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Constraining Interseismic Strain Release at the Cascadia Subduction Zone with Geodetic Data

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A dissertation

submitted in partial fulfillment of the
requirements for the degree of

Doctor of Philosophy

University of Washington

2022

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Program Authorized to Offer Degree:

Earth and Space Sciences

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Abstract

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In this dissertation, I study the style and distribution of interseismic strain release along a major convergent plate boundary to better understand how faults accommodate the relative motion between tectonic plates. Focusing on the Cascadia Subduction Zone, I utilize geodetic data, specifically in the form of Global Navigation Satellite System (GNSS) data, to measure crustal deformation and relate this deformation to slip processes along the megathrust and crustal faults. In particular, I focus on characterizing the spatiotemporal pattern of slow slip events (SSEs) across the margin and examine how conditions along the plate interface may influence the distribution of these events. I also analyze how strain is accommodated by crustal forearc faults in southern Cascadia near the Mendocino triple junction (MTJ) and consider how slip on these faults reduces the buildup of strain along the megathrust fault. The release of interseismic

strain via SSEs and slip on crustal faults impacts the distribution of accumulated strain within the ‘locked zone’, a portion of the megathrust fault that is capable of producing large catastrophic earthquakes. Therefore, by characterizing SSE behavior and slip rates on crustal faults, this work provides important insights on the state of strain along the megathrust and how it relates to seismic hazards in Cascadia. Furthermore, this work serves as a case study for strain release in subduction zone systems and broadens our understanding of seismic and aseismic deformational processes within these environments. Chapter 1 provides an introduction to the themes and concepts that are discussed in subsequent chapters.

In Chapter 2, I focus on a mode of strain release associated with long-duration SSEs. I conduct a systematic analysis of 13 years of GNSS time series data from 2006 to 2019 and present evidence of at least one low-amplitude long-term SSE on the Cascadia subduction zone, with the possibility of others that are less resolved. This 1.5-year transient is observed in southern Cascadia, and the data are modeled as a Mw 6.4 slow slip event occurring at 15–35 km depth on the plate interface, just updip of previously recognized short-term slow slip and tremor. The event shares many characteristics with similar long-term transient events on the Nankai subduction zone. However, the maximum horizontal surface displacements and total fault slip amplitudes for this event are an order of magnitude smaller than most other long-term SSEs at other subduction zones. Therefore, I propose that the frequency and size of long-term SSEs in Cascadia is limited by the width of the semi-frictional zone along the plate interface.

In Chapter 3, I examine how plate boundary strain is partitioned across shallow crustal faults in the forearc of the Cascadia subduction zone. I construct elastic block models of the MTJ region and leverage GNSS velocity data to constrain deformation on the Little Salmon fault, Mad River fault zone (MRFZ) and Grogan fault, which are all identified as quaternary-active

crustal structures. I evaluate models with various forearc fault structures and apply a bootstrap analysis to provide histograms of the long-term slip rate for each fault. Model results indicate that the Little Salmon fault zone is an important structure that accommodates both thrust and right-lateral shear motion in the forearc. Appreciable right-lateral slip rates on the MRFZ indicate that this system plays an important role in facilitating translation of the forearc and may serve as an extension of the northern San Andreas fault system. In contrast, the models place very little strike-slip motion on the Grogan fault and prefer this structure to mainly host reverse slip. The highest slip rates are observed on fault structures immediately to the north of the MTJ, indicating significant strain on the Bear River fault zone or nearby fault strands. Overall, these results help to constrain the seismic hazards associated with crustal faults in this region.

In Chapter 4, I characterize strain release along the plate boundary in a special setting where slow slip and tremor are observed simultaneously during episodic tremor and slip (ETS) events. I use tremor and GNSS time series data to identify nineteen of the largest ETS events in southern Cascadia between 2016.5-2022 and document source properties. Distributed slip models for these events show that cumulative fault slip along the megathrust reaches a maximum near 40.5° N latitude and that large ETS events accommodate up to 80% of plate convergence at this location on the plate interface. However, ETS fault slip and tremor terminate farther to the south near ~40° N latitude, some 50 km before the southern lateral edge of the subducting Gorda plate. I explore possible controls on the distribution of strain release from ETS in southern Cascadia, including changes in the slab geometry and thermal gradient near the southern edge of the subduction zone. After exploring possible controls on the distribution of ETS, I propose that the heating of the downgoing slab edge and complex slab geometry inhibit ETS behavior in southernmost Cascadia. In particular, I demonstrate that seismic deformation in the form of

tremor does not take place along the plate interface in the southernmost 50 km of Cascadia below 35 km depth, which is distinct from the rest of the subduction zone.

In Chapter 5, I summarize the primary results and highlight the significance of my work. I also identify what future study is needed to resolve any lingering scientific questions and test outstanding hypotheses.

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ACKNOWLEDGEMENTS

I would like to thank my friends and family for supporting me throughout my graduate career. I am grateful to my committee for all of their guidance and feedback on this dissertation. Thank you to the ESS Department, USGS, and NSF for providing the financial support to conduct this research; specific grants are noted within each chapter.

For Toula

Who taught me to never stop chasing

Chapter 1. INTRODUCTION

The Cascadia subduction zone constitutes a thousand-kilometer-long fault system that extends from the Queen Charlotte fault in northern Vancouver Island, British Columbia to the San Andreas fault in Cape Mendocino, California. In this region, the Juan de Fuca plate system subducts beneath the North American plate at 3-4 cm/year with a subduction angle that increases in obliquity from north to south (DeMets et al., 2010). Coupling between these two plates along the megathrust fault results in the accumulation of elastic strain that is oriented roughly parallel to the subduction direction of the Juan de Fuca plate (Savage, 1983). This elastic strain is eventually released in large megathrust earthquakes, which represents a major hazard to the Pacific Northwest (Wang and Tréhu, 2016; Walton et al., 2022). The deformation signal from the elastic strain build-up is observed along the entire Cascadia margin and decays in magnitude away from the plate boundary (e.g. Li et al., 2018).

Long-term permanent deformation and smaller scale crustal faulting in the North American plate also accommodates plate convergence (e.g. McCrory et al. 2000, Padgett et al., 2019). In addition to the strains formed by subduction, permanent crustal strain is also driven by large-scale crustal block motions. Specifically, the current northwest migration of the Sierra Nevada Block (SNB) at ~11 mm/yr and the extension of the Basin and Range province is forcing the Oregon Coast Block (OCB) to translate north and rotate counter clockwise (Wells et al., 1998). This process helps drive the overall clockwise rotation of the forearc, which is captured by a regional network of GNSS stations (e.g. McCaffrey et al. 2013).

The elastic strain stored around the Cascadia megathrust is released by a spectrum of slip modes along the plate boundary interface (Peng & Gomberg, 2010). At shallow depths (top tens of kms), strain is released during earthquakes via seismic slip with slip rates on the order of kilometers per second. This zone is commonly referred to as the seismogenic zone or locked zone. At deeper depths (several tens of kms and greater), relative motion between the two plates is facilitated by continuous aseismic creep with slip rates on the order of centimeters per year. In between the locked and creeping portions of the fault, strain is released along the plate interface by slow slip events (SSEs). These events occur periodically and exhibit slip rates on the order of centimeters per day. SSEs have been observed on various subduction zones around the world

including southwest Japan (Hirose et al., 1999), New Zealand (Wallace & Beaven, 2010), Alaska (Ohta et al., 2006), Costa Rica (Outerbridge et al., 2010), Mexico (Radiguet et al., 2012), and Chile (Kato & Nakagawa, 2014), indicating that they are an inherent part of subduction zone systems. Additionally, large SSEs are capable of releasing the same amount of energy as Mw 6 earthquakes or larger, and their location relative to the locked zone suggests that they could have a significant impact on the timing of large megathrust earthquakes (Shwartz & Rokosky, 2007; Segall & Bradley, 2012).

The continuum of slip modes observed along the plate interface, which varies systematically as a function of depth, is thought to be controlled by various fault properties that impact the stability of slip along the megathrust (Peacock, 2009). Laboratory experiments suggest that slow slip behavior emerges along the plate interface near the threshold between stable and unstable failure, which is controlled by frictional dynamics along the fault (Leeman et al., 2016). Numerical models relate the spectrum of deformation along the megathrust to changes in fault zone rheology, which controls whether the dominant mode of shear along the fault is frictional, viscous, or somewhere in-between (Gao & Wang, 2017). These models predict that short-term SSEs arise in a zone that is frictionally unstable due to high pore-fluid pressures (Gao & Wang, 2017). The presence of high pore-fluid pressures along the plate interface is supported by geophysical imaging of a low-velocity zone (LVZ) near the top of numerous megathrust faults (e.g. Kim et al., 2014; Audet & Shcaefffer, 2018; Akuhara et al., 2020). Near-lithostatic pore pressures along the megathrust likely act to lower the effective normal stress along the fault, making it more susceptible to failure from very small changes in stress (Hawthorne & Rubin, 2010; Kobayashi & Sato, 2021). Therefore, high pore-fluid pressures are often invoked as a prerequisite for possible ETS mechanisms (Audet & Kim, 2016).

In Cascadia, SSEs are observed throughout the margin at 30-50 km depth along the plate interface (Gomberg et al., 2010). These events last on the order of weeks, and have recurrence intervals that vary along strike, with northern, central, and southern Cascadia having recurrence intervals of 14 ± 2 months, 19 ± 4 months, and 10 ± 2 months, respectively (Brudzinski & Allen, 2007). Large SSEs in Cascadia are recorded in GNSS time series data as trenchward displacements that are opposite to regular inter-SSE deformation patterns. Slow slip in Cascadia is also associated spatially and temporally with low frequency, low amplitude seismic tremor (Rogers & Dragert, 2003). Therefore, tremor can serve as a proxy for the timing and spatial

extent of slow slip, and can be used to identify small SSEs that are not resolvable with GNSS data. Together, slow slip and tremor are commonly referred to as episodic tremor and slip (ETS).

The southern margin of the Cascadia subduction zone experiences especially complex deformation due to its proximity to the Mendocino Triple Junction (MTJ). This junction represents the intersection of the North American, Pacific, and Juan de Fuca (or Gorda) plates and signifies the transition from convergent tectonics associated with Cascadia subduction to strike-slip tectonics associated with the San Andreas fault (SAF) system. The MTJ is currently migrating north relative to North American, facilitating the northward propagation of the SAF system (Kelsey & Carver, 1988) and leaving behind a slab window in its wake (Dickinson & Snyder, 1979a, 1979b). Growth of the SAF system is likely accommodated by discrete faults, whereas long wavelength deformation signals may be imposed by the passage of this slab window. Furlong & Govers (1999) propose the “crustal conveyor belt” model, whereby viscous coupling between the Gorda plate and North American lithospheres causes patterns of migrating uplift and subsidence following the course of the MTJ. In addition to all of the processes listed above, southern Cascadia is also the most seismically active region in the subduction zone. All of the processes outlined above are superimposed in the southern Cascadia GNSS time series data with varying timescales and amplitudes, making the interpretation of the surface deformation challenging.

Overall, the Cascadia subduction zone represents a dynamic system wherein interseismic strain is released by a spectrum of deformational modes on various spatiotemporal scales. Understanding how and where strain is released within the Cascadia margin is important for understanding seismic hazards associated with subduction. Furthermore, identifying the dominant mode of deformation can shed light on the deep plate interface environment, which cannot be studied through direct observations. Therefore, the goal of this work is to utilize GNSS and seismic data to better understand how strain is accommodated within the Cascadia subduction zone, both along the megathrust and on crustal faults in the forearc. In particular, this dissertation seeks to answer three questions regarding strain release along the margin: Does long-term slow slip occur along the Cascadia margin? How is interseismic strain partitioned along crustal forearc faults in southern Cascadia? Does ETS behavior change at the southern edge of the subduction zone where the thermal and structural properties are complex?

In Chapter 2, I analyze GNSS time series data to determine if long-term slow slip has taken place in Cascadia from 2006-2019. The lack of long-term slow slip in Cascadia has been enigmatic, as this mode of slow slip has been observed on other comparable subduction systems but has not yet been documented in Cascadia. By thoroughly cleaning the data, I show that no large-amplitude signal associated with long-term slow slip is found in Cascadia during the observational period. However, I find evidence for a low-amplitude long-term slow slip event in southern Cascadia from mid-2012 to 2014. The fault slip amplitudes and horizontal surface displacements for this event are an order of magnitude smaller than other long-term events observed at other subduction zones. Therefore, long-term slow slip events in Cascadia are relatively small and infrequent. I suggest that the size of long-term slow slip events in Cascadia are limited by the relatively narrow down-dip width of the semi-frictional zone along the megathrust fault. The contents of this chapter are published in *Geochemistry, Geophysics, Geosystems* under the title “Filling the Gap in Cascadia: The Emergence of Low-Amplitude Long-Term Slow Slip” (Nuyen & Schmidt, 2021).

In Chapter 3, I examine the complex surface deformation that surrounds the MTJ, which represents the nexus of three plate boundaries. I create elastic block models and regional strain rate maps of the MTJ region constrained by GNSS velocity data to study how interseismic strain is partitioned in the southern Cascadia forearc. Modeling results, together with the strain rate maps, show that the crustal Little Salmon fault, Mad River fault zone, and Grogan fault actively accommodate contraction and right-lateral shear in the forearc. Additionally, I find that slip on these crustal faults lowers the slip deficit rate along portions of the megathrust fault. The contents of this chapter are currently in revision in *Journal of Geophysical Research: Solid Earth* under the title “Strain Partitioning among Forearc Faults in Southern Cascadia Inferred from GNSS”.

Finally, in Chapter 4, I use tremor and GNSS time series data to estimate the timing and horizontal surface displacements for nineteen of the largest ETS events in southern Cascadia between 2016.5-2022. By calculating distributed slip models for these events, I find that ETS fault slip and tremor terminate near $\sim 40^\circ$ N latitude and do not extend to the southern lateral edge of the subduction zone. I explore possible controls on the distribution of ETS in this region and conclude that the complex geometry and heating of the subducting slab inhibit ETS behavior near the edge of the subduction zone. Therefore, I hypothesize that seismic deformation (in the

form of tremor) does not take place along the plate interface in southernmost Cascadia below 35 km depth. This work is currently being prepared for submission to a peer-reviewed geophysics journal.

Chapter 2. FILLING THE GAP IN CASCADIA: THE EMERGENCE OF LOW-AMPLITUDE LONG-TERM SLOW SLIP

The content of this chapter is published in:

Nuyen, C.P. & Schmidt, D.A. (2021). Filling the gap in Cascadia: The emergence of low-amplitude long-term slow slip. *Geochemistry, Geophysics, Geosystems*, 22(3).
<https://doi.org/10.1029/2020GC009477>

2.1 INTRODUCTION

Slow slip falls within a spectrum of slip modes that serve to release and redistribute elastic strain along the subduction plate interface (Peng & Gomberg, 2010). Representing transient episodes of aseismic slip, slow slip events (SSEs) have durations that are longer than normal earthquakes (seconds-to-minutes) and shorter than continuous creep. Long-term SSEs are typically found immediately down-dip from the seismogenic zone and exhibit durations of months-to-years, while short-term SSEs occur over days-to-months, and are accompanied by non-volcanic tremor (Obara & Kato, 2016). The spectrum of fault behavior along the megathrust, which varies systematically as a function of depth, is thought to be controlled by fault properties that affect the stability of fault slip (Peacock, 2009; Leeman et al., 2016). While SSEs serve to accommodate relative plate motion, they also transfer stress updip, incrementally loading the seismogenic zone and bringing it closer to failure (Segall & Bradley, 2012; Uchida et al., 2016). Therefore, characterizing SSEs at subduction zones is crucial for our understanding of subduction zone slip budgets and megathrust earthquake nucleation.

Long-term SSEs have been observed at numerous subduction zones around the world, including Japan (e.g. Ozawa et al., 2001; Ozawa et al., 2007; Ozawa et al., 2013; Ozawa et al., 2016; Takagi et al., 2016), Mexico (e.g. Lowry et al., 2001; Larson et al., 2007; Radiguet et al., 2012), New Zealand (e.g. Wallace & Beavan, 2010; Wallace et al., 2012), Alaska (e.g. Ohta et al., 2006; Wei et al., 2012; Fu & Freymueller, 2013; Li et al., 2016), and Costa Rica (Jiang et al., 2012). The Nankai subduction zone in southwest Japan has a particularly rich history of long-term SSEs, which fill the spatial gap between short-term SSEs and the locked zone throughout most portions of the margin. These long-term events last on the order of months-to-years, occur

at depths of 20-40 km, and have equivalent moment magnitudes of M_w 6-7.2 (Ozawa, 2017; Takagi et al., 2019). However, the spatial relationship that exists between long- and short-term SSEs at Nankai is not found at all subduction zones. For example, along the Hikurangi subduction zone, short-term SSEs occur at shallower depths (5-20 km) than long-term events (25-75 km) (Wallace & Beavan, 2010). Furthermore, distinct short-term SSEs have not been observed updip or downdip of long-term SSEs in south central Alaska; however, it has been proposed that these long-term events are partially or completely composed of short-term SSEs (Rousset et al., 2019).

Nankai is often considered directly comparable to Cascadia, as they both represent relatively young, warm subduction zones (Oleskevich et al., 1999) and exhibit similar short-term slow slip behavior (Obara, 2011). However, the similarities between these two subduction zones begin to break down when long-term slow slip is considered. In the Cascadia subduction zone (Figure 2.1), short-term SSEs have been observed along the entire margin at depths of 25-45 km (Brudzinski & Allen, 2007), while various studies have placed the locked zone predominantly offshore at depths shallower than 20 km (Schmalzle et al., 2014; Li et al., 2018). Despite stress transfer (Wech & Creager, 2011) and numerical models predicting long-term SSEs at depths between the locked zone and short-term slow slip (Matsuzawa et al., 2013), such events have not yet been observed in Cascadia. Thus, an anomalous gap in seismic and aseismic activity exists in Cascadia between the locked zone and the short-term SSE region (Hyndman, 2013). The apparent absence of long-term slow slip in Cascadia is surprising given that this behavior has been observed downdip of the seismogenic zone in numerous other subduction zones, particularly Nankai.

Long-term SSEs are typically represented by large (>1 cm) transients in surface deformation time series data that last many months and whose direction of motion is opposite the local convergence direction. A first-order examination of GNSS time series in Cascadia does not reveal any obvious transient signals of this amplitude, duration, and direction, suggesting that large long-term SSEs with recurrence intervals within the record of contemporary GNSS time series are non-existent. However, this simple inspection does not eliminate the possibility that low-amplitude long-term SSEs exist within the data at or near the level of noise. Therefore, we conduct a margin-wide analysis of Cascadia GNSS time series data to investigate whether or not low amplitude long-term transients exist. In this work, we show that by carefully removing the

signal imposed by noise, secular tectonics, short-term SSEs and non-tectonic processes, we are able to identify low-amplitude, long-term transient signals that are consistent with a long-term SSE.

2.2 METHODS AND DATA

We examine thirteen years of continuous GNSS time series data from 2006 to the beginning of 2019 for over 150 stations throughout Cascadia (Figure 2.1). The primary data consist of the horizontal east and north component GNSS time series. In order to isolate long-period transient signals, we reduced the data with five primary steps illustrated in Figures 2.2 and 2.3 for station P332:

- Step 1)** Outliers (i.e. positions that strayed >50 mm from adjacent epochs) and steps due to known earthquakes and equipment changes were removed (Figure 2.2a). The timing of steps was determined by the GAGE offsets from the Central Washington University (found at <ftp://data-out.unavco.org/pub/products/offset/>). Step offsets were estimated by differencing the average positions 21 days before the event and 21 days after the event.
- Step 2)** A 30-day moving average was applied to filter out unwanted high-frequency noise (Figure 2.2a).
- Step 3)** Each time series was decomposed into seasonal (S), trend (T), and remainder (R) components using a seasonal trend decomposition method based on loess (STL) (Cleveland et al., 1990) (Figures 2.2a and 2.3). STL consists of an algorithm that employs smoothing and locally weighted linear least squares regression. The parameters for this method were set so that the S component estimates signals with annual periodicities by averaging the data on common days of the year through a weighted algorithm. Short-term SSEs are successfully isolated in the R component, meaning that short-term SSEs have little to no effect on the other STL components (Figures 2.3c and A1.1). Since we are interested in finding

long-term changes in the GNSS time series trends, the remainder of our cleaning process only deals with the T component. More details about the STL parameters used in this procedure can be found in the Supporting Information.

Step 4) The T component was detrended to remove the secular signal imposed by deformation due to interseismic strain accumulation and crustal block rotation (Wells et al., 1998) (Figure 2.2b). We individually detrended each time series by removing the best-fit line for the entire 2006.0-2019.0 time period.

Step 5) Common mode signals were removed to eliminate spatially coherent noise amongst stations that span the geodetic network (Figure 2.2b). We assume that GNSS stations in eastern California, Oregon, and Washington (i.e. east of the Cascadia volcanic arc) should not be sensitive to deformation on the plate interface. Therefore, we binned the eastern stations by latitude and stacked 30 time series to produce a common mode time series that varies as a function of latitude (Figure A1.2). We applied a yearly moving average to each common mode signal and detrended T component, in order to remove any leftover noise. The common mode signal was then removed from the GNSS time series depending on each site's latitude. All of the removed common mode signals had maximum amplitudes less than 1 mm (Figure A1.2). We also tested a 6-month moving average to ensure that we were not erasing 6-month-long transients. However, we did not find any significant difference between the two resulting products.

Every GNSS processing center employs a unique methodology for producing positional time series data. The Nevada Geodetic Laboratory (NGL) and the Pacific Northwest Geodetic Array (PANGA) at Central Washington University both utilize GIPSY/OASIS software, which uses a precise point positioning (PPP) approach, while the Scripps Orbit and Permanent Array Center (SOPAC) relies on a relative positioning methodology with GAMIT/GLOBK software.

Each software package employs different a priori assumptions, filtering methods, and other factors that can produce slight variations in positional time series data. Therefore, we applied our cleaning methodology to network solutions from NGL, PANGA and SOPAC to test whether residual patterns were consistent. For reducing the NGL and PANGA datasets, we used the five steps outlined above. For reducing the SOPAC dataset, we used data that already had outliers and steps due to earthquakes and equipment changes removed. Thus, we only used Steps 2-5 for the SOPAC time series data. The NGL and PANGA solutions were in the NA12 and NAM14 stable North American reference frames, respectively, while the SOPAC solutions were in the earth-centered, earth-fixed WGS84 reference frame.

2.2.1 Sensitivity Analysis

The seasonal trend decomposition (STL) algorithm requires several mutable variables that are determined by the user and does not have a predetermined error level. Therefore, we employ a suite of tests to evaluate our choice of variables, the overall sensitivity of this method to small transients, and the average error. To estimate the error associated with STL, we created 1,000 synthetic ~13.5 year GNSS time series with random slopes, seasonal signals, white noise, flicker noise and random-walk noise. The range of amplitudes used for each of these components is listed in Table 1, and the values for white noise, flicker noise, and random-walk noise were taken from Langbein (2008). An identical common mode signal with quarterly and biannual periodicities was also added to each time series with a maximum amplitude of 1 mm. We then ran these thousand randomly-generated time series through our 5-step cleaning process, removing the known common mode signal in Step 5 (Figure A1.4). The standard deviation of the final thousand time series was calculated for each day in the time series, and then averaged across the entire epoch. The average standard deviation, and the error margin that we reference in this work, is 2.6 mm.

To test the sensitivity of the STL method, which relies on our chosen variables, we created synthetic time series with various hypothetical transient signals and evaluated the ability of our algorithm to extract the input transient signal. Random slopes and noise were superimposed on these time series, and then ran through our 5-step cleaning process, as described above. Statistical analysis of the final time series reveals that on average, the STL method, and thus our choice of variables, is able to capture the SSEs hidden within the noise.

However, our methods tend to overestimate the amplitude of these events by $\sim 50\%$. Thorough visual inspection of the final time series shows that although smaller one-year transients (1 mm) are resolved by our methods on average, they often do not stand out relative to the noise. In comparison, larger one-year transients (2 mm) regularly stand above the noise and are more distinguishable. Therefore, we are most confident in transients that have amplitudes of ≥ 2 mm in the final time series. Nevertheless, it is still possible that < 2 mm coherent displacements between neighboring stations are indicative of transients that are below the level of noise. More information about the methods and details of our STL analysis can be found in the Supporting Information.

2.3 RESULTS

Our final time series product reveals how surface displacements have deviated from the long-term regional deformation pattern after interseismic strain, forearc rotation, short-term SSEs, seasonal signals and noise are removed. Here we present our results for the NGL dataset (Figure 2.4), while the results for the PANGA and SOPAC datasets can be found in Appendix 1 (Figures A1.6 and A1.7). We have separated the data into yearlong epochs that start at the beginning of the water year (October 1st) to minimize any seasonal biases due to water-loading (Fu et al., 2015). The residual surface displacements should appear spatially random if no coherent signal remains and the deformation is steady state. Therefore, we focus our attention on groups of neighboring stations with uniform patterns of westward surface displacement that are ≥ 2 mm and seen in all three datasets.

In the NGL dataset, we identified four instances wherein groups of stations exhibited coherent westward motions. Of these events, there are two that occur over consecutive epochs in northern Oregon (2007.75-2009.75) and southern Cascadia (2012.75-2014.75). We merged consecutive years to better estimate the total surface displacement for a long-duration event. Combining years 2007.75-2009.75 did not yield larger cumulative displacements and resulted in a more chaotic pattern of station motions. In contrast, combining years 2012.75-2014.75 did yield larger cumulative westward displacements in southern Cascadia and the coherency between stations was retained. By adjusting the limits of the epoch to 2012.45-2014, we achieved maximum displacements, with several stations rising above the STL error margin of 2.6 mm (Figure 2.5a). The last two events, occurring in southern Washington and northern California,

happened during the most recent 2017.75-2018.75 epoch. We found that the displacements for these events were maximized if we extended their epochs to 2017.6-2019 for southern Washington and 2017.2-2019 for northern California. However, neither event exhibits maximum displacements larger than 2 mm, making the 2012.45-2014 epoch the only period in which surface displacements exceed 2 mm.

Focusing in on the region north of the Mendocino Triple Junction (MTJ), where the largest displacements occurred during the 2012.45-2014 epoch, we observed similar longitudinal motions at GNSS stations near the coast (Figures 2.5b and 2.5c). Specifically, coastal stations exhibited disparate patterns of deformation until 2012.45, at which point they all began moving westward through 2014. After 2014, the majority of these stations rotated eastward until 2017.2 (Figure 2.5b). For the remainder of the time series, the northern coastal stations maintained a random state, whereas the southern stations resumed westward motions. This overall pattern of motion is not seen at the more inland stations, which displayed unsystematic displacements for the entirety of the 2006-2019 epoch (Figure 2.5d).

2.4 DISCUSSION

Our final GNSS time series results reveal four instances wherein coherent, transient westward motion is observed at stations along the coast of the Cascadia subduction zone. We are most confident in the transient signal observed in southern Cascadia from 2012.45-2014, as it is clearly seen in all three data sets (Figures 2.5, A1.6, A1.7) and has surface displacements ≥ 2 mm. In the following sections, we examine possible sources for this westward transient and consider its implications for the Cascadia subduction zone. We are less confident in the remaining three transients in southern Washington, northern Oregon, and northern California, as they have relatively small surface displacement values. However, we also explore these additional smaller signals for completeness.

2.4.1 Non-tectonic Sources

The coherent westward displacements we observed in our final GNSS time series data could be explained by a number of tectonic and non-tectonic causes. We first considered four potential non-tectonic sources: processing artifacts, regional changes in soil moisture, time-dependent variations in snowpack, and common mode signals. The small scale at which these

deformation signals are observed opens up the possibility that they are artifacts imposed by the original processing of the position time series. We applied our GNSS time series analysis to the independent GNSS solutions provided by the Nevada Geodetic Laboratory (NGL) (Blewitt et al., 2018), the Pacific Northwest Geodetic Array (PANGA) (PANGA, 1996), and the Scripps Orbital and Permanent Array Center (SOPAC) (Jamason et al., 2004). The final time series for these three sources reveal similar results on annual (Figures 2.4, A1.6, A1.7) and multi-year timescales (Figures 2.5b and A1.8), with all three solutions containing the transient patterns discussed in Section 2.3. Therefore, the consistency between the residual time series results from these three independent processing centers suggests that the signals in our residual time series data are not biased by the choice of network processing software.

Long-term changes in regional soil moisture content or snowpack, which provide a surface load, could also be responsible for the patterns we observed in our data. To test this possibility, we acquired monthly soil moisture and snow water equivalent (SWE) data for Cascadia from 2006-2019 from the land surface model for Phase 2 of the North American Land Data Assimilation System (NLDAS-2) (Mocko et al., 2012; Xia et al., 2012). For each dataset, we summed the monthly data for each year starting on October 1st. The yearly totals were then normalized by subtracting the study period's thirteen-year average. When examining these datasets, we focused on identifying localized regions of extremely wet or dry conditions that would impose a signal on the GNSS time series data, while escaping the common mode signal. However, after analyzing the spatiotemporal patterns of soil moisture and SWE volumes throughout Cascadia from 2006-2019 (Figure A1.9 and A1.10), we did not identify any regional changes in soil moisture or SWE that correlated in space and time with the residual surface deformations shown in Figure 2.4.

Lastly, it is possible that the transient signals we observed were imposed onto the time series by the removal of the common mode signal. However, we do not believe that this is the case for two reasons. Firstly, the common mode signals that we calculated for GNSS stations west of the Cascadia volcanic arc show similar trends throughout the entire margin (Figure A1.2). Therefore, if the common mode signal were responsible for the local transients that we observe in the final time series, then we would expect to see these transients at most GNSS stations throughout the margin. In contrast, the transients that we observe are within a limited range of latitudes in Cascadia. Secondly, the relatively small amplitude of the common mode

signal (<1 mm) makes it an unlikely source for the transients, as the majority of the total surface displacements that we focus on are over 1 mm. Therefore, it may be possible that the common mode signal is enhancing some transient signals, but it is not possible for the common mode signal to be generating the transients entirely (Figure A1.11). Therefore, having ruled out these non-tectonic deformation sources, we are left with the possibility that the trenchward transients in the GNSS data are of tectonic origin.

2.4.2 Long-term Slow Slip in Cascadia

The most reasonable tectonic source for the westward transients that we observe in the GNSS time series data is long-duration slow slip on the megathrust. Therefore, we invert the total GNSS displacements for the four possible events and create slip models on the plate interface to evaluate the plausibility of long-term slow slip in Cascadia.

2.4.2.1 Inversion Methods and Data

Having shown that the 2012.45-2014 event is visible in all three independent network solutions that we processed, we opted to use a fourth solution from the Plate Boundary Observatory (PBO) to constrain the inversion. PBO solutions represent a combination solution based on individual PANGA and New Mexico Tech (NMT) solutions through September of 2018. The PBO solutions were reduced using the cleaning methods described in Section 2.2. We did not include any stations north of the Washington-Oregon border in the inversion, since no notable coherent trenchward motion was identified in northern Cascadia (Figure 2.5a).

We utilized a linear non-negative least-squares inversion wherein horizontal surface displacements at GNSS stations were approximated using an elastic half-space dislocation model (Okada, 1992) with a shear modulus of 40 MPa. Our fault model was based on the plate contours from McCrory et al. (2006) and the rake for each fault element was fixed to be antiparallel to the local convergence direction. The convergence direction at each element was determined using a Juan de Fuca-Oregon forearc Euler pole (Wells & Simpson, 2001). The inversion was regularized by smoothing the fault slip with a smoothing parameter that minimized data misfit and roughness of the slip distribution, as determined from the L-curve (Figure A1.12). Roughness is determined by the Laplacian of the fault slip model, while the data misfit is defined by the chi-square statistic:

$$\chi^2 = \frac{1}{n_{data}} \sum_{i=1}^{n_{data}} \frac{(d_{obs}(i) - d_{model}(i))^2}{\sigma(i)^2} \quad (1.1)$$

where n_{data} is the number of time series, d_{obs} and d_{model} are the data and modeled displacements, respectively, and σ is the standard deviation of the observations (the quadrature sum of the 1σ uncertainties at the first and last days of the inversion period). The average 1σ uncertainty for the PBO dataset was 2.1 mm.

We determined the updip and downdip depth limits of our fault model by performing a series of sensitivity tests (Figure A1.13). For one set of tests, we fixed the updip limit of the fault model at the trench (5 km depth) and performed the inversion with various downdip depth limits. Calculating chi-square values for each inversion, we found that surface displacements were not sensitive to slip below 40 km depth. Similarly, we performed a second set of tests where we held the downdip limit of the fault model fixed at 80 km depth, evaluated various updip depth limits, and determined that the surface displacements were not sensitive to slip above 10 km depth. Therefore, the final fault model extended from 10-40 km depth and a slip minimization was applied to fault elements at the perimeter of the model.

2.4.2.2 Slip Model for the 2012.45-2014 event

The slip model produced by this inversion is dominated by a large, well resolved slip patch at 30 km depth with a maximum fault slip of 18 mm (Figure 2.6a). There is also a string of smaller slip patches that extends up to northern Oregon along the 15-20 km depth contour. Along this string, we are most confident about the presence of fault slip at 43°N and 42°N, since there are clusters of coastal GNSS stations at these locations. It is possible that these smaller slip patches are connected by a continuous band of fault slip, but this feature is poorly resolved because of limited station coverage. We are not confident in the small northernmost slip patch in our distributed slip model. This slip patch has small slip amplitudes, is entirely driven by deformation at a few stations in northern Oregon with a small signal-to-noise ratio, and is located offshore where there is limited resolution (Figures 2.6a and 2.6b). Although our model consistently underpredicts surface displacements in southern Cascadia, only two stations are outside of the uncertainty ellipse (Figure 2.6c). Additionally, our STL analysis showed that our cleaning method tended to overpredict transient surface displacements. Therefore, the

underfitting of our slip model and the overestimation of the STL analysis are likely offsetting effects. The equivalent moment magnitude for this long-term SSE along the margin is Mw 6.4.

2.4.2.3 Additional Small Displacement Transients

In addition to the 2012.45-2014 event, we observed additional small displacement transients in the final NGL GNSS time-series data (Figure 2.4). In the end, there are three events that are seen in all three of the datasets: northern Oregon in 2008.75-2009.75, northern California in 2017.2-2019 and southern Washington in 2017.6-2019. Surface displacements for these three events do not exceed 2 mm and fall below our estimated STL error margin, making it difficult to discern whether or not these transients are real tectonic events. Nevertheless, we inverted the final GNSS time series data for these three events to explore the possible slip models that could represent smaller long-term SSEs in Cascadia.

Using the same methods described in Section 2.4.2.1, we inverted the final PBO displacements for the 2008.75-2009.75 epoch in the region surrounding northern Oregon. The slip model produced for this inversion includes a large slip patch located on the updip edge of tremor, with a maximum fault slip of 5 mm (Figure 2.7). Additional patches with smaller slip amounts are also located offshore where there is questionable resolution. The equivalent moment magnitude for this slip event is a Mw 6.2.

The inversions for the 2017.2-2019 and 2017.6-2019 epochs were constrained with our final NGL GNSS time series data, as the combination solutions from PBO were not available after mid-September of 2018. Additionally, our final PBO time series results for the 2017.75-2018.75 epoch differ from the results of the three other processing centers, showing larger and more widespread westward displacements throughout northern Cascadia. It is unclear whether this signal is real or perhaps a consequence of the PBO combined solutions ending in late 2018; thus, we opted to use the more consistent NGL time series results. Our slip model for the northern California 2017.2-2019 epoch is similar to our slip model for the earlier event in 2012.45-2014. Specifically, slip is concentrated in a patch located directly updip from tremor with a maximum fault slip of 18 mm (Figure A1.14). This event has an equivalent moment magnitude of Mw 6.1. Our slip model for the southern Washington 2017.6-2019 epoch includes a single patch located predominantly offshore, with a maximum slip of 9 mm, and an equivalent moment magnitude of Mw 6.1 (Figure A1.15). It should be noted that our inversions for the

latter two events provide *minimum* slip and moment magnitude estimates, as we only inverted displacement up until 2019. Additional analysis of post-2019 time series is necessary to capture the end of these events and provide final slip and moment magnitude estimates.

2.4.3 Comparison with Nankai

The 2012.45-2014 southern Cascadia long-term SSE is comparable to events in Nankai in terms of its duration and depth on the plate interface (Ozawa, 2017; Takagi et al., 2019). The majority of slip in our model, including the patch of maximum slip in the south, fills the gap between short-term SSE tremor and the 20% locking boundary (Schmalzle et al., 2014; Figure 2.6a). This location of the long-term slow slip relative to the updip seismogenic zone and downdip short-term slow slip region is consistent with the spatial pattern of slow slip in Nankai. However, the maximum slip amount (18 mm) and surface displacement (3.7 mm) for this Cascadia event are an order of magnitude smaller than most well-documented Nankai events. As a result, the moment magnitude for this event (Mw 6.4) is also relatively small. In order to calculate the stress drop for this event, we assumed a circular fault geometry for the southernmost slip patch. We selected a radius that encompassed the region where the fault slip was greater than 1 mm. We calculated the average slip on this circular patch by averaging the fault slip on all fault elements that fell within the circular bounds. This method provided a stress drop of 27 kPa. This stress drop value falls on the low end of the stress drop spectrum for most long-term SSEs in Nankai (5-200 kPa) (Ozawa et al., 2017). Nevertheless, relatively small long-term SSEs with $\leq$ Mw 6.5 have been observed underneath Nankai's Bungo Channel (Takagi et al., 2016), Shima Peninsula (Kobayashi & Tsuyuki, 2019), and western Shikoku (Takagi et al., 2019), confirming that a lower range of slip and moment magnitudes has been observed for long-term SSEs.

2.4.4 Relocking and Recurrence Intervals

Although our slip model can account for the westward deformation seen at stations in southern Cascadia from 2012.45-2014, it cannot explain the subsequent eastward deformation at these stations from 2014-2017.2 (Figures 2.5b and 2.5c). Materna et al. (2019) was the first to recognize similar transients in southern Cascadia and concluded that they were caused by dynamic stress changes from regional earthquakes, which triggered fluid migration and changes

in the level of coupling along the plate interface. Therefore, Materna et al. (2019) attributes this west-east shift as a recoupling of the plate interface triggered by the 2014 offshore earthquake (Figure 2.1). Alternatively, this shift could be signaling the rapid relocking of the plate interface after the end of the long-term SSE, as has been observed for short-term SSEs in Cascadia (Haines et al., 2019). Rapid relocking of the megathrust is also supported by laboratory experiments, which observe the commencement of fault healing within minutes of slow slip termination (Kaproth & Marone, 2013).

