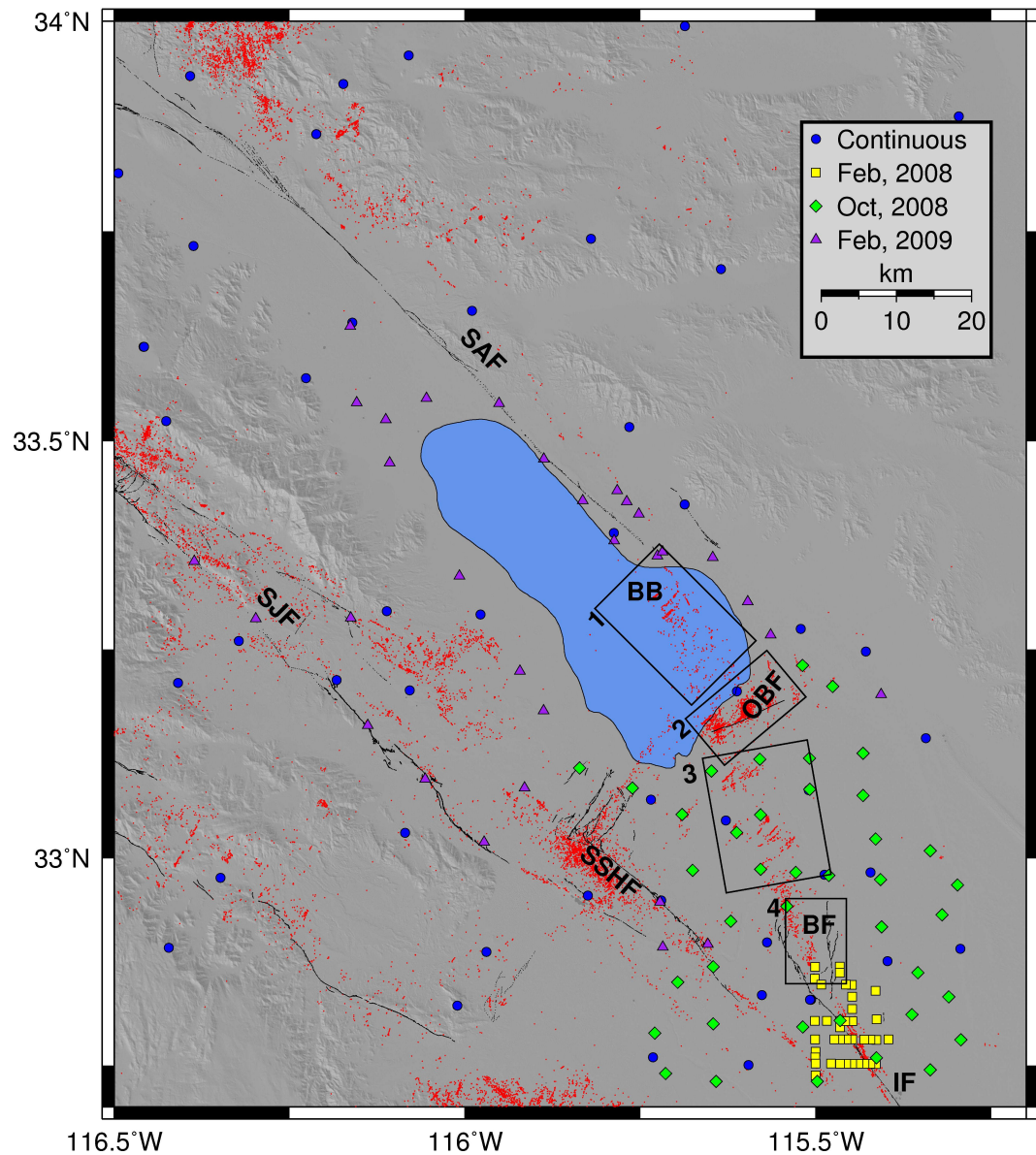


Figure 4.1: Locations of continuous and campaign GPS monuments in the Salton Trough. The boxes represent the four main study areas: the southern Salton Sea (1), Obsidian Buttes (2), the Central Brawley Seismic Zone (3) and the Mesquite Basin (4). Abbreviations: SAF, San Andreas fault; SJF, San Jacinto fault; SSHF, Superstition Hills fault; OBF, Obsidian Buttes fault; BF, Brawley fault; IF, Imperial fault; BB, Bombay Beach. The red dots are all earthquakes greater than M 2 from the *Lin et al.* [2007] catalog.

occurred in 1991, 1993, 1999 and 2000 [Genrich *et al.*, 1997; Lyons *et al.*, 2002]. The surveys in 1999 and 2000 [Lyons *et al.*, 2002] were processed through the method of instantaneous positioning with respect to the ITRF 1996 reference frame [Sillard *et al.*, 1998] and the earlier surveys were processed in GAMIT [Herring *et al.*, 2010]. For the purposes of this study, we only use the 2000 results from Lyons *et al.* [2002] since the earlier studies would be affected by coseismic motions from the 1999 Hector Mine earthquake. We transformed these positions into the ITRF 2005 reference frame [Altamimi *et al.*, 2007] prior to computing velocities.

In October, 2008, a survey of stations throughout the Imperial Valley was performed, encompassing the Imperial fault and the Brawley Seismic Zone. Of the 43 monuments surveyed, 23 were NGS and 20 were monuments installed in 2004 by the California Department of Transportation (Caltrans) and only surveyed once before. The Caltrans stations greatly increased the density throughout the Brawley Seismic Zone. The survey in February, 2009 extended the October, 2008 survey from the Brawley Seismic Zone into the Coachella Valley along both sides of the Salton Sea and consisted of 28 stations, 18 from the NGS and 10 from the USGS. The survey in October, 2008, consisted of stations surveyed previously in 2004. The 2004 occupations were done at a 15 second sampling rate, and generally occupations were between 10 and 15 minutes. The survey in February, 2009 occupied sites that were previously surveyed in 2000, 2003, 2004 or 2005. There were 5 NGS monuments surveyed in 2000 in the Coachella Valley and 3 NGS monuments surveyed in 2003 near the southern terminus of the SAF that we resurveyed in 2009. The 2000 and 2003 data were at a 15 second sampling rate, and consisted of longer occupations (15-30 minutes) than the 2004 data. There were 10 NGS stations that we surveyed in February, 2009 that were previously surveyed in 2004. These stations were ones that we were not able to survey during the October, 2008 survey. We resurveyed 10 monuments installed in 2005 by the USGS that are within a few km of the San Andreas, San Jacinto and Superstition Hills faults. Data from these monuments were at a sampling rate of once per second and, in many cases, consisted of occupations many hours long. We post-processed the 2003, 2004 and 2005 data through instantaneous positioning and combined the occupations using DeLorme's GeoSpider

network adjustment program using the true-of-date positions of nearby continuous GPS stations as reference stations.