After ~ 2.2 years of eastward deformation, the stations in southernmost Cascadia revert back to westward orientations, indicating the possible start of another long-term SSE that was ongoing as of 2019. Thus, we may be observing the sawtooth pattern associated with repeating slow slip cycles (Szeliga et al., 2004). If we define a recurrence interval as the difference in time between the start of consecutive SSEs, then the timing of these two events provides a recurrence interval of 4.75 years. Assuming an average fault slip of 18 mm per event and a convergence rate of 2.9 cm/yr for southern Cascadia (McCorry et al., 2006), the long-term SSEs in southern Cascadia have accommodated $\sim 13\%$ of plate motion since 2012 based on one cycle. Since we do not see clear evidence for a similar long-term SSE from 2006 to 2012, this percentage appears reasonable. Therefore, long-term slow slip is responsible for only a small percentage of the strain release at this location on the megathrust.

It is worth noting that these interpretations, and others in this work, assume that an accurate steady-state interseismic velocity has been removed from the GNSS time series data. In Cascadia, this interseismic rate is likely some combination of strain accumulated through megathrust locking and crustal block rotation (Wells et al., 1998). However, it is possible that other processes, such as yet-unidentified longer-duration SSEs or transients, are contributing to the overall baseline trend in Cascadia. Since this thirteen-year study only spans a small fraction of the earthquake cycle, it is difficult to discern the true GNSS background velocity in Cascadia.

2.4.5 Relationship with Short-Term SSEs and Regional Earthquakes

We do not find an interpretable temporal relationship between the 2012.45-2014 long-term SSE and down-dip tremor activity. Short-term SSEs (as indicated by elevated tremor) occurred right before and right after the onset of the long-term SSE in southern Cascadia, while several other short-term events occurred throughout the duration of the long-term event. In

Nankai, interactions between short- and long-term SSEs, such as triggering, changes in recurrence interval, and accelerated slip rates, have been observed in the Bungo Channel (Hirose & Obara, 2005); however, we do not observe any first order changes in short-term SSE behavior after the southern Cascadia long-term event, nor do we see any obvious connection between the timing of the short- and long-term events.

In addition to short-term SSEs, three regional strike-slip earthquakes occurred in the subducting Gorda slab updip of the long-term slow slip patch in January 2010 (Mw 6.5), March 2014 (Mw 6.8), and December 2016 (Mw 6.6) (Figure 2.1). The 2014 and 2016 earthquakes took place near the beginning and end of the relocking period, respectively (Figure 2.5c). However, the nature of our cleaning method results in smoothed time series that lack precise temporal resolution, making it difficult to determine the exact timing of the earthquakes relative to the start and end of the relocking period. Previous geodetic work by Materna et al. (2019) asserts that the 2014 and 2016 earthquakes are tied to the transients observed in the southern Cascadia GNSS data, as they triggered increased and decreased levels of coupling along the plate interface, respectively. Our final GNSS time series results generally agree with those of Materna et al. (2019). In particular, both results reveal an increase in eastward displacements near the time of the 2014 earthquake and a subsequent increase in westward displacements near the time of the 2016 earthquake. However, our time series do not show a systematic change near the time of the 2010 earthquake, as shown in the results of Materna et al. (2019), but rather in 2012.45. Thus, we attribute transients in the 2012.45-2014 epoch to a long-term SSE and the 2014-2017.2 period to rapid and sustained relocking after the long-term SSE. After 2017.2, we suspect that another long-term SSE may have commenced, but the analysis of more recent GNSS time series data must be done before we can confirm this hypothesis.

2.4.6 Relation to Interseismic Locking

The 2012.45-2014 long-term SSE in southern Cascadia fills the gap between the updip locked zone and the downdip zone of short-term SSEs and suggests that a small fraction of the plate rate is released as long-term slip transients. We also find evidence for possible long-term SSEs in northern and central Cascadia, but these events are poorly resolved and are expressed as very small surface displacements. Thermal numerical models for Nankai, Mexico, and the Hikurangi trench in New Zealand suggest that long-term slow slip occurs within a semi-frictional

zone along the plate interface that fills the gap between the locked and short-term slow slip zones (Gao & Wang, 2017). However, this semi-frictional zone is not present in similar models for northern Cascadia due to higher subduction zone temperatures. Therefore, it may be the case that the width of this semi-frictional zone dictates whether or not long-term slow slip is possible within a subduction zone and that southern Cascadia offers the only location where this zone is wide enough to host sizable long-term events.

In addition to the width of the semi-frictional zone, the size and frequency of long-term SSEs may also be impacted by the degree of locking along the plate interface. Takagi et al. (2019) found that long-term SSEs in Nankai were smaller and less frequent in regions with strong updip locking due to stress shadowing effects. A similar anticorrelation between locking and the magnitudes of adjacent long-term SSEs is observed in Mexico, where large $>M_w 7.4$ long-term SSEs occur directly downdip of the weakly coupled Guerrero seismic gap (e.g. Radiguet et al., 2012). Following this hypothesis, we might expect to see the largest and most frequent long-term SSEs in central Cascadia, where the shallow plate is assumed to be partially creeping due to low levels of coupling (Schamlzle et al. 2014). However, this is not what we observed, perhaps due to the short observation period or perhaps suggesting that Cascadia does not fit into the framework proposed by Takagi et al. (2019). Additional subduction zones, such as Alaska and New Zealand, also do not clearly fit into this model (Rousset 2019). For example, large ($> M_w 7$) long-term SSEs in Alaska occur directly downdip of the asperity that ruptured in the 1964 Prince William Sound earthquake (e.g. Ohta et al. 2006). Future improved locking models of Cascadia that employ offshore geodetic data will allow us to further evaluate the connection between up-dip locking and long-term slow slip.

2.5 CONCLUSIONS

We analyzed thirteen years of Cascadia GNSS time series data and found evidence for a small long-term SSE in southern Cascadia from 2012.45-2014. We inverted the GNSS displacements for this event, and the resulting slip model contained a maximum 18 mm of slip concentrated at 15-35 km depth and had an equivalent moment magnitude of $M_w 6.4$. We also found evidence for additional smaller transient events; however, we are not confident in the source of these transients on account of their relatively small (<2 mm) surface displacements. The discovery of a long-term SSE in southern Cascadia solves an observational conundrum and

better aligns Cascadia's slow slip behavior with other subduction zones by filling the gap between the locked and short-term slow slip zones. We suspect that the spatial extent of the semi-frictional zone limits the size, location, and recurrence interval of long-term SSEs in southern Cascadia and across the entire margin.

2.6 ACKNOWLEDGEMENTS

We would like to thank our two anonymous reviewers and the associate editor at G^3 for their valuable feedback. Funding for this work has been provided by the National Science Foundation (EarthScope-1251954). The authors would also like to thank the multiple processing centers that provided GNSS data for this work (PANGA, PBO, NGL, and SOPAC).

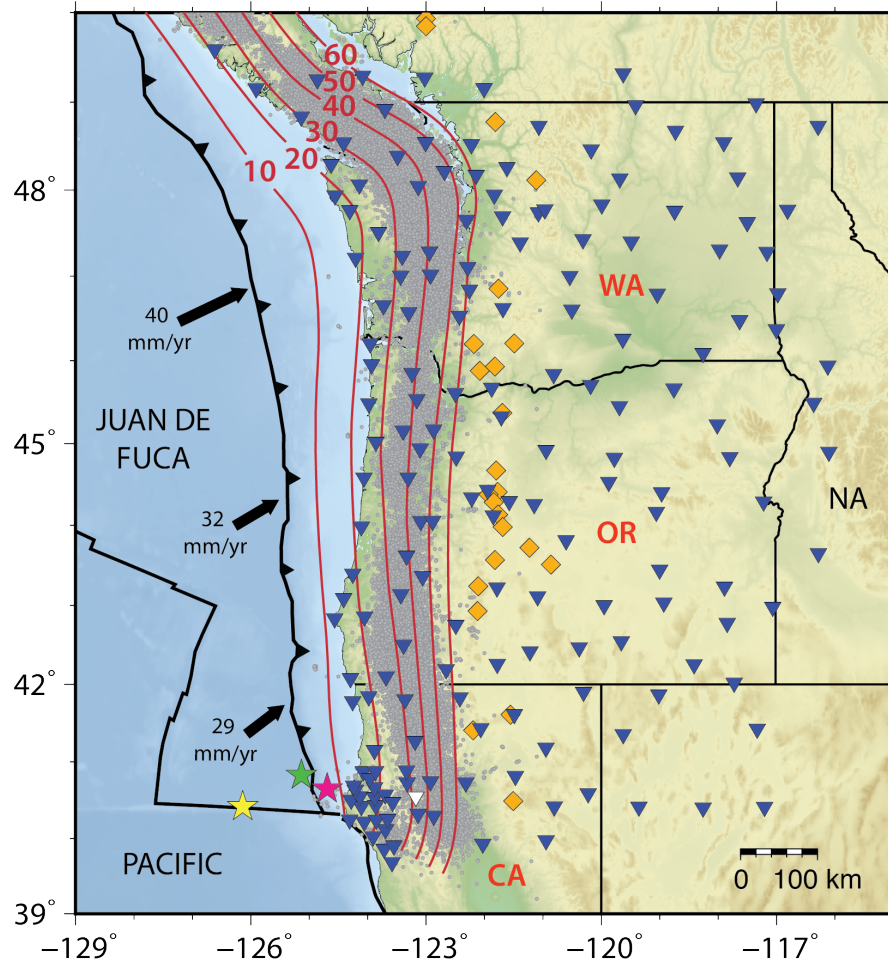


Figure 2.1 Map of the Cascadia subduction zone. Black arrows show Juan de Fuca plate convergence direction and rate relative to North America (NA), red lines delineate slab depth contours from McCrory et al. (2006), orange diamonds mark the locations of Holocene volcanoes in the Cascadia arc, grey circles display the location of tremor, a spatiotemporal proxy for short-term SSEs, from 2010-2019, inverted triangles indicate the locations of GNSS stations used in this study, inverted white triangle notes the location of station P332, and the locations of the 2010, 2014 and 2016 offshore earthquakes in southern Cascadia from the USGS Earthquake Catalog (<https://earthquake.usgs.gov/earthquakes/search/>) are shown with magenta, green, and yellow stars, respectively. State names: WA - Washington, OR - Oregon, and CA - California.

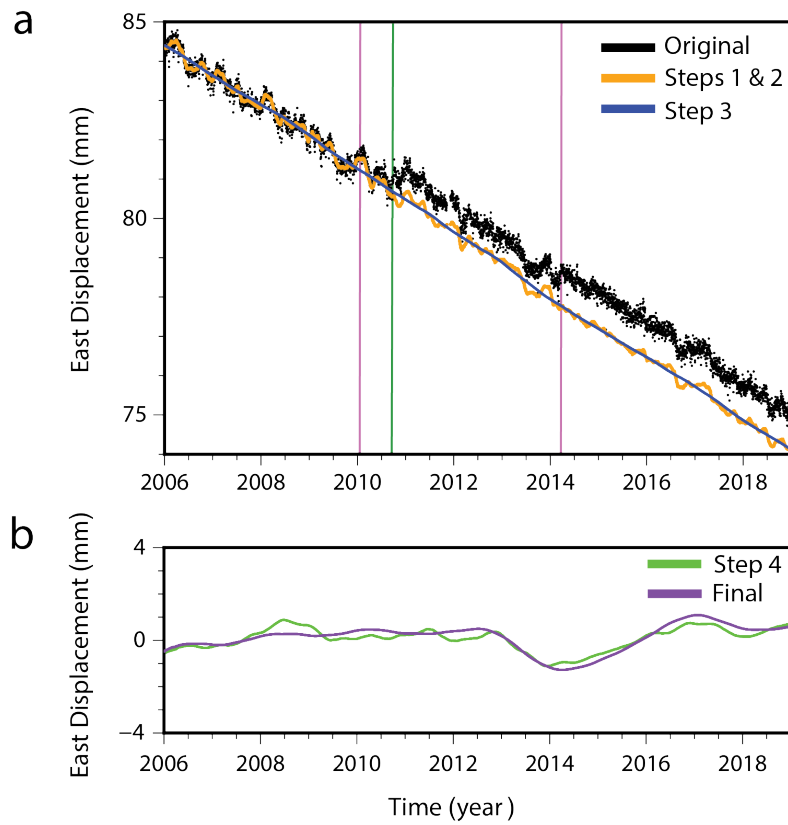


Figure 2.2 Time series plots for station P332 outlining the 5-step cleaning procedure for the GNSS time series data. **a** The original GNSS time series data (black), and the resulting time series once Steps 1 and 2 (orange) and Step 3 (blue) have been applied. Vertical bars represent the timing of steps that were removed due to earthquakes (purple) and equipment changes (green). **b** The resulting time series after Steps 4 (green) and 5 (purple) have been applied.

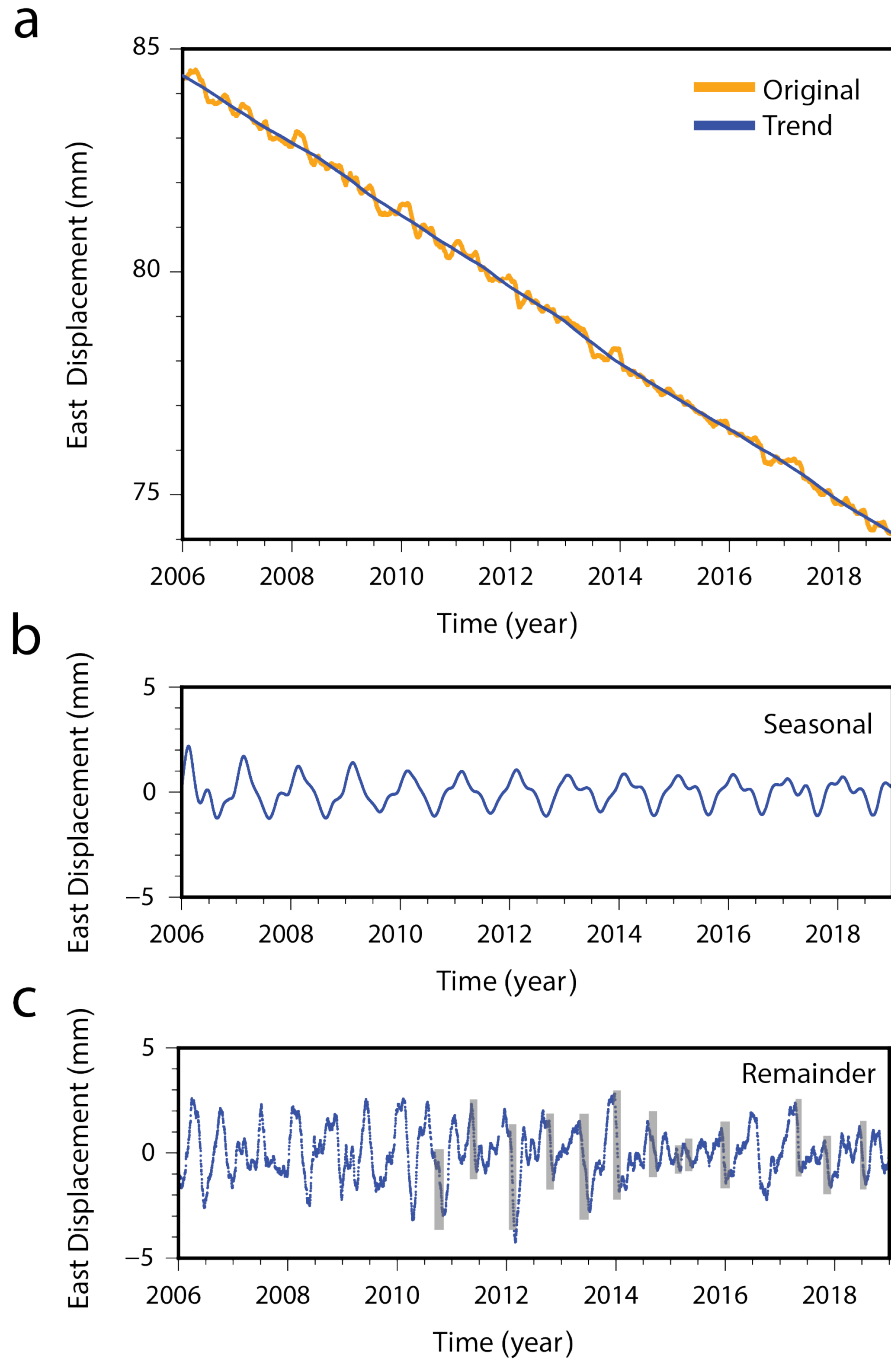


Figure 2.3 Time series plots for station P332 outlining the results from the STL algorithm. **a** The original GNSS time series data (orange) and the trend (T) component from STL (blue). **b** The seasonal (S) component from STL. **c** The remainder (R) component from STL. Grey bars denote the timing of large SSEs that correlate with elevated levels of local tremor (Figure A1.1).

	Minimum	Maximum
Slope (mm/yr)	-10	10
Seasonal (mm)	0	2
Common Mode (mm)	N/A	1
White Noise (mm)	0.55	0.86
Flicker Noise (mm/a ^{0.25})	0.3	1.6
Random-Walk Noise (mm/a ^{0.5})	0	1.4

Table 2.1 Minimum and maximum values for the different components of the synthetic time series used for STL error analysis.

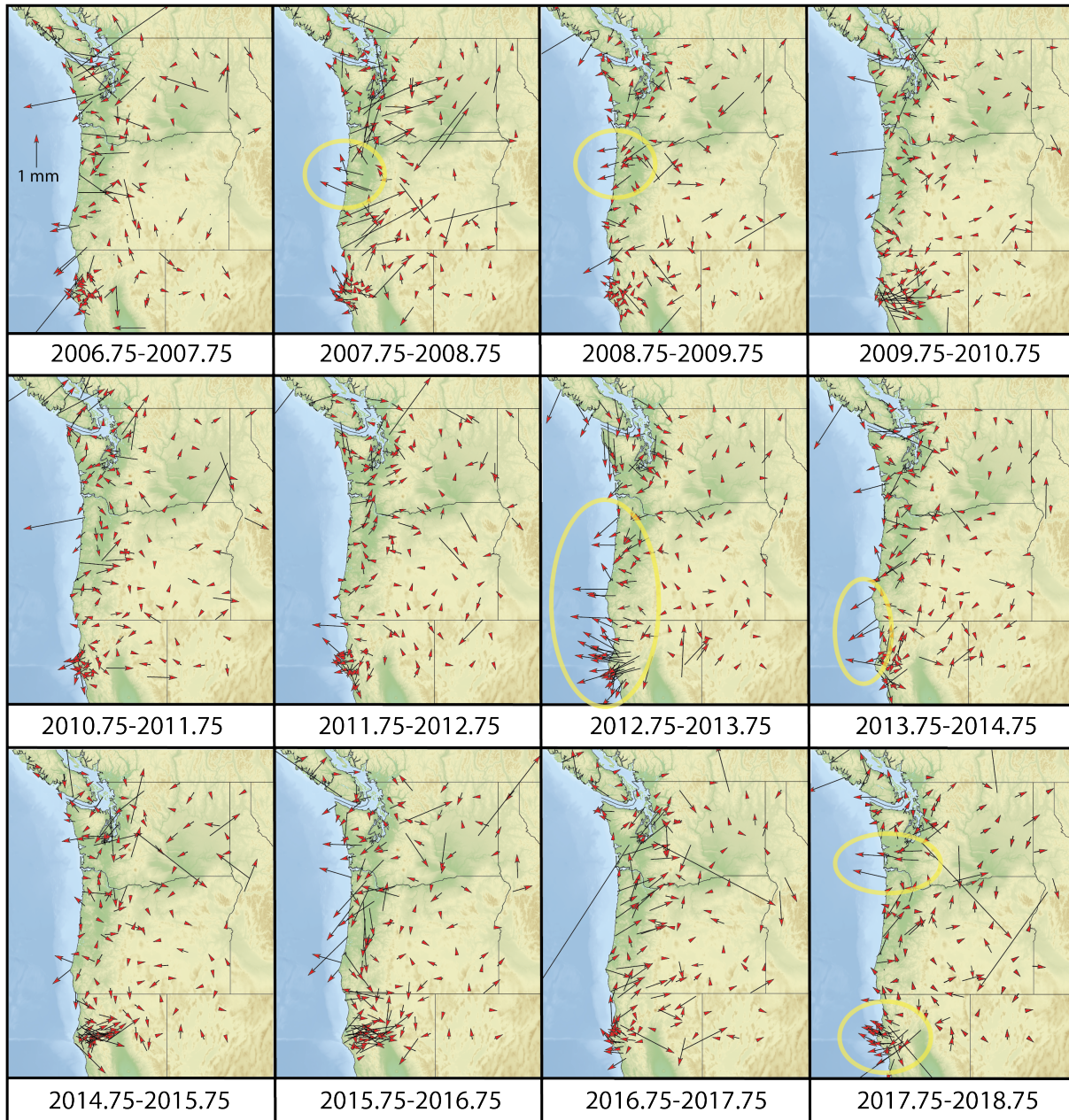


Figure 2.4 Horizontal displacements from our residual time-series data at GNSS stations during one-year epochs. Each epoch begins at the start of the water year (October 1st) to minimize seasonal signals. Original time series solutions are provided by the Nevada Geodetic Laboratory (NGL), which are then cleaned to produce residual displacements. Yellow ellipses outline neighboring subsets of stations with coherent westward motion.

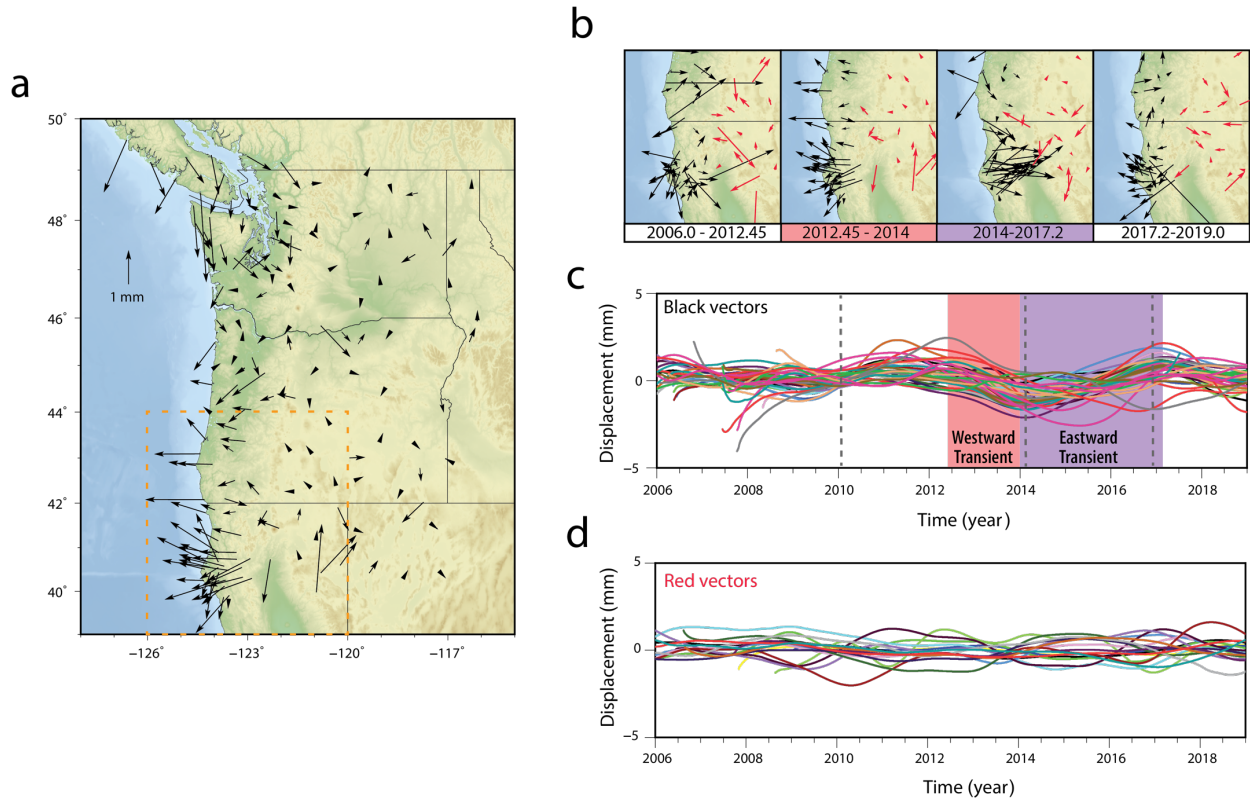


Figure 2.5 Residual vector and time series results for the reduced GNSS time series data, using solutions provided by the Nevada Geodetic Laboratory. All products represent the deviation from the long-term trend after the removal of linear and seasonal components. **a** Vector map showing horizontal displacements at GNSS stations from 2012.45-2014. Dashed orange box outlines the region displayed in panel **b**. **b** Vector maps for southern Cascadia showing horizontal displacements at GNSS stations during designated time periods displayed at the bottom of each panel. Black and red vectors indicate stations that do and do not move coherently together, respectively. **c** Reduced longitudinal GNSS time series results for the black vectors in panel **b**. Red and purple time periods denote westward and eastward transients, respectively, and dashed dark grey lines show the timing of the 2010, 2014 and 2016 earthquakes. **d** Reduced longitudinal GNSS time series results for the red vectors in panel **b**.

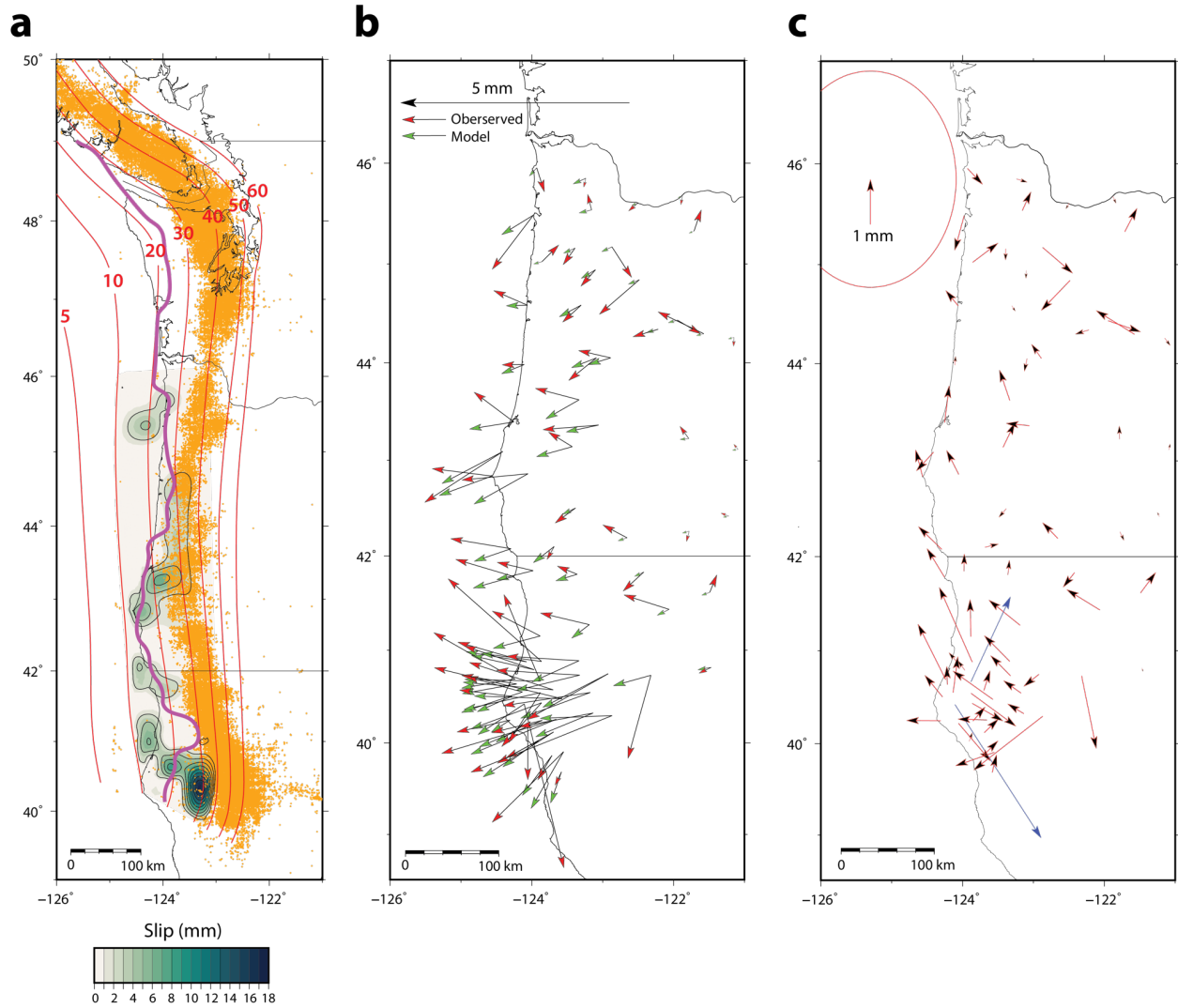


Figure 2.6 Inversion results from horizontal GNSS displacements from 2012.45-2014. **a** Final slip model from our inversion with contours (black lines with green shading) outlining every 2 mm of fault slip, slab depth contours (red lines, km), tremor locations during the event (orange dots), and the contour of 20% locking (Schmalzle et al., 2014), representing the base of the seismogenic zone (thick purple line). **b** Vector map comparing observed (red) and modelled (green) GNSS displacements. **c** Vector map showing misfit between observed and modelled GNSS displacements. The average 1σ error ellipse for all stations is shown in the upper left-hand corner with the average east and west error being 1.9 and 2.4 mm, respectively. Only two stations (blue vectors) exceed this error. Vectors in panels **b** and **c** share the same scale, shown in the upper left-hand corner of each panel.

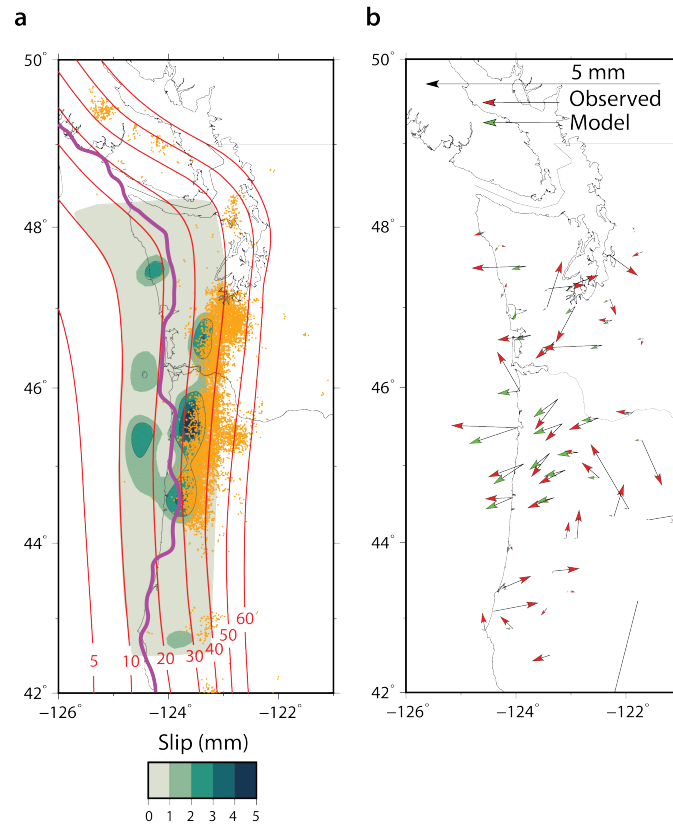


Figure 2.7 Inversion results from horizontal PBO GNSS displacements from 2008.75-2009.75 in northern Oregon. **a** Final slip model from our inversion with contours (black lines with green shading) outlining every 2 mm of fault slip, slab depth contours (red lines, km), tremor locations during the event (orange dots), and the contour of 20% locking (Schmalzle et al., 2014), representing the base of the seismogenic zone (thick purple line). **b** Vector map comparing observed (red) and modelled (green) GNSS displacements.

Chapter 3. STRAIN PARTITIONING AMONG FOREARC FAULTS IN SOUTHERN CASCADIA INFERRED FROM GNSS

The content of this chapter is published in:

Nuyen, C.P & Schmidt, D.A. (2022). Strain partitioning among forearc faults in southern Cascadia inferred from GNSS. *Journal of Geophysical Research: Solid Earth*, 127(8). <https://doi.org/10.1029/2022JB024236>

3.1 INTRODUCTION

Deformation in southern Cascadia is a complex overprinting of various tectonic processes, including subduction, migration of the Mendocino triple junction (MTJ), San Andreas strike-slip faulting, and crustal block translation. Together, these processes produce compressional and translational strains that are partially accommodated along active crustal faults in the forearc. In particular, the Little Salmon fault, Mad River fault zone (MRFZ), and Grogan fault have been identified as Quaternary active structures (Kelsey & Carver, 1988; McCrory, 2000), with the Little Salmon and MRFZ having hosted several large earthquakes in the past two thousand years (Carver & Clarke, 1992). Estimating slip rate values for these forearc faults is important, as they 1) help indicate the hazards associated with future crustal earthquakes, 2) shed light on how slip is balanced between crustal faults and the megathrust, and 3) provide a better understanding of how crustal faults accommodate strain in the broader MTJ region.

Previous geologic and paleoseismic work (see citations in Section 3.2.2) provide slip rates for the Little Salmon fault, MRFZ, and Grogan fault; however, these values are still fairly unconstrained due to discrepancies between different studies and field sites, as well as the uncertainties associated with the methods used to estimate fault slip rates. Fortunately, increased spatial and temporal coverage of geodetic instrumentation in the MTJ region over the last several decades provides another lens through which these faults can be studied. Nevertheless, modern geodetic data has not yet been used to investigate the full spectrum of interseismic deformation on these specific fault structures north of the MTJ. Therefore, this study presents elastic block

models of the MTJ region to provide independent slip rate estimates and offer insights into how these faults accommodate strain in the forearc.

3.2 BACKGROUND

3.2.1 Tectonic Setting

The MTJ represents the meeting of the Pacific, North American, and Gorda plates offshore of Cape Mendocino, CA (Figure 3.1). Crustal deformation in the region surrounding the MTJ is largely impacted by the interactions of these three plates. North of the MTJ, the Gorda slab subducts obliquely beneath the North American plate at the southern terminus of the Cascadia subduction zone (CSZ) at 32 mm/yr (McCaffrey et al., 2007). However, there is some uncertainty about the convergence rate (e.g. Bartlow, 2020), as the Euler pole model fails to incorporate the internal deformation of the Gorda plate (Wilson, 1989; Chaytor et al., 2004). Interseismic coupling between the Gorda and North American plates along the CSZ megathrust fault results in compressional strain within the North American plate that is stored elastically along the megathrust (e.g. Savage, 1983) and along crustal faults within the forearc (e.g. Clarke & Carver, 1992). Megathrust locking models based on GNSS data indicate that the partially locked portion of the plate interface extends onshore in the MTJ region (Li et al., 2018). However, this locking pattern may be time dependent and influenced by dynamic stressing from nearby earthquakes (Materna et al., 2019).

South of the MTJ, the northern San Andreas fault (SAF) system accommodates ~40 mm/yr of dextral shear between the Pacific and North American plates (Freymueller et al., 1999; Murray et al., 2014). Motion across the SAF system results in ~11 mm/yr of northwest translation of the Sierra Nevada-Great Valley (SNGV) block, a rigid portion of the crust that experiences relatively little internal deformation (Dixon et al., 2000; McCaffrey, 2005). The northern end of the SNGV block neighbors the Oregon Coast block, which comprises the Cascadia forearc in northern California and Oregon. The northwest migration of the SNGV block, coupled with Basin and Range extension, causes the Oregon Coast block to translate north and rotate clockwise, driving the overall rotation of the Cascadia forearc (Wells et al., 1998; McCaffrey et al., 2013). The exact location of the boundary between the SNGV and Oregon

Coast blocks is unclear, but it is likely represented by a diffuse zone of deformation (Williams et al., 2006; Unruh and Humphreys, 2017).

In addition to contributions from the CSZ and SAF, crustal deformation in the forearc is further complicated by the northward migration of the MTJ. Geometric analysis of the MTJ transform-transform-subduction system predicts the creation of a slab window south of the MTJ, wherein hot mantle upwelling can occur (Atwater, 1970; Dickinson & Snyder, 1979a, 1979b). Deformation associated with the passage of the MTJ and slab window is commonly attributed to the Mendocino “crustal conveyor” model, which predicts regional changes in shallow crustal strain rates and velocities arising from mantle-lithosphere interactions (Furlong & Govers, 1999). In this model, hot mantle rises to fill the slab window and consequently experiences cooling, creating viscous coupling between the southern edge of the Gorda plate and the base of the North American crust. Therefore, as the Gorda plate migrates to the north, it drags the North American crust along with it, causing the North American crust to contract and thicken in front of the MTJ. In contrast, the crust behind the MTJ experiences extension and thinning on account of the crust being pulled to the north. Support for the Mendocino crustal conveyor model is provided by spatiotemporal patterns of uplift from marine terraces (Merritts & Bull, 1989), geodetic vertical surface velocities (Peng et al., 2017), the migration of knickpoints and landscape response (e.g. Bennett et al., 2016), and tomography (e.g. Liu et al., 2012).

3.2.2 Active Forearc Faults

Geologic work suggests that the Mad River fault zone (MRFZ) and the Little Salmon fault are the two major thrust systems in the southern Cascadia forearc (Kelsey & Carver, 1988). The 15-km-wide MRFZ consists of five main imbricate thrust faults that strike NW-SE and dip to the northeast, while the Little Salmon fault is a 100-km-long arcuate thrust system that consists of shallowly north-to-northeast dipping faults. Paleoseismic trenching across the Little Salmon fault and MRFZ reveals that these systems have hosted multiple earthquakes in the Late Holocene. Clarke & Carver (1992) found evidence for three slip events on the Little Salmon fault in the past ~1,700 years, each resulting in 3.6 to 4.5 meters of reverse slip. Trenching across the McKinleyville and Mad River faults of the MRFZ revealed at least two displacement events during that same time period, each with a minimum of 3.5 meters of slip. Three to four slip events in the past 1,800-2,300 years have also been identified on the western Little Salmon fault,

with 1.3 - 4.1 meters of slip per event (R.C. Witter, personal communication, 2022). Estimates for past earthquake magnitudes on the Little Salmon fault, based on approximate fault rupture area and slip amount, come out to $M_w \sim 7.6-7.8$ (Clarke & Carver, 1992; PGE, 2002). Current slip rate estimates for the MRFZ and Little Salmon fault are 3.7-6.4 mm/yr and 1.3-12 mm/yr, respectively (Kelsey & Carver, 1988; Clark & Carver, 1992; McCrory, 2000; Kelsey, 2001).