Velocities were computed with respect to the ITRF 2005 reference frame [Altamimi *et al.*, 2007] and then corrected for North American plate motion through an Euler pole calculation. The one-sigma uncertainties on the velocity measurements are simply defined as

$$\sigma_{N,E} = \frac{\sqrt{\sigma_{N_1,E_1}^2 + \sigma_{N_2,E_2}^2}}{t} \quad (4.1)$$

where $\sigma_{N,E}$ are the velocity uncertainties, σ_{N_1,E_1} and σ_{N_2,E_2} are the individual positional uncertainties, and t is the time between observations in years [Lyons *et al.*, 2002]. Although the cross-covariances between north and south components have been neglected, the positional variances have been adjusted to account the temporal correlations in the GPS observations [Genrich and Bock, 2006].

4.3.2 Continuous GPS Processing

Over 1800 stations in western North America were, at the time of this study, processed daily at SOPAC using the GAMIT/GLOBK package [Herring *et al.*, 2010] with respect to the ITRF 2005 reference frame. For this study, we utilize 74 continuous GPS stations around the Imperial Valley and estimate velocities through least squares between 2000 and 2010. We exclude modeling postseismic terms and fit only a linear trend, annual, and semiannual terms since we are interested in the actual velocities during the time period, rather than the estimated interseismic velocities that are computed through a superposition of many model terms. This also produces velocities that are similar to the campaign velocities since only a simple linear trend is fit to that data. After velocities are computed, we perform the same Euler pole correction used for the campaign measurements to put the velocities in the local reference frame.

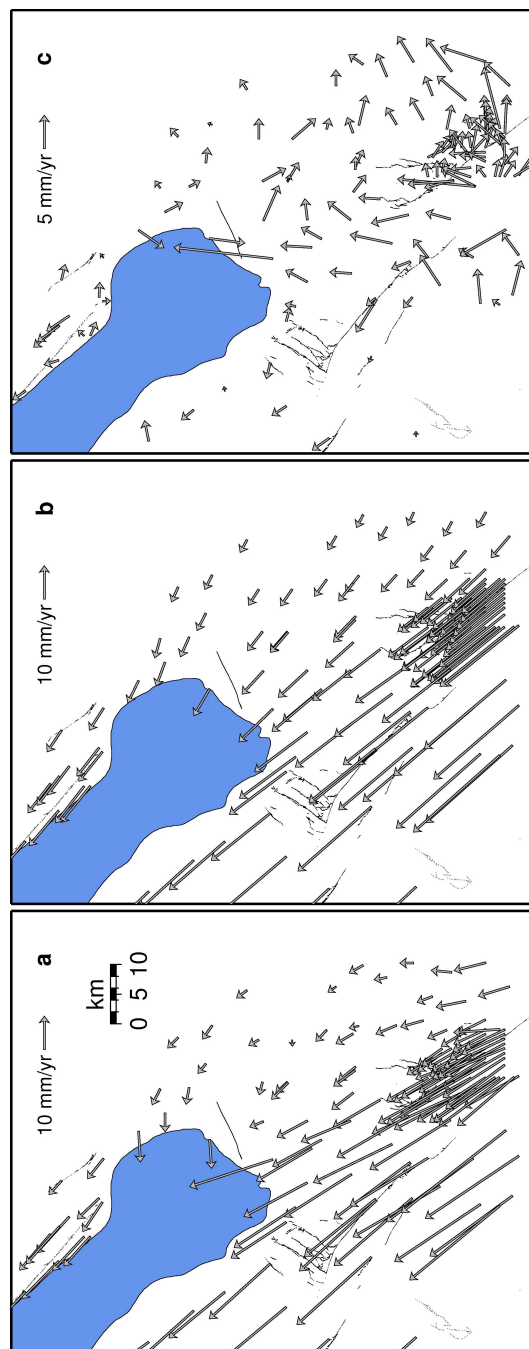


Figure 4.2: Interseismic velocity fields from campaign and continuous GPS measurements (a), the block model of *Smith-Konter and Sandwell* [2009] (b), and the difference between the two (c).

4.4 Methods

4.4.1 Extensional Baseline Measurements

The observed velocity field can be represented as the superposition of a regional velocity field and a local velocity field. The regional field is due to large scale tectonics and is dominated by the relative motion between the Pacific and North American plates. The local field is the result of conjugate faulting, fault edge effects, hydrological signals and diffuse (i.e., not related to any major faults) deformation. Since the velocity field is dominated by the regional field, in order to isolate the local component, we remove the regional field from the total velocity field. We take the block model solution of *Smith-Konter and Sandwell* [2009] for southern California as a proxy for the regional velocity field. The block model is a good approximation for the regional velocity field because it incorporates only the large scale features, leaving features associated with conjugate faulting, extension and crustal thinning in the Salton Trough. Figure 4.2 shows the total, regional, and local velocity fields. Of note is that velocities in the far-field go to zero in the local velocity field, indicating that the block model is able to remove large scale tectonic features while leaving the more complicated local effects.

Horizontal baselines are computed between pairs of stations that traverse lines of seismicity and known faulting using the local velocity field. The four major regions we are concerned about are the southern Salton Sea, Obsidian Buttes, the Central Brawley Seismic Zone, and the Mesquite Basin (Figure 4.1, regions 1-4 respectively). These four regions make up the Brawley Seismic Zone. We rotate the baselines into fault parallel and perpendicular directions to obtain the strike-slip and extensional motions throughout the region. In the southern Salton Sea, the Bombay Beach seismic swarms of 2001 and 2009 were along planes striking $N43^{\circ}E$ to $N57^{\circ}E$, respectively [*Chen and Shearer*, 2011]. Seismic reflection data obtained along the floor of the Salton Sea shows the existence of $N15^{\circ}E$ striking faults, however, events on these structures are thought to be infrequent [*Brothers et al.*, 2009]. *Lohman and McGuire* [2007] show from seismic relocations during the 2005 Obsidian Buttes seismic swarm that the predominant strike

on the Obsidian Buttes fault is N35°E with a dip of 72°, which is similar to the migration pattern from *Chen and Shearer* [2011] (N20-66°E). However, field observations of surface offsets indicate a strike of N65°E (fault trace on Figure 4.1 near OBF), while the strike of geothermal plants is between N50-55°E. For computing baseline length changes, we choose N65°E to correspond with our field observations. Predominant fault planes in the Central Brawley Seismic Zone (swarms in 1983, 1986, 1999, 2008, and 2012) and Mesquite Basin (swarms in 1983, 2000, and 2003) are highly variable, although the most common orientation is about N40°E for the Central Brawley Seismic Zone and N50°E for the Mesquite Basin [*Chen and Shearer*, 2011].