The north-northwest striking Grogan fault lies to the north of the Little Salmon fault and MRFZ, and has facilitated translation of the forearc through large amounts of dextral slip (Kelsey & Carver, 1988; Kelsey, 2001). Although the displacement of corollary units across the Grogan fault suggests that it accommodated upwards of 75 km of slip in the late Tertiary (Kelsey & Hagans, 1982), more precise slip rate estimates cannot be made from geologic observations alone. While the majority of the Grogan fault is inferred to have nearly vertical dip, exposures of this fault near the coast show slightly shallower dips of 55° and indications of reverse slip motion (Kelsey, 2001). Reverse slip on the Grogan fault is also supported by vertical offsets observed in offshore seismic reflection profiles along the Grogan fault and parallel fault structures (McCrory, 1996; Gulick & Meltzer, 2002). Therefore, the Grogan fault is thought to be a predominantly high-angle strike-slip fault that also has some component of reverse slip.

The MRFZ, Little Salmon fault, and Grogan fault extend offshore and likely connect with structures in the accretionary prism to the north. The offshore geometry of these structures, and their mode of slip, remains an area of active research. Seismic reflection studies reveal a complex pattern of active faulting throughout the offshore portion of the forearc in the Eel River Basin, wherein several different deformational regimes exist (Clarke, 1992; Gulick & Meltzer, 2002; Watt & Brothers, 2020). For example, the western margin of the offshore forearc basin is dominated by east-west oriented compression due to subduction, while the eastern margin represents a more transpressional regime, resulting from the combination of subduction forces and strike slip tectonics (Gulick & Meltzer, 2002). Therefore, as the Little Salmon fault, Grogan fault, and MRFZ continue offshore, it is possible that their slip sense and/or slip rates may evolve to adjust to the changing tectonic environments.

Paleoseismic work and regional mapping have helped to characterize these structures and their significance; however, less work has been done to better constrain and verify these geological estimates with geodetic observations. In particular, the geologic slip rate on the Little Salmon fault spans a wide range (1.3-12 mm/yr), and the slip rate on the Grogan fault is fairly

unknown. Placing tighter constraints on the slip rates of these forearc faults is important, as they may be the source of future large crustal earthquakes. Additionally, slip on these faults is likely accommodating subduction convergence and transferring slip away from the megathrust. Therefore, seismic hazards in this region may be shifted off of the megathrust and onto crustal forearc faults (McCrory, 2000).

3.2.3 Previous Geodetic Work

A decade of GNSS time-series data afforded by the Plate Boundary Observatory (Herring et al., 2016) is available at stations throughout the MTJ and has the potential to help clarify deformational patterns and slip rates in this area. Williams et al. (2006) used GNSS velocity data to investigate the various tectonic processes that contribute strain to the MTJ region and constrain slip rates on the northern SAF system. This work notes meaningful changes in deformation associated with the translation of the SNGV block and the SAF system, and successfully recreates SAF-parallel GNSS velocities by modeling slip on the northern SAF strands with 2-D dislocation models. However, Williams et al. (2006) does not explicitly consider slip on any faults north of the MTJ, particularly the MRFZ, Little Salmon fault, and Grogan fault. Furthermore, this work assumes all fault motion to be strike slip and does not incorporate any transpressional or thrust component.

Block models are another tool that can leverage GNSS velocity data to study crustal deformation near the MTJ. These models divide the crust into fault-bounded blocks and assume that interseismic surface velocities are a superposition of 1) block rotation, 2) homogeneous strain within the blocks, and 3) elastic strain imparted by locked faults on the block edges (McCaffrey, 2002; Meade & Loveless, 2009). By simultaneously solving for 1), 2) and 3), block models are able to estimate fault slip rates and offer insights into how slip is partitioned. Previous block modeling efforts near the MTJ either focus solely on Cascadia (e.g. McCaffrey et al., 2013) or the SAF system (e.g. McCaffrey, 2005) and exclude the MTJ, or focus too broadly on the western US and approximate the structures around the MTJ (e.g. Parsons et al., 2013; Pearson and Snay, 2014; Evans et al., 2015). Furthermore, strain contributions from the southern Cascadia subduction zone are largely ignored in many of these previous block models. Deformation in the forearc due to locking on the megathrust has been explored through various viscoelastic models of the Cascadia subduction zone (Pollitz & Evans, 2017; Li et al., 2018).

However, very few of these studies examine how slip may be partitioned between the megathrust and crustal faults near the MTJ (e.g. Pollitz et al., 2008). Therefore, we present a block model of the MTJ region that fills a gap in the coverage of existing models, incorporates refined forearc fault structures, and includes the Cascadia megathrust.

3.3 METHODS AND DATA

We use an open source package called TDEFNODE to perform the block modeling (McCaffrey, 2009). Using this code, we model GNSS surface velocities as a combination of block motion, elastic strain accumulation, and uniform internal block deformation. The motion of each block is defined by rotation about a Euler pole, with the cartesian rotation vector intersecting the center of the block. To accommodate relative block rotations, slip is projected along the faults that constitute the block boundaries. Elastic strain accumulation within the blocks is attributed to interseismic locking along these fault boundaries, which can be modeled as a coseismic slip deficit rate resulting from steady slip on the fault below a predetermined locking depth (Savage & Buford, 1973). The coseismic slip deficit is modeled by applying backslip to the faults, and the resulting surface deformation is estimated with an elastic, half-space dislocation model (Okada, 1985; Okada 1992). Since we do not incorporate transient events into our modeling, TDEFNODE assumes that all strain is due to steady tectonic motions. Readers are referred to the Supplementary Information of McCaffrey (2009) for further details on TDEFNODE.

We constrain the block model with horizontal GNSS velocities from 76 stations across northern California. We use published velocities from the Pacific Northwest Geodetic Array (PANGA) at Central Washington University in the NAM14 stable North American reference frame (PANGA, 1996). Velocity estimates in this region may be biased by slow slip events (SSEs) in southern Cascadia and by large offshore earthquakes in the Gorda plate. Therefore, we calculate independent velocity estimates from the original CWU GNSS time series data to verify that neither SSEs nor earthquakes influence the published site velocities. Details on how we calculate the GNSS surface velocities can be found in the Supplemental Information. Ultimately, our independent velocities match the published velocities well, with only one station having a difference ≥ 1 mm/yr. Therefore, we are confident that the published GNSS velocities that we use to constrain the block model are not biased by transient tectonic processes.

The vertical velocity field around the MTJ is dominated by large signals that represent a complex combination of tectonic and non-tectonic processes, including subduction, migration of the slab window, and sustained groundwater extraction (Hammond et al., 2016; Argus et al., 2017; Peng et al., 2017). Although it is possible to incorporate these velocities into our block modeling, we found that TDEFNODE was not able to successfully model the vertical component for several reasons. Firstly, TDEFNODE does not include vertical strain in the internal deformation of blocks. Therefore, vertical deformation can only be modeled by fault slip. As a result, blocks that are bounded by vertical strike-slip faults in our models are not able to recreate vertical signals without placing unreasonably high slip rates on the block edges. Additionally, block models are not able to represent vertical deformation in the MTJ region that is due to processes outside of fault slip, such as crustal thickening and thinning associated with the crustal conveyor model (Furlong & Govers, 1999; Peng et al., 2017) or changes in groundwater levels (Hammond et al., 2016; Argus et al., 2017). Therefore, after evaluating the advantages of using this additional data, we opted to not include the vertical component in our block modeling.

3.3.1 Strain Rate Analysis

To help guide the creation of the block model geometries, we calculate the regional strain rate field using the horizontal GNSS velocities to better understand the distribution of strain around the MTJ and identify important fault structures (Figure 3.2). We calculate the strain rate field using an open source package called PyTAGS (Crowell, 2019) and employ a gridded method that calculates the strain rate values at predetermined grid nodes throughout the study region (Almendinger et al., 2007). We use a grid spacing of 15km and apply a distance weighting so that the strain at each grid node is not affected by distant GNSS stations. We calculate strain rates for the broader MTJ region using all GNSS station velocities, but choose to focus our analysis on the onshore portions of the forearc and northern SAF system.

Our regional strain rate maps show contraction north of the MTJ along the coast (Figures 3.2A, 3.2B). This signal encompasses the Little Salmon fault, MRFZ, and Grogan fault, suggesting that they are actively facilitating shortening in the forearc. Additionally, we observe a change in the azimuth of the most compressive principal strain rate from south to north. Specifically, azimuths rotate counter-clockwise from a north-south orientation south of the MTJ to an east-west orientation in northern California (Figure 3.2B). Unsurprisingly, high shear strain

rates are observed along the northern termini of the San Andreas, Maacama, and Bartlett Springs faults (Figure 3.2C). This broad zone of high shear strain rates, which is associated with the northern SAF system, wraps around the coast and terminates at Cape Mendocino. However, an additional band of high shear strain rates is observed in the vicinity of the MRFZ and Grogan fault, indicating that these faults are also accommodating shear deformation. Overall, these strain rate maps reinforce the notion that the Little Salmon fault, MRFZ, and Grogan fault are actively facilitating deformation in the forearc and that these faults should be incorporated into our block model geometries in order to more fully represent regional deformation.

3.3.2 Fault Geometries

Our objective when creating the block model geometries is to 1) subdivide the crust into a series of realistic blocks, where the margins of the blocks coincide with primary faults, and 2) keep the geometry as simple as possible, i.e. using the fewest blocks, while still representing the observed deformation. We test a total of eight different block geometries to explore various forearc configurations and fault interactions. Geometry 1 represents the simplest geometry, wherein the Little Salmon fault, MRFZ, and Grogan fault are absent (Figure 3.3). Geometries 2-8 then build off of geometry 1, incorporating different combinations of these three forearc faults.

3.3.2.1 Onshore

The onshore block boundaries follow the mapped traces of known faults (Figure 3.1B) and are adapted from the USGS Quaternary Fault and Fold Database for the United States (<https://www.usgs.gov/natural-hazards/earthquake-hazards/faults>) and the third Uniform California Earthquake Rupture Forecast (UCERF3) (Field et al., 2013). For the MRFZ, we use the trace of the northernmost Trinidad fault as a block boundary. In some locations where blocks need to connect, block boundaries cross a region without a known mapped fault. We refer to these unmapped boundaries as pseudo-faults (Figure A2.2). These pseudo-faults do not contribute any deformation to the model and serve to connect known fault structures in order to create closed polygonal blocks. The outer perimeters of the block model consist entirely of pseudo-faults and are placed far enough away from the MTJ region so as to avoid edge effects, while still maintaining model simplicity. Lastly, we infer the presence of a fault between the Maacama fault and Garberville fault zone, as these two fault systems align geographically and

have the same sense of slip. Additionally, our regional strain rate analysis shows high maximum shear strain rates in this area, which supports the presence of a through-going fault structure (Figure 3.2C). This inferred fault is modeled as an extension of the Maacama fault and shares the same fault characteristics, i.e. dip and locking depth.

3.3.2.2 *Offshore Fault Structures*

It is difficult to constrain the offshore deformation in southern Cascadia with a block model due to the lack of offshore geodetic instrumentation and our limited understanding of active fault structures offshore. Therefore, we opt to incorporate a simplified offshore (OFFS) block into our models that encompasses the westernmost portion of the offshore forearc (Figure 3.3). Most of this region is dominated by east-west compression due to the subduction of the Gorda plate, which is evidenced by numerous north-south trending thrust faults and folds (Clarke, 1992; Gulick & Meltzer, 2002; Watt & Brothers, 2020). However, farther to the east, deformation in the forearc is more transpressional due to the northward propagation of the SAF system (Gulick & Melzer, 2002). Therefore, we choose the eastern boundary of the OFFS block to coincide with this transition between compression near the trench and transpression farther east. This eastern boundary is defined by a 45° eastward-dipping pseudo-fault, which we refer to as the ‘offshore fault’. To the east of this boundary, we extend the onshore forearc faults offshore and have them curve to the north to eventually connect up with the offshore fault. This northward bending of the offshore forearc faults is consistent with previous offshore seismic mapping (Clarke, 1992; Gulick & Meltzer, 2002).

We model the southern end of the Cascadia megathrust fault as the eastern boundary of the GORD block. We assume that slow slip and continuous creep at depths greater than 40 km along the megathrust do not contribute any net long-term surface deformation. Therefore, we only model the shallowest 40 km of the megathrust based on the fault geometry from Blair et al. (2013). We note that the eastern border of the GORD block, which represents the surface boundary between the Gorda and North American plates, does not perfectly align with the deformation front delineated by bathymetric data. This inconsistency reflects a simplified geometry along the shallow megathrust fault and is a result of the limited amount of offshore seismic data available to constrain the model (McCrory et al., 2012).

3.3.3 Fault Locking

Each fault consists of rectangular fault patches that are defined in 3-D space by a set of nodes. These nodes are equally spaced at 5 to 10 km along strike and are placed every 5 km at depth. We construct each fault down to a predetermined locking depth adapted from UCERF3 (Field et al., 2013), Murray et al. (2014), and Lienkaemper et al. (2014), and assume that the fault is fully slipping below this depth. Locking is controlled at each fault node by the ϕ parameter, which varies between 0 (fully slipping) and 1 (fully locked). We force all nodes at a given depth on a given fault to have the same ϕ value. We then allow the model to solve for locking at all depths, with the exception of the offshore and megathrust faults. We force ϕ to be zero at all depths along the offshore fault, so as to avoid possible trade-offs with slip deficit on the megathrust fault. For the megathrust fault, we impose full locking ($\phi=1$) at the up-dip edge, and then linearly decrease ϕ to fully unlocked ($\phi=0$) at 35 km depth. We assume this megathrust locking structure for our initial model runs, but explore various other locking scenarios in Section 4.1.

3.3.4 Fault Slip Rate and Sense

To ensure that the San Andreas, Maacama, and Bartlett Springs faults conform with previously published estimates by Field et al. (2013) and Murray et al. (2014), we constrain the slip rates on these faults by penalizing modeled slip rates with high residual values. The slip rates are penalized according to:

$$penalty = \left(\frac{\frac{(v_2 + v_1)}{2} - v_{model}}{(v_2 - v_1)/2} \right)^2 \quad (3.1)$$

where v_2 is the maximum reported velocity, v_1 is the minimum reported velocity, and v_{model} is the modeled velocity. We provide maximum slip rates of 24, 14, and 8 mm/yr and minimum slip rates of 21, 9, and 6 mm/yr for the San Andreas, Maacama, and Bartlett Springs faults, respectively (Field et al., 2013; Murray et al., 2014). We also give a maximum slip rate of 1 mm/yr for the offshore fault to avoid excessively high slip rates on this inferred fault. Due to the absence of stations on the OFFS block, we also apply additional constraints to this portion of the model. Specifically, we set the model so that it applies a severe penalty according to McCaffrey

(2009) if 1) the OFFS block does not move relative to the GORD block at an azimuth of 180-300°, 2) the offshore fault has any component of normal slip or 3) the portion of the Mendocino Transform fault along the southern edge of the OFFS block does not exhibit right lateral strike-slip motion. Lastly, we apply the same severe penalty to ensure that the Eaton Roughs fault zone exhibits right-lateral slip, the MRFZ and Little Salmon fault exhibit reverse slip, and the Grogan fault exhibits slip that ranges from thrust to right lateral.

3.3.5 Block Poles and Internal Strain

We fix the Euler pole and rotation rate for the Pacific (PAC) block according to Argus and Gordon (2001). Since there is currently no established rotation pole for the Gorda plate, we assign the Juan de Fuca - North American pole and rotation rate from Wang et al. (2003) to the GORD block. The poles and rotation rates for the remainder of the blocks are solved for by the model. We also allow the model to solve for uniform internal strain within all of the blocks except for the PAC, GORD, and OFFS blocks. Although strain is likely occurring in all three of these blocks, the lack of offshore instrumentation in these regions does not allow for any resolution of internal strain in these regions. Additionally, we do not allow strain within the block that is bounded by the MRFZ and Grogan fault in geometries 6 and 8, since model runs allowing for strain in this block produce unrealistic results, such as unfeasibly high extensional strain rate values. We attribute this inconsistency to the fact this block only contains two GNSS stations, making it difficult to resolve strain.

3.3.6 Modeling Approach

We run the model for each block geometry starting with $\phi=0.5$ for all of the faults except for the megathrust and offshore faults. In order to solve for the unknown parameters, the model uses the down-hill simplex algorithm followed by a grid search. Once the first model run is complete, we then run the model again using the output parameters from the first model run as the input parameters for the second model run. We iterate this process until the model has been run ten times to ensure that all of the model parameters converge.

3.4 RESULTS

The initial results provide block motions, block strains, and fault slip rates for all of the eight block geometries. However, these results incorporate an *a priori* assumption about the distribution of locking along the megathrust. Therefore, we assess these initial results to determine which model geometries are the most successful and warrant further exploration with different distributions of ϕ on the megathrust. Firstly, we evaluate the models by calculating their respective χ^2 misfit values according to:

$$\chi^2 = \sum_{i=1}^{n_{data}} \frac{(v_{obs}(i) - v_{model}(i))^2}{\sigma(i)^2} \quad (3.2)$$

where σ is the standard deviation of the velocity observations, n_{data} is the number of velocity observations, and v_{obs} and v_{model} are the data and modeled velocities, respectively. A comparison of these misfit values in Table 1 shows that geometry 1 does the worst job of fitting the data, with a χ^2 value that is an order of magnitude larger than all of the other geometries. Geometries 2-4, which include only the Little Salmon fault, MRFZ, or Grogan fault individually, generally perform worse than geometries 5-8, which include some combination of these faults. However, geometry 6 is an exception, as it has a higher misfit value than geometries 2 and 3. Overall, these results suggest that the simpler geometries perform worse than the more complicated geometries. However, it is possible that the increased complexity of geometries 5-8 (more blocks, more faults, and more overall parameters) allow for overfitting of the data. Therefore, we also consider the Bayesian Information Criterion (BIC) (Schwarz, 1978) to address parameter variability between models. The BIC takes both model fit and model complexity into account, allowing for the direct comparison of models with varying numbers of model parameters. Ultimately, the model geometry associated with the lowest BIC score is considered the best. We calculate the BIC score for each model geometry using a modified BIC formula adapted from Spinler et al. (2010):

$$BIC = N \ln \left(\frac{\chi^2}{N} \right) + k \ln (N) \quad (3.3)$$

where N is the degrees of freedom, χ^2 is the chi-squared statistic defined by equation 3.2, and k is the number of model parameters. These BIC scores are listed in Table 1 and show similar results as the χ^2 misfit values. In particular, geometry 1 has the highest BIC score, geometry 7 has the lowest BIC score, and geometries 2-4 have higher BIC scores than geometries 5, 7, and 8. This consistency between the two metrics supports the notion that more appropriate fault geometries, not additional parameters alone, contribute to the higher fit of models using geometries 5, 7, and 8.

In terms of model selection, geometries 5, 7, and 8 have the three lowest χ^2 values and BIC scores, indicating that they are the three best geometries for fitting the data. Additionally, all three of these model geometries include at least two of the MRFZ, Little Salmon and Grogan faults, which provides insights into possible tradeoffs between these fault structures. However, without geometry 6, these three geometries do not explore all possible combinations of these three forearc faults. Therefore, we opt to include geometries 5-8 in the following megathrust and bootstrap analyses, since these four geometries allow us to test the full spectrum of forearc fault combinations.

3.4.1 Optimization of Megathrust Locking Parameters

We repeat the steps outlined in Section 3.6 for geometries 5-8 with different megathrust locking scenarios to determine which distribution of the locking fraction ϕ best fits the data. We adjust the updip limit of full locking and the down dip limit of full slip for each geometry to evaluate a range of locking depths. Through this analysis, we find that an updip limit of 10km depth for complete locking ($\phi=1$) and a downdip limit of 25km depth for zero locking ($\phi=0$) fits the data best for all four of the geometries (Figure A2.4). Therefore, for our final model runs, we employ a locking pattern wherein the megathrust is fully locked from the trench down to 10 km depth, with locking then linearly decreasing to zero at 25 km depth (Figure 3.4).

3.4.2 Bootstrap Analysis

The block models presented in this work run the risk of being biased by the GNSS network geometry, specific site velocities, and the choice of block geometry. Therefore, we conduct a bootstrapping analysis to evaluate the models' sensitivity to the GNSS velocity data and determine the full range of possible fault slip rate values. To do this, the GNSS velocity data

are randomly resampled so that at least 80% of the original stations are retained. This resampling technique is repeated to produce 400 unique GNSS velocity datasets. Each reduced dataset is then used to estimate the model parameters for geometries 5-8, following the steps outlined in Section 3.6 and using the megathrust parameters found in Section 4.1.

3.4.3 Bootstrap Results

After performing the bootstrap analysis for geometries 5-8, we are left with 400 results for each block model geometry. We calculate the χ^2 misfit and BIC score for each geometry's 400 results and evaluate the model geometries according to their average value for each metric (Figure A2.4). In the end, geometry 6 has the highest χ^2 misfit and the highest BIC score, meaning that it consistently performs the worst out of all four of the block geometries. In contrast, geometry 8 has the lowest χ^2 misfit and geometry 5 has the lowest BIC score, leading to a discrepancy between the two metrics and which geometry is “best”. This discrepancy stems from the fact that the BIC score incorporates the number of parameters in each model, while the χ^2 misfit does not. Since geometry 8 includes all three forearc faults and requires the largest number of parameters, it experiences the greatest penalty for model complexity and receives a relatively high BIC score. However, we believe that these additional parameters are warranted, since activity on all three of the forearc faults is evidenced by geologic work (e.g. Kelsey & Carver, 1988; Clarke & Carver, 1992; McCrory 2000) and regional strain rate analyses (Figure 3.2). Therefore, geometry 8 is our preferred geometry, as it exhibits the lowest average χ^2 misfit and incorporates all three forearc faults.

Inspection of modeled velocities also shows that geometry 8 is able to successfully capture the overall velocity field in the MTJ region (Figure 3.5). Visual examination of the modeled velocities from individual bootstrap runs shows that the pattern of velocities is fairly consistent and that the average is representative of the total 400 runs. Therefore, we present the average modeled velocity and residual value for each GNSS station in Figure 3.5. Aside from a group of stations in the center of the easternmost block with coherent westward motions, we do not observe any significant spatial patterns in the residuals for geometry 8. Similar residual maps for geometries 5, 6, and 7 can be found in the Supplementary Information.

3.4.3.1 *Fault Slip Rates*

In the following sections, we report the slip rate distributions for the onshore portions of the Little Salmon Fault, MRFZ, and Grogan fault from the bootstrap analysis. Similar slip rate results for other crustal faults can be found in the Supplementary Information. Although the majority of the bootstrap slip rate estimates resemble a Gaussian distribution, there are several instances wherein the distributions are left or right skewed. For these cases, the mean is not a representative value, as it is biased by the tails and does not reflect the peak of the distribution. In contrast, the median of a skewed distribution still tends to align with the peak, as it is not impacted by outlier values. Therefore, we present the median slip rate value, as opposed to the mean slip rate value, to more fully represent the slip distributions. Additionally, we provide the range of slip values that encompass the middle 68% of the distribution. This method allows us to provide slip rate ranges that are similar to the common 1-sigma approach, while still incorporating the skew of certain distributions. However, it is important to note that for cases with nearly normal distributions, our results are similar to the equivalent mean $\pm$ 1-sigma results. Slip rate distributions for geometries 5-8 are provided in Figures 3.6-3.8 and Tables 2-4. These results describe the long-term slip rates along the shallow portions of the faults, which are also representative of the long-term slip rates at depth. We highlight the slip rates for geometry 8 in these tables to emphasize that these are our preferred slip rates. We also summarize the slip rate results for geometry 8 in Figure 3.9, which shows a map view of the median slip rate results from the bootstrap analysis. Together, these figures provide context for how slip rate predictions vary along strike and offer location-based estimates for comparison with future paleoseismic studies. Lastly, we present median distributed slip deficit rate results for the megathrust fault.

3.4.3.2 *Little Salmon Fault*

Slip rate results for the Little Salmon fault are displayed in Figure 3.6 and Table 2 for geometries 5, 7, and 8. All three of the geometries show a component of reverse and right-lateral strike-slip motion on the Little Salmon fault, and they all share similar patterns in terms of along-strike slip rates. In particular, the thrust component of slip increases along the fault from east to west until longitude -124.17° to -124.1°, at which point it begins to decrease towards the coast. In contrast, the strike-slip component of slip decreases from east to west until longitude -123.09°, at which point it begins to increase towards the coast.

3.4.3.3 *Mad River Fault Zone*

Slip rate results for the MRFZ are displayed in Figure 3.7 and Table 3 for geometries 6, 7, and 8. All three of the geometries show components of reverse and right-lateral strike slip on the MRFZ, with the strike slip component comprising the majority of the total slip rate. All of the geometries also show fairly consistent right-lateral slip rates along strike, with values increasing slightly from east to west. For reverse slip, values remain fairly steady throughout the length of the fault for geometry 8, while geometries 6 and 7 show slip values increasing from east to west.

3.4.3.4 *Grogan fault*

Slip rate results for the Grogan fault are displayed in Figure 3.8 and Table 4 for geometries 5, 6, and 8. All three of the geometries show that the reverse slip component dominates the total slip rate, with the strike-slip component being near zero. All of the geometries also show the reverse component of slip increasing monotonically from east to west, while the strike-slip component remains fairly constant along strike.

3.4.3.5 *Additional Crustal Faults*

The Bear River fault zone (BRFZ) has the highest slip rate results (~ 28 mm/yr) of any crustal fault in terms of both the reverse and right-lateral slip components (Figure 3.9, A2.11; Table A2.6). All four of the geometries show the right-lateral component of slip on the BRFZ increasing from east to west, while the dip-slip component transitions from normal slip in the east to reverse slip in the west. In our models, the Garberville fault zone connects to the southern end of the BRFZ and displays right-lateral slip rates of ~ 17 mm/yr (Figure A2.9; Table A2.4).

3.4.3.6 *Megathrust Fault*

Figure 3.10 shows the median slip deficit rate at all nodes along the megathrust fault for geometries 5-8. All four geometries share a fairly similar slip deficit rate structure at depth, with relatively low values (10-15 mm/yr) reaching the coast and even extending slightly offshore below latitude 41° . All of the geometries also show relatively high slip deficit rates (>35 mm/yr) concentrated along the portion of the Mendocino transform fault that separates the OFFS and PAC blocks. North of this region, offshore patterns of slip deficit rate begin to differ between geometries. For example, geometry 5 shows a relatively smooth transition from higher values in

the south to lower values in the north, while geometry 6 shows a more abrupt change north of latitude 41°.

3.5 DISCUSSION

Our bootstrap analysis provides slip rate distributions for various faults in the MTJ region. For this analysis, we test four different geometries that each incorporate different combinations of the Little Salmon fault, MRFZ, and Grogan fault. After analyzing the bootstrap results, we conclude that geometry 8 is our preferred geometry, since it yields relatively low misfit values and incorporates all three forearc faults that align with zones of high strain rates in Figure 3.2. In the following sections, we discuss the important limitations and assumptions associated with our models and interpret the modeling results in terms of deformation and seismic hazards along crustal forearc faults. Lastly, we examine the pattern of slip deficit rate along the megathrust fault and consider how it relates to processes along the plate interface. These results are mainly discussed in the context of geometry 8; however, we also explore certain tradeoffs between other forearc geometries.

3.5.1 Model Limitations and Assumptions

Block models provide important insights into regional deformation patterns and fault characteristics. However, there are certain limitations and assumptions associated with these models that must be considered. Firstly, block models are not able to reproduce crustal deformation resulting from mantle-lithosphere interactions. For example, crustal thickening and thinning predicted by the crustal conveyor model is not incorporated into the block model framework (Furlong & Govers, 1999). As a result, distributed deformation in the MTJ region stemming from dynamic mantle processes is not directly captured by the models in this study. Secondly, these models incorporate simplified fault geometries with relatively homogenous fault properties. Therefore, they are unable to model spatial variations in the deformation field stemming from possible fault and rock heterogeneities along strike, such as changes in the dip, locking depth or elastic modulus. These models also do not consider fault-normal heterogeneities in the surrounding rheology, such as the presence of a compliant fault zone, which is characterized by a reduced shear modulus near the fault plane (Barbot et al., 2008). Lastly, the simplified fault geometries in our models lead to certain blocks having large aspect ratios, which

may cause biased surface deformation near the block extremities. Together, these factors likely contribute to un-modeled residual values as shown in Figure 3.5C.

The simplified fault geometries used in block models can also lead to trade-offs between localized and distributed deformational processes. For example, numerous unmodeled crustal faults and folds that accommodate notable amounts of contraction in the forearc are located within the blocks in our models (Kelsey & Carver, 1988; McCrory, 2000; Padgett et al., 2019). Ideally, distributed deformation on these systems is incorporated into the internal block strains. However, the proximity of some of these structures to the block boundaries can make it difficult for the models to discern whether deformation is due to slip on a fault boundary or is attributed to a nearby, unmodeled structure. As a result, these models run the risk of projecting distributed strain within the block onto nearby block boundaries as fault slip.

Finally, the block models in this study are constrained by a GNSS network that can be relatively sparse between certain forearc faults. As a result, certain blocks contain only a few GNSS stations and the free parameters for these blocks may not be well resolved. Additionally, the GNSS velocities in this study are representative of short-term crustal deformation over the past several decades, which is much shorter than the average earthquake cycle. This is in contrast to previous geologic work in the MTJ region, which leverages processes and observations that span much longer timescales. Therefore, it is worth noting that the findings of this work build upon observations that reflect fairly recent deformational patterns.

3.5.2 Forearc Faults

3.5.2.1 *Little Salmon*

Block modeling results from this work indicate that the Little Salmon fault plays an important role in facilitating deformation in the forearc. Firstly, geometries that do not include the Little Salmon fault tend to do a poorer job of fitting the data. For example, geometry 6 performs the worst out of the four bootstrapped geometries, with the highest χ^2 misfit and BIC scores (Figure A2.4). Secondly, out of the three forearc faults, the highest total slip rates are placed on the Little Salmon fault. In particular, the Little Salmon fault experiences higher reverse slip rates than the MRFZ, especially towards the western end of the fault (Figure 3.6D). Although both of these fault systems are regarded as predominant thrust structures in the forearc

(Kelsey & Carver, 1988), this result suggests that the Little Salmon fault plays a more significant role in accommodating compression in the crust.

In addition to reverse slip, the block models also place a component of right-lateral slip on the Little Salmon fault. Geologic evidence for strike-slip motion on the Little Salmon fault has not been observed; however, right-lateral slip has been noted on the Goose Lake fault directly to the south of the Little Salmon fault (Woodward-Clyde Associates, 1980). Therefore, although the block models place a combination of thrust and shear slip on the Little Salmon fault, these two components of slip may be partitioned on different strands within the broader Little Salmon fault system (Ladinsky et al., 2020). However, regardless of the structure on which the slip is occurring, our models suggest that this region of the forearc is facilitating some degree of right-lateral shear.

Although we choose to interpret the deformation along the Little Salmon fault in the context of geometry 8, it is worth noting that the dominant component of slip on the Little Salmon fault varies across the model geometries. Specifically, geometries 5 and 7 prefer a higher right-lateral slip component, while geometry 8 prefers a higher reverse slip component. This variability highlights how slip can be transferred between the different forearc faults depending on the model geometry. For example, the Little Salmon fault has higher slip rates when the MRFZ is not present in geometry 5, since the Little Salmon fault is forced to accommodate the deformation that would otherwise be shared by these two fault systems. Therefore, variability in the slip rate results across the different geometries for the Little Salmon fault can be attributed to logical tradeoffs in slip between the three forearc faults and distributed strain within the blocks.

3.5.2.2 *Mad River Fault Zone*

Previous geologic work provides evidence for reverse slip throughout the MRFZ and emphasizes the role that this system plays in facilitating forearc contraction. Slip rate estimates from our work confirm a component of reverse slip along the MRFZ block boundary; however, relatively high right-lateral slip rate results indicate that this system also accommodates translation in the forearc. In particular, the right-lateral component of slip on the MRFZ is nearly 8x larger than the reverse component of slip. Elevated strike-slip values are consistent with our regional strain rate analysis, which shows the MRFZ falling within a zone of relatively high shear strain rates (Figure 3.2C). Additionally, the northernmost portion of the MRFZ, which

includes the Trinidad fault, merges with the Eaton Roughs fault zone (ERFZ) (Padgett et al., 2019) (Figure 3.1B). The ERFZ is considered part of the northern SAF system and accommodates right-lateral strike-slip motion in the forearc (Kelsey & Carver, 1988). Therefore, faults in the MRFZ may represent northern extensions of the ERFZ and SAF system.

It is worth noting that the block models in this study use a simplified geometry for the MRFZ, wherein the Trinidad fault serves to represent the entire system. The remaining four faults in the MRFZ fall to the south of this block boundary, and the slip on each of these faults is either captured as internal block strain or projected onto the Trinidad fault boundary. Therefore, it is unclear what proportion of the total fault slip within the MRFZ is being captured by the block models in this study. Additionally, the relatively high right-lateral slip rate results for the MRFZ block boundary may not be representative of all of the faults in the system. For example, strain partitioning within this 15 km-wide fault zone may cause some of the faults to experience more right-lateral slip than others. Nevertheless, our results, together with ample evidence of reverse motion on the MRFZ (e.g. Clarke & Carver, 1992), suggest that this is an overall transpressional system.

3.5.2.3 *Grogan Fault*

The translation of corollary units across the Grogan fault suggests that it hosted up to 75 km of right-lateral slip in the Late Tertiary (Kelsey & Hagans, 1982). However, slip rate results from our block models indicate that the Grogan fault accommodates very little contemporary strike-slip motion, suggesting that this fault no longer facilitates large amounts of translation within the forearc (Figure 3.8B). Additional evidence for decreased strike-slip motion on the Grogan fault is provided by Pliocene-Pleistocene sediments that overlie the fault near the coast. Although they provide no constraints on vertical slip, these sediments exhibit negligible signs of lateral motion and suggest that dextral slip on the Grogan fault terminated around 2 Ma (Kelsey & Hagans, 1982). Therefore, the Grogan fault may represent a reactivated strike-slip feature that now experiences reverse slip. Previous geodetic work proposes that the Grogan and other parallel fault structures undergo dextral slip due to the translation of the rigid Klamath terrane to the north (McKenzie & Furlong, 2021). However, our results suggest that this translation may be accommodated more so by the faults in the MRFZ region to the south than by the Grogan fault.

3.5.2.4 *Bear River & Garberville Fault Zones*

The high right-lateral and reverse slip rates that the models estimate on the Bear River fault zone (BRFZ) suggest that this region of the forearc is undergoing significant deformation. High strain rates near the BRFZ are supported by our regional strain rate analysis, which shows a band of high shear strain rates encompassing the San Andreas fault (SAF), BRFZ, and Garberville fault zone (Figure 3.2C). This band of high shear strain rates broadens south of latitude 40° to cover the three main strands of the northern SAF system. This transition from a relatively narrow (~30 km) to a relatively wide (>60 km) band of high shear strain rates may indicate a redistribution of strain within the northern SAF system. Specifically, high shear strain rates diminish on the Bartlett Springs fault north of latitude 40° and may signal that strain is being transferred onto the neighboring BRFZ and Garberville fault zone, and eventually offshore to the Mendocino transform fault. Our models show right-lateral slip rates increasing on the Garberville fault zone from south to north (Figure 3.9, A2.9), which supports the idea that slip is gradually transferred onto the Garberville fault zone and then propagated onto the BRFZ. Therefore, high slip rates on the Bear River and Garberville fault zones, along with high strain rates in the surrounding region, may result from strain being concentrated along fewer structures in the northern SAF system near the MTJ.

Although geologic fault slip rates are currently not available for the BRFZ, evidence of thrust and right-lateral slip in this region is provided by the nearby Russ fault. Located just north of the BRFZ, the Russ fault offsets multiple marine terrace platforms along the coast, and age estimates from these platforms provide an uplift rate of 0.2 mm/yr (Kelsey & Carver, 1988; McCrory, 2000). Additionally, a ~10 km offset in the continental shelf aligns with the offshore extension of the Russ fault and is compatible with right-lateral motion across the fault (McCrory, 2000). Active deformation in this region is also evidenced by the rapid uplift of the King Range, a terrain along the coast that has experienced upwards of 4 mm/yr of Holocene uplift (Merritts & Bull, 1989). Right-lateral slip on the Garberville fault zone is inferred from its alignment with the Maacama fault to the south, but there is no geologic evidence of Holocene activity on this fault (Kelsey & Carver, 1998). However, geodetic analysis shows that surface velocities in the MTJ region are consistent with 14 mm/yr of right-lateral slip on the Garberville-Maacama system (Williams et al., 2006).

3.5.2.5 *Seismic Hazards*

The occurrence of multiple earthquakes on the Little Salmon fault and MRFZ during the Late Holocene suggests that these faults are capable of hosting future earthquakes and pose a future seismic hazard. Indeed, regional strain rate analysis shows that these faults coincide with regions of relatively high strain rates (Figure 3.2), indicating that crustal deformation is being accommodated on these faults. Furthermore, block model results from this work provide non-zero slip rates on the Little Salmon fault and MRFZ, providing further evidence that these structures are actively facilitating forearc deformation.

We estimate the earthquake recurrence interval for the Little Salmon fault, using slip offset values from previous studies and slip rate estimates from our preferred model. We only consider the thrust component of slip in these estimates, since previous work only details the vertical offsets along the fault. Assuming a range of 1.3 – 4.5 m for slip offset (Clarke & Carver, 1992; R.C. Witter, personal communication, 2022) and using the median slip rate of 4 mm/yr from our analysis, we estimate a recurrence interval of 325 - 1125 years on the primary strand of the Little Salmon fault. Estimating earthquake recurrence intervals for the MRFZ is more challenging, because this fault system consists of five major strands. Since our block model geometry is only based on the Trinidad fault, it is unclear if our slip rate results represent the entire system or just this one fault. Additionally, slip offset estimates from past earthquakes are not available for several faults in the MRFZ and it is unclear whether all of the faults in this system rupture simultaneously. Due to the uncertainty of our models and the paleoseismic history of the MRFZ, we do not provide a recurrence interval for the MRFZ. We also do not provide a recurrence interval for the Grogan fault, since there are currently no observations of discrete slip events on this fault.

3.5.3 *Megathrust Slip Deficit Rate*

The slip deficit rate is the product of the locking fraction (ϕ) and the estimated fault slip rate. The bootstrap results for slip deficit rate along the megathrust fault are largely controlled by the optimized locking pattern found in Section 4.1, which includes zero locking below 25 km depth (Figure 3.4). This depth roughly coincides with the updip location of tremor in southern Cascadia (Figure 3.11A), which can be used as a spatial proxy for short-term slow slip events (SSEs) along the megathrust (e.g. Wech et al., 2009). Analysis of GNSS time series data

indicates that short-term SSEs take up 100% of the slip budget on the megathrust at ~35km depth in southern Cascadia, assuming a convergence rate of 32 mm/yr (Bartlow, 2020). Therefore, no locking and zero slip deficit rate below 25 km depth in our models is consistent with short-term SSEs facilitating the entire slip budget during the time spanned by the GNSS velocity measurements. Directly updip of the tremor and short-term SSE region, our results show slip deficit rates remaining fairly low (5-10 mm/yr). It is less clear how plate convergence is facilitated on this portion of the plate interface; however, evidence for various slip processes has been observed in this region (Figure 3.11A). For example, dynamic changes in plate coupling and long-term slow slip have been inferred from transients in the GNSS time series data near the MTJ (Materna et al., 2019; Nuyen & Schmidt, 2021). Additionally, a patch of near-continuous tremor on the southern edge of the Gorda slab suggests that there may be near-continuous slip occurring on that portion of the plate interface (Wech, 2021). Therefore, a combination of slip processes is likely contributing to the relatively low slip deficit rates that we observe directly updip of tremor and short-term slow slip.

At shallower depths of 10-20 km, the overriding forearc faults begin to influence the pattern of slip deficit rate along the megathrust (Figure 3.10). We analyze the impact of these forearc faults by comparing the slip deficit rates between the two end-member block model geometries: 1 and 8 (Figure 3.11B). This is accomplished by differencing the slip deficit rates of the two models, where the locking parameters have been set according to Section 4.1 and all of the GNSS velocity data have been used to estimate the model parameters. Any variability in slip deficit rates owing to the motion of the OFFS block is also removed, since this block is relatively unconstrained. Comparing these two results shows that the presence of active faults in the forearc decreases the nearby slip deficit rates along the megathrust by up to several mm/yr (Figure 3.11B). This tradeoff between the model geometries suggests that slip on the Little Salmon fault, MRFZ, and Grogan fault removes slip from the megathrust fault by accommodating plate convergence in the crust. As a result, megathrust locking models that do not incorporate slip on forearc faults may be overestimating the amount of locking along the coast in southern Cascadia. Additionally, this transfer of slip may mean that seismic hazards in this region are shifted away from the megathrust and onto crustal forearc faults, as suggested by McCrory (2000). Seismic moment outputs from TDEFNODE indicate that slip on the forearc faults, together with internal block strain, removes 7.1×10^{17} N·m/yr from the megathrust. Using

the moment magnitude scale from Hanks & Kanamori (1979), this is equivalent to a Mw 5.9 earthquake per year.

Relatively high slip deficit rates along the trench indicate that the shallowest portion of the megathrust is fully locked, assuming a convergence rate of 32 mm/yr (McCaffrey et al., 2007). Along the Mendocino transform fault, we observe a band of especially high slip deficit rates that exceed the expected convergence rate. These high values are an artifact that result from interactions between the Gorda and Pacific plates and do not contribute significant amounts of surface deformation to the overall model. Although full locking at the trench is consistent with previous megathrust locking studies (e.g. Li et al., 2018), we are not confident in the details of the offshore locking patterns presented in this work for several reasons. In particular, the slip deficit patterns along the shallowest portion of the megathrust fault are controlled by the overriding OFFS block. This block represents a simplified offshore geometry and does not incorporate structural changes that occur near the trench (Watt & Brothers, 2020). Additionally, a lack of offshore instrumentation means that the model parameters for this block are not well constrained. As a result, internal strain is not allowed in the OFFS block, and the rotation parameters have relatively high uncertainties. Therefore, the assumptions and uncertainties associated with the OFFS block make it difficult to confidently interpret the offshore locking results.

3.6 CONCLUSIONS

We constructed detailed block models of the Mendocino triple junction region to provide insights on how strain is accommodated along crustal faults in the southern Cascadia forearc. Through a bootstrap analysis, we provided a distribution of slip rates for the Little Salmon fault, Mad River fault zone (MRFZ), and Grogan fault for various forearc block geometries. Our results indicate that the Little Salmon fault accommodates compressional and translational strains within the forearc; however, these two components may be partitioned along different fault strands within the broader Little Salmon fault system. Slip rate results for the MRFZ show relatively high right-lateral slip rates, suggesting that the MRFZ is a transpressional system that may serve as a northern extension of the northern San Andreas fault system. However, the resolution of our models is not able to determine whether these slip rates are representative of the entire MRFZ or specific faults within the system. In contrast, the Grogan fault has relatively high

reverse slip rates and does not facilitate much strike-slip motion. Therefore, the Grogan fault may represent a reactivated strike-slip feature that now experiences reverse slip.

3.7 ACKNOWLEDGEMENTS

We would like to thank Kathryn Materna, an anonymous reviewer, the associate editor, and the editor at JGR for their valuable feedback. We would like to thank Harvey Kelsey and Janet Watt for sharing their insights about deformation and active faulting in southern Cascadia. We would also like to thank the PANGA processing center and PBO for providing the GNSS and velocity data for this work. Research supported by the U.S. Geological Survey (USGS), Department of the Interior, under USGS award number (G21AP10043). The views and conclusions contained in this document are those of the authors and should not be interpreted as necessarily representing the official policies, either expressed or implied, of the U.S. Government.

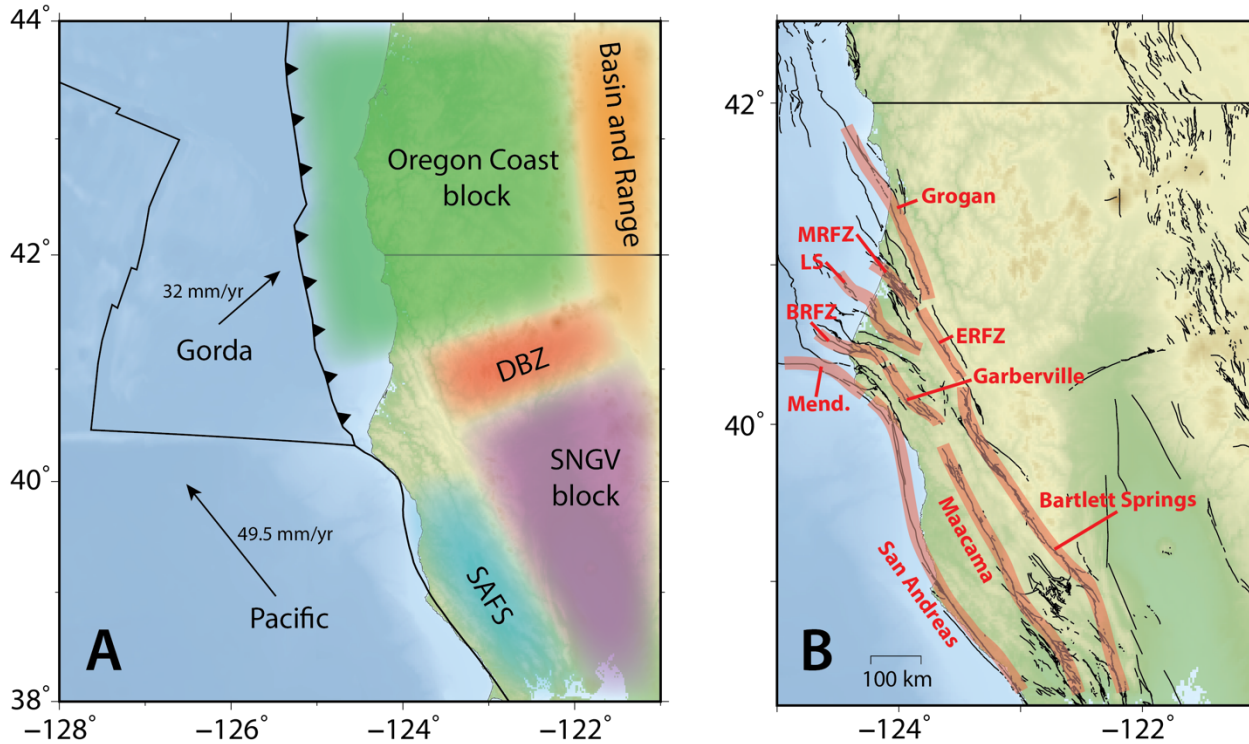


Figure 3.1 Maps of the Mendocino triple junction region. **A** Tectonic map where black lines delineate the boundaries between the Pacific, Gorda, and North American plates. Colored regions designate different deformational regimes from Williams et al. (2006): DBZ - diffuse boundary zone, SAFS - San Andreas fault system, SNGV - Sierra Nevada-Great Valley. Black vectors show plate motions relative to North America, with velocity values from Argus & Gordon (2001) and McCaffrey et al. (2007). **B** Fault map where black lines denote Quaternary faults, and red shading and labels identify important structures relevant to this work: BRFZ - Bear River fault zone, ERFZ - Eaton Roughs fault zone, LS - Little Salmon, MRFZ - Mad River fault zone, Mend. - Mendocino transform fault.

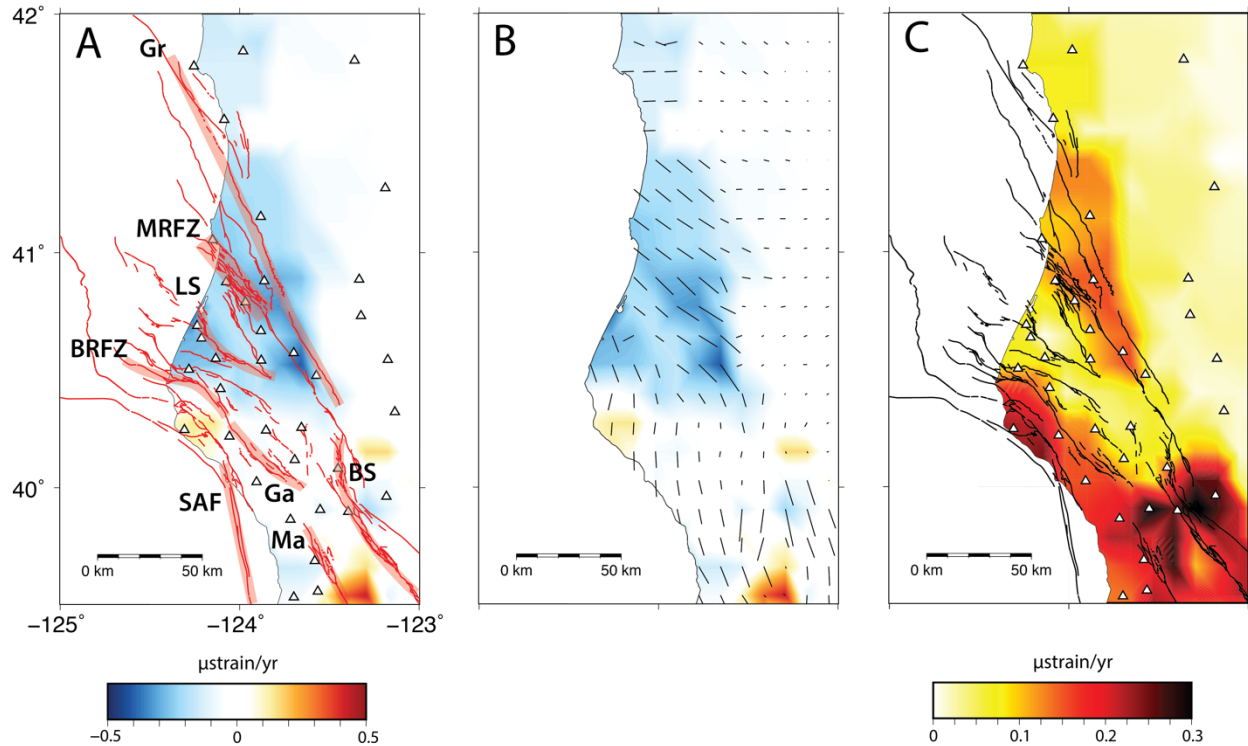


Figure 3.2 Regional strain rate maps for the Mendocino triple junction. **A** Map of the first invariant of strain rate, where colder and warmer colors denote areas of contraction and extension, respectively. Red lines outline Quaternary faults and white triangles show GNSS station locations. Red shading outlines labeled faults: BRFZ - Bear River fault zone, BS - Bartlett Springs, Ga - Garberville, Gr - Grogan, LS - Little Salmon, Ma - Maacama, MRFZ - Mad River Fault Zone, SAF - San Andreas. **B** Map of the first invariant of strain using the same scale as panel A, with black bars showing the azimuth of the most compressive principal strain rate. Bars are located at grid nodes and are scaled by magnitude. **C** Map of maximum shear strain rate, where darker and lighter colors denote areas of high and low shear strain rates, respectively. Black lines outline Quaternary faults and white triangles show GNSS station locations.

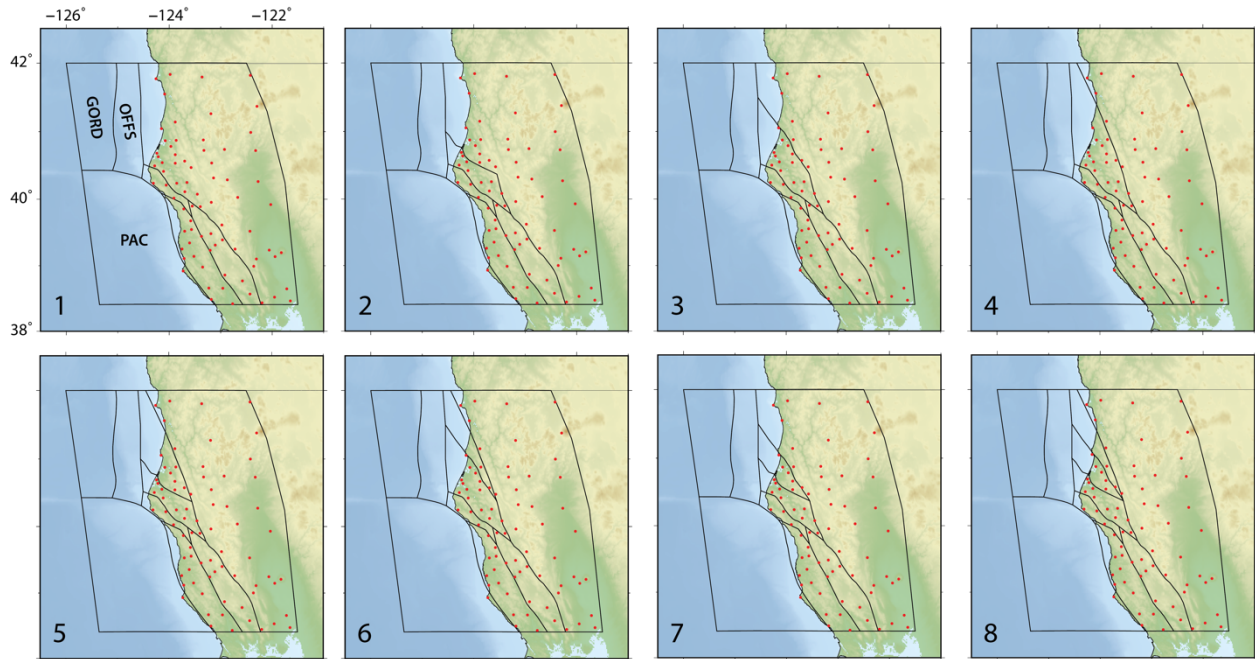


Figure 3.3 Maps of the eight different block geometries tested in this work. Black lines outline the block boundaries with the geometry number shown in the bottom left-hand corner of each panel, and red dots show the location of GNSS stations. Geometry 1 shows the names of the offshore blocks: GORD - Gorda, OFFS - offshore, PAC - Pacific.

Geometry	χ^2	BIC	N	k
1	19256	766	107	45
2	6992	667	97	55
3	6891	668	93	59
4	8868	690	97	55
5	6236	662	87	65
6	7126	674	83	69
7	6030	661	83	69
8	6410	666	73	79

Table 3.1 Table containing model misfits (χ^2), BIC scores, degrees of freedom (N), and number of model parameters (k) for each of the fault geometries shown in Figure 3.3.

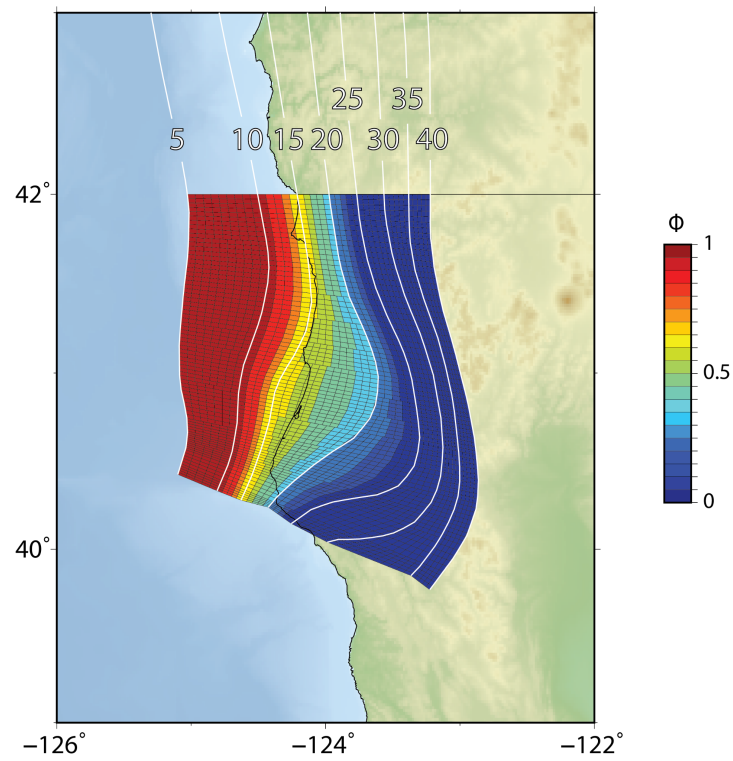


Figure 3.4 Map showing the optimal distribution of locking fraction on the megathrust fault that is prescribed in the block model. Colors indicate the locking fraction, ϕ , white lines and corresponding labels delineate plate depth contours in km from Blair et al. (2013).

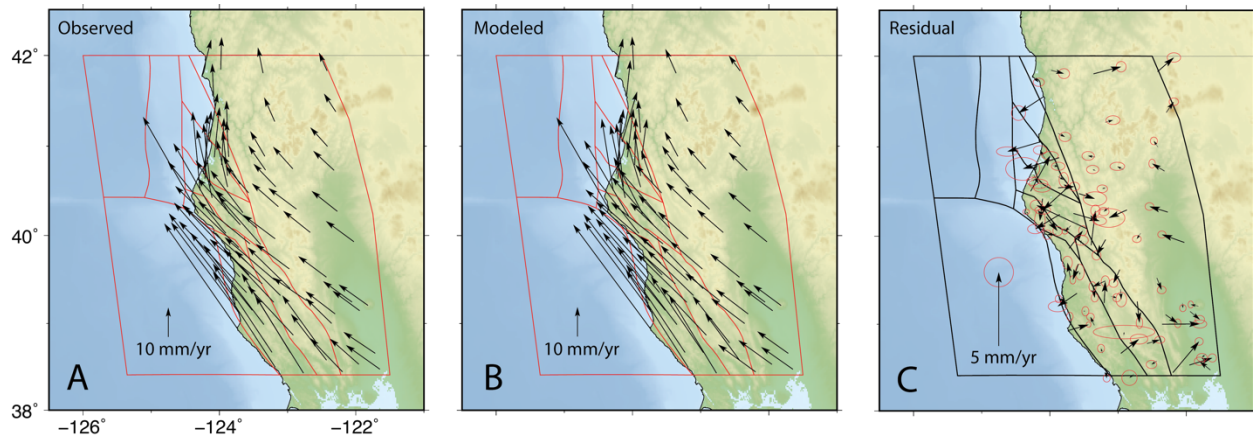


Figure 3.5 Maps of observed and modeled GNSS surface velocities in southern Cascadia. **A** Vector map showing the observed GNSS velocities from PANGA used to constrain the block models in this work (PANGA, 1996). Red lines outline the block boundaries for geometry 8. **B** Vector map showing the average modeled GNSS velocities from the bootstrap analysis for geometry 8. Red lines outline the block boundaries for geometry 8. **C** Vector map showing the average residuals between observed and modeled GNSS velocities from the bootstrap analysis for geometry 8. Red ellipses represent 2-sigma uncertainty, with the scale vector having a 1 mm/yr error ellipse. Black lines outline the block boundaries. Relatively small uncertainties arise from long decadal-length GNSS time series data.

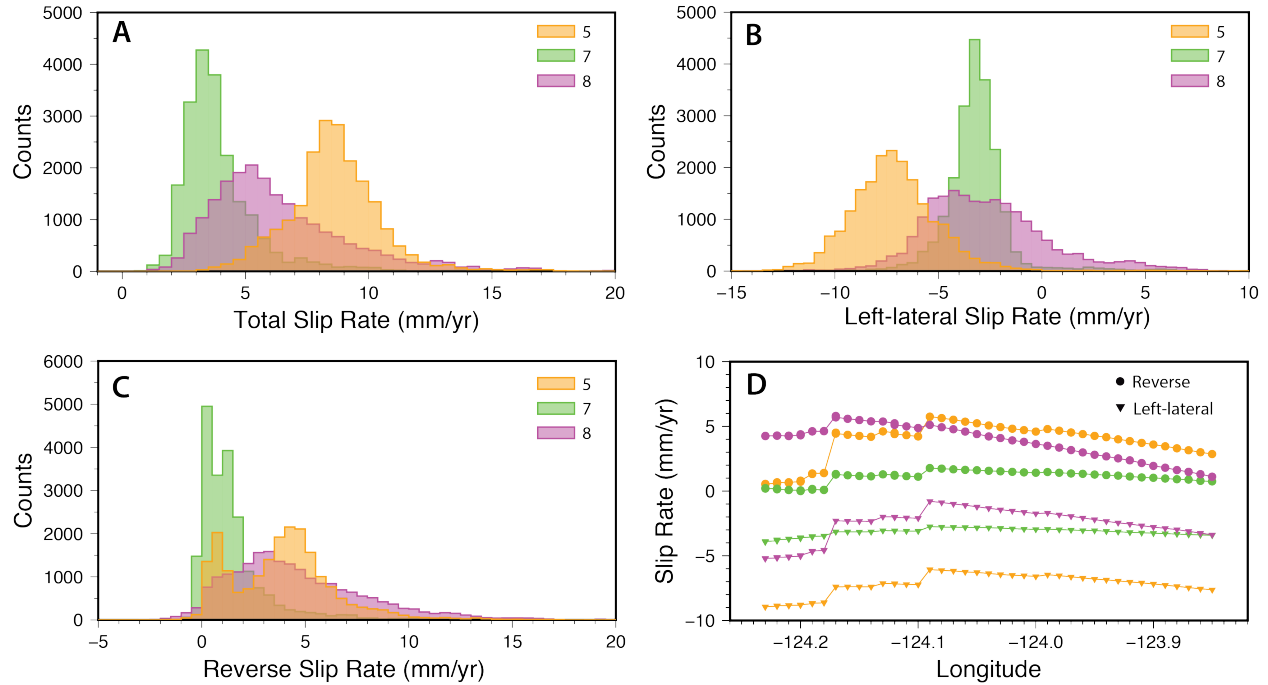


Figure 3.6 Bootstrap fault slip rate results for the Little Salmon fault. **A** Histograms of total slip rate. Histograms are colored by geometry, shown in the upper-right-hand of the panel. **B** Histograms of the left-lateral component of slip rate. Negative values indicate right-lateral slip rates. **C** Histograms of the reverse component of slip rate. Negative values indicate normal slip rates. **D** Line plots showing the median slip rate values at nodes along the fault. Inverted triangles and circles delineate left-lateral and reverse slip rates, respectively. Lines are colored by geometry and match the coloring in panels A, B and C.

	5	6	7	8
Total Slip (mm/yr)	8.5 (6.9 - 10.1)	-	3.5 (2.7 - 5.0)	5.8 (3.9 - 8.9)
Reverse Slip (mm/yr)	4.0 (0.9 - 5.9)	-	1.0 (0.1 - 2.4)	4.0 (1.4 - 7.8)
Left-lateral Slip (mm/yr)	-7.3 (-9.0 to -5.2)	-	-3.2 (-4.3 to -2.2)	-3.0 (-5.4 to -0.0)

Table 3.2 Bootstrap slip rate results for the Little Salmon fault. Model geometry is indicated in the top row and all ranges in parentheses represent the 68% confidence interval. Red values indicate slip rate values from our preferred geometry. Geometry 6 does not include the Little Salmon fault.

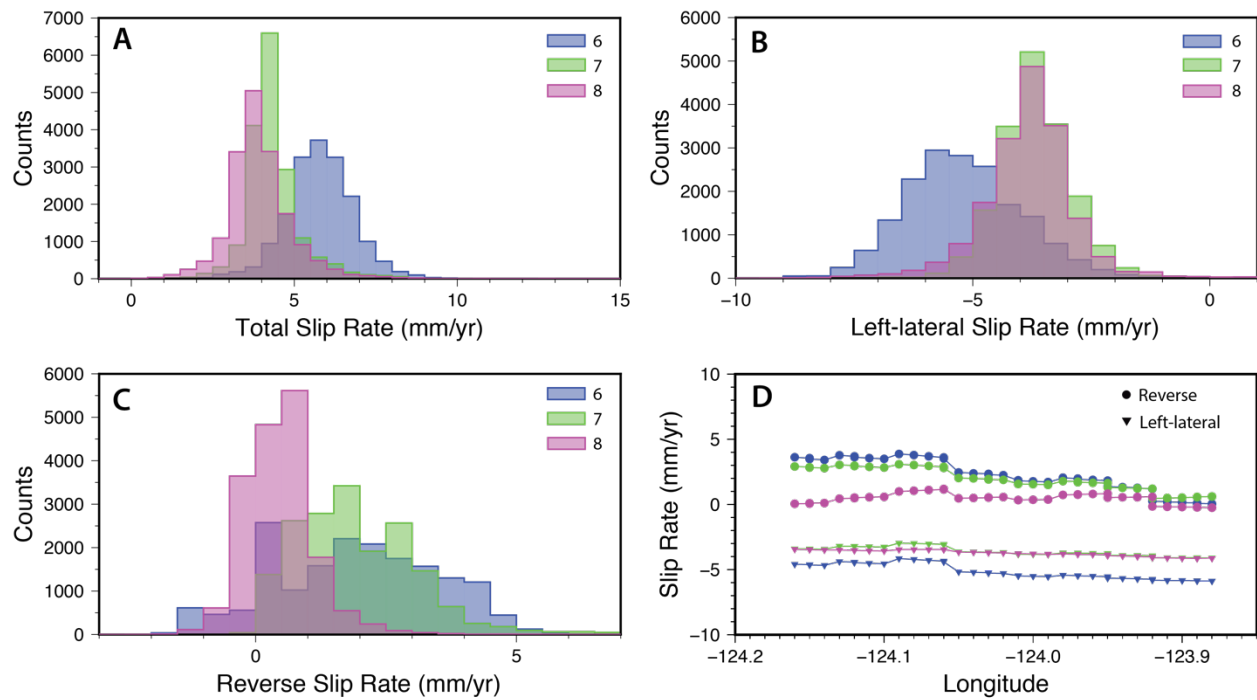


Figure 3.7 Bootstrap fault slip rate results for the Mad River fault zone. Subplots are the same as in Figure 3.6.

	5	6	7	8
Total Slip (mm/yr)	-	5.8 (4.9 - 6.7)	4.2 (3.7 - 4.9)	3.8 (3.1 - 4.7)
Reverse Slip (mm/yr)	-	1.9 (0.3 - 3.6)	1.7 (0.6 - 3.0)	0.5 (-0.2 to 1.0)
Left-lateral slip (mm/yr)	-	-5.3 (-6.4 to -4.0)	-3.7 (-4.4 to -2.9)	-3.8 (-4.6 to -3.0)

Table 3.3 Bootstrap slip rate results for the Mad River fault zone (MRFZ). Model geometry is indicated in the top row and all ranges in parentheses represent the 68% confidence interval. Red values indicate slip rate values from our preferred geometry. Geometry 5 does not include the MRFZ.

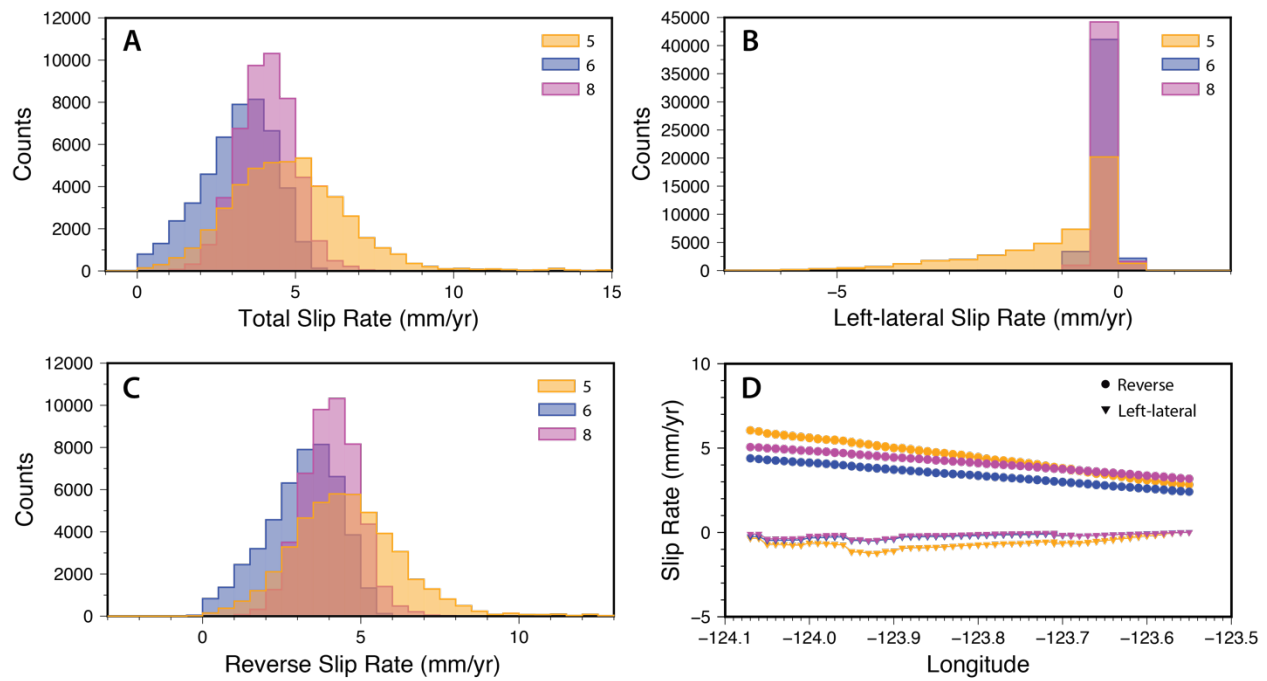


Figure 3.8 Bootstrapping fault slip rate results for the Grogan fault. Subplots are the same as in Figure 3.6.

	5	6	7	8
Total Slip (mm/yr)	4.7 (3.1 - 6.5)	3.3 (2.0 - 4.3)	-	4.1 (3.2 - 4.9)
Reverse Slip (mm/yr)	4.5 (3.0 - 6.1)	3.3 (1.9 - 4.3)	-	4.1 (3.2 - 4.9)
Left-lateral slip (mm/yr)	-0.6 (-2.4 to -0.1)	-0.2 (-0.4 to -0.1)	-	-0.1 (-0.3 to 0.0)

Table 3.4 Bootstrap slip rate results for the Grogan fault. Model geometry is indicated in the top row and all ranges in parentheses represent the 68% confidence interval. Red values indicate slip rate values from our preferred geometry. Geometry 7 does not include the Grogan fault.

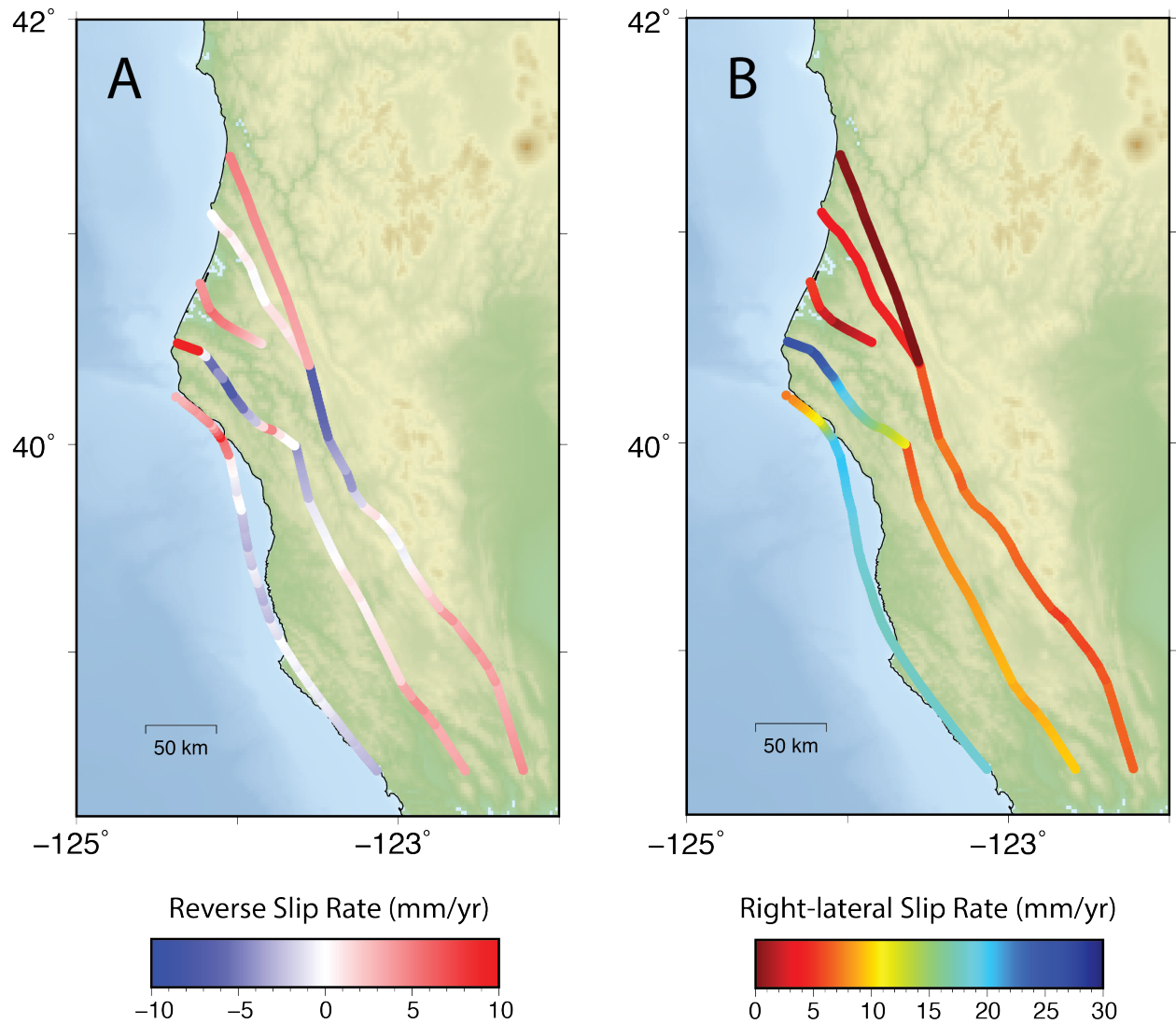


Figure 3.9 Maps of the median slip rate results from the bootstrap analysis for geometry 8. **A** Heavy lines show surface traces of faults colored by the median reverse slip rate for the top 5 km. Negative slip rates indicate normal slip. **B** Heavy lines show surface traces of faults colored by the median right-lateral slip rate for the top 5 km.

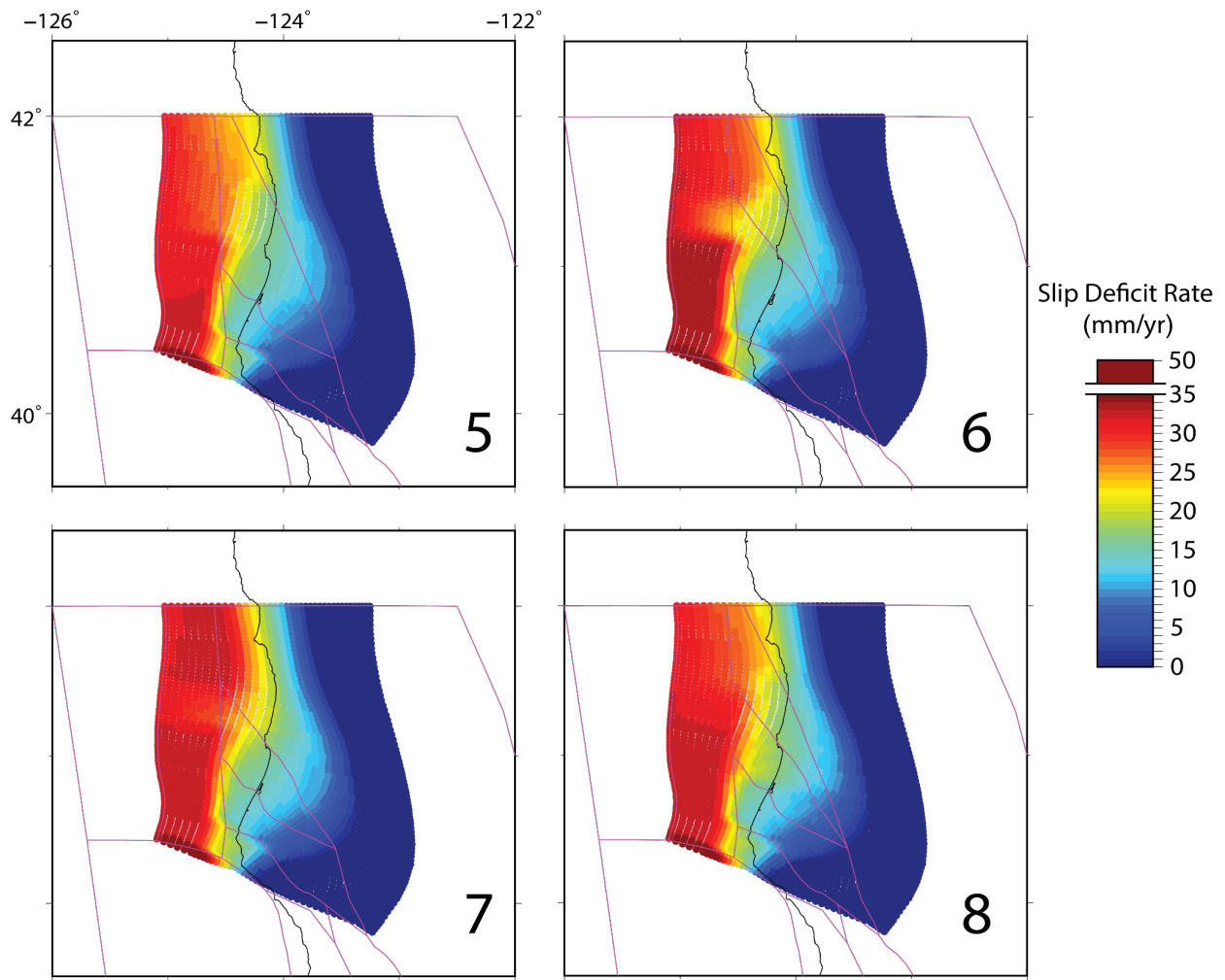


Figure 3.10 Median slip deficit rate at all nodes along the megathrust for geometries 5-8 calculated from the bootstrapping runs. Geometry number is shown in the bottom right-hand corner of each panel, magenta lines delineate block geometries, black lines delineate the coast, and background colors show slip deficit rate with shared scale on the right. A discontinuous scale is provided so as not to skew the color distributions.