After the baselines are computed and rotated into their fault-parallel and perpendicular directions, we compute a weighted mean of the baselines to determine the deformation of the area. The weighted mean is defined as

$$\bar{b} = \frac{\sum_{i=1}^n \sigma_i^{-1} b_i}{\sum_{i=1}^n \sigma_i^{-1}} \quad (4.2)$$

where b are the baseline measurements and σ are the baseline velocity uncertainties (the sum of the individual station uncertainties in the baseline). This in essence gives greater weight to baselines with two continuous GPS stations which are more reliable than baselines between campaign GPS sites. Uncertainties in subsidence and dip-slip rates are computed using an assumed 2° uncertainty on the dip angle.

4.4.2 Strain Rate Computation

In addition to baseline velocity changes, we also compute the velocity gradient tensor over a 0.1° grid following a least squares solution using both the block model velocity field (interpolated onto observation points) and our velocity field from campaign and continuous GPS measurements. These two strain rate fields are then differenced which allows us to appropriately locate where differential strain is being accumulated throughout the Salton Trough. GPS velocity at a given point can be written as

$$u_i = t_i + \frac{\partial u_i}{\partial x_j} \Delta x_j = t_i + L_{ij} \Delta x_j \quad (4.3)$$

where indexes i and j correspond to spatial coordinates, t_i is translation with respect to the reference frame, L_{ij} is the velocity gradient tensor, u_i are the individual GPS velocities and Δx_j is the baseline between each station and the grid point [Allmendinger *et al.*, 2007; Shen *et al.*, 1996]. The velocity gradient tensor in two dimensions can be decomposed into a symmetric and anti-symmetric component such that

$$L = \begin{bmatrix} \frac{\partial u_x}{\partial x} & \frac{1}{2} \left(\frac{\partial u_x}{\partial y} + \frac{\partial u_y}{\partial x} \right) \\ \frac{1}{2} \left(\frac{\partial u_y}{\partial x} + \frac{\partial u_x}{\partial y} \right) & \frac{\partial u_y}{\partial y} \end{bmatrix} + \begin{bmatrix} 0 & \frac{1}{2} \left(\frac{\partial u_x}{\partial y} - \frac{\partial u_y}{\partial x} \right) \\ \frac{1}{2} \left(\frac{\partial u_y}{\partial x} - \frac{\partial u_x}{\partial y} \right) & 0 \end{bmatrix} \quad (4.4)$$

$$L = E + \Omega = \begin{bmatrix} e_{xx} & e_{xy} \\ e_{yx} & e_{yy} \end{bmatrix} \quad (4.5)$$

In Equation 4.5, E is the strain rate tensor, Ω is the rotation rate tensor, and e_{ij} are components of the velocity gradient tensor, $e_{ij} = \frac{\partial u_i}{\partial x_j}$. To solve for the velocity gradient tensor, we set up the following inverse problem

$$u = \begin{bmatrix} u_x^1 \\ u_y^1 \\ \vdots \\ u_x^n \\ u_y^n \end{bmatrix} = Gl = \begin{bmatrix} 1 & 0 & \Delta x_1 & \Delta y_1 & 0 & 0 \\ 0 & 1 & 0 & 0 & \Delta x_1 & \Delta y_1 \\ & & & \vdots & & \\ 1 & 0 & \Delta x_n & \Delta y_n & 0 & 0 \\ 0 & 1 & 0 & 0 & \Delta x_n & \Delta y_n \end{bmatrix} \begin{bmatrix} t_x \\ t_y \\ e_{xx} \\ e_{xy} \\ e_{yy} \\ e_{yx} \end{bmatrix} \quad (4.6)$$

where G is the design matrix, l is a vector containing the velocity gradient and translation terms from Equation 4.3, and n is the number of GPS stations. We follow the example of Allmendinger *et al.* [2007] in Gaussian distance weighting each station-grid pair by

$$W = \exp \left(\frac{-d^2}{2\alpha^2} \right) \quad (4.7)$$

where α is a constant that controls the decay and d is the distance between the grid point and the station. Grid distance weighting gives a higher weight to stations that are close to the grid point and effectively ignores stations far away. For this study, we use $\alpha = 7$ km, which is the average distance of the closest station to each grid point. Additionally, we also weigh continuous stations by a factor of two over campaign stations since we

are more confident in the continuous velocities. We then solve for the vector l (Equation 4.6) through linear least squares

$$l = [G^T W G]^{-1} G^T W u \quad (4.8)$$

The shear strain rate is defined as the off-diagonal term of the strain rate tensor (Equation 4.4)

$$E_{xy} = \frac{1}{2}(e_{xy} + e_{yx}) \quad (4.9)$$

and the principal components of strain rate are

$$E_{1,2} = \frac{(e_{xx} + e_{yy})}{2} \pm \sqrt{\frac{(e_{xx} + e_{yy})^2}{4} + (E_{xy})^2} \quad (4.10)$$

The maximum shear strain rate is given by the difference between the principal strain rates,

$$E_{xy}^{max} = \frac{E_1 - E_2}{2} \quad (4.11)$$

The dilatation rate, δ , is simply the trace of the strain rate tensor

$$\delta = E_1 + E_2 \quad (4.12)$$

We use the dilatation rate as a proxy for extension even though we are only looking in two dimensions (in three dimensions, dilatation is the volumetric change). The rotation rate, ω , comes from the rotation rate tensor in Equation 4.4 and is defined as

$$\omega = -\frac{1}{2}(e_{xy} - e_{yx}) \quad (4.13)$$

where positive values of ω correspond to clockwise rotations.

4.5 Results

4.5.1 Baseline Results

Southern Salton Sea

To look at the component of motion in the southern Salton Sea (region 1 in Figure 4.1, Figure 4.3a), we compute baselines between continuous GPS station P507

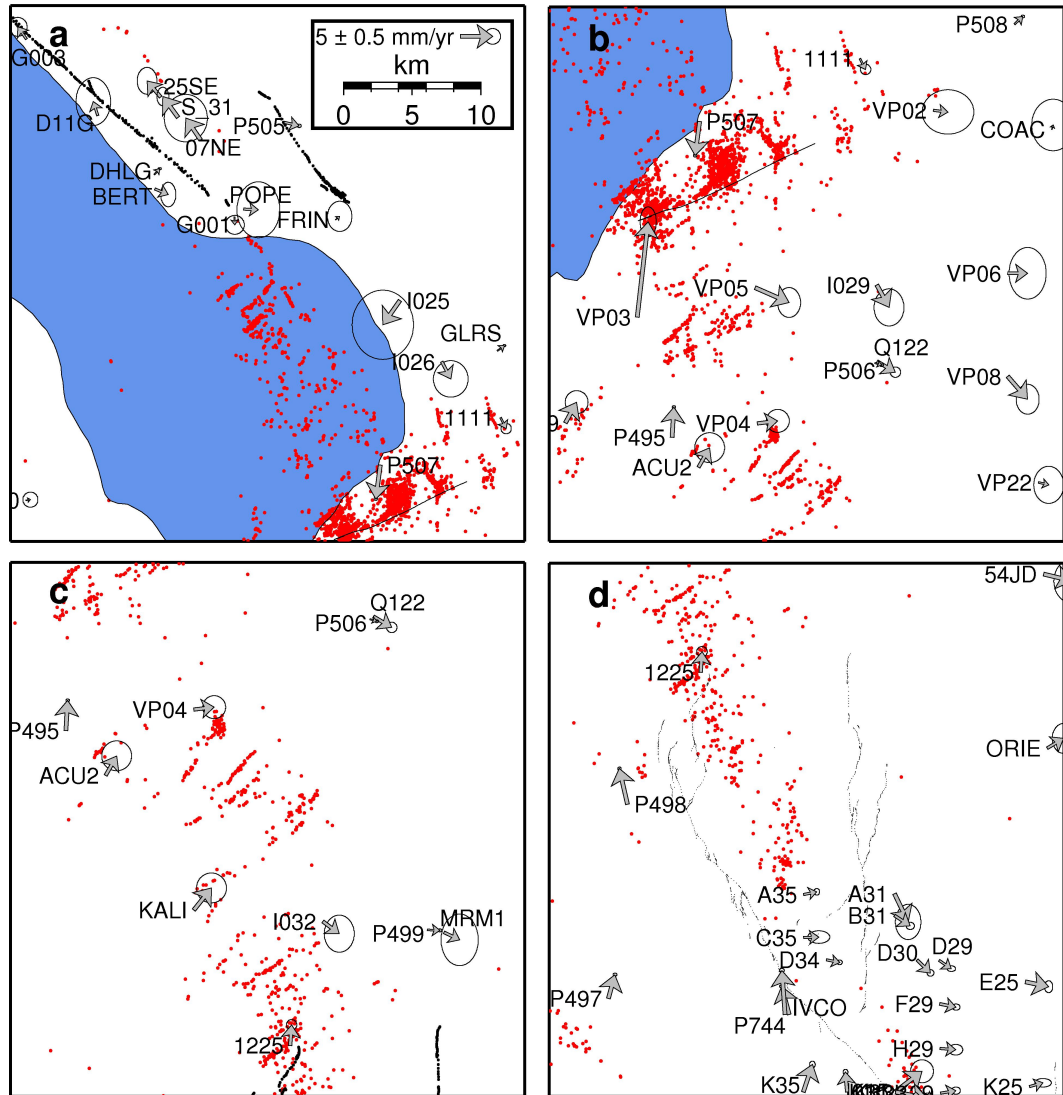


Figure 4.3: The local velocity field (total field minus block model) used for computing baseline length changes for the four regions of interest: (a) The Southern Salton Sea, (b) Obsidian Buttes, (c) the Central Brawley Seismic Zone, and (d) the Mesquite Basin. The red dots are all earthquakes greater than M 2 from the *Lin et al.* [2007] catalog. Error ellipses are plotted at the 95% confidence level.

and stations near Bombay Beach (from north to south, G003, D11G, DHLG, BERT, G001, POPE, FRIN and I025). We rotate the baselines into N50°E (the mean orientation from *Chen and Shearer* [2011]) and N40°W to obtain the fault-parallel and fault-perpendicular directions. The results are shown in Table 4.1. Excluding the baselines to FRIN and I025, we obtain a weighted mean fault-parallel velocity of 3.65 ± 0.89 mm/yr (left-lateral) and extension of 4.72 ± 0.92 mm/yr. We exclude the baselines to FRIN and I025 because they only extend across part of the zone of deformation, and therefore give a constraint as to the southern extent of motion. The baseline between FRIN/I025 and P507 indicates that most of the deformation across the southern Salton Sea is accommodated between FRIN and I025, with a small amount of extension occurring south of I025. This shows that the deformation of the southern Salton Sea is almost completely localized in the region encompassed by the 2009 Bombay Beach seismic swarm. The small amount of extension that exists between I025 and P507 can possibly be explained by northwest trending lineaments (i.e. Wister fault) along the eastern edge of the Salton Sea, an area dominated by mud volcanoes and mud pots [*Lynch and Hudnut*, 2008]. This is generally thought of as a diffuse continuation of the San Andreas fault that has become somewhat dormant but still probably accommodates a small amount of strike-slip motion on the order of a few mm per year as opposed to $\sim 15\text{-}25$ mm/yr along the southern San Andreas fault [*Fialko*, 2006; *Lindsey and Fialko*, 2013].

Obsidian Buttes

The Obsidian Buttes fault (region 2 in Figure 4.1, Figure 4.3b) is dominated by mostly left-lateral shear with a small amount of normal motion, leading to subsidence near P507. We investigate the baseline length change between P507 and VP05, Q122, I029 and VP06 (Table 4.1). The weighted mean velocities are 4.38 ± 0.80 mm/yr of fault-parallel motion and 2.62 ± 0.91 mm/yr of extension. Projecting this extension onto a fault dipping 72° to the north as indicated by *Lohman and McGuire* [2007] from seismic relocations gives 8.1 ± 3.0 mm/yr of subsidence and 8.5 ± 3.1 mm/yr of dip-slip motion (in the normal direction). This amount of interseismic loading is high compared

to the 15 year seismic recurrence interval of M5 earthquakes on the Obsidian Buttes fault [Reasenbergs and Jones, 1994]; although, as suggested by Lohman and McGuire [2007], a significant portion of slip in this region is aseismic. In the case of the 2005 Obsidian Buttes swarm, Lohman and McGuire [2007] inferred that the geodetic moment release exceeded the seismic moment release by a factor of 5, which has significant implications for seismic hazard near the southern San Andreas fault. Our estimated loading rates across a 10 km by 5 km segment of the Obsidian Buttes fault would indicate a geodetic moment release equivalent to a M5.5 earthquake every 15 years.