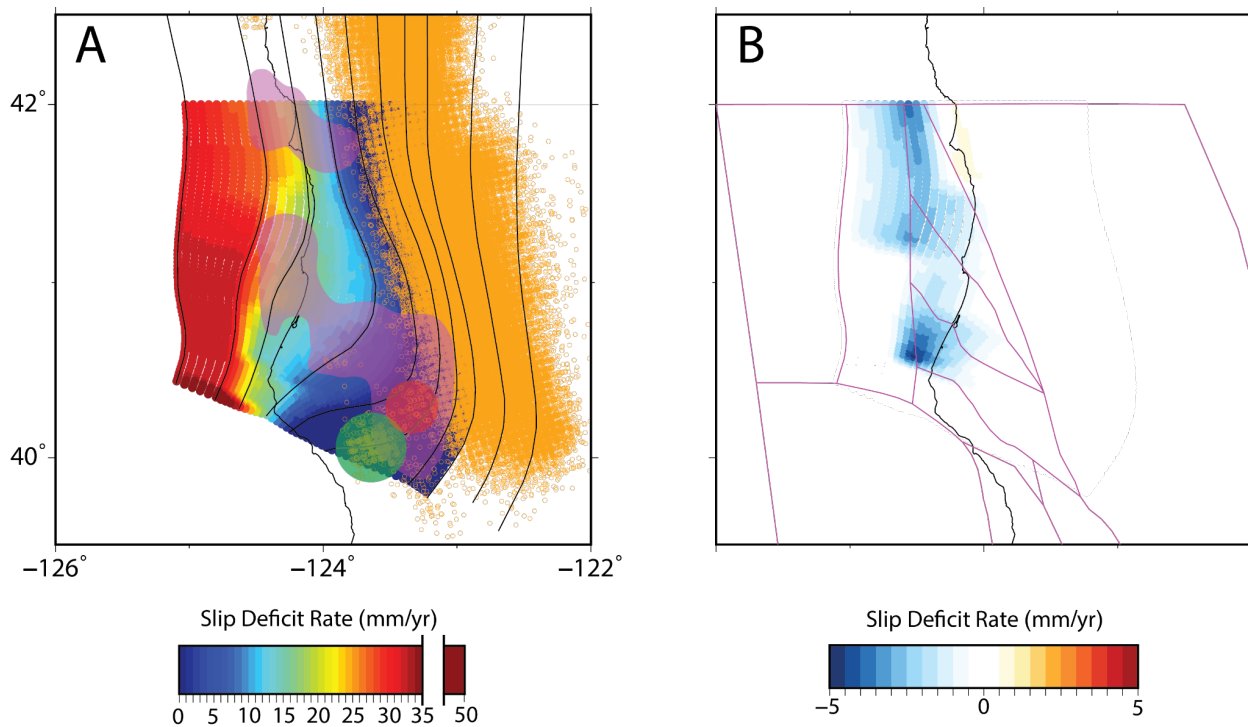


Figure 3.11 The estimated slip deficit rate for the megathrust fault. **A** Map of the MTJ region where background colors show the median slip deficit rate at all nodes along the megathrust for geometry 8 calculated from the bootstrap runs. Black lines denote plate contours, orange circles show tremor locations, red circle designates the location of plate coupling changes from Materna et al. (2019), purple polygons outline the locations of long-term slow slip from Nuyen & Schmidt (2021), and green circle displays location of near-continuous tremor from Wech (2021). **B** Map of the MTJ region showing the difference in slip deficit rates between geometries 1 and 8, where negative values indicate lower slip rates in geometry 8 compared to geometry 1. Changes in slip deficit rate due to the OFFS block are not included. Magenta lines delineate the block boundaries for geometry 8.

Chapter 4. WHAT HAPPENS TO EPISODIC TREMOR AND SLIP AT THE SOUTHERN EDGE OF THE CASCADIA ZONE?

4.1 INTRODUCTION

In subduction zone systems, interseismic strain along the plate interface is periodically released at certain depths by slow slip events (SSEs; Schwartz & Rokosky, 2007). These events typically occur along the plate boundary downdip of the seismogenic locked zone and updip of the creeping portion of the fault (Ito et al., 2007). SSEs involve a component of accelerated aseismic slip along the plate interface and manifest as transient surface displacements in GNSS time series data (Rogers & Dragert, 2003). These events are also associated with seismic shear slip along the plate interface, which produces non-volcanic tremor and low-frequency earthquakes (LFEs) that coincide spatially and temporally with the SSEs (Ide et al., 2007; Shelley et al., 2007). Together, these periodic seismic and aseismic phenomena are termed episodic tremor and slip (ETS).

Ample work has been done to study ETS in the Cascadia subduction zone (Gomberg et al., 2010), with margin-wide studies providing insights on the broad patterns of tremor and slow slip along the margin (e.g. Bartlow, 2020) and more localized studies shedding light on the characteristics of individual events, such as moment magnitude and fault slip amplitude (e.g. Schmidt & Gao, 2010). However, only a small portion of this work has focused exclusively on southern Cascadia, where ETS encounters the southern edge of the subduction zone (Figure 4.1). Although several studies analyze the characteristics of individual ETS events in this region, such as slip amplitudes on the megathrust and tremor migration patterns (Boyarko & Brudzinski, 2010; Schmalzle et al., 2014; McKenzie et al. 2020), the various methods and *a priori* assumptions used by these studies make it difficult to directly compare their results and discern if the behavior of ETS events changes at the edge of the subduction zone.

This study examines nineteen of the largest SSEs in Southern Cascadia from mid 2016 to 2022. Using tremor and GNSS time series data, we determine the timing and horizontal surface

displacements associated with these events, and invert for the slip distribution on the megathrust fault. These slip models allow us to compare patterns of tremor and fault slip between numerous ETS events and analyze how transient strain release is distributed near the southern edge of the subduction zone. Our results show that ETS terminates >50 km north of the Gorda slab edge and suggest that ETS behavior in southernmost Cascadia may be influenced by slab geometry and the presence of an inferred slab window near the triple junction.

4.2 BACKGROUND

4.2.1 Gorda Slab Structure

Southern Cascadia is characterized by the subduction of the Gorda slab beneath the North American plate in southern Oregon and northern California (Figure 4.1). While the northern portion of the Gorda slab is planar between 42° and 43° N latitude, the southernmost portion exhibits a warped geometry beneath northern California. Specifically, the slab flattens between 15 and 25 km depth, creating a broad high centered at 41° N latitude (McCrory et al., 2012; Guo et al., 2021). Down dip of 25 km depth, the slab arcs around this shallow region to create a bend that extends to the edge of the subduction zone. This distortion can be attributed to the northward impingement of the Pacific plate on the Gorda slab, which results in the internal deformation of the slab offshore (Chaytor et al., 2004; McCrory et al., 2012; Ismat et al., 2022). Seismic velocity models and the location of earthquake hypocenters in southern Cascadia also suggest that the edge of the slab is dipping south-to-southeast by $6\text{--}12^\circ$ (Verdonk & Zandt, 1994; Furlong & Schwartz, 2004). This complex slab geometry in southernmost Cascadia likely interacts with mantle flow patterns near the edge of the subduction zone (Schellart, 2004; Bodmer et al., 2018). Offshore of Cape Mendocino, the southern boundary of the Gorda plate is clearly delineated by the Mendocino fracture zone. However, following subduction, the location of the slab edge becomes less obvious. Therefore, the southern terminus of the Blair et al. (2013) slab model in Figure 4.1 is inferred from the first-motion readings at seismic stations after the 1994 Mw 7.1 Mendocino earthquake (McCrory et al., 2006). Specifically, McCrory et al. (2006) observe a change in the first-motion recordings at seismic stations in northern California across a boundary that intersects the coast at the MTJ with an azimuth of $\sim 120^\circ$. McCrory et al. attribute this first-motion pattern to a seismic wavefront that diffracted into the Gorda slab and interpret this first-

motion boundary as the slab edge. This slab edge location is also supported by subsurface models that are constrained by gravity data (Jachens & Griscom, 1983).

4.2.2 Mendocino Triple Junction and Slab Window

Subduction in southern Cascadia ends at the Mendocino triple junction (MTJ), where the Pacific, Gorda, and North American plates meet offshore of Cape Mendocino, CA. Over time, the MTJ steadily migrates northward relative to North America, extending the San Andreas transform boundary by shifting the southern limit of subduction to the north (Atwater, 1970). This northward migration of the MTJ promotes the growth of a slab window directly south of the Gorda slab (Atwater, 1970; Dickinson & Snyder, 1979a, 1979b). As this slab window forms, mantle material flows to fill the space that was previously occupied by the Gorda slab. Various 2-D and 3-D models of the MTJ region predict that this emplaced mantle material cools over time as it equilibrates with the surrounding lithosphere (Goes et al., 1997; van Wijk et al., 2001; Popov et al., 2012). Heat flow measurements in northern California show a pronounced increase in heat flow from north to south, with maximum heat flow values occurring ~200 km south of the MTJ (Lachenbruch & Sass, 1980). Thermal models that include conductive and/or advective heat transfer between the slab window and North American lithosphere (e.g. Guzowski & Furlong, 2002; Popov et al., 2012) are able to recreate this high heat flow anomaly and suggest that the slab window is actively warming the overlying crust. Generic models for ridge subduction and subsequent slab window generation suggest that this warming can lead to rheological weakening, causing the transition from brittle to ductile deformation to become shallower (Groom & Thorkelson, 2007).

Seismic velocity and thermomechanical models indicate that the northern edge of the slab window abuts the southern edge of the Gorda slab (Benz et al., 1992; Liu et al., 2012; Stanciu & Humphreys, 2021). Therefore, in addition to the continental lithosphere, hot mantle material in the slab window is likely transferring heat to the edge of the descending Gorda slab as well. Goes et al. (1997) presents 3-D thermal models for the MTJ region that incorporate the subducting Gorda slab and allow for the thermal structure to be impacted by variations in the slab's age. Inspection of the figure from these models indicates that temperatures at 40km depth along the plate interface located 20 km north of the slab edge are ~100 °C warmer than temperatures at 40km depth located 100 km north of the slab edge. Therefore, this suggests that temperatures

along the plate interface at the slab edge at 40km depth may have increased by more than 100 °C due to heating from mantle material in the adjacent slab window.

4.2.3 Previous ETS Observations in Southern Cascadia

ETS events occur frequently in southern Cascadia with recurrence intervals of 10 ± 2 months (Szeliga, 2004; Brudzinski & Allen, 2007). These events occur along the plate interface at 30-60 km depth and combine to create a continuous band of tremor and transient fault slip that broadly aligns with the depth contours of the warped Gorda slab (Figure 4.1, Michel et al., 2019; Bartlow, 2020; Wech, 2021). An anomalous near-continuous patch of tremor is also located near the slab edge and offset ≥ 40 km to the west of the main tremor zone (Wech, 2021; Figure 4.1). However, the source of this tremor patch is unclear and could be related to slip on a transform fault within the San Andreas Fault system, slip on the plate interface, or transform slip on the edge of the Gorda slab (Wech, 2021).

Distinct LFE families are also observed in northern California and are attributed to slip along the San Andreas fault system and ETS on the megathrust (Plourde et al., 2015; Figure 4.1). The ETS families exhibit depth-dependent recurrence intervals, with deeper families having shorter recurrence intervals than shallower families (Ducellier & Creager, 2022). This behavior is also observed for LFEs and tremor in northern Cascadia, where smaller, more frequent events tend to occur downdip of larger, less frequent events (Wech & Creager, 2011; Sweet et al., 2019). Wech & Creager (2011) attribute this spatial relationship to changes in fault strength, suggesting that the megathrust weakens with depth due to decreases in friction. However, this trend does not apply to the southernmost ETS families (overlapping pair of cyan dots in Figure 4.1), which exhibit relatively short recurrence intervals (< 50 days) despite being located farther updip (Ducellier & Creager, 2022). Therefore, Ducellier & Creager (2022) propose that this family may be located on a crustal fault above the megathrust and not associated with ETS on the plate interface.

Changes in the slab geometry and thermal gradient at the edge of the subduction zone may impact the behavior of ETS in southern Cascadia. However, assessing possible changes in ETS behavior is challenging without the compilation and direct comparison of numerous events. Boyarko & Brudzinski (2010) analyze the spatiotemporal patterns of tremor in southern Cascadia between 2005-2007 and present observations on event distribution and migration style. However,

this work relies on a semiautomated tremor location algorithm that only considers the most active nighttime hour (in terms of tremor) for every day of a given tremor episode. As a result, this work only considers the most active times of tremor and may not capture the full spatial extent of tremor events. Schmalzle et al. (2014) and McKenzie et al. (2020) present distributed slip models for twelve ETS events in southern Cascadia from 2005.5-2011 and 2014-2017, respectively. Although these studies provide valuable insights on fault slip magnitudes and the orientation of strain accumulation along the megathrust, they represent margin-wide studies that do not focus on characterizing ETS in southern Cascadia. Therefore, we identify and calculate distributed slip models for nineteen ETS events in southern Cascadia to determine if ETS behavior changes at the edge of the Cascadia subduction zone.

4.3 DATA AND METHODS

We utilize a combination of publicly available tremor and GNSS time series data to characterize ETS events in southern Cascadia. The tremor data are sourced from the Pacific Northwest Seismic Network (PNSN) tremor catalog (<https://pnsn.org/tremor>), and the GNSS time series data are obtained from the Pacific Northwest Geodetic Array (PANGA) at Central Washington University (PANGA, 1996). We use horizontal east and north component GNSS time series data from 137 stations throughout central and southern Cascadia in the NAM14 stable North American reference frame. GNSS time series data capture signals from various tectonic and non-tectonic sources that can obscure the timing of ETS events. This is particularly true in southern Cascadia, where elevated levels of seismicity indicate active deformation in the region (McCrory et al., 2012). Therefore, we process the GNSS time series data to remove unwanted tectonic signals and non-tectonic noise to make the timing of ETS events more clear. We process the data using the following steps:

- 1) We remove outliers (i.e. positions that stray >20 mm from adjacent observations).
- 2) We remove steps that are due to known earthquakes and equipment changes. The timing of steps is determined by the GAGE offsets from the Central Washington University (<ftp://data-out.unavco.org/pub/products/offset/>). We estimate the step offsets by differencing the average position for the 21 days before the event and the average position for the 21 days after the event.

- 3) We remove the annual and biannual seasonal components and detrend the time series data. We assume that the annual and biannual signals are a combination of sine and cosine waves and model the time series data as a linear combination of seasonal, slope, and offset terms. We solve for the coefficients of these terms using a linear least-squares inversion, and then remove the scaled seasonal and slope terms from the GNSS time series data.
- 4) We smooth the time series data by applying a seven-day moving average, which helps remove high-frequency (daily) noise. The initiation and cessation phases of ETS events typically produce small daily displacements that can easily be obscured by noise. Therefore, removing high frequency noise helps to elucidate the starting and stopping times of ETS events.

4.3.1 Event Timing and Displacements

We identify the timing of ETS events in southern Cascadia by visually inspecting the spatiotemporal patterns of tremor (Figure 4.2). When examining the PNSN tremor catalog (<https://pnsn.org/tremor>), we focus on continuous tremor events where the majority of tremor occurs below 44° N latitude and the total duration of tremor associated with an individual ETS event is at least 70 hours. Using these criteria, we identify nineteen ETS events in southern Cascadia between 2016.5 and 2022 (Figure 4.2). The tremor monitoring system for the PNSN was updated in 2017, resulting in more accurate tremor detection in southern Cascadia (Wech, 2021; K.C. Creager, personal communication, 2022). Therefore, aside from two events at the end of 2016, we do not consider any events prior to 2017 given the changes in the tremor detection algorithm and seismic station network at this time (Wech, 2021). After identifying the nineteen events with the tremor data, we then use the GNSS time series data to estimate the timing and offset amplitude of the associated surface displacements.

ETS events in southern Cascadia manifest as westward displacement transients in GNSS time series data (Szeliga et al., 2004). Therefore, we develop a simple algorithm to identify negative displacements in the GNSS east time series data and estimate the timing of ETS events (Figure 4.3). For each event, the algorithm starts by locating the GNSS stations that are closest to the tremor in Figure 4.2. This is accomplished by finding all stations that have at least 100 epicenters of tremor (where an epicenter represents a detection within a five-minute window)

within 30 km of the station. For smaller events with less tremor (e.g. event 16 indicated in Figure 4.2), we reduce the number of tremor counts to twenty. We assume that these GNSS stations nearest the tremor experience larger transients than stations farther away and thus offer a clearer picture of when surface deformation is occurring. Therefore, we utilize these stations to estimate the timing of transients and refer to them as ‘index’ stations. For each index station, the algorithm computes a time series of daily displacements from the horizontal east component GNSS time series data. The algorithm then applies a centered 14-day moving sum to the daily displacement time series data and flags all days where the sum falls below a given threshold (Figure 4.3A, 4.3B). For this work, we find that a threshold of -0.5 mm is large enough to avoid false-detections due to leftover noise in the data and low enough to capture relatively small transients associated with ETS events. The algorithm then establishes transient start and end times by going through the time series data and looking for consecutively flagged days. When the algorithm finds a succession of flagged days, it marks the first flagged day as a start date and the last flagged day as an end date. We limit our analysis to transients that last at least seven days, so the algorithm ignores any consecutively flagged days that do not span at least a week (Figure 4.3C). The algorithm also only considers flagged days that fall within two weeks of the beginning or end of the event tremor shown in Figure 4.2. Once the algorithm has found the timing of transients for all of the index stations, it then assigns every non-index station the timing of the closest index station by latitude. Finally, the algorithm calculates the east and north component horizontal surface displacements for each station by differencing the average position for the seven days before the transient start date and the average position for the seven days after the transient end date. If a station has multiple start and end dates, then the algorithm sums the displacements from the multiple epochs. Smoothing the GNSS time series data in Step 4 runs the risk of dampening the amplitude of ETS transients. Therefore, we calculate the horizontal surface displacements from GNSS time series data that is processed without applying the final smoothing step (Figure 4.3C).

We manually inspect the results of our algorithm for each ETS event to ensure consistency between the transient times at different stations and to check that the results are compatible with the tremor data. Overall, we find that the algorithm is able to accurately identify transients in the GNSS time series data that align well with the timing of tremor (Figure 4.4). However, there are several ETS events where the timing of transients at certain stations can be

improved by removing or adding index stations. For example, removing an index station can be helpful when the index station's GNSS data is particularly noisy and the transient timing is impacted by non-tectonic signals. Conversely, it can be beneficial to add index stations when there are clear transients in the GNSS data (Figure 4.4A). For example, station YBHB consistently shows surface displacements associated with multiple ETS events, but it is not close enough to tremor to be considered as an index station. Therefore, we manually remove and/or add index stations for certain ETS events to improve the timing of transient signals in the GNSS time series data. Details for each ETS event, including which index stations are used and the transient start and stop times for each station, can be found in the Supplementary Information.

For ETS Event 9 (November 2018), we observe a large-scale regional pattern of westward surface displacements at GNSS stations throughout Cascadia (Figure A3.1). The tremor during this event is concentrated on the Oregon-California border and the GNSS stations nearest the tremor exhibit larger westward transients than the surrounding stations. Therefore, we believe that the displacements for this ETS event are over-printed by a widespread common-mode signal. To remove this signal, we average the displacements at all stations located south of 40° latitude or east of -121° longitude, since we believe that these stations are far enough away from the event tremor that they should not experience any displacements. This average displacement was then removed from all of the stations to create the final displacement field shown in Figure A3.11 (Dong et al., 2006).

4.3.2 Inversion

We invert the horizontal surface displacements for each of the nineteen ETS events to create distributed fault slip models. We utilize a non-negative least-squares inversion, so that slip on each fault patch can only have components of reverse and right lateral slip. Surface displacements at GNSS stations are approximated using an elastic half-space dislocation model (Okada, 1992) with a shear modulus of 40 MPa. Each inversion is regularized by applying a smoothing parameter that minimizes the roughness of the slip distribution and data misfit, as determined by the L-curve. Roughness is determined by the Laplacian of the fault slip model, and the data misfit is defined by the chi-square statistic:

$$\chi^2 = \frac{1}{n_{obs}} \sum_{i=1}^{n_{obs}} \frac{(d_{obs}(i) - d_{pred}(i))^2}{\sigma(i)^2} \quad (4.1)$$

where n_{obs} is the number of observations, d_{obs} and d_{pred} are the observed and predicted displacements, respectively, and σ is the uncertainty of the observations. We estimate the uncertainty for each observation by averaging the standard deviations for the seven days before the transient start date(s) and the seven days after the transient start date(s). We force fault slip to be zero along the perimeter of the fault model. However, since we are exploring ETS at the southern edge of the subduction zone, we do not damp slip on fault patches at the edges of the fault below 40.6° N latitude.

We use a megathrust fault model based on the plate contours from Blair et al. (2013) and determine the updip and downdip limit of our fault model by performing a series of sensitivity tests (Figure A3.2). We perform these tests with Event 1, since surface displacements during this event span all of southern Cascadia (<43° N latitude). For these tests, we fix the down-dip limit of the fault model to 80 km depth and perform the inversion with fault geometries that have progressively shallower up-dip depth limits. We run a similar set of tests where we fix the up-dip depth limit of the fault model at the trench (5km depth) and perform the inversion with progressively deeper down-dip depths limits. We calculate the data misfit for each inversion according to the chi-square statistic above (Equation 1). The results of our sensitivity tests show that surface displacements in southern Cascadia are not sensitive to slip on the plate interface above 15 km depth or below 55 km depth. However, a comparison of tremor locations and plate depth contours shows that tremor in southernmost Cascadia extends to 75 km depth (Figure 4.1). If we assume that tremor locations serve as a proxy for slip on the plate interface, then tremor in this region indicates that slip extends past 55 km depth. Therefore, we consider both our sensitivity results and tremor observations to establish a downdip limit of 65 km for our faults model.

4.4 RESULTS

We calculate distributed slip models for nineteen ETS events in southern Cascadia (south of 46° N latitude) from mid 2016 to 2022. These events range in moment magnitude from Mw 6.3 to 6.9 (Table A3.1). Figure 4.5 shows our slip model for ETS Event 15 (June 2020), which constitutes one of the southernmost events and has an equivalent moment magnitude of Mw 6.5. This slip model indicates that large amounts of slip are not required along the southern edge of

the slab to recreate the observed surface displacements. Instead, the maximum slip contour for 22 mm is located >50 km north of the slab edge. It is important to note that the damping of the fault patches at the perimeter of the fault model is not responsible for this lack of slip at the slab edge, as we purposefully do not apply any slip minimization along the southern edges of the fault model (Section 4.3.2). When projecting the fault rake to the surface, the overall azimuth for this patch of high slip is roughly 270° , which corresponds to pure reverse slip on the plate interface (Figure 4.5D). Along the slab edge where there is a small patch of fault slip (-124° E longitude, 40° N latitude), the slip azimuth rotates north and approaches 360° . However, due to the slab bend, this azimuth also corresponds to nearly pure reverse slip. The residual vectors for this event show a coherent westward pattern around the event tremor, indicating that our predicted surface velocities are underfitting the observations (Figure 4.5E). We attribute this underfitting to the spatial smoothing that we impose in the inversion and acknowledge that slip amplitudes may be higher along certain portions of the fault. Similar results for the other eighteen events can be found in the Supplementary Information.

Overall, our results for individual ETS events in southern Cascadia are consistent with the distributed slip models from previous works. In particular, Schmalzle et al. (2014) estimates a similar moment magnitude range of M_w 6.4 to 7.0 for nine ETS events in southern Cascadia. These nine slip models also show comparable slip amplitudes on the megathrust, with the majority of events having maximum slip values of ~ 20 mm. McKenzie et al. (2020) analyzes three additional ETS events in southern Cascadia and also finds similar slip amplitudes.

To investigate the total transient deformation field, we sum the slip distribution for all nineteen ETS events and observe a maximum of 140 mm of total fault slip on the plate interface centered around 40.5° N latitude between 30 to 50 km depth (Figure 4.6A). This patch of high slip is located down-dip of the slab high along the northern portion of the slab bend. Furthermore, ETS does not follow the depth contours to the south and thus wrap around the southern portion of the slab bend, nor does it continue towards the southern edge of the slab. Rather, ETS appears to migrate down-dip, cutting across depth contours on the plate interface and terminating at $\sim 40^\circ$ N latitude.

For each ETS event that we identified in the PNSN tremor catalog (Figure 4.2), we found corresponding surface displacements at GNSS stations that are consistent with slow slip on the megathrust. Therefore, slow slip and tremor appear to be coincident in space and time for each of

the large ETS events that we studied, within the resolution of the tremor and slip detection. To evaluate the spatial correlation of tremor and slow slip, we segregate the tremor counts by ETS and inter-ETS time periods. The locations of tremor during the nineteen largest ETS events align with the depths of high cumulative fault slip, with the majority of event tremor occurring between 43° and 40° N latitude at 30-55 km depth (Figure 4.6B). Tremor that occurs outside of the largest ETS events is located at similar latitudes but is concentrated in a narrower and deeper depth range (Figure 4.6C). In particular, inter-ETS tremor below 42° N latitude occurs between 40 and 60 km depth. Similar to slip on the plate interface, neither ETS nor inter-ETS tremor extends to the southern edge of the Gorda slab.

4.5 DISCUSSION

Our results show that the cumulative fault slip of nineteen slow slip events along the megathrust reaches a maximum of 140 mm near 40.5° N latitude (Figure 4.6A). This suggests that large ETS events accommodate up to 80% of the plate convergence in southern Cascadia within the ETS zone, assuming a convergence rate of 32 mm/yr (McCaffrey et al., 2007). Bartlow (2020) uses tremor to identify the timing of ETS in Cascadia and estimates the time-averaged ETS slip rate on the megathrust fault. Results from this work also show relatively high slip rates between 40° and 41° N latitude, confirming that the majority of strain release in this region between 30 and 50 km depth is accommodated by ETS. However, our results indicate that at least 20% of the plate convergence is accounted for by smaller ETS events (periods of tremor marked by blue dots in Figure 4.2) and/or by steady aseismic slip. In contrast, Bartlow (2020) estimates that the entire slip budget in this region is facilitated by ETS events, and thus suggests that steady aseismic slip is not needed.

A comparison of the tremor locations in Figure 4.6 shows that the more frequent inter-ETS tremor occurs slightly downdip of the less frequent ETS tremor south of 42° N latitude. This spatial relationship is observed for tremor and LFEs elsewhere in Cascadia, and is attributed to fault strength decreasing as a function of depth due to changes in friction along the plate interface (Wech & Creager, 2011; Sweet et al., 2019; Ducellier & Creager, 2022). Therefore, our results are consistent with other observations from Cascadia and support the notion that the megathrust weakens with depth in southern Cascadia.

The depth of maximum fault slip and tremor as a function of latitude increases from north to south in southern Cascadia, with relatively high cumulative slip values (>100 mm) occurring at 35 km depth at $\sim 41.5^\circ$ N latitude and continuing to 60 km depth at $\sim 40.25^\circ$ N latitude. This suggests that ETS gradually transitions to deeper depths in southern Cascadia, or that our fault model overestimates the depth to the slab in this region. Delph et al. (2021) offer an alternate slab model for southern Cascadia, wherein the McCrory et al. (2012) model is adjusted to fit the location of LFEs. In this model, the slab bend is accentuated below 42° N latitude at depths ≥ 40 km, leading to an overall shallower slab model in this region. Therefore, if we use the Delph et al. (2021) slab model, then high cumulative slip values (>100 mm) do not migrate past 50 km depth at $\sim 40.25^\circ$ N and the apparent deepening of ETS is less pronounced. ETS fault slip and tremor terminate farther to the south near $\sim 40^\circ$ N latitude and do not extend to the southern edge of the Gorda slab. LFEs associated with slow slip along the megathrust exhibit a similar pattern and are not observed south of 40° N latitude (Figure 4.7A; Plourde et al., 2015; Ducellier & Creager, 2022). Together, these observations suggest that ETS does not occur along the southernmost edge in Cascadia.

In the following sections, we explore why ETS may be absent in this region and present possible controls on the southern extent of ETS in Cascadia. In particular, we discuss how the non-planar geometry of the Gorda slab may influence fluid migration along the plate interface, the width of the ETS zone, and shear-strength along the megathrust fault. Additionally, we analyze how thermal interactions between the subduction zone and the slab window may impact the location of the aseismic-seismic boundary along the megathrust and the depth at which fluids are released from the Gorda slab.

4.5.1 Gorda Slab Geometry

4.5.1.1 Location of the Slab Edge

The distance between the southern extent of ETS and the edge of the Cascadia subduction zone is dependent on the location of the southern edge of the Gorda slab. In this work, we reference the slab edge from McCrory et al. (2006), which is based on the pattern of first-motion readings at seismic stations in northern California. A similar slab edge location is also inferred by Jachens & Griscom (1983) from gravity data. The agreement between these two independent studies, which employ distinct datasets and methods, gives us some confidence in the location of

the slab edge. However, several other studies have focused on locating the slab edge beneath Northern California and the results from these works reveal a range of interpretations with slab edges farther to the north (Figure 4.7B).

Liu et al. (2012) identifies the southern edge of the slab using a 3D S-wave velocity model from the joint inversion of ambient noise data, ballistic Rayleigh wave dispersion curves, and receiver functions. In this model, the Gorda slab is associated with relatively high V_s (>4.4 km/s), and the slab edge is delineated as the southern limit of these high velocities. This inferred edge also aligns with the termination of positive amplitude conversions in the P_s receiver functions, which are interpreted as the Gorda slab Moho. Beaudoin et al. (1998) approximates the southern edge of the Gorda slab using a 2-D seismic P-wave velocity model and gravity data. The velocity model is created by inverting the travel times of refracted and reflected waves from an active source survey that collected data along a north-south trending profile near the MTJ (Godfrey et al., 1995). Beaudoin et al. (1998) interprets high velocities (>7.5 km/s) at the base of their velocity model as the Gorda crust and mantle. Beaudoin et al. also utilizes the same gravity data as Jachens & Griscom (1983) to create a subsurface model wherein the slab edge is located at $\sim 40^\circ$ latitude and $\sim 123.3^\circ$ longitude. By shifting the location of the slab edge in their gravity models, they find that their slab edge could be in error of ± 25 km. Ultimately, Beaudoin et al. defines the slab edge as the line that projects from the triple junction through their interpreted slab edge along the 2-D profile. Verdonk & Zandt (1994) locate the edge of the Gorda slab using a 3-D velocity model inverted from $\sim 19,000$ P-wave arrivals from >1400 local earthquakes. Using this model, they interpret the Gorda slab as a high V_p (>7 km/s) region that coincides with high concentrations of earthquake hypocenters and suggests that the slab edge is dipping $6-12^\circ$ to the south. Lastly, Furlong & Schwartz (2004) estimate the position of the slab edge by analyzing the location of earthquake hypocenters from the Northern California Earthquake Data Center from 1969 to 2003. By studying the distribution of hypocenters along northwest-trending cross-sectional profiles, Furlong & Schwartz observes a clear break in seismicity and interprets this break as the slab edge. Similar to Verdonk & Zandt (1994), Furlong & Schwartz also finds evidence for a southeastward dip to the slab edge.

The discrepancies in the slab edge estimates outlined above and summarized in Figure 4.7B likely stem from the various methods used in each study, since they all image the underlying structure at different scales and levels of resolution. For example, reflection and

refraction studies are good for distinguishing sharp boundaries within the lithosphere, whereas tomographic models excel at recognizing larger-scale spatial variations in seismic velocity. The dimensionality of these methods also varies between 1-dimensional profiles (e.g. receiver functions, dispersion curves), 2-dimensional cross-sections (e.g. reflection, refraction), and 3-dimensional models (e.g. tomography). Additionally, several of the studies utilize gravity data, which is most sensitive to density contrasts and changes in rock composition. Therefore, seismic and gravity data may resolve different boundaries within the lithosphere. However, subsurface structural models constrained by gravity data are non-unique and add further uncertainty to the position of the slab edge. This is exemplified by the gravity models presented in Beaudoin et al. (1998), which suggests an uncertainty of ± 25 km in the position of the slab edge. Together, these differences in the resolution and scope of different methods likely account for some of the variation in the estimates presented in Figure 4.7B. Nevertheless, the majority of these estimates place the slab edge at least 50 km south of the maximum slip contour and confirm that ETS does not occur at the edge of the subduction zone.

4.5.1.2 Geometric Controls on Slip Amplitude, Width of the ETS Zone, and Fluid Migration

In northern Cascadia, the Juan de Fuca slab exhibits a bend beneath northern Washington and southern British Columbia that is comparable to the bend in the Gorda slab. Observations of ETS events in northern Cascadia indicate that the axis of this bend, located in northwestern Washington, exhibits higher amounts of ETS fault slip per event than the portions of the slab located directly to the north or south (Schmidt & Gao, 2010). A similar trend is also observed at the Nankai subduction zone, where higher cumulative slip rates are found near slab bends beneath Shikoku and Ise Bay (Nishimura et al., 2013; Kano & Kato, 2020). Our results also fit into this pattern, as we observe a patch of high cumulative fault slip along the northern limb of the bend in the Gorda slab at 40° - 41° latitude. Together, these observations suggest that the slab bend in the Gorda plate may influence the distribution of slip amplitude in southern Cascadia.

However, the mechanism that promotes higher slip amplitudes per event near a slab bend is unclear. Li & Liu (2016) present numerical models of SSEs in northern Cascadia that are able to recreate higher moment rates and cumulative slip amounts in the central portion of the slab bend. These simulations suggest that slip rates are proportional to the width of the ETS zone, which is controlled by the dip of the slab (Perez-Silva et al., 2022). Specifically, lower dip angles

lead to a wider ETS zone, and the higher slip amplitudes would scale with the spatial dimensions of the slipping patch. However, this finding does not appear to apply to southern Cascadia, as our predicted patch of high slip is located on a portion of the plate interface with relatively steep dip according to the Blair et al. (2013) model (Figure 4.7A). However, Li & Liu (2016) present an alternative model, where the width of the ETS zone is held constant. These model results show that slip amplitudes are still inversely proportional to fault dip and are highest on fault patches that strike perpendicular to the convergence direction. Therefore, higher slip rates near the bend in the Gorda slab may indicate that this portion of the fault is more optimally oriented relative to convergence.

The curved geometry of the Gorda slab may also impact the migration of fluids along the plate interface. Geophysical evidence for a low-velocity zone (LVZ) along numerous megathrust faults suggests that high pore-fluid pressures exist along the plate interface at ETS depths (e.g. Audet et al., 2009). As a result, high pore-fluid pressures are often invoked as a prerequisite for possible ETS mechanisms, such as dilatant strengthening (e.g. Segall et al., 2010). Morishige & van Keken (2017) presents 3-D numerical models of fluid migration along curved slab geometries and finds that fluid paths concentrate in regions where the slab has a convex shape, leading to increased porosity values near slab bends. Therefore, the relatively high slip amplitudes around the bend in the Gorda slab may relate to the focusing of fluids in this region. However, it is still unclear if and how fault slip amplitudes scale with the magnitude of pore pressure along the plate interface.

High fault slip amplitudes do not extend around the southern portion of the slab bend in southernmost Cascadia (Figure 4.7). A possible explanation for this observation is an increased warping of the plate interface near the slab edge. Several studies propose that the edge of the Gorda slab dips to the south-southeast (sub-orthogonal to subduction direction) with dip angles of up to 12° (Verdonk & Zandt, 1994; Furlong & Schwartz, 2004). This southward dip, together with the slab bend, may lead to a steeper plate interface and a narrower ETS zone near the slab edge. Therefore, low fault slip rates in southernmost Cascadia may be due to a narrow ETS zone that is incapable of producing large ETS events. In addition to the width of the ETS zone, a complex geometry near the slab edge may also influence how far slow slip can propagate across the fault. Global comparisons of subduction zone systems show that the along-dip curvature of megathrust faults generally scales with maximum earthquake magnitude (Bletery et al., 2016;

Plescia & Hayes, 2020). Bletery et al. (2016) attribute this relationship to the shear-strength heterogeneity across a fault, proposing that shear-strength heterogeneity increases as a function of fault curvature and thus limits the magnitude of megathrust earthquakes by acting as a natural barrier for earthquake slip propagation (Bletery, et al., 2017). If we assume that this theory also applies to slow slip propagation, then an increasingly curved slab geometry in southernmost Cascadia may produce high shear-strength heterogeneity near the slab edge, which limits the propagation and size of ETS events.

4.5.2 Subduction Zone Heating

4.5.2.1 *Seismic Aseismic Transition*

Deformation along the plate interface is accommodated by a continuum of seismic and aseismic slip modes along the megathrust (Peng & Gomberg, 2010). In general, ETS occurs in the seismic-aseismic transition zone that is located updip of the aseismic creeping portion of the fault and downdip of the seismogenic locked zone (Schwartz & Rokosky, 2007). At warm subductions zones, such as Cascadia, the depth of this seismic-aseismic transition is thought to be largely controlled by temperature (e.g. Hyndman & Wang, 1993; Oleskevich et al., 1999). Therefore, changes in a subduction zone's thermal gradient may impact the distribution of different deformational regimes along the plate interface. Specifically, higher temperatures in southernmost Cascadia due to heating from the adjacent slab window may shift the seismic-aseismic transition updip along the megathrust. As a result, deformation within the typical ETS zone (i.e. along the plate interface at 35-60 km depth) may be dominated by aseismic slip and not ETS. This shift in deformational regimes may also explain why LFEs along the megathrust are not observed south of 40° N latitude (Figure 4.7A), since these events are associated with seismic shear slip (Ide et al., 2007; Shelley et al., 2007; Plourde et al., 2015). Furthermore, the relatively short recurrence intervals for the southernmost megathrust LFE family may reflect north-to-south (along-strike) weakening of the plate interface due to increasing temperatures along the plate interface (Ducellier & Creager, 2022).

Following this reasoning, we might expect to observe ETS at shallower depths (<30 km) in southernmost Cascadia near the slab edge, where lower temperatures along the megathrust facilitate seismic deformation in the form of tremor. Although shallower ETS events are not observed, Wech (2021) identifies a temporally near-continuous patch of tremor in this region

with an unknown source (Figure 4.7A). If this tremor does originate from the plate interface, then it may signal the presence of ETS at ~30 km depth and support the notion that ETS is transferred updip due to a higher thermal gradient in southern Cascadia that warms the slab edge. Furthermore, the near continuous activity of this tremor patch suggests that it would represent the downdip limit of the ETS zone, where the fault cannot accumulate large amounts of stress and fails via small, frequent slip events (Wech & Creager, 2011).