Central Brawley Seismic Zone

The Central Brawley Seismic Zone is a rather broad region of deformation that stretches between the northern termini of the Imperial and Brawley faults and the Obsidian Buttes fault. There are three distinct zones of seismicity (region 3 in Figure 4.1, Figure 4.3c); however, we only focus on the zone where the 2012 Brawley seismic swarm was located due to the lack of baseline measurements across the southern (between 1225 and KALI) and northern (between VP04, VP05, and VP03) zones. Baselines are computed between stations KALI, I032, P499, and MRM1 on the southern end of the Central Brawley Seismic Zone and stations ACU2, P495, and VP04 on the northern end of the Central Brawley Seismic Zone (Table 4.1). We find the baseline length changes between KALI and the three northern stations (ACU2, P495, and VP04) are drastically different from the baseline changes between the northern stations and the other three southern stations, indicating that the majority of deformation is localized to the area just south of VP04. We also see a drop off in the fault perpendicular motion for the baselines between VP04/P499 and VP04/MRM1, signifying that those baselines pass through the eastern extent of deformation. Excluding KALI and the two eastern baselines, we obtain a weighted mean left-lateral motion of 2.69 ± 1.46 mm/yr and extension of 4.04 ± 1.43 mm/yr.

Mesquite Basin

The Mesquite Basin is bounded by the Imperial fault to the west and the Brawley fault to the east. The interactions between these two faults cause a natural extensional basin due to the relative right-lateral motions. We have three baselines between 1225 and A35, B35 and D34 (Table 4.1), traversing the entire north-south extent of the basin (region 4 in Figure 4.1, Figure 4.3d). We find left-lateral strike-slip motion of 0.61 ± 0.64 mm/yr and extension of 3.69 ± 0.62 mm/yr. The relative absence of strike-slip faulting is seen in moment tensor solutions for larger events in the Mesquite Basin [Chen and Shearer, 2011]. The dominance and magnitude of extension in the Mesquite Basin can be explained simply by the relative motion of the Brawley fault to the Imperial fault. Lyons *et al.* [2002] estimated that fault creep on the southern Imperial fault is ~ 9 mm/yr but drops to ~ 6 mm/yr in the vicinity of the Brawley fault. The remaining fault creep is accommodated by the Brawley fault, which is roughly the amount of extension we see in the Mesquite Basin. When we look at baselines that traverse the Brawley Fault directly (A35, C35 and D34 to A31, B31, D30 and D29), we see 1.92 ± 0.69 mm/yr of right-lateral fault creep, directly in line with estimates from Lyons *et al.* [2002]. Moreover, baselines that traverse the Imperial fault (A35, C35 and D34 to P498, P744 and IVCO) indicate 6.34 ± 0.40 mm/yr of right-lateral fault creep, further verifying the Lyons *et al.* [2002] results, and indicating that the measured interseismic deformation is steady state. It should be noted that the baselines traversing the Brawley and Imperial faults indicate no fault-perpendicular motion. We estimate that the extension within the Mesquite Basin corresponds to 13.8 ± 3.0 mm/yr of subsidence (using a 75° dip angle estimated from Chen and Shearer [2011]), which is roughly in line with estimates of vertical creep since the 1970s along the Brawley fault from fault trenching by Meltzner *et al.* [2006].

4.5.2 Strain Rate Results

Maximum shear strain, dilatation, and rotation rate changes are shown in Figure 4.4 using the total velocity field and the local velocity field (difference of total velocity field and block model of Smith-Konter and Sandwell [2009]). The most prominent

feature in the strain rate fields is the large shear strain ($0.9 \mu\text{strain/yr}$) in the total field associated with the strain localization on the Imperial fault and a large dilatation (0.2 to $0.5 \mu\text{strain/yr}$) in the Central Brawley Seismic Zone. The strain becomes diffuse as it is partitioned between the San Andreas and San Jacinto faults, down to between 0.25 and $0.5 \mu\text{strain/yr}$. This is roughly the amount of total engineering strain found through triliteration by *Anderson et al.* [2003]. Also of note is that *Anderson et al.* [2003] reported a dilatation of $0.18 \mu\text{strain/yr}$ near Obsidian Buttes, well in the range of values obtained in our study. Strong negative dilatational lobes are present near the Chocolate Mountains to the east of the Salton Sea and the Fish Creek Mountains to the west of the Salton Sea, encompassing the Superstition Hills fault in both the total and local velocity fields, indicative of compression. In the Central Brawley Seismic Zone, 72% of the shear strain rate is explained by the block model, the lowest rate in the region, whereas on the outskirts of the region, more than 90% of the shear strain rate is explained by the block model.

Using the total velocity field, rotation rate throughout the Salton Trough varies rapidly through the Brawley Seismic Zone from $\sim 10\text{-}20^\circ/\text{Ma}$ of clockwise rotation west of the Brawley Seismic Zone to $\sim 10^\circ/\text{Ma}$ of counterclockwise rotation to the east. This is clearly demonstrated when looking at the total velocity field in Figure 4.2. This transition is in general agreement with the observed right-lateral motion across the San Andreas fault and other faults striking northwest. Looking just at the local velocity field, clockwise rotation of $5^\circ/\text{Ma}$ is centered on the Mesquite Basin and the Central Brawley Seismic Zone. This signal is qualitatively confirmed when looking at the ratio of extension to shear in the region. The highest ratio of extension to shear is in the southern Brawley Seismic Zone and tapers off to more shear than extension in the northern Brawley Seismic Zone. The high dilatation rate from the local field also hints at this higher extension to shear ratio in the Central Brawley Seismic Zone. Finally, we want to emphasize that all of the strain rate features persist using both the total and the local velocity fields, indicating that these signals are not due to an anomaly from deviations in the assumed fault positions from the true positions in the block model.

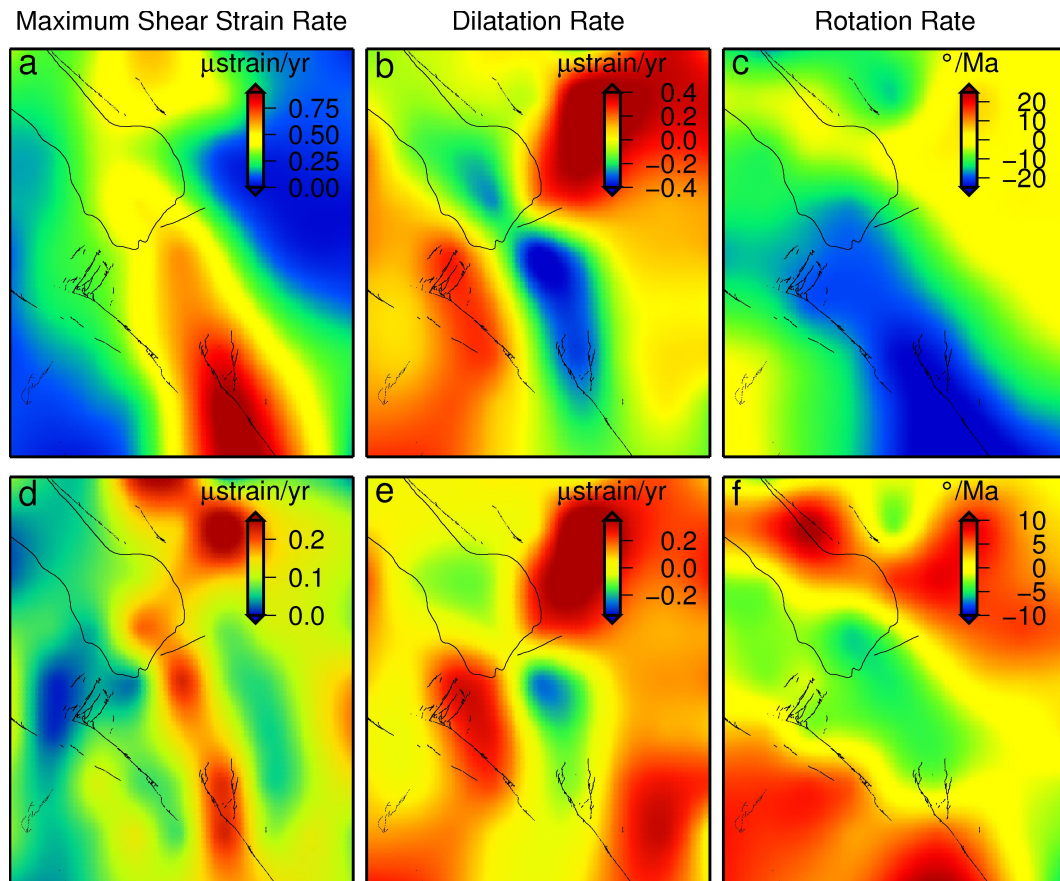


Figure 4.4: The maximum shear strain (**left**), dilatation (**center**), and rotation (**right**) rates for the total velocity field (**top**; **a**, **b**, and **c** respectively) and the local velocity field (**bottom**; **d**, **e**, and **f** respectively).

4.5.3 Vertical Deformation Results

From the baseline measurements, we obtained 4.72 ± 0.92 , 2.62 ± 0.91 , 4.04 ± 1.43 and 3.69 ± 0.62 mm/yr of extension in Regions 1-4 respectively. Figure 4.5 shows the vertical interseismic velocity field for the continuous GPS stations in the Salton Trough. We ignore the vertical signals for the campaign measurements since they are much noisier than those for the continuous stations. A first order observation is that the majority of vertical signals are localized within the Brawley Seismic Zone, and are indicative of subsidence. The maximum subsidence signal is at site P507 (-13.68 ± 0.32 mm/yr), which is located on a graben between the Obsidian Buttes fault and faults in the southern Salton Sea. Subsidence at P507 is likely due to a combination of normal faulting to the north and south of the station and possibly water extraction at the nearby geothermal plants [Chen and Shearer, 2011]. Furthermore, using the permanent scatterers InSAR technique yields a large spatially coherent subsidence signal along the Obsidian Buttes fault between 10 and 20 mm/yr [Eneva *et al.*, 2009]. Stations P506, P499, and P502 are outside of the zone of seismicity through the Central Brawley Seismic Zone and their vertical velocities are -2.12 ± 0.31 , -2.34 ± 0.35 , and -1.02 ± 0.32 mm/yr respectively. Station P495 is between the Central Brawley Seismic Zone and the Obsidian Buttes fault, which explains the slightly higher vertical velocity of -4.00 ± 0.33 mm/yr. Due to this difference, we can attribute ~ 2 mm/yr of subsidence to the faults in the Central Brawley Seismic Zone, an understandable conclusion given the diffuse nature of seismicity throughout this zone. This is in general agreement with InSAR line-of-sight velocities in the Central Brawley Seismic Zone (~ 5 mm/yr, Eneva and Shanker [2007]). While the extension through the Central Brawley Seismic Zone is about 4 mm/yr, it is probably distributed across many faults that each accommodate only a fraction of that amount, lending to smaller localized subsidence as in the case of P495.

The vertical deformation along the Obsidian Buttes fault is sharp as indicated by the large subsidence at P507. If we assume the zone-wide subsidence is ~ 2 mm/yr (roughly the amount at P506, P499 and P502), then P507 is undergoing ~ 11 -12 mm/yr

of subsidence due to faulting on the Obsidian Buttes fault and faults in the southern Salton Sea. From the horizontal baseline changes, we estimated 8.1 ± 3.0 mm/yr of subsidence along the Obsidian Buttes fault, which leaves between 3 and 4 mm/yr of subsidence accommodated along faults in the Salton Sea. However, from the baseline changes, we show that most of the deformation in the southern Salton Sea is occurring north of I025. Since deformation attenuates at approximately $1/r^2$, where r is distance to the slipping fault, one would expect a factor of 2 or 3 more vertical deformation localized under the sea, consistent with findings from other studies [Brothers *et al.*, 2009]. The lack of vertical deformation at DHLG indicates that the effects of faulting cease to exist well before the start of the San Andreas fault, and deformation is much more strongly localized than previously estimated.

4.6 Physical Model of Rotating Brawley Seismic Zone

Maintaining the counterclockwise rotation to the east of the Brawley Seismic Zone while also accommodating left-lateral slip along conjugate faults is counterintuitive, but can be explained in terms of internal deformation of discrete blocks in combination with oblique hinged extension and volcanism. The western side of the Brawley Seismic Zone can be envisioned to be "locked" due to the interactions with the Superstition Hills fault, but the eastern side has no such locking mechanism except diffuse or extinct faults. The clockwise rotation of the western Brawley Seismic Zone provides the locking mechanism and explains the existence of the Elmore Ranch fault zone striking roughly perpendicular to the Superstition Hills fault. Figure 4.6 shows the beginning and final states of this model without internal deformation for clarity. Of note from Figure 4.6 is that a simple rotation of two blocks, one clockwise (eastern Brawley Seismic Zone) and one counterclockwise (western Brawley Seismic Zone), can qualitatively explain the existence of compression in the Superstition Hills, motion along the Elmore Ranch fault, strong dilatation from the Mesquite Basin through the Central Brawley Seismic Zone, large shear strain through the entire Brawley Seismic Zone, a propagat-