4.5.2.2 *Fluid Release*

As described in Section 4.5.1.2, high pore-fluid pressures are thought to exist along portions of the plate interface at subduction zones and play a critical role in the mechanisms for ETS (e.g. Audet & Kim, 2016). Fluids at the plate interface are sourced from the crust and mantle of the subducting slab, which release mineral-bound water through a variety of dehydration reactions (Peacock, 1993; Rupke et al., 2004). However, the depth at which these reactions occur is dependent on the pressure-temperature (P-T) path of the slab, which can vary substantially between subduction zones (Peacock, 2009). Thermodynamic models for southern Cascadia at ~42° N latitude indicate that average mid-ocean ridge basalts (MORB) dehydrate 1-1.5 wt% H₂O at ~35 km depth and suggest that in-situ fluid release is the dominant water source for ETS in this region (Condit et al., 2020). These models also show that higher slab temperatures lead to the slab releasing fluids at shallower depths. For example, increasing the temperature path of the slab by 50 °C results in fluids being released at ~5 km shallower depths (Condit et al., 2020). Therefore, heating of the Gorda slab from the adjacent slab window may result in fluids being released at shallower depths along the megathrust in southernmost Cascadia. Depending on the thermal gradient of the slab, fluids may be released updip of typical ETS depths, limiting high pore-fluid pressures from developing along the plate interface in the expected ETS zone. Therefore, ETS events in southernmost Cascadia may be inhibited by low pore-fluid pressures along the plate interface caused by insufficient in-situ fluid release from the warm Gorda slab at typical ETS depths.

Fluid release is not a single event and can occur gradually and at various temperatures and depths along the plate interface (van Keken et al., 2011). Therefore, increases in slab temperature do not preclude the possibility of fluids migrating into the ETS zone from other portions of the subduction zone. In particular, fluids released deeper on the plate interface from

dehydration reactions that take place at higher P-T conditions could travel updip to ETS depths (Katayama et al., 2012; Hyndman et al., 2015). Additional fluids may also be sourced from the serpentinized mantle wedge. Specifically, Kirby et al. (2014) shows that heating of the relict mantle wedge in the slab window can lead to the release of water through the dehydration of serpentine. This dehydration may also occur in the current subduction zone mantle wedge if temperatures have increased to the point of making serpentine thermally unstable. However, it is unclear whether fluids released from the relict or current mantle wedge could supply fluids to the plate interface or would migrate upwards into the overlying crust. Nevertheless, high pore-fluid pressures generated by the updip migration of fluids may not be enough to facilitate ETS if the appropriate fault rheology is not present. For example, heating from the slab window may promote ductile deformation that prohibits shear faulting and tremor along the plate interface even if high pore-fluid pressures are present (Shelly et al., 2007).

4.5.3 Synthesis and Implications

An analysis of nineteen ETS events suggests that fault slip associated with ETS does not extend to the southern edge of the Gorda slab. Therefore, we hypothesize that the southern termination of ETS marks a major shift in megathrust properties that is tied to changes in the geometry of the Gorda slab and/or thermal conditions near the edge of the subduction zone, which may in turn alter the deformation modes or pore pressure conditions along the plate interface. In terms of slab geometry, a proposed southward dip at the southern edge of the Gorda slab may combine with the slab bend to create a three-dimensional curling of the slab (Verdonk & Zandt, 1994; Furlong & Schwartz, 2004). This complex geometry could lead to (1) a narrower ETS zone that does not host large amounts of slip (Li & Liu, 2016; Perez-Silva et al., 2022) or (2) heterogeneous shear-strength across the fault that acts as a barrier for ETS propagation (Bletery et al., 2016, 2017). In terms of subduction zone temperatures, numerous models suggest that the slab window south of the MTJ actively transfers heat to the adjacent Gorda slab and North American lithosphere (Goes et al., 1997; van Wijk et al., 2001; Guzowski & Furlong, 2002; Popov et al., 2012). Increases in subduction zone temperatures in Cascadia may (1) shift the depth of the aseismic-seismic transition updip (Groom & Thorkelson, 2007) or (2) cause the Gorda slab to release fluids at shallower depths above the ETS zone (Condit et al., 2020).

Overall, these proposed changes in subduction zone properties suggest that seismic deformation does not take place along the plate interface below 35 km depth in southernmost Cascadia.

It is important to note that the processes presented above are likely interconnected. For example, the three-dimensional geometry of the slab edge should influence how slab temperatures evolve in response to the slab window, which in turn impacts the location of different deformational modes along the plate interface. Therefore, better constraints on the individual components of the subduction zone system, e.g. the geometry of the slab edge, and future modeling of the interactions between the subduction zone and slab window are necessary to determine the relative importance of each process. Outstanding questions that future models could address include (1) how are slab temperatures impacted by different slab edge geometries, (2) what is the spatiotemporal scale of subduction zone heating and how is it influenced by different mantle flow patterns in the slab window (e.g. upwelling vs. toroidal flow), and (3) what is the P-T path for the non-uniform portion of the Gorda slab and what does that mean for the depth of fluid release? Furthermore, additional work on the near-continuous tremor patch found by Wech (2021) would also help illuminate the state of ETS in southernmost Cascadia. For example, a moment tensor analysis of individual LFEs could help to determine the source fault of the tremor and clarify if ETS is shifted updip near the edge of the Gorda slab.

The results from this work are relevant to other systems that have a complex geometry and/or slab window near the lateral edge of a subducting plate. For instance, the northern terminus of the Cascadia subduction zone is defined by the lateral edge of the Explorer plate beneath northern Vancouver Island. The distribution of earthquake hypocenters in this region reveals significant warping of the subducted Explorer slab, wherein the plate interface deepens ~15 km along strike over a distance of <100 km (Hutchinson et al., 2020). Geochemical analysis of basalts from the northern Cascade Arc also suggest that a slab gap or tear exists between the Juan de Fuca and Explorer plates, allowing for enriched asthenosphere to flow into the Cascadia mantle wedge (Mullen & Weis, 2015). Observations of ETS in northernmost Cascadia show a non-uniform distribution of tremor, including decreased tremor activity at the northern end of the subduction zone and a prominent gap in tremor near 49.5° N latitude (Kao et al., 2009). Therefore, ETS behavior in northernmost Cascadia may be influenced by the warped geometry of the Explorer plate, similar to what we observe in southernmost Cascadia and the Gorda plate. In southern Chile, the varying dips and depth extents of the subducting Nazca and Antarctic

plates beneath the South American plate help to form a slab window at the Chile triple junction (Russo et al., 2010). In the southernmost portion of the Chilean subduction zone, the depth of tremor along the Nazca plate shallows ~20 km as it approaches the edge of the subduction zone and the slab window (Ide, 2012). This shallowing of tremor may correlate with plate age and temperature (Ide, 2012), or it may represent another locale where increased temperatures from a slab window are impacting the depth of ETS at the lateral edge of a subduction zone.

4.6 CONCLUSIONS

We analyze 5.5 years of tremor and GNSS time series data and identify 19 large ETS events in southern Cascadia from 2016.5-2022. Distributed slip models indicate that fault slip and tremor during these events are spatially correlated, with tremor during the largest ETS events falling slightly updip of inter-ETS tremor. Our models also show that ETS terminates near 41 °N latitude and does not extend to the southern edge of the Gorda slab. Therefore, we hypothesize that seismic slip (i.e. tremor) does not take place along the plate interface below 35 km depth in southernmost Cascadia and that the southern boundary of ETS reflects changes in megathrust properties that are linked to changes in slab geometry and/or subduction zone temperatures. Possible factors that influence ETS in southern Cascadia include the depth at which the Gorda slab releases fluids, the width of the ETS zone, the optimal orientation of the plate interface to local convergence, the depth of the seismic-aseismic transition, the migration patterns of fluids within the subduction zone, and the pattern of shear-strength across the megathrust fault. Future modeling studies that explore the geometric and thermal structure of the Gorda slab and southernmost Cascadia subduction zone are needed to discern the primary controls on ETS in this region.

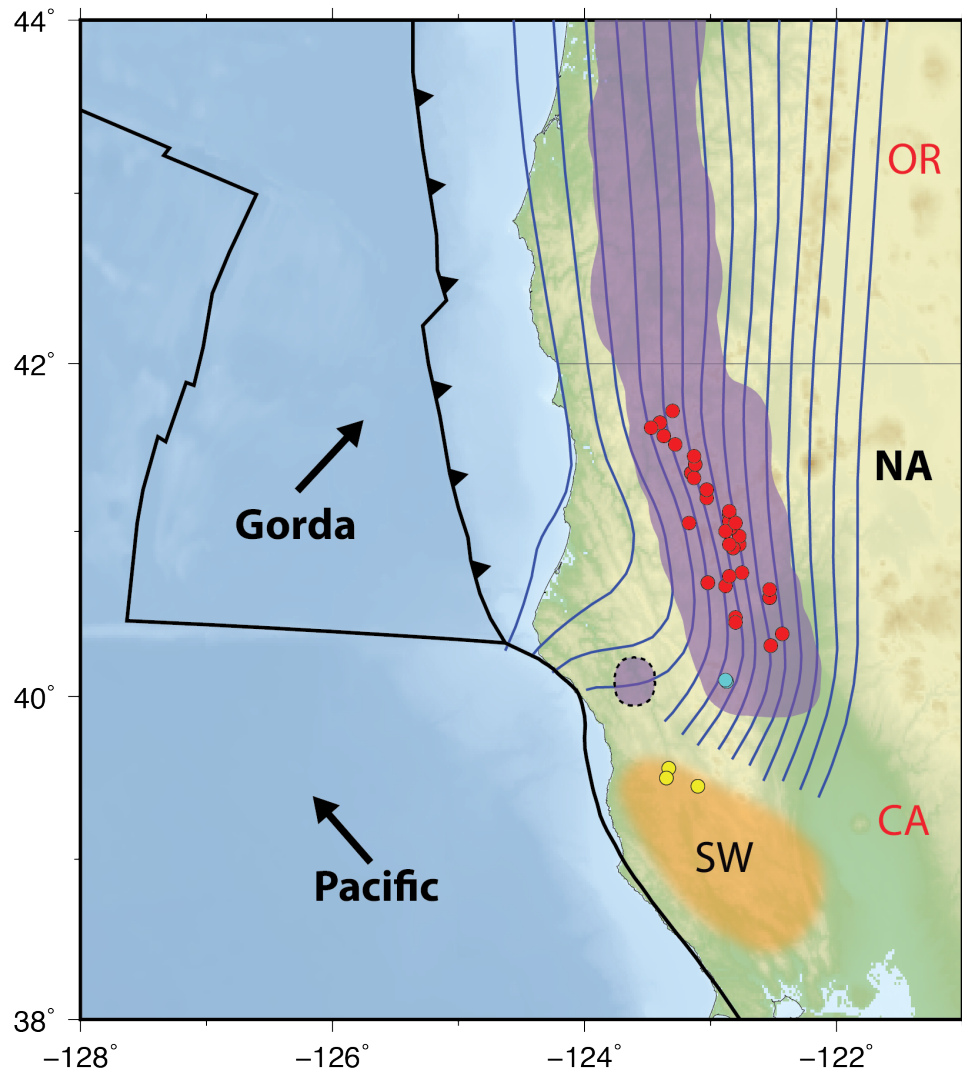


Figure 4.1 Map of the southern Cascadia subduction zone. Black arrows show plate motion relative to North America (NA), blue lines delineate slab depth contours from Blair et al. (2013) every 5 km starting at 15 km depth, purple polygon shows the outline of tremor locations from the Pacific Northwest Seismic Network tremor catalog (<https://pnsn.org/tremor>) with the updip tremor patch from Wech (2021) outlined with a dashed black line. Yellow and red circles show the location of LFE families associated with the San Andreas fault system and megathrust, respectively (Plourde et al. 2015; Ducellier & Creager, 2022). Overlapping cyan circles show an anomalous pair of LFE families with near-continuous temporal activity. Orange region shows the location of the inferred slab window (SW). State names (red lettering): OR, Oregon; CA, California.

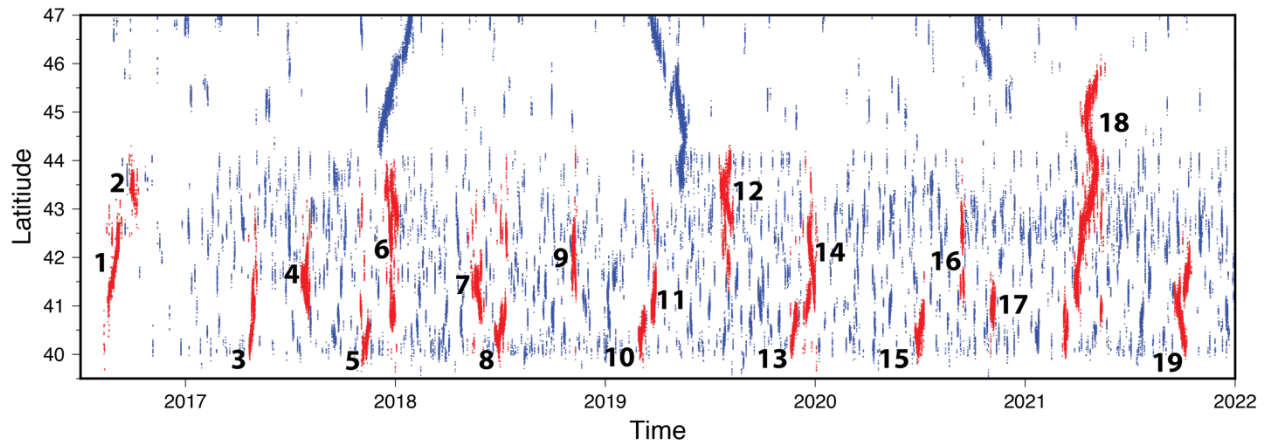


Figure 4.2 Map of tremor locations in southern Cascadia from the Pacific Northwest Seismic Network tremor catalog (<https://pnsn.org/tremor>). Red dots show the tremor that is included in the nineteen labeled ETS events. The California-Oregon border is at a latitude of 42° .

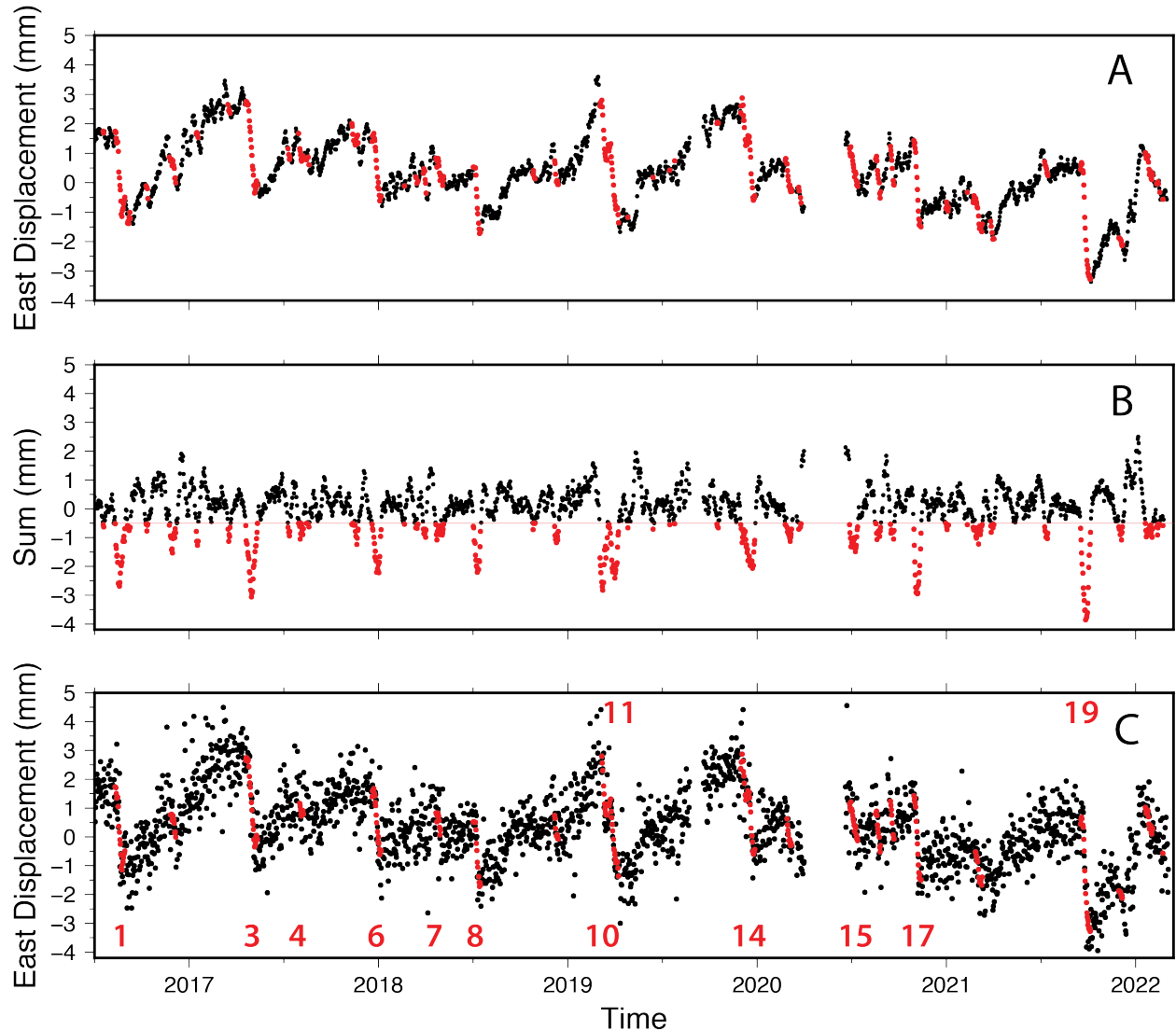


Figure 4.3 Time series plots for station P343 showing the results of the algorithm used to estimate the timing of ETS transients. **A** Fully processed east component time series data (black dots) with all of the initial days flagged by the algorithm shown in red as representing transient displacements. **B** Results from the centered 14-day moving sum of the daily displacements (black dots) with all days falling below the -0.5 mm threshold (dashed red line) shown in red. **C** Processed east component time series data without smoothing (black dots) with all of the transients from panel A with durations longer than week shown in red. Displacements associated with ETS events in this study are labeled with red numbers and correlate with the event numbers from Figure 4.2.

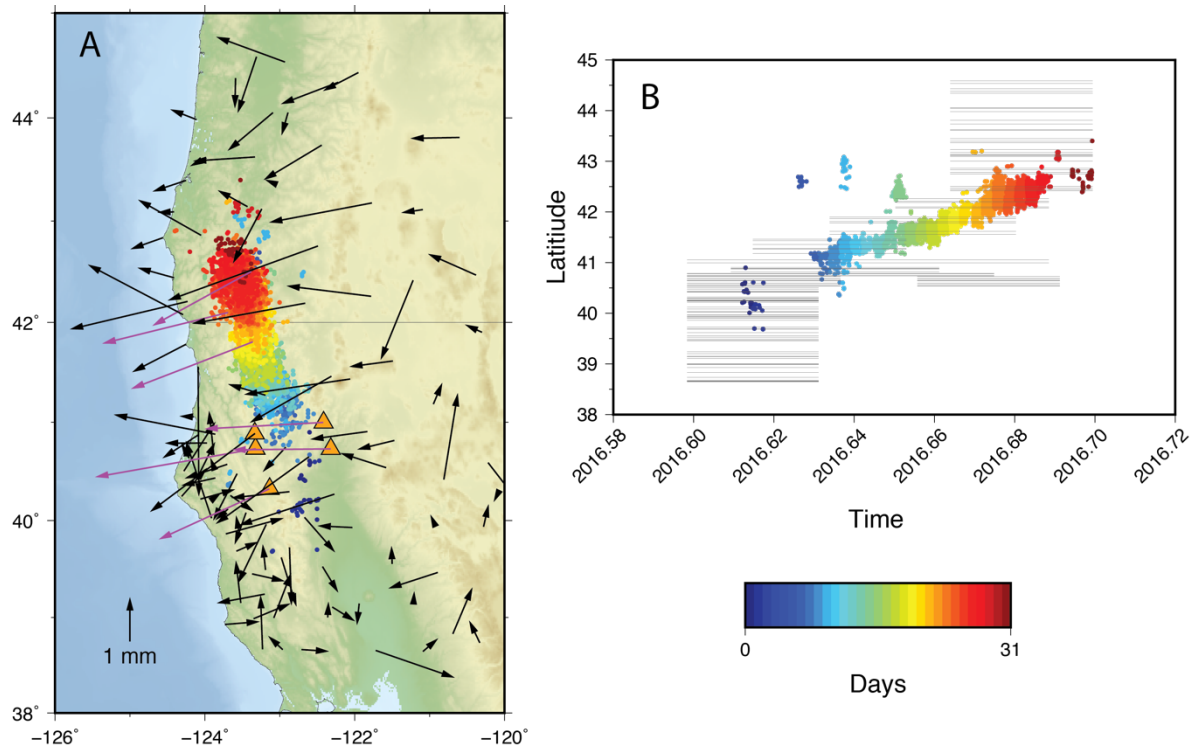


Figure 4.4 GNSS surface displacements and transient times for ETS Event 1 (August 2016) indicated in Figure 2. **A** Map showing tremor locations (colored dots) and GNSS surface displacements (black vectors) for ETS Event 1. Purple vectors show displacements at index stations and orange triangles indicate the locations of manually added index stations. Tremor is colored according to the number of days since the first tremor occurrence, with the scale provided in the bottom right-hand corner. This event exhibits diminished tremor activity to the south despite a clear signal at the GNSS stations. **B** Plot showing the time periods used to calculate the surface displacements for each GNSS in panel A. Tremor for Event 1 is shown with colored dots and uses the same scale as panel A. Each horizontal bar indicates the time spanned by an individual GNSS station and the vertical placement of the bar corresponds to the station's latitude.

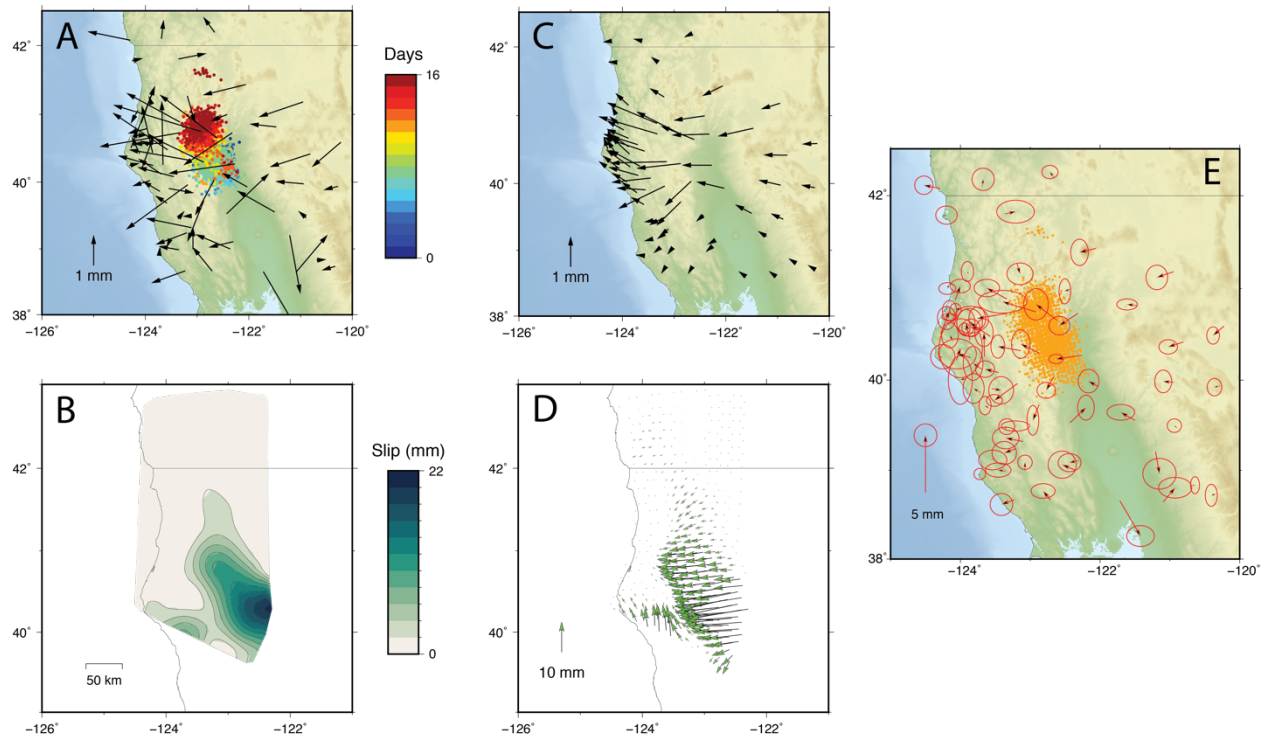


Figure 4.5 Model results for ETS event 15 (June 2020). **A** Vector map showing horizontal surface displacements for ETS event 15. Circles show the location of tremor during the event and are colored by the number of days that passed since the first instance of tremor (June 23, 2020). **B** Distributed slip model from our inversion with the contours (black lines with green shading) outlining every 2 mm of fault slip. **C** Vector map showing the predicted surface displacements resulting from the slip model in panel B. **D** Vector map showing the surface projection of fault slip on the megathrust fault. Arrows are scaled by magnitude. **E** Vector map showing the misfit between the observed and predicted surface displacements and tremor locations with orange circles. Red ellipses show 1σ error.

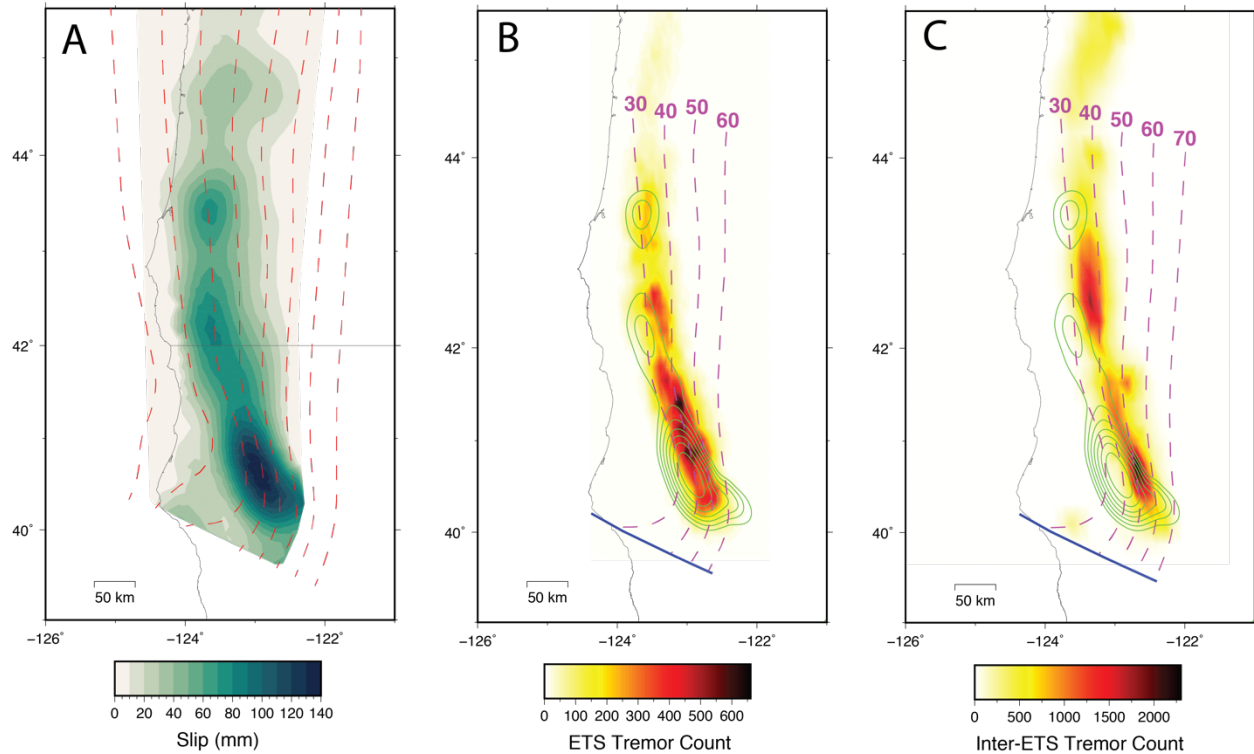


Figure 4.6 Maps of cumulative slip and tremor results for all of the nineteen ETS events in southern Cascadia. **A** Colored contours show the total fault slip magnitudes on the plate interface from summing all nineteen slip models. Dashed red lines delineate slab depth contours at 10km intervals. **B** Background colors show the distribution of tremor during the nineteen ETS events, overlain green contours outline total ETS fault slip above 60 mm from panel A with 20 mm intervals, dashed purple lines and blue line delineate plate depth contours and the southern edge of the Gorda slab from Blair et al. (2013), respectively. **C** Background colors show the distribution of inter-ETS tremor from time periods outside of the nineteen ETS events, overlain green contours outline total slip above 60 mm from panel A with 20 mm intervals, and dashed purple and blue lines delineate plate depth contours and the southern edge of the Gorda slab from Blair et al. (2013), respectively.

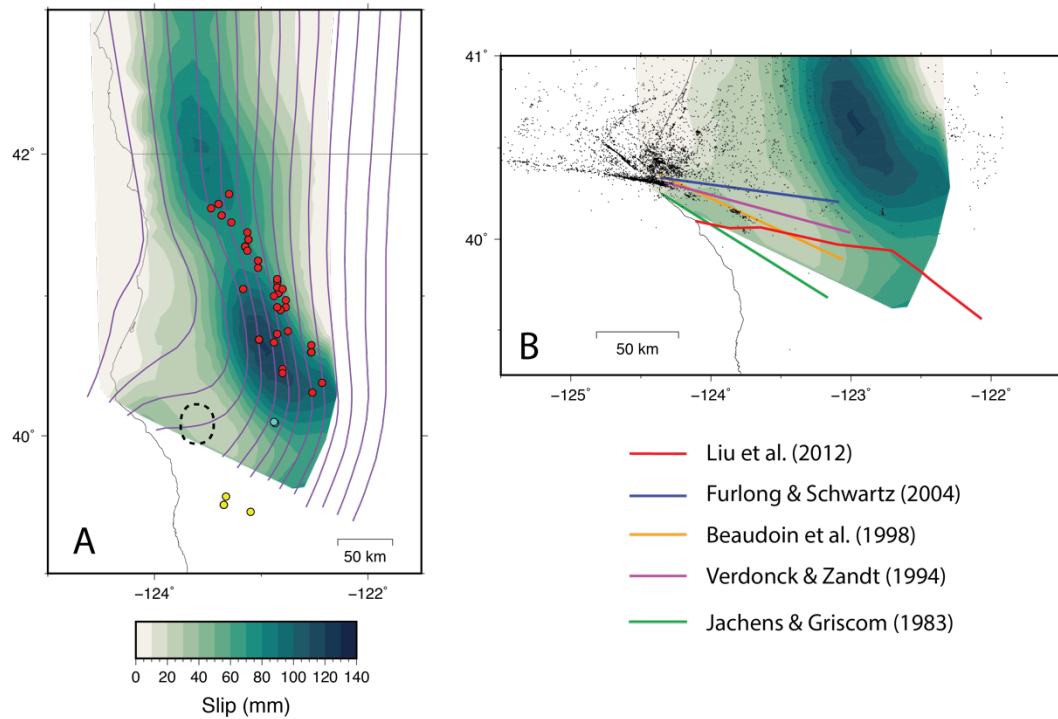


Figure 4.7 Maps of fault slip in southern Cascadia, with colored contours showing the cumulative fault slip magnitudes from the nineteen ETS events in this study. **A** Yellow and red circles show the location of LFE families associated with the San Andreas fault system and megathrust, respectively (Plourde et al., 2015; Ducellier & Creager, 2022). Cyan circles show anomalous LFE families with short recurrence intervals. Black dashed circle outlines the region of near-continuous tremor from Wech (2021). Purple lines delineate slab depth contours from Blair et al. (2013) every 5 km starting at 15 km depth. **B** Colored lines delineate the estimated location of the slab edge from the five studies listed below the panel. Black dots show earthquake locations from Guo et al. (2021).

Chapter 5. SUMMARY AND FUTURE WORK

The Cascadia subduction zone is a well-instrumented plate margin with a dense regional array of seismic and GNSS stations, making this region a prime location to study strain release at a convergent margin. SSEs are recognized as an important mode of strain release in subduction zone systems and are associated with specific conditions along the plate interface, such as high pore-fluid pressures. Therefore, studying the spatiotemporal patterns of SSEs in Cascadia provides insights on the plate interface environment and how it changes along-strike and -dip. Important insights on seismic hazards at convergent margins are provided by the direct observation of earthquakes. However, the Cascadia subduction zone has not hosted a large megathrust earthquake in over three centuries, making it difficult to study past earthquake behavior and characterize seismic hazard in this region. Strain release via SSEs and slip on crustal faults impacts the distribution of strain along the megathrust and influences the generation of megathrust earthquakes and the overall seismic cycle. Therefore, studying the behavior of SSEs and crustal faulting in Cascadia also sheds light on the state of strain along the megathrust and informs our understanding of seismic hazards in the region.

In Chapter 2, I analyzed 13 years of GNSS time series data and discovered evidence for a small long-term SSE in southern Cascadia from 2012.45 to 2014. Furthermore, I found that the maximum horizontal surface displacements and fault slip amplitudes for this event were an order of magnitude smaller than most long-term SSEs at other subduction zones. This result suggests that conditions along the Cascadia subduction zone are somehow different from other systems that are capable of hosting multiple large long-term SSEs. I propose that the size and frequency of long-term SSEs in Cascadia are limited by the down-dip width of the semi-frictional zone, and that southern end of Cascadia may be the only location where this zone is wide enough to host resolvable long-term SSEs for this subduction zone. Future work could test this hypothesis by estimating the width of the semi-frictional zone in southern Cascadia using thermal numerical models similar to Gao and Wang (2017). In this chapter, I also found potential evidence for a second long-term SSE in southern Cascadia that started in early 2017 and was ongoing as of 2019. With more data having been collected since this study, future work could focus on analyzing more recent GNSS time series data from southern Cascadia to test whether this signal

is due to long-term slow slip or some other process. Overall, future studies that focus on longer time series can help identify more long-term SSEs and constrain the recurrence interval for these events. Furthermore, this work would help to better understand what proportion of the slip budget is accommodated by transient slow slip in southern Cascadia.

In Chapter 3, I constructed block models of the Mendocino triple junction (MTJ) region to analyze how crustal faults facilitate interseismic strain in the southern Cascadia forearc. This is the first study to use detailed elastic block models to explore strain partitioning among crustal faults in this region. Furthermore, this study will help constrain fault slip rates that are incorporated into the national seismic hazard maps. Results from this work suggested that the Little Salmon and Mad River fault zones both accommodate some degree of right-lateral slip in the forearc. However, these fault zones are typically regarded as thrust structures, and previous geologic work has focused on measuring the vertical slip rates along these faults. Therefore, future work could involve searching for field evidence of lateral slip along these fault structures to provide independent geologic slip rates for comparison with the slip rate estimates from this chapter. Similar work could also be done for vertical slip rates along the Grogan fault. The block model results from this chapter also show that the Bear River fault zone (BRFZ) has the highest right-lateral and reverse slip rates of any crustal fault in the model. Regional strain rate maps from this work also showed high shear strain rates around the BRFZ, indicating that this region north of the MTJ is experiencing significant strain. However, it is unclear whether the BRFZ experiences high slip rates, or if slip on other nearby structures is collapsed onto the BRFZ in the block models. Therefore, future work could focus on constraining slip rates along the BRFZ and looking for other potential active structures in this region to clarify how strain is accommodated directly north of the MTJ. Lastly, the block models in this chapter were limited in the range of geodynamical processes that were considered. In particular, these elastic models did not consider mantle-lithosphere interactions associated with the slab window. Crustal deformation predicted by the crustal conveyor model was not incorporated into the block model framework (Furlong & Govers, 1999). Therefore, future work could include building detailed finite element models that incorporate dynamic mantle processes and far-field stresses to provide additional independent fault slip estimates.

In Chapter 4, I analyzed 5.5 years of tremor and GNSS time series data and identified 19 large ETS events in southern Cascadia from 2016.5-2022. Distributed slip models for these

events revealed that ETS fault slip and tremor terminate near 41 °N latitude and do not extend to the southern edge of the subduction zone. Therefore, I hypothesized that ETS does not occur in southernmost Cascadia due to the geometry of the Gorda slab and increased temperatures near the edge of the subduction zone. However, there is still uncertainty surrounding the structure of the Gorda slab at the edge of the subduction zone and how the thermal gradient of the slab changes as a function of distance from the slab window. As a result, potential avenues for future research include 1) refining the geometry of the Gorda slab, 2) modeling the thermal state of the slab in response to the slab window, and 3) testing how different slab geometries impact the thermal evolution of the slab. Overall, this work elucidates the complex nature of the lateral slab edge, and provides an avenue to separate out the dominant controls on ETS behavior.

This work focuses specifically on the Cascadia subduction zone, but the results presented in this dissertation can be applied and tested at other subduction zone systems around the world. For instance, this work demonstrates that crustal faults impact the accumulation of strain along the megathrust in southern Cascadia. Future work at other subduction zones could extend this finding by analyzing the characteristics of past megathrust earthquakes and seeing if their behavior is influenced by the distribution of active crustal faults along the margin. Also, many subduction zones contain lateral slab edges that mark the end of the convergent margin. Therefore, future studies could analyze ETS near these edges to help refine our understanding of what properties influence ETS behaviors in subduction zone systems.