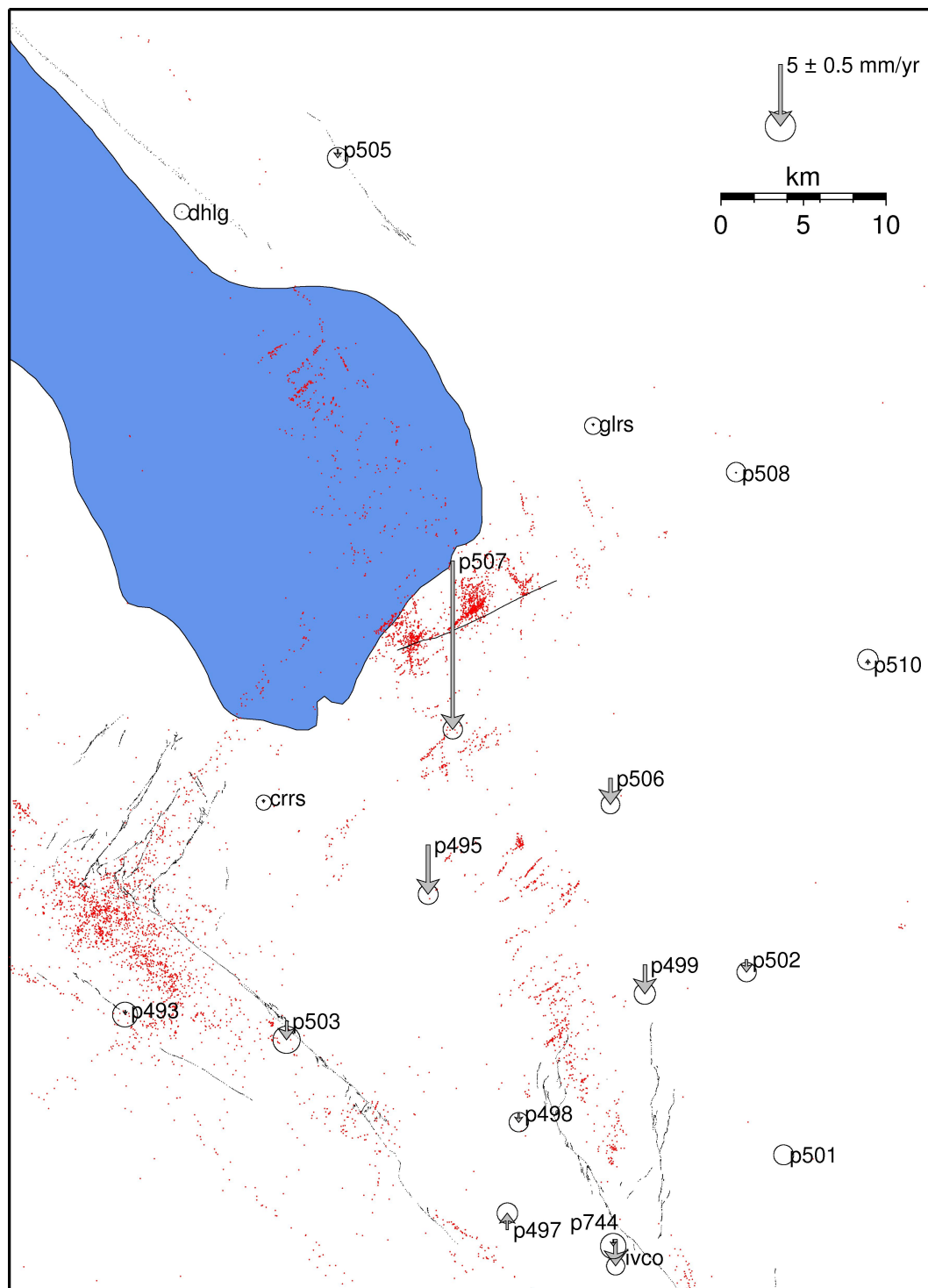


Figure 4.5: Interseismic vertical velocities for the continuous GPS stations in the Salton Trough. Error ellipses are plotted at the 95% confidence level.

ing Brawley fault, and the termination of the southernmost strands of the San Andreas fault by shifting the fault zone westward. The schematic deformation pattern shown in Figure 4.6 is of course a simplification, and the upper crust in the Imperial Valley might consist of many smaller blocks that would accommodate the relative motion through left-lateral faulting and extension. However, even this simple schematic can explain the general features of deformation in the region. The observed rotation of the Brawley Seismic Zone disappears abruptly approaching the San Andreas fault, indicating that the San Andreas and San Jacinto faults are not creating differential extension across the Salton Sea as evidenced by similar geologic and geodetic slip rates along those two faults [Behr *et al.*, 2010; Janecke *et al.*, 2010; Lindsey and Fialko, 2013; Meade and Hager, 2005; Rockwell *et al.*, 1990]. During counterclockwise rotation of the eastern part of the Brawley Seismic Zone, discrete faults may open and accommodate the rotation through extension and left-lateral shear. This model is supported by evidence for northwesterly migration of the Brawley Seismic Zone from basement morphology [Larsen and Reilinger, 1991]. In short, rotation of the Brawley Seismic Zone would need to be accommodated along further northwestward parallel faults to the Brawley fault over time given topographic constraints (i.e. no along fault compression along the Imperial fault). Coincidentally, the existence of diffuse or extinct lineaments southeast of the San Andreas fault near Bombay Beach [Lynch and Hudnut, 2008] further supports this model for northwestward propagation of the Brawley Seismic Zone. Finally, Fuis *et al.* [1984] shows from seismic reflection data cutting across the central part of the Brawley Seismic Zone that the basement rocks become shallower towards the east, which would be a direct result of hinged extension across the area (i.e. more extension to the east than to the west).

4.7 Conclusions

Using a combination of campaign and continuous GPS measurements and a block model for regional tectonics constrained by the continuous GPS data [Smith-

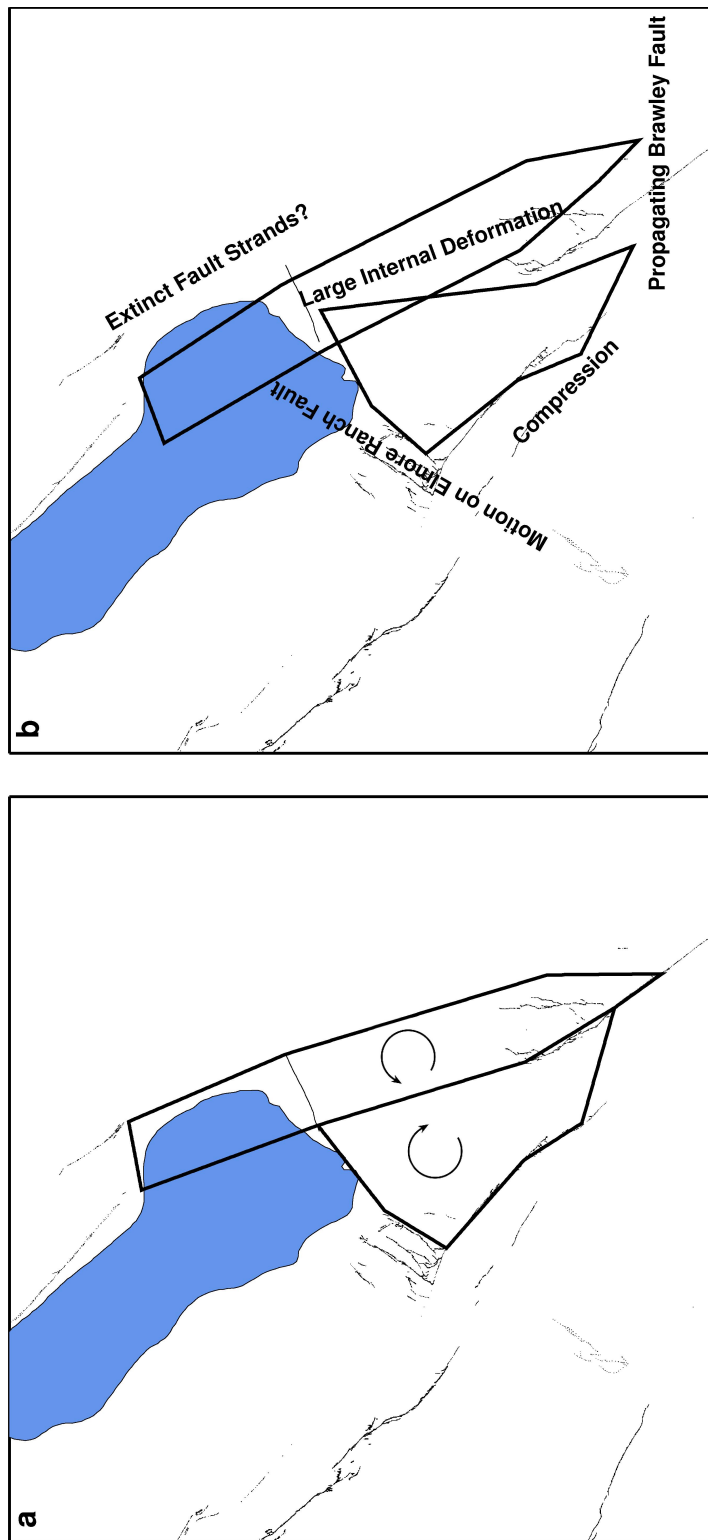


Figure 4.6: Schematic of the rotating extensional model in the Brawley Seismic Zone. The blocks in (b) are simply rotated from (a) and include no internal deformation. Important features of the model are shown on (b).

Konter and Sandwell, 2009], we investigated deformation through the Brawley Seismic Zone from the Imperial fault in the south to the San Andreas fault in the north. Differential strain is localized completely within a 10 km wide band that accommodates 11 mm/yr of left-lateral shear and 15 mm/yr of extension. Rotating the shear and extension measurements into the strike of the San Andreas fault yields 17 mm/yr of dextral shear, roughly equivalent to geologic and geodetic slip rates near the southern San Andreas fault. Counterclockwise rotation in the Imperial Valley can be explained through hinged oblique extension, and may be evidence for northwestward propagation of the Brawley Seismic Zone and the San Andreas fault system. Finally, deformation south of the San Andreas fault appears to be localized, leading to greater hazard and the possibility of sizeable earthquakes that might act as triggers for great earthquakes on the southern San Andreas fault.

4.8 Acknowledgements

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Table 4.1: Baseline velocity changes throughout the Brawley Seismic Zone

Site 1	Site 2	SS (mm)	±	Ext. (mm)	±
Southern Salton Sea (N50°E)					
G003	P507	4.13	0.71	5.91	0.70
D11G	P507	4.78	1.38	5.69	1.45
DHLG	P507	5.15	0.20	3.76	0.19
BERT	P507	4.95	0.71	1.60	0.76
G001	P507	3.18	0.71	2.59	0.70
POPE	P507	5.59	1.64	2.04	1.71
FRIN	P507	4.75	0.96	2.60	0.99
I025	P507	-0.53	2.16	2.39	2.20
Obsidian Buttes Fault (N65°E)					
VP05	P507	6.67	0.88	0.15	1.01
Q122	P507	4.71	0.44	1.75	0.43
P506	P507	3.70	0.21	3.74	0.20
I029	P507	3.13	1.10	0.51	1.24
VP06	P507	5.65	1.37	3.17	1.66
Central Brawley Seismic Zone (N40°E)					
KALI	ACU2	0.98	1.90	-0.51	1.90
I032	ACU2	3.26	2.03	3.80	2.00
P499	ACU2	2.24	1.06	2.15	1.05
MRM1	ACU2	2.76	2.40	3.23	2.32
KALI	P495	0.65	1.05	-2.69	1.05
I032	P495	3.58	1.18	5.98	1.15
P499	P495	2.57	0.21	4.33	0.21
MRM1	P495	3.09	1.55	5.41	1.47
KALI	VP04	-2.07	1.68	-2.29	1.68
I032	VP04	2.16	1.81	1.00	1.78
P499	VP04	-1.14	0.84	0.65	0.84
MRM1	VP04	1.66	2.18	0.43	2.10
Mesquite Basin (N50°E)					
A35	1225	0.41	0.52	3.31	0.53
C35	1225	0.38	0.91	3.81	0.85
D34	1225	0.92	0.49	3.96	0.49
Northern Imperial Fault (N35°W)					
A35	P498	6.10	0.30	3.31	0.30
C35	P498	6.60	0.60	3.81	0.71
D34	P498	6.80	0.26	3.96	0.27
A35	P744	6.72	0.34	3.31	0.33
C35	P744	7.22	0.64	3.81	0.75
D34	P744	7.42	0.29	3.96	0.31
A35	IVCO	5.08	0.30	3.31	0.31
C35	IVCO	5.58	0.60	3.81	0.72
D34	IVCO	5.78	0.26	3.96	0.28
Brawley Fault (N2°W)					
A35	A31	4.83	1.33	0.08	0.99
C35	A31	4.46	1.52	0.42	1.51
D34	A31	4.00	1.27	0.08	0.98
A35	B31	2.05	0.44	-0.34	0.44
C35	B31	1.68	0.63	0.00	0.96
D34	B31	1.22	0.38	-0.34	0.43
A35	D30	2.48	0.42	0.06	0.40
C35	D30	2.11	0.61	0.39	0.92
D34	D30	1.65	0.36	0.06	0.39
A35	D29	1.72	0.39	0.21	0.45
C35	D29	1.35	0.58	0.55	0.97
D34	D29	0.89	0.33	0.21	0.44

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