BIBLIOGRAPHY

- Almendinger, R.W., Reilinger, R., & Loveless, J. (2007), Strain and rotation rate from GPS in Tibet, Anatolia, and the Alitplano. *Tectonics*, 26(3).
<https://doi.org/10.1029/2006TC002030>
- Atwater, T. (1970), Implications of Plate Tectonics for the Cenozoic tectonic evolution of western North America. *GSA Bulletin*, 81(12), 3513-3536. [https://doi.org/10.1130/0016-7606\(1970\)81\[3513:IOPTFT\]2.0.CO;2](https://doi.org/10.1130/0016-7606(1970)81[3513:IOPTFT]2.0.CO;2)
- Argus, D.F. & Gordon, R.G. (2001), Present tectonic motion across the Coast Ranges and San Andreas fault system in central California. *Geological Society of America Bulletin*, 113(12), 1580-1592. [https://doi.org/10.1130/0016-7606\(2001\)113<1580:PTMATC>2.0.CO;2](https://doi.org/10.1130/0016-7606(2001)113<1580:PTMATC>2.0.CO;2)
- Argus, D.F., Landerer, F.W., Wiese, D.N., Martens, H.R., Fu, Y., Famiglietti, J.S., Thomas, B.F., Farr, T.G., Moore, A.W., & Watkins, M.W. (2017), Sustained water loss in California's mountain ranges during severe drought from 2012 to 2015 inferred from GPS. *Journal of Geophysical Research: Solid Earth*, 122(12), 10559-10585.
<https://doi.org/10.1002/2017JB014424>
- Audet, P., Bostock, M.G., Christensen, N.I. & Peacock, S.M. (2009). Seismic evidence for overpressured subducted oceanic crust and megathrust fault sealing. *Nature*, 457, 76-78.
<https://doi.org/10.1038/nature07650>
- Audet, P. & Kim, Y. (2016). Teleseismic constraints on the geological environment of deep episodic slow earthquakes in subduction zone forearcs: A review. *Tectonophysics*, 670(22), 1-15. <https://doi.org/10.1016/j.tecto.2016.01.005>
- Audet, P. & Schaeffer, A.J. (2018). Fluid pressure and shear zone development over the locked to slow slip region in Cascadia. *Science Advances*, 4(3). doi:10.1126/sciadv.aar2982
- Barbot, S., Fialko, Y., & Sandwell, D. (2008). Effect of a compliant fault zone on the inferred earthquake slip distribution. *Journal of Geophysical Research*, 113.
doi:10.1029/2007JB005256
- Bartlow, N.M. (2020), A Long-term View of Episodic Tremor and Slip in Cascadia. *Geophysical Research Letters*, 47(3). <https://doi.org/10.1029/2019GL085303>.

- Beaudoin, B.C., Hole, J.A., Klemperer, S.L. & Trehu, A.M. (1998). Location of the southern edge of the Gorda slab and evidence for an adjacent asthenospheric window: Results from seismic profiling and gravity. *Journal of Geophysical Research*, 103, 30101-30115. doi:10.1029/98JB02231.
- Beeson, J.W., Johnson, S.Y., & Goldfinger, C. (2017), The transtensional offshore portion of the northern San Andreas fault: Fault zone geometry, late-Pleistocene to Holocene sediment deposition, shallow deformation patterns, and asymmetric basin growth. *Geosphere*, 13(4), 1173-1206. doi:10.1130/GES01367.1
- Bennett, G. L., Miller, S. R., Roering, J. J., & Schmidt, D. A. (2016), Landslides, threshold slopes, and the survival of relict terrain in the wake of the Mendocino Triple Junction. *Geology*, 44(5), 363-366. <https://doi.org/10.1130/G37530.1>
- Benz, H.M., Zandt, G., & Oppenheimer, D.H. (1992). Lithospheric structure of northern California from teleseismic images of the upper mantle. *Journal of Geophysical Research: Solid Earth*, 97, 4791-4807. <https://doi.org/10.1029/92JB00067>
- Blair, J.L., McCrory, P.A., Oppenheimer, D.H., & Waldhauser, F. (2013), A Geo-referenced 3D model of the Juan de Fuca Slab and associated seismicity, U.S. Geological Survey Data Series 633, v.1.2. <https://pubs.usgs.gov/ds/633/>
- Bletery, Q., Thomas, A.M., Rempel, A.W., Sladen, A., & de Barros, L. (2016). Mega-earthquakes rupture flat megathrusts. *Science*, 354(6315), 1027-1031. DOI: 10.1126/science.aag0482
- Bletery, Q., Thomas, A.M., Rempel, A.W., & Hardebeck, J.L. (2017). Imaging shear strength along subduction faults. *Geophysical Research Letters*, 44(22), 11329-11339. <https://doi.org/10.1002/2017GL075501>
- Blewitt, G., Hammond, W.C. & Kreemer, C. (2018). Harnessing the GPS data explosion for interdisciplinary science, *Eos*, 99, <https://doi.org/10.1029/2018EO104623>
- Bodmer, M., Toomey, D.R., Hooft, E.E.E. & Schmandt, B. (2018). Buoyant asthenosphere beneath Cascadia influences megathrust segmentation. *Geophysical Research Letters*, 45(14), 6954-6962. <https://doi.org/10.1029/2018GL078700>
- Boyarko, D.C. & Brudzinski, M.R. (2010). Spatial and temporal patterns of nonvolcanic tremor along the southern Cascadia subduction zone. *Journal of Geophysical Research: Solid Earth*, 115. <https://doi.org/10.1029/2008JB006064>

- Brudzinski, M.R. & Allen, R.M. (2007). Segmentation in episodic tremor and slip all along Cascadia. *Geology*, 35(10), 907-910, <https://doi.org/10.1130/G23740A.1>
- Chaytor, J.D., Goldfinger, C., & Dziak, R.P. (2004), Active deformation of the Gorda plate: Constraining deformation models with new geophysical data. *Geology*, 32(4), 353-356, doi:10.1130/G20178.1.
- Clarke, S.H. (1992), Geology of the Eel River Basin and Adjacent Region: Implications for Late Cenozoic tectonics of the southern Cascadia Subduction Zone and Mendocino Triple Junction. *AAPG Bulletin*, 76, 199-224.
- Clarke, S.H., & Carver, G.A. (1992), Late Holocene tectonics and Paleoseismicity, southern Cascadia subduction zone. *Science*, 225(5041), 188-192. doi:10.1126/science.255.5041.188
- Cleveland, R.B., Cleveland, W.S., McRae, J.E., & Terpenning, I. (1990). STL: A seasonal-trend decomposition procedure based on loess. *Journal of Official Statistics*, 6(1), 3-73.
- Condit, C.B., Guevara, V.E., Delph, J.R., & French, M.E. (2020). Slab dehydration in warm subduction zones at depths of episodic slip and tremor. *Earth and Planetary Science Letters*, 15. <https://doi.org/10.1016/j.epsl.2020.116601>
- Crowell, B.W. (2019), PyTAGS: A Python package for the temporal analysis of GNSS strains, V1.0, doi: 10.5281/zenodo.2634525
- Delph, J.R., Thomas, A.M., & Levander, A. (2021). Subcretionary tectonics: Linking variability in the expression of subduction along the Cascadia forearc. *Earth and Planetary Science Letters*, 556. <https://doi.org/10.1016/j.epsl.2020.116724>
- Dickinson, W.R., & Snyder, W.S. (1979a), Geometry of triple junctions related to San Andreas transform. *Journal of Geophysical Research*, 84, 561-572. <https://doi.org/10.1029/JB084iB02p00561>
- Dickinson, W.R., & Snyder, W.S. (1979b), Geometry of subducted slabs related to San Andreas transform. *The Journal of Geology*, 87(6), 609-627. <https://doi.org/10.1086/628456>
- Dixon, T.H., Miller, M., Farina, F., Wang, H., & Johnson, D. (2000), Present-day motion of the Sierra Nevada block and some tectonic implications for the Basin and Range province, North American Cordillera. *Tectonics*, 19, 1-24. <https://doi.org/10.1029/1998TC001088>

- DeMets, C., Gordon, R.G., & Argus, D.F. (2010). Geologically current plate motions. *Geophysical Journal International*, 118(1), 1-80. <https://doi.org/10.1111/j.1365-246X.2009.04491.x>
- Dong, D., Fang, P., Bock, Y., Webb, F., Prawirodirdjo, L., Kedar, S., & Jamason, P. (2006). Spatiotemporal filtering using principal component analysis and Karhunen-Loeve expansion approaches for regional GPS network analysis. *Journal of Geophysical Research: Solid Earth*, 111. <https://doi.org/10.1029/2005JB003806>
- Ducellier, A. & Creager, K.C. (2022). An 8-year-long low-frequency earthquake catalog for southern Cascadia. *Journal of Geophysical Research: Solid Earth*, 127(4). <https://doi.org/10.1029/2021JB022986>
- Evans, E.L., Loveless, J.P., & Meade, B.J. (2015). Total variation regularization of geodetically and geologically constrained block models for the Western United States. *Geophysical Journal International*, 202(2), 713-727. <https://doi.org/10.1093/gji/ggv164>
- Field, E.H., Biasi, G.P., Bird, P., Dawson, T.E., Felzer, K.R., Jackson, D.D., Johnson, K.M., Jordan, T.H., Madden, C., Michael, A.J., Milner, K.R., Page, M.T., Parsons, T., Powers, P.M., Shaw, B.E., Thatcher, W.R., Weldon, R.J., II, & Zeng, Y. (2013), Uniform California earthquake rupture forecast, version 3 (UCERF3) - The time-independent model, U.S. Geological Survey Open-File Report 2013–1165, 97 p., California Geological Survey Special Report 228, and Southern California Earthquake Center Publication 1792. <http://pubs.usgs.gov/of/2013/1165/>
- Freymueller, J.T., Murray, M.H., Segall, P., & Castillo, D. (1999), Kinematics of the Pacific-North America plate boundary zone, northern California. *Journal of Geophysical Research: Solid Earth*, 104, 7419-7441. <https://doi.org/10.1029/1998JB900118>.
- Fu, Y. & Freymueller, J.T. (2013). Repeated large slow slip events at the southcentral Alaska subduction zone. *Earth and Planetary Science Letters*, 375, 303-311. <https://doi.org/10.1016/j.epsl.2013.05.049>
- Fu, Y., Argus, D.F., & Landerer, F.W. (2015). GPS as an independent measure to estimate terrestrial water storage variations in Washington and Oregon. *Journal of Geophysical Research Solid Earth*, 120, 552-566. doi:10.1002/2014JB011415

- Furlong, K.P., & Govers, R. (1999), Ephemeral crustal thickening at a triple junction: The Mendocino crustal conveyor. *Geology*, 27(2), 127-130. doi:10.1130/0091-7613(1999)027<0127:ECTAAT>2.3CO;2.
- Gao, X. & Wang, K. (2017). Rheological separation of the megathrust seismogenic zone and episodic tremor and slip. *Nature*, 543, 416-419. <https://doi.org/10.1038/nature21389>
- Godfrey, N.J., Beaudoin, B.C., Lendl, C., Meltzer, A. & Luetgert, J. (1995). Data report for the 1993 Mendocino triple junction seismic experiment. *U.S. Geological Survey Open-File Report*, 95-275. doi:10.3133/ofr95275
- Goes, S., Govers, R., Schwartz, S. & Furlong, K. (1997). Three-dimensional thermal modeling for the Mendocino Triple Junction area. *Earth and Planetary Science Letters*, 148, 45-47. [https://doi.org/10.1016/S0012-821X\(97\)00044-7](https://doi.org/10.1016/S0012-821X(97)00044-7)
- Gomberg, J. & Cascadia 2007 and Beyond Working Group (2010). Slow-slip phenomena in Cascadia from 2007 and beyond: A review. *Geological Society of America Bulletin*, 122(7-8), 963-978. <https://doi.org/10.1130/B30287.1>
- Groome, W.G. & Thorkelson, D.J. (2007). The three-dimensional thermos-mechanical signature of ridge subduction and slab window migration. *Tectonophysics*, 464, 70-83. doi:10.1016/j.tecto.2008.07.003
- Gulick, S.P.S. & Meltzer A.S. (2002), Effect of the northward-migrating Mendocino triple junction on the Eel River forearc basin, California: Structural evolution. *GSA Bulletin*, 114, p. 1505 - 1519.
- Guo, H., McGuire, J.J., & Zhang, H. (2021). Correlation of porosity variations and rheological transitions on the southern Cascadia megathrust. *Nature Geoscience*, 14, 341-348. <https://doi.org/10.1038/s41561-021-00740-1>
- Guzofski, C.A. & Furlong, K.P. (2002). Migration of the Mendocino triple junction and ephemeral crustal deformation: Implications for California Coast Range heat flow. *Geophysical Research Letters*, 29(1). <https://doi.org/10.1029/2001GL013614>
- Haines, J., Wallace, L.M., & Dimitrova, L. (2019). Slow slip event detection in Cascadia using vertical derivatives of horizontal stress rates. *Journal of Geophysical Research: Solid Earth*, 124, 5153-5173. <https://doi.org/10.1029/2018JB016898>

- Hammond, W.C., Blewitt, G. & Kreemer, C. (2016), GPS Imaging of vertical land motion in California and Nevada: Implications for the Sierra Nevada uplift. *Journal of Geophysical Research Solid Earth*, 121, 7681-7703. doi:10.1002/2016JB013458
- Hanks, T.C. & Kanamori, H. (1979), A moment magnitude scale. *Journal of Geophysical Research Solid Earth*, 84, 2348-2350. <https://doi.org/10.1029/JB084iB05p02348>
- Hawthorne, J.C. & Rubin, A.M. (2010). Tidal modulation of slow slip in Cascadia. *Journal of Geophysical Research: Solid Earth*, 115. <https://doi.org/10.1029/2010JB007502>
- Herring, T.A., Melbourne, T.I., Murray, M.H., Floyd, M.A., Szeliga, W.M., King, R.W., Phillips, D.A., Puskas, C.M., Santillan, M. & Wang, L. (2016), Plate Boundary Observatory and related networks: GPS data analysis methods and geodetic products. *Reviews of Geophysics*, 54(4), 759-808. <https://doi.org/10.1002/2016RG000529>
- Hirose, H., Hirahara, K., Kimata, F., Fujii, N., & Miyazaki, S. (1999). A slow thrust slip event following the two 1996 Hyuganada earthquakes beneath the Bungo Channel, southwest Japan. *Geophysical Research Letters*, 26(21), 3237-3240. <https://doi.org/10.1029/1999GL010999>
- Hirose, H. & Obara, K. (2005). Repeating short- and long-term slow slip events with deep tremor activity around the bungo channel region southwest Japan. *Earth, Planets and Space*, 57, 961-972 .
- Hutchinson, J., Kao, H., Riedel, M., Obama, K., Wang, K., Kodaira, S., Takahashi, T., & Yamamoto, Y. (2020). Significant geometris variation of the subducted plate beneath the northernmost Cascadia subduction zone and its tectonic implications as revealed by the 2014 Mw 6.4 earthquake sequence. *Earth and Planetary Science Letters*, 555. <https://doi.org/10.1016/j.epsl.2020.116569>
- Hyndman, R.D. & Wang, K. (1993). Thermal constraints on the zone of major thrust earthquake failure: The Cascadia Subduction Zone. *Journal of Geophysical Research: Solid Earth*, 98, 2039-2060. <https://doi.org/10.1029/92JB02279>
- Hyndman, R.D. (2013). Downdip landward limit of Cascadia great earthquake rupture. *Journal of Geophysical Research Solid Earth*, 188, 5530-5549. <https://doi.org/10.1002/jgrb.50390>
- Hyndman, R.D., McCrory, P.A., Wech, A., Kao, H., & Ague, J. (2015). Cascadia subducting plate fluids channeled to fore-arc mantle corner: ETS and silica deposition. *Journal of*

- Geophysical Research: Solid Earth*, 120(6), 4344-4358.
<https://doi.org/10.1002/2015JB011920>
- Ide, S. (2012). Variety and spatial heterogeneity of tectonic tremor worldwide. *Journal of Geophysical Research*, 117, B03302. doi:10.1029/2011JB008840
- Ide, S., Shelly, D.R., & Beroza, G.C. (2007). Mechanism of deep low frequency earthquakes: Further evidence that deep non-volcanic tremor is generated by shear slip on the plate interface. *Geophysical Research Letters*, 34(4). <https://doi.org/10.1029/2006GL028890>
- Ismat, Z., Putera, H., & Patzkowski, S. (2022). Internal deformation of the Gorda plate and its tectonic significance within the Cascadia subduction zone. *Journal of Structural Geology*, 161. <https://doi.org/10.1016/j.jsg.2022.104643>
- Ito, Y., Obara, K., Shiomi, K., Sekine, S., & Hirose, H. (2007). Slow earthquakes coincident with episodic tremors and slow slip events. *Science*, 315(5811), 503-506.
doi:10.1126/science.1134454
- Jachens, R.C. & Griscom, A. (1983). Three-dimensional geometry of the Gorda Plate beneath northern California. *Journal of Geophysical Research: Solid Earth*, 88, 9375-9392.
<https://doi.org/10.1029/JB088iB11p09375>
- Jamason, P., Yehuda, B., Peng, F., Gilmore, B., Malveaux, D., Prawirodirdjo, L., & Scharber, M. (2004). SOPAC Web site (<http://sopac.ucsd.edu>). *GPS Solutions*, 8, 272-277.
<https://doi.org/10.1007/s10291-004-0118-2>
- Jiang, Y., Wdowinski, S., Dixon, T.H., Hackl, M., Protti, M. & Gonzalez, V. (2012). Slow slip events in Costa Rica detected by continuous GPS observations, 2002-2011. *Geochemistry, Geophysics, Geosystems*, 13(4), Q04006.
<https://doi.org/10.1029/2012GC004058>
- Kano, M. & Kato, A. (2020). Detailed spatial slip distribution for short-term slow slip events along the Nankai subduction zone, southwest Japan. *Journal of Geophysical Research: Solid Earth*, 125(7). <https://doi.org/10.1029/2020JB019613>
- Kao, H., Shan, S.-J. Dragert, H. & Rogers, G. (2009). Northern Cascadia episodic tremor and slip: A decade of tremor observations from 1997 to 2007. *Journal of Geophysical Research*, 114. <https://doi.org/10.1029/2008JB006046>

- Kaproth, B.M. & Marone, C. (2013). Slow earthquakes, preseismic velocity changes, and the origin of slow frictional stick-slip. *Science*, *341*, 1229-1232.
doi:10.1126/science.1239577
- Katayama, I., Terada, T., Okazaki, K., & Tanikawa, W. (2012). Episodic tremor and slow slip potentially linked to permeability contrasts at the Moho. *Nature Geoscience*, *5*, 731-734.
<https://doi.org/10.1038/ngeo1559>
- Kato, A. & Nakagawa, S. (2014). Multiple slow-slip events during a foreshock sequence of the 2014 Iquique, Chile Mw 8.1 earthquake. *Geophysical Research Letters*, *41*(15), 5420-5427. <https://doi.org/10.1002/2014GL061138>
- Kelsey, H.M. & Hagens, D.K. (1982). Major right-lateral faulting in the Franciscan assemblage of northern California in late Tertiary time. *Geology*, *10*(7), 387-391.
[https://doi.org/10.1130/0091-7613\(1982\)10<387:MRFITF>2.0.CO;2](https://doi.org/10.1130/0091-7613(1982)10<387:MRFITF>2.0.CO;2)
- Kelsey, H.M. & Carver, G.A. (1988). Late Neogene and Quaternary tectonics associated with northward growth of the San Andreas Transform Fault, northern California. *Journal of Geophysical Research: Solid Earth*, *93*, 4797-4819.
<https://doi.org/10.1029/JB093iB05p04797>
- Kelsey, H.M. (2001). Active faulting associated with the southern Cascadia subduction zone in northern California. *Engineering Geology Practice in Northern California, Division of Mines and Geology Bulletin 210, Association of Engineering Geologists Special Publication 12*, p. 259-274.
- Kirby, S.H., Wang, K., & Brocher, T.M (2014). A large mantle water source for the northern San Andreas fault system: a ghost of subduction past. *Earth, Planets and Space*, *66*(67).
<https://doi.org/10.1186/1880-5981-66-67>
- Kobayashi, A. & Tsuyuki, T. (2019). Long-term slow slip event detected beneath the Shina Peninsula, central Japan, from GNSS data. *Earth, Planets, and Space*, *70*(60).
<https://doi.org/10.1186/s40623-019-1037-3>
- Kobayashi, T. & Sato, T. (2021). Estimating effective normal stress during slow slip events from slip velocities and shear stress variations. *Geophysical Research Letters*, *48*(20).
<https://doi.org/10.1029/2021GL095690>

- Lachenbruch, A.H. & Sass, J.H. (1980). Heat flow and energetics of the San Andreas Fault Zone. *Journal of Geophysical Research: Solid Earth*, 85, 6185-6222.
<https://doi.org/10.1029/JB085iB11p06185>
- Ladinsky, T., Kelsey, H., Michalak, M., Witter, R., Sherrod, B., Bold, S., Cashman, S., & Buck, J. (2020). In southern Cascadia, do upper plate faults rupture in concert with subduction zone earthquakes: A paleoseismic investigation of the Little Salmon fault zone, National Earthquake Hazards Reduction Program Final Technical Report, USGS.
- Langbein, J. (2008). Noise in GPS displacement measurements from southern California and southern Nevada. *Journal of Geophysical Research Solid Earth*, 113, B05405.
[doi:10.1029/2007JB005247](https://doi.org/10.1029/2007JB005247)
- Larson, K.M., Kostoglodov, V., Miyazaki, S., & Santiago, J.A.S. (2007). The 2006 asiesmic slow slip event in Guerrero, Mexico: New results from GPS. *Geophysical Research Letters*, 34, L13309. [doi:10.1029/2007GL029912](https://doi.org/10.1029/2007GL029912)
- Leeman, J.R., Saffer, D.M., Scuderi, M.M. & Marone, C. (2016). Laboratory observations of slow earthquakes and the spectrum of tectonic fault slip modes. *Nature Communications*, 7, 11104. <https://doi.org/10.1038/ncomms11104>
- Lowry, A.R., Larson, K.M., Kostoglodov, V. & Bilham, R. (2001). Transient fault slip in Guerrero, southern Mexico. *Geophysical Research Letters*, 28(19), 3753-3756.
<https://doi.org/10.1029/2001GL013238>
- Li, D. & Liu, Y. (2016). Spatiotemporal evolution of slow slip events in a nonplanar fault model for northern Cascadia subduction zone. *Journal of Geophysical Research: Solid Earth*, 121(9), 6828-6845. <https://doi.org/10.1002/2016JB012857>
- Li, S., Freymueller, J., & McCaffrey, R. (2016). Slow slip events and time-dependent variations in locking beneath Lower Cook Inlet of the Alaska-Aluetian subduction zone. *Journal of Geophysical Research: Solid Earth*, 121, 1060-1079. [doi:10.1002/2015JB012491](https://doi.org/10.1002/2015JB012491)
- Li, S., Wang, K., Wang, Y., Jiang, Y., & Dosso, S.E. (2018). Geodetically inferred locking locking state of the Cascadia megathrust based on a viscoelastic earth model. *Journal of Geophysical Research Solid Earth*, 123(9), 8056-8072.
<https://doi.org/10.1029/2018JB015620>
- Lienkaemper, J.J., McFarland, F.S., Simpson, R.W., & Caskey, S.J. (2014). Using surface creep rate to infer fraction locked for sections of the San Andreas fault system in northern

- California from alignment and GPS data. *Bulletin of the Seismological Society of America*, 104(6). doi:10.1785/0120140117
- Liu, K., Levander, A., Zhai, Y., Porritt, R. W., & Allen, R. M. (2012), Asthenospheric flow and lithospheric evolution near the Mendocino Triple Junction. *Earth and Planetary Science Letters*, 323, 60-71. <https://doi.org/10.1016/j.epsl.2012.01.020>
- Materna, K., Bartlow, N., Wech, A., Williams, C. & Roland, B. (2019). Dynamically triggered changes of plate interface coupling in southern Cascadia. *Geophysical Research Letters*, 46(22), 12890-12899. doi:10.1029/2019GL084395
- Matsuzawa, T., Shibazaki, B., Obara, K. & Hirose, H. (2013). Comprehensive model of short- and long-term slow slip events in the Shikoku region of Japan, incorporating a realistic plate configuration. *Geophysical Research Letters*, 40(19), 5125-5130. <https://doi.org/10.1002/grl.51006>
- McCaffrey, R. (2002), Crustal Block Rotations and Plate Coupling, in *Plate Boundary Zones, Geodynamic Series*, 30, Stein S. Freymueller J.T. (Editors), 101-119. <https://doi.org/10.1029/GD030p0101>
- McCaffrey, R. (2005), Block kinematics of the Pacific-North America plate boundary in the southwestern United States from inversion of GPS, seismological, and geologic data. *Journal of Geophysical Research: Solid Earth*, 110, B07401. doi:10.1029/2004JB003307.
- McCaffrey, R., Qamar, A.I., King, R.W., Wells, R., Khazaradze, G., Williams, C.A., Stevens, C.W., Vollick, J.J., & Zwick, P.C. (2007), Fault locking, block rotation and crustal deformation in the Pacific Northwest. *Geophysical Journal International*, 169(3), 1315-1340. <https://doi.org/10.1111/j.1365-246X.2007.03371.x>
- McCaffrey, R. (2009), Time-dependent inversion of three-component continuous GPS for steady and transient sources in northern Cascadia. *Geophysical Research Letters*, 36(7), L07304. doi:10.1029/2008GL036784
- McCaffrey, R., King, R.W., Payne, S.J., & Lancaster, M. (2013), Active tectonics of northwestern U.S. inferred from GPS-derived surface velocities. *Journal of Geophysical Research: Solid Earth*, 118(2), 709-723. doi: 10.1029/2012JB009473.
- McCrory, P.A. (1996), Evaluation of fault hazards, northern coastal California. *USGS Open-File Report 96-656*, 87 pp. doi:10.3133/ofr96656

- McCrory, P.A. (2000), Upper plate contraction north of the migrating Mendocino triple junction, northern California: Implications for strain partitioning. *Tectonics*, 19, 1144-1160.
- McCrory, P. A., Blair, J. L., Oppenheimer, D. H. & Walter S. R. (2006). Depth to the Juan De Fuca Slab beneath the Cascadia Subduction Margin—A 3-D model for sorting earthquakes, *U.S. Geologic Survey, Data Series 91*.
- McCrory, P.A., Blair, J.L., Waldhauser, F., Oppenheimer, D.H. (2012), Juan de Fuca slab geometry and its relation to the Wadati-Benioff zone seismicity. *Journal of Geophysical Research: Solid Earth*, 117, B09306. doi:10.1029/2012JB009407
- McKenzie, K.A., Furlong, K.P., & Herman, M.W. (2020). Bidirectional loading of the subduction interface: Evidence from the kinematics of slow slip events. *Geochemistry, Geophysics, Geosystems*, 21(9). <https://doi.org/10.1029/2020GC008918>
- McKenzie, K. A., & Furlong, K. P. (2021), Isolating non-subduction-driven tectonic processes in Cascadia. *Geoscience Letters*, 8(1), 1-12. <https://doi.org/10.1186/s40562-021-00181-z>
- Meade, B.J. & Loveless, J.P. (2009), Block modeling with connected fault-network geometries and a linear elastic coupling estimator in spherical coordinates. *Bulletin of the Seismological Society of America*, 99(6), 3124-3119. <https://doi.org/10.1785/0120090088>
- Merritts, D., & Bull, W.B. (1989), Interpreting Quaternary uplift rates at the Mendocino triple junction, northern California, from uplifted marine terraces. *Geology*, 17(11), 1020-1024. [https://doi.org/10.1130/0091-7613\(1989\)017<1020:IQRAT>2.3.CO;2](https://doi.org/10.1130/0091-7613(1989)017<1020:IQRAT>2.3.CO;2)
- Michel, S., Gualandi, A., & Avouac, J. (2018). Interseismic coupling and slow slip events on the Cascadia megathrust. *Pure and Applied Geophysics*, 176, 3867-3891. <https://doi.org/10.1007/s00024-018-1991-x>
- Mocko, D. & NASA/GSFC/HSL (2012). NLDAS Noah Land Surface Model L4 Monthly 0.125 x 0.125 degree V002, Greenbelt, Maryland USA, Goddard Earth Sciences Data and Information Services Center (GES DISC). doi:10.5067/NOXZSD0Z6JGD
- Morishige, M. & van Keken, P.E. (2017). Along-arc variation in short-term slow slip events caused by 3-D fluid migration in subduction zones. *Journal of Geophysical Research: Solid Earth*, 122(2), 1434-1448. <https://doi.org/10.1002/2016JB013091>
- Mullen, E.K. & Weis, D. (2015). Evidence for trench-parallel mantle flow in the northern Cascade Arc from basalt geochemistry. *Earth and Planetary Science Letters*, 414, 100-107. <https://doi.org/10.1016/j.epsl.2015.01.010>

- Murray, J.R., Minson, S.E. & Svarc, J.L. (2014), Slip rates and spatially variable creep on faults of the northern San Andreas system inferring through Bayesian inversion of Global Positioning System data. *Journal of Geophysical Research: Solid Earth*, 119, 6023-6047. doi:10.1002/2014JB010966
- Nishimura, T., Matsuzawa, T., & Obara, K. (2013). Detection of short-term slow slip events along the Nankai Trough, southwest Japan, using GNSS data. *Journal of Geophysical Research: Solid Earth*, 118(6), 3112-3125. <https://doi.org/10.1002/jgrb.50222>
- Nuyen, C.P. & Schmidt, D.A. (2021), Filling the gap in Cascadia: the emergence of low-amplitude long-term slow slip. *Geochemistry, Geophysics, Geosystems*, 22(3). <https://doi.org/10.1029/2020GC009477>
- Obara, K. (2011). Characteristics and interactions between non-volcanic tremor and related slow earthquakes in the Nankai subduction zone, southwest Japan. *Journal of Geodynamics*, 52, 229-248. <https://doi.org/10.1016/j.jog.2011.04.002>
- Obara, K. & Kato, A. (2016). Connecting slow earthquakes to huge earthquakes. *Science*, 353(6296), 253-257. doi: 10.1126/science.aaf1512
- Ohta, Y., Freymueller, J.T., Hreinsdottir, S. & Suito, H. (2006). A large slow slip event and the depth of the seismogenic zone in the south central Alaska subduction zone. *Earth and Planetary Science Letters*, 247, 108-116. <https://doi.org/10.1016/j.epsl.2006.05.013>
- Okada, Y. (1985), Surface deformation due to shear and tensile faults in a half-space. *Bulletin of the Seismological Society of America*, 75(4), 1135-1154.
- Okada, Y. (1992). Internal deformation due to shear and tensile faults in a half-space, *Bulletin of the Seismological Society of America*, 82(2), 1018-1040.
- Oleskevich, D.A., Hyndman, R.D. & Wang, K. (1999). The updip and downdip limits to great subduction earthquakes: Thermal and structural models of Cascadia, south Alaska, SW Japan and Chile. *Journal of Geophysical Research Solid Earth*, 104, 14965-14991. <https://doi.org/10.1029/1999JB900060>
- Outerbridge, K.C., Dixon, T.H., Schwartz, S.Y., Walter, J.I., Protti, M., Gonzalez, V., Biggs, J., Thorwart, M., & Rabbel, W. (2010). A tremor and slip event on the Cocos-Caribbean subduction zone as measured by a global positioning system (GPS) and seismic network on the Nicoya Peninsula, Costa Rica. *Journal of Geophysical Research: Solid Earth*, 115. <https://doi.org/10.1029/2009JB006845>

- Ozawa, S., Murakami, M., & Tada, T. (2001). Time-dependent inversion study of the slow thrust event in the Nankai trough subduction zone, southwestern Japan. *Journal of Geophysical Research*, 106(B1), 787-802. <https://doi.org/10.1029/2000JB900317>
- Ozawa, S., Suito, H., Imakiire, T., & Murakami, M. (2007). Spatiotemporal evolution of aseismic interplate slip between 1996 and 1998 and between 2002 and 2004, in Bungo channel, southwest Japan. *Journal of Geophysical Research: Solid Earth*, 112(B5). <https://doi.org/10.1029/2006JB004643>
- Ozawa, S., Yarai, H., Imakiire, T., & Tobita, M. (2013). Spatial and temporal evolution of the long-term slow slip in the Bungo Channel, Japan. *Earth, Planets and Space*, 65, 67-73. doi:10.5047/eps.2012.06.009
- Ozawa, S., Tobita, M., & Yarai, H. (2016). A possible restart of an interplate slow slip adjacent to the Tokai seismic gap in Japan. *Earth, Planets and Space*, 68. <https://doi.org/10.1186/s40623-016-0430-4>
- Ozawa, S. (2017). Long-term slow slip events along the Nankai trough subduction zone after the 2011 Tohoku earthquake in Japan. *Earth, Planets and Space*, 69. doi:10.1186/s40623-017-0640-4
- Pacific Gas and Electric (PGE) (2002), Seismic Assessment for the Humboldt Bay ISFSI Project, Technical Report, TR-HBIP-2002-01.
- Pacific Northwest Geodetic Array (PANGA) (1996). GPS/GNSS Network and Geodesy Laboratory: Central Washington University, Int. Fed. of Digital Seismograph Networks. doi:70.7914/SN/PW
- Padgett, J.S., Kelsey, K.M., & Lamphear, D. (2019), Upper-plate deformation of Late Pleistocene marine terraces in the Trinidad, California, coastal area, southern Cascadia subduction zone. *Geosphere*, 15(4), 1323-1341. <https://doi.org/10.1130/GES02032.1>.
- Parsons, T., K. M. Johnson, P. Bird, J.M. Bormann, T.E. Dawson, E.H. Field, W.C. Hammond, T.A. Herring, R. McCaffrey, Z.-K. Shen, W.R. Thatcher, R.J. Weldon II, & Y. Zeng, (2013), Appendix C—Deformation models for UCERF3, *USGS Open-File Report 2013–1165*, 66 pp.
- Peacock, S.M. (1993). The importance of the blueschist → eclogite dehydration reactions in subducting oceanic crust. *Geological Society of America Bulletin*, 105(5), 684-694. [https://doi.org/10.1130/0016-7606\(1993\)105<0684:TIOBED>2.3.CO;2](https://doi.org/10.1130/0016-7606(1993)105<0684:TIOBED>2.3.CO;2)

- Peacock, S.M. (2009). Thermal and metamorphic environment of subduction zone episodic tremor and slip. *Journal of Geophysical Research*, 114, B00A08.
<https://doi.org/10.1029/2008JB005978>
- Pearson, C.F. & Snay, R.A. (2014), Strain partitioning along the western margin of North America. *Journal of Structural Geology*, 64, 67-78.
<https://doi.org/10.1016/j.jsg.2014.02.012>
- Peng, Z. & Gomberg, J. (2010). An integrated perspective of the continuum between earthquakes and slow-slip phenomena. *Nature Geoscience*, 3, 599-607
<https://doi.org/10.1038/ngeo940>
- Peng, Y., D. Dong, J. Yan, and W. Chen (2017), Geodetic constraint on the motion of a slab window: Implication for the Mendocino Crustal Conveyor model. *Geophysical Research Letters*, 44(14), 7187–7193. doi:10.1002/2017GL074149.
- Perez-Silva, A., Kaneko, Y., Savage, M. Wallace, L., Li., D., & Williams, C. (2022). Segmentation of shallow slow slip events at the Hikurangi subduction zone explained by along-strike changes in fault geometry and plate convergence rates. *Journal of Geophysical Research: Solid Earth*, 127(1). <https://doi.org/10.1029/2021JB022913>
- Plescia, S.M. & Hayes, G.P. (2020). Geometric controls on megathrust earthquakes. *Geophysical Journal International*, 222(2), 1270-1282. <https://doi.org/10.1093/gji/ggaa254>
- Plourde, A.P., Bostock, M.G., Audet, P. & Thomas, A.M. (2015). Low-frequency earthquakes at the southern Cascadia margin. *Geophysical Research Letters*, 42, 4849-4855.
doi:10.1002/2015GL064363
- Pollitz, F. F., McCrory, P., Svarc, J., & Murray, J. (2008), Dislocation models of interseismic deformation in the western United States, *Journal of Geophysical Research*, 113, B04413. doi:10.1029/2007JB005174
- Pollitz, F.F. & Evans, E.L. (2017). Implications of the earthquake cycle for inferring fault locking on the Cascadia megathrust. *Geophysical Journal International*, 209(1), 167-185.
<https://doi.org/10.1093/gji/ggx009>
- Popov, A.A., Sobolev, S.V., & Zoback, M.D. (2012). Modeling evolution of the San Andreas Fault system in northern and central California. *Geochemistry, Geophysics, Geosystems*, 13(8). doi:10.1029/2012GC004086

- Radiguet, M., Cotton, F., Vergnolle, M., Campillo, M., Walpersdorf, A., Cotte, N., & Kostoglodov, V. (2012). Slow slip events and strain accumulation in the Guerrero gap, Mexico. *Journal of Geophysical Research*, 117, B04305. doi:10.1029/2011JB008801
- Rogers, G. & Dragert, H. (2003). Episodic tremor and slip on the Cascadia subduction zone: chatter of silent slip. *Science*, 300(5627), 1942-1943. doi:10.1126/science.1084783
- Rousset, B. (2019). Months-long subduction slow slip events avoid the stress shadows of seismic asperities. *Journal of Geophysical Research: Solid Earth*, 124, 7227–7230.
<https://doi.org/10.1029/2019JB018037>
- Rousset, B., Fu, Y., Bartlow, N., & Burgmann, R. (2019). Weeks-long and years-long slow slip and tectonic tremor episodes on the south central Alaska megathrust. *Journal of Geophysical Research: Solid Earth*, 124, 13392-13403.
<https://doi.org/10.1029/2019JB018724>
- Rüpke, L.H., Morgan, J.P., Hort, M., & Connolly, J.A.D. (2004). Serpentine and the subduction zone water cycle. *Earth and Planetary Science Letters*, 223(1-2), 17-34.
<https://doi.org/10.1016/j.epsl.2004.04.018>
- Russo, R.M., VanDecar, J.C., Comte, D., Mocanu, V.I., Gallego, A., & Murdie, R.E. (2010). Subduction of the Chile Ridge: Upper mantle structure and flow. *GSA Today*, 20(9), 4-10. doi:10.1130/GSATG61A.1
- Savage, J.C. & Buford, R.O. (1973), Geodetic determination of relative plate motion in central California. *Journal of Geophysical Research*, 78(5), 832-845.
<https://doi.org/10.1029/JB078i005p00832>
- Savage, J.C., (1983), A dislocation model of strain accumulation and release at a subduction zone. *Journal of Geophysical Research*, 88, 4984-4996.
<https://doi.org/10.1029/JB088iB06p04984>
- Schellart, W. P. (2004). Kinematics of subduction and subduction-induced flow in the upper mantle. *Journal of Geophysical Research: Solid Earth*, 109.
<https://doi.org/10.1029/2004JB002970>
- Schmalzle, G.M., McCaffrey, R. & Creager, K.C. (2014). Central Cascadia subduction zone creep. *Geochemistry, Geophysics, Geosystems*, 15(4), 1515-1532.
<https://doi.org/10.1002/2013GC005172>

- Schmidt, D.A. & Gao, H. (2010). Source parameters and time-dependent slip distributions of slow slip events on the Cascadia subduction zone from 1998 to 2008. *Journal of Geophysical Research: Solid Earth*, 115. <https://doi.org/10.1029/2008JB006045>
- Schwartz, S.Y. & Rokosky, J.M. (2007). Slow slip events and seismic tremor at circum-Pacific subduction zones. *Reviews of Geophysics*, 45(3). <https://doi.org/10.1029/2006RG000208>
- Schwarz, G. (1978), Estimating the dimension of a model, *Annals of Statistics*, 6(2), 461–464. doi:10.1214/aos/1176344136
- Segall, P., Rubin, A.M., Bradley, A.M., & Rice, J.R. (2010). Dilatant strengthening as a mechanism for slow slip events. *Journal of Geophysical Research: Solid Earth*, 115. <https://doi.org/10.1029/2010JB007449>
- Segall, P. & Bradley, A.M. (2012). Slow-slip evolves into megathrust earthquakes in 2D numerical simulations. *Geophysical Research Letters*, 39, L18308. <https://doi.org/10.1029/2012GL052811>
- Shelly, D.R., Beroza, G.C., & Ide, S. (2007). Non-volcanic tremor and low-frequency earthquake swarms. *Nature*, 446, 305-307. <https://doi.org/10.1038/nature05666>
- Spinler, J.C., Bennett, R.A., Anderson, M.L., McGill, S.F., Hreinsdottir, S., & McCallister, A. (2010), Present-day strain accumulation and slip rates associated with southern San Andreas and eastern California shear zones. *Journal of Geophysical Research: Solid Earth*, 115(B11), B11407. doi:10.1029/2010JB007424
- Stanciu, A.C. & Humphreys, E.D. (2021). Seismic architecture of the upper mantle underlying California and Nevada. *Journal of Geophysical Research: Solid Earth*, 126(12). <https://doi.org/10.1029/2021JB021880>
- Sweet, J.R., Creager, K.C., Houston, H., & Chestler, S.R. (2019). Variations in Cascadia low-frequency earthquake behavior with downdip distance. *Geochemistry, Geophysics, Geosystems*, 20(2), 1202-1217. <https://doi.org/10.1029/2018GC007998>
- Szeliga, W., Melbourne, T. I., Miller, M. M., & Santillan, V. M. (2004). Southern Cascadia episodic slow earthquakes. *Geophysical Research Letters*, 31(16), L16602. doi:10.1029/2004GL020824
- Takagi, R., Obara, K. & Maeda, T. (2016). Slow slip event within a gap between tremor and locked zones in the Nankai subduction zone. *Geophysical Research Letters*, 43, 1066-1074. <https://doi.org/10.1002/2015GL066987>

- Takagi, R., Uchida, N. & Obara, K. (2019). Along-strike variation and migration of long-term slow slip events in the western Nankai subduction zone, Japan. *Journal of Geophysical Research*, 124, 3853-3880. <https://doi.org/10.1029/2018JB016738>
- Uchida, N., Iinuma, T., Nadeau, R.M., Burgmann, R., & Hino, R. (2016). Periodic slow slip triggers megathrust zone earthquakes in northeastern Japan. *Science*, 351(6272), 488-492. doi:10.1126/science.aad3108
- Unruh, J. & J. Humphreys (2017), Seismogenic deformation between the Sierran microplate and Oregon Coast block, California, USA. *Geology*, 45(5), doi:10.1130/G38696.
- van Keken, P.E., Hacker, B.R., Syracuse, E.M., Abers, G.A. (2011). Subduction factory: 4. Depth-dependent flux of H₂O from subducting slabs worldwide. *Journal of Geophysical Research: Solid Earth*, 116. <https://doi.org/10.1029/2010JB007922>
- van Wijk, J.W., Govers, R., & Furlong K.P. (2001). Three-dimensional thermal modeling of the California upper mantle: a slab window vs. stalled slab. *Earth and Planetary Science Letters*, 186, 175-186. [https://doi.org/10.1016/S0012-821X\(01\)00243-6](https://doi.org/10.1016/S0012-821X(01)00243-6)
- Verdonck, D. & Zandt, G. (1994). Three-dimensional crustal structure of the Mendocino Triple Junction region from local earthquake travel times. *Journal of Geophysical Research: Solid Earth*, 99, 23843-23858. <https://doi.org/10.1029/94JB01238>
- Wallace, L.M. & Beavan, J. (2010). Diverse slow slip behavior at the Hikurangi subduction margin, New Zealand. *Journal of Geophysical Research*, 115, B12402. <https://doi.org/10.1029/2010JB007717>
- Wallace, L.M., Beavan, J., Bannister, S., & Williams, C. (2012). Simultaneous long-term and short-term slow slip events at the Hikurangi subduction margin, New Zealand: Implications for the processes that control slow slip event occurrence, duration, and migration. *Journal of Geophysical Research*, 117, B11402. doi:10.1029/2012JB009489
- Wallace, L.M., Bartlow, N., Hamling, I. & Fry, B. (2014). Quake clamps down on slow slip. *Geophysical Research Letters*, 41, 8840-8846. <https://doi.org/10.1002/2014GL062367>
- Walton, M.A., Staisch, L.M., Dura, T., Pearl, J.K., Sherrod, B., Gomberg, J., Engelhart, S., Tréhu, A., Watt, J., Perkins, J. and Witter, R.C., Bartlow, N., Goldfinger, C., Kelsey, H., Morey, A.E., Sahakian, V.J., Tobin, H., Wang, K., Wells, R., & Wirth, E. (2021). Toward an integrative geological and geophysical view of Cascadia subduction zone

- earthquakes. *Annual Review of Earth and Planetary Sciences*, 49, 367-398.
<https://doi.org/10.1146/annurev-earth-071620-065605>
- Wang, K., Wells, R., Mazzotti, S., Hyndman, R.D., & Sagiya, T. (2003), A revised dislocation model of interseismic deformation of the Cascadia subduction zone. *Journal of Geophysical Research*, 108(B1). doi:10.1029/2001JB001227
- Wang, K., & Tréhu, A.M. (2016). Invited review paper: Some outstanding issues in the study of great megathrust earthquakes—The Cascadia example. *Journal of Geodynamics*, 98, 1-18. <https://doi.org/10.1016/j.jog.2016.03.010>
- Watt, J.T. & Brothers, D.S. (2020), Systematic characterization of morphotectonic variability along the Cascadia convergent margin: Implications for shallow megathrust behavior and tsunami hazards. *Geosphere*, 17(1), 95-117. <https://doi.org/10.1130/GES02178.1>
- Wech, A.G. (2010). Interactive tremor monitoring. *Seismological Research Letters*, 81(4), 664-669. <https://doi.org/10.1785/gssrl.81.4.664>
- Wech, A.G., Creager, K.C., & Melbourne, T.I. (2009), Seismic and geodetic constraints on Cascadia slow slip. *Journal of Geophysical Research: Solid Earth*, 114.
<https://doi.org/10.1029/2008JB006090>
- Wech, A.G. & Creager, K.C. (2011). A continuum of stress, strength and slip in the Cascadia subduction zone. *Nature Geoscience*, 4, 624-628. <https://doi.org/10.1038/ngeo1215>
- Wech, A.G. (2021), Cataloging tectonic tremor energy radiation in the Cascadia subduction zone. *Journal of Geophysical Research: Solid Earth*, 126, e2021JB022523.
<https://doi.org/10.1029/2021JB022523>
- Wei, M., McGuire, J.J., & Richardson, E. (2012). A slow slip event in the south central Alaska Subduction Zone and related seismicity anomaly. *Geophysical Research Letters*, 39, L15309. doi:10.1029/2012GL052351
- Wells, R.E., Weaver, C.S. & Blakely, R.J. (1998). Fore-arc migration in Cascadia and its neotectonic significance. *Geology*, 26(8), 759-762. [https://doi.org/10.1130/0091-7613\(1998\)026<0759:FAMICA>2.3.CO;2](https://doi.org/10.1130/0091-7613(1998)026<0759:FAMICA>2.3.CO;2)
- Wells, R.E. & Simpson, R.W. (2001). Northward migration of the Cascadia forearc in the northwestern U.S. and implications for subduction deformation. *Earth, Planets and Space*, 53, 275-283. <https://doi.org/10.1186/BF03352384>

- Williams, T.B., Kelsey, H.M., & Freymueller, J.T., (2006), GPS-derived strain in northwestern California: Termination of the San Andreas fault system and convergence of the Sierra Nevada-Great Valley block contribute to southern Cascadia forearc contraction. *Tectonophysics*, 413, 171-184. doi:10.1016/j.tecto.2005.10.047
- Wilson, D.S. (1989), Deformation of the so-called Gorda plate. *Journal of Geophysical Research: Solid Earth*, 94, 3065-3075. <https://doi.org/10.1029/JB094iB03p03065>
- Woodward-Clyde Associates (1980), Evaluation of the potential for resolving the geologic and seismic issues at the Humboldt Bay Power Plant Unit no. 3, Appendices, Woodward-Clyde Consultants, Walnut Creek, California.
- Xia, Y., Mitchell, K., Ek, M., Sheffield, J., Cosgrove, B., Wood, E., et al. (2012). Continental-scale water and energy flux analysis and validation for the North American Land Data Assimilation System project phase 2 (NLDAS-2): 1. Intercomparison and application of model products. *Journal of Geophysical Research*, 117, D03109. <https://doi.org/10.1029/2011JD016048>

APPENDIX 1: SUPPLEMENTARY MATERIALS FOR CHAPTER 2

Text A1.1 Leveraging the Pacific Northwest Seismic Network (PNSN) tremor catalogue (from <https://pnsn.org/tremor>), we have collected the timing of all tremor events within 50 km of station P332 from 2009-2019. Figure A1.1 shows these tremor counts binned into three-day intervals and normalized by the highest tremor count, along with the STL remainder (R) component for station P332. The majority of tremor spikes in Figure A1.1 correlate with eastward transients in the data, representing nearby SSEs. The few spikes that do not line up with transients, e.g. late 2011, likely represent more distal SSEs that did not impart a signal on station P332. Furthermore, we believe that the reduced tremor amplitude after 2016 is due to PNSN network changes and does not reflect a lack of SSEs in the R component time series.

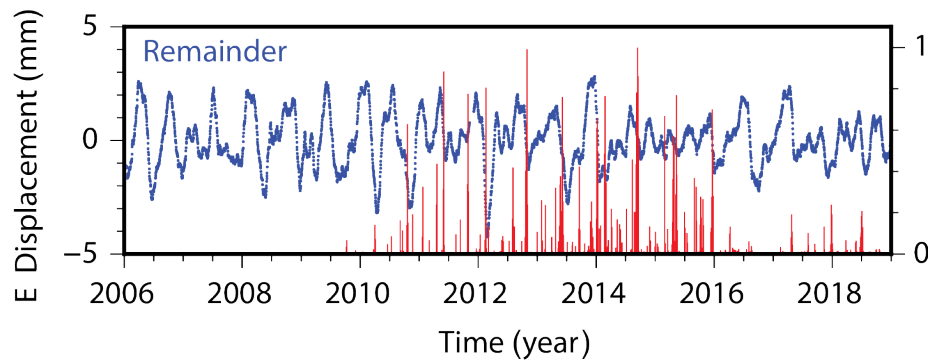


Figure A1.1 Comparison of the remainder east component (blue) for station P332 and local tremor activity (red). The tremor count includes any tremor located within 50 km of station P332 and is binned into three-day intervals and normalized. Tremor data are sourced from the Pacific Northwest Seismic Network tremor catalogue (Wech, 2010). Spikes in tremor indicate the timing of nearby short-term slow slip events and coincide with the grey bars in Figure 2.3c.

Text A1.2 The deconstruction of the GNSS time series is controlled by 6 parameters in the STL analysis, which are summarized in **Table S3**. Please refer to Cleveland et al. (1990) for more details about how STL is implemented and how these parameters impact the data results.

Parameter Name	Parameter Description	Parameter Value
n_p	number of observations in each cycle of the seasonal component	365
n_i	number of passes through the inner loop	1
n_o	number of robustness iterations of the outer loop	10
n_l	smoothing parameter for the low-pass filter	365
n_t	smoothing parameter for the trend component	633
n_s	smoothing parameter for the seasonal component	11

Table A1.1 Summary of the parameters used in the STL analysis.

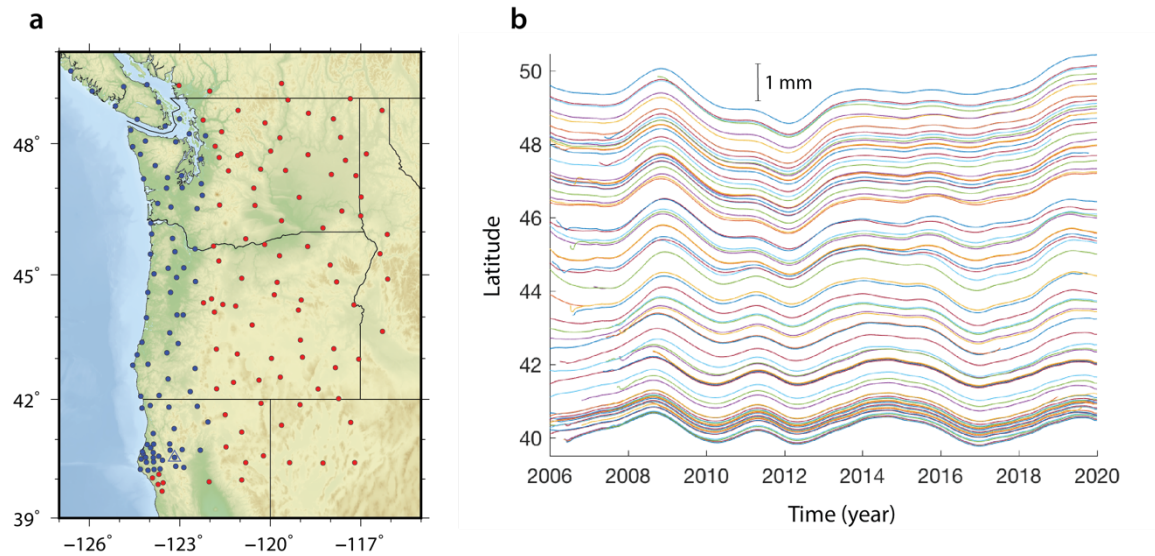


Figure A1.2 Map of stations used for calculating common mode signals and the resulting common mode stations by latitude. **a** Map of GNSS stations (dots), showing the location of stations (red) that were used to calculate common mode signals. Station P332 (in northern California) is enclosed by a triangle. **b** The resulting common mode signals for blue stations in panel **a**, ordered by latitude.

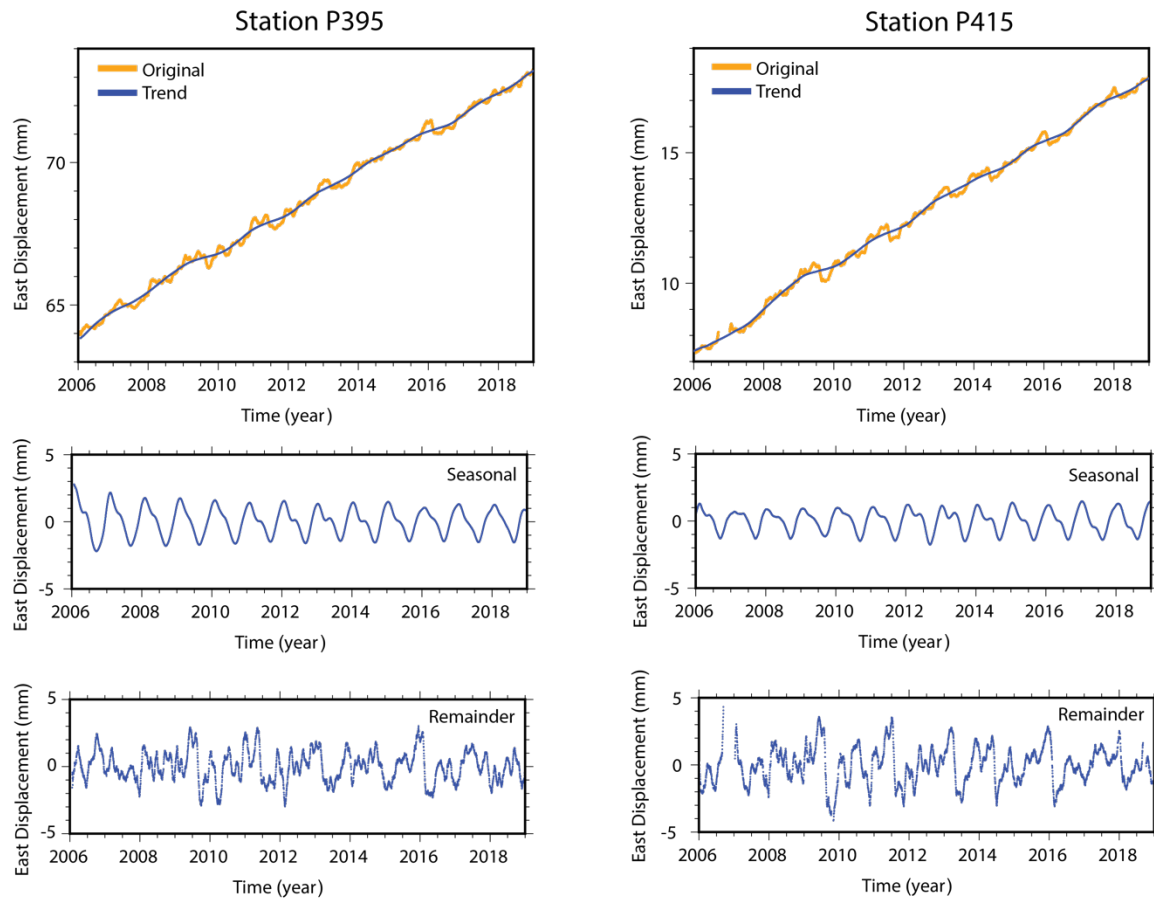


Figure A1.3 STL results for GNSS stations P395 (left column) and P415 (right column). **Top row** shows the original GNSS time series data (orange) and the trend (T) component from STL (blue). **Middle row** shows the seasonal (S) component from STL. **Bottom row** shows the remainder (R) component from STL. Stations P395 and P415 are in the vicinity of the 2008.75-2009.75 and 2017.6-2019 transients, respectively.

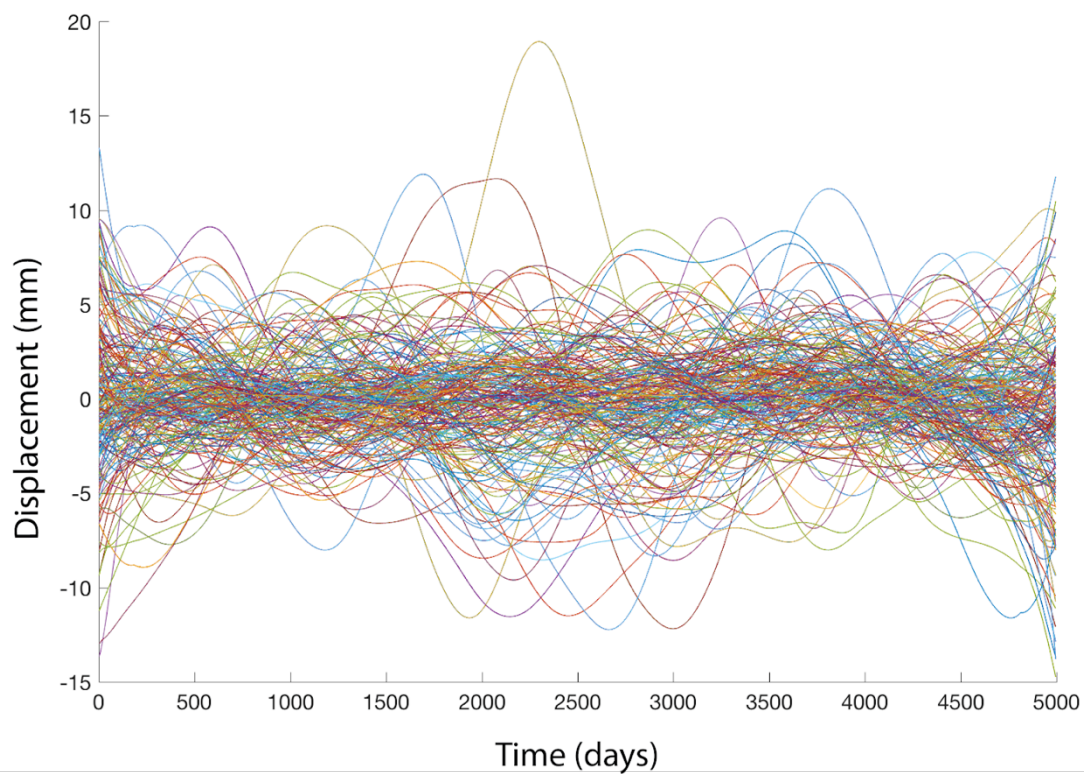


Figure A1.4 Resulting 1,000 synthetic time series after our cleaning method was applied. These traces were ultimately used to calculate the average error margin for our STL processing technique.

Text A1.3 In order to test the sensitivity of our STL processing technique, we created synthetic time series that contained long-term transients with 1 or 2 mm of negative offset over a one-year period. In certain scenarios, we also included subsequent long-term transients with positive offset of 1, 2, or 3 mm. We created 1,000 time series for each scenario, and slopes and noise were added to each synthetic time series as described in Section 2.2.1 of the main text. All of the time series were then reduced using our five-step cleaning method. Figure A1.5 shows the results for each scenario, including the transients before the addition of noise (red), an example of one of the final reduced time series after cleaning (green), and the average product for all of the 1,000 final reduced time series (blue). Panels a-c show how the 1 mm negative transient and subsequent positive transients can easily be lost within the noise in both the example and average time series. In comparison, the 2 mm negative transients and the following positive transients are relatively well defined in panels d-f.

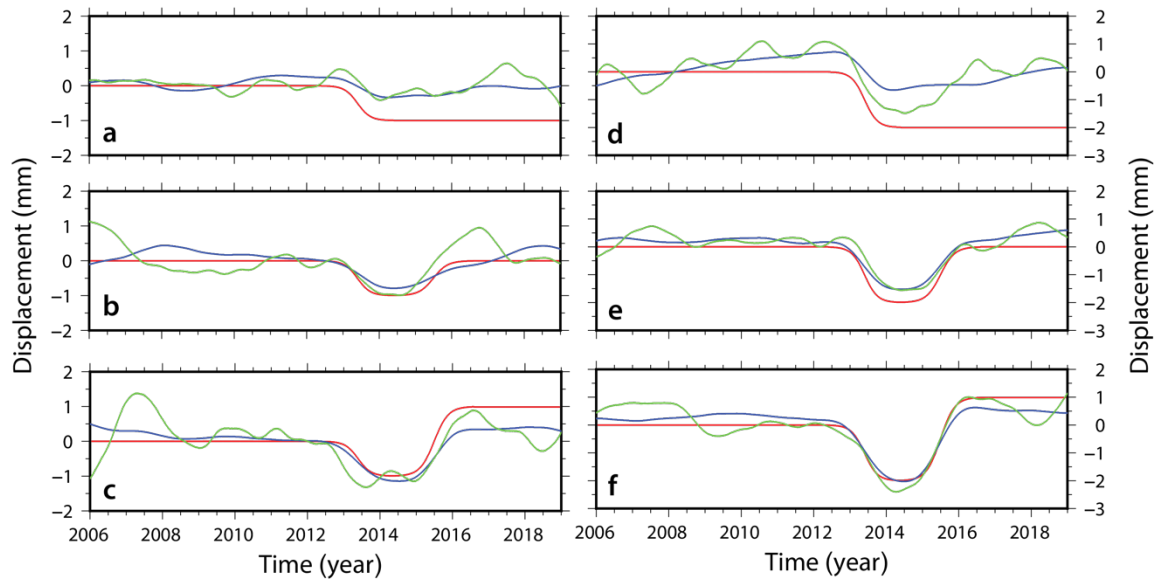


Figure A1.5 Results for the different transient scenarios. Each panel shows the transients before the addition of noise (red), an example of one of the final reduced time series after cleaning (green), and the average product for all of the 1,000 final reduced time series (blue). **a** 1 mm negative offset and no positive offset. **b** 1 mm negative offset followed by 1 mm positive offset. **c** 1 mm negative offset followed by 2 mm positive offset. **d** 2 mm negative offset and no positive offset. **e** 2 mm negative offset followed by 2 mm positive offset. **f** 2 mm negative offset followed by 3 mm positive offset.

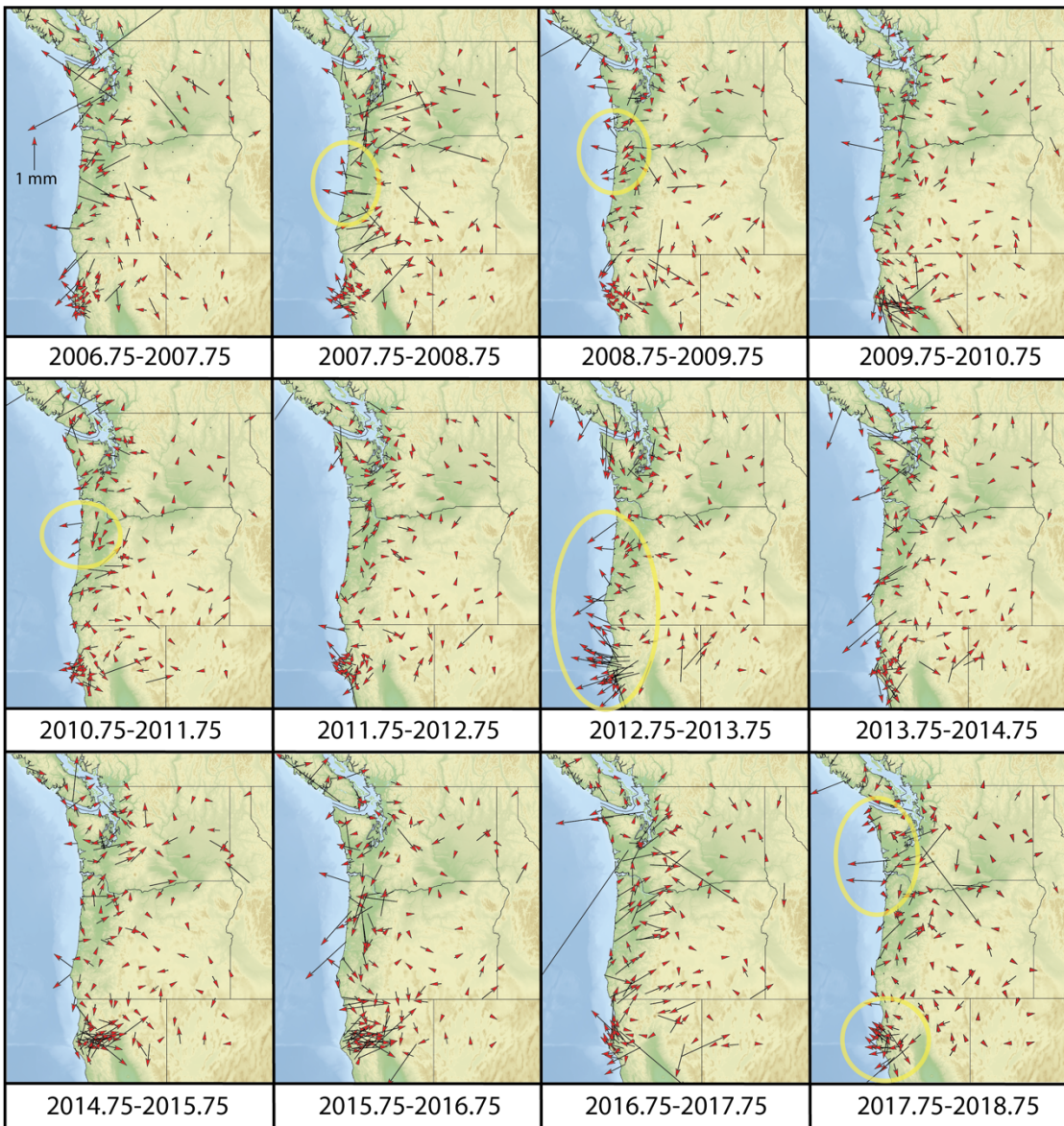


Figure A1.6 Horizontal displacements from our residual time-series data at GNSS stations during one-year epochs. Each epoch begins at the start of the water year to minimize seasonal signals. Original time series solutions are provided by the Pacific Northwest Geodetic Array (PANGA), which are then cleaned to produce residual displacements. Yellow ellipses outline sets of stations with coherent motion.

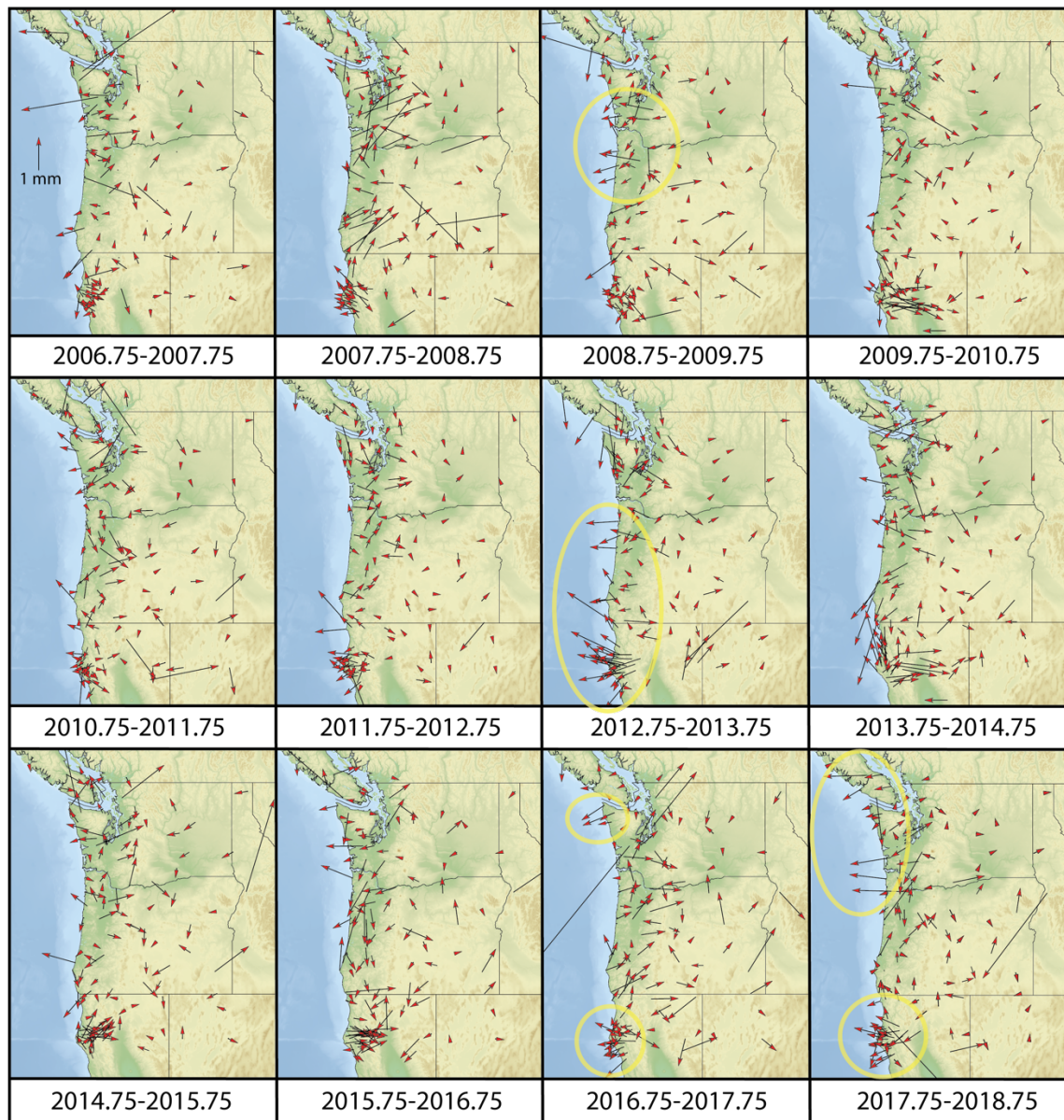


Figure A1.7 Horizontal displacements from our residual time-series data at GNSS stations during one-year epochs. Each epoch begins at the start of the water year to minimize seasonal signals. Original time series solutions are provided by the Scripps Orbit and Permanent Array Center (SOPAC), which are then cleaned to produce residual displacements. Yellow ellipses outline sets of stations with coherent motion.

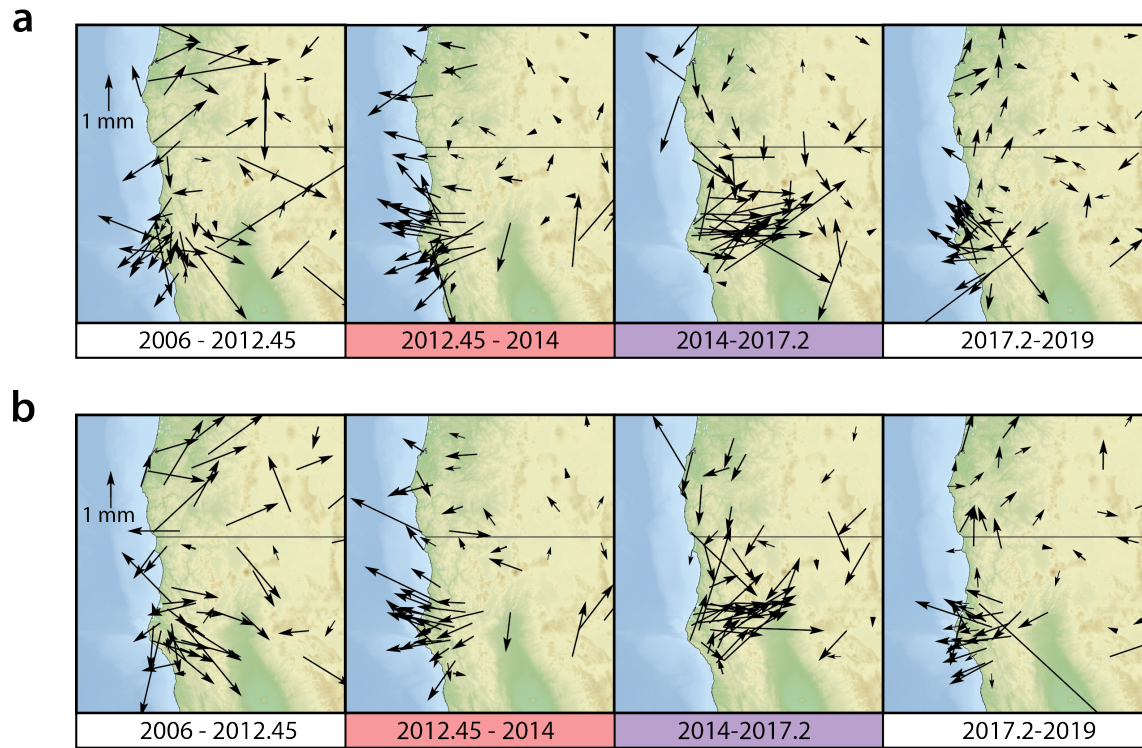


Figure A1.8 Comparison of residual data from different processing centers. **a** Residual horizontal displacements using PANGA solutions. **b** Residual horizontal displacements using SOPAC solutions. Similar results using NGL solutions are shown in Figure 2.5b.

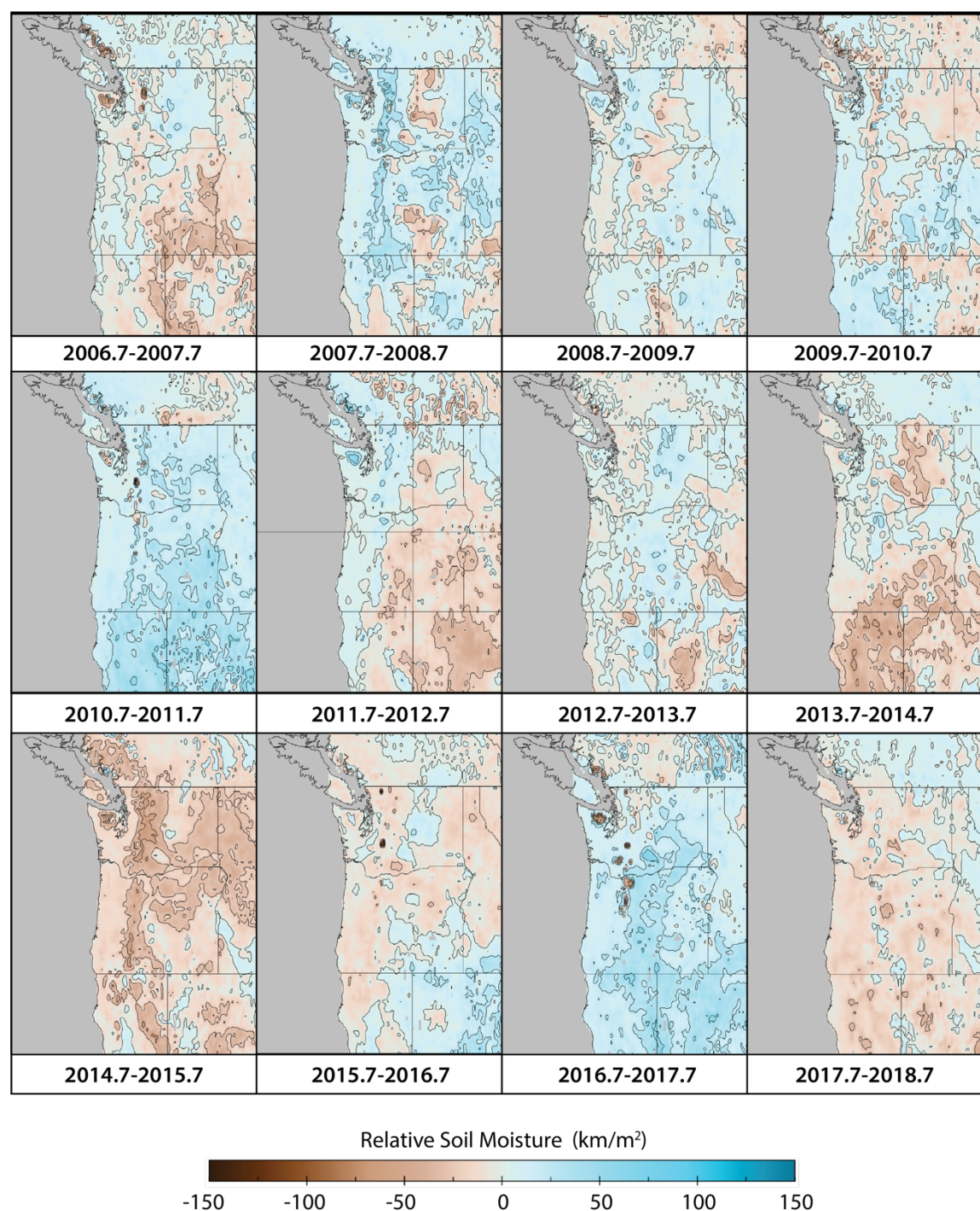


Figure A1.9 Yearly snapshots of variable soil moisture content in Cascadia relative to the 2006-2019 average. The thirteen-year mean has been removed from each epoch. Soil moisture data was obtained from the land surface model for Phase 2 of the North American Land Data Assimilation System (NLDAS-2) (Mocko et al., 2012; Xia et al., 2012).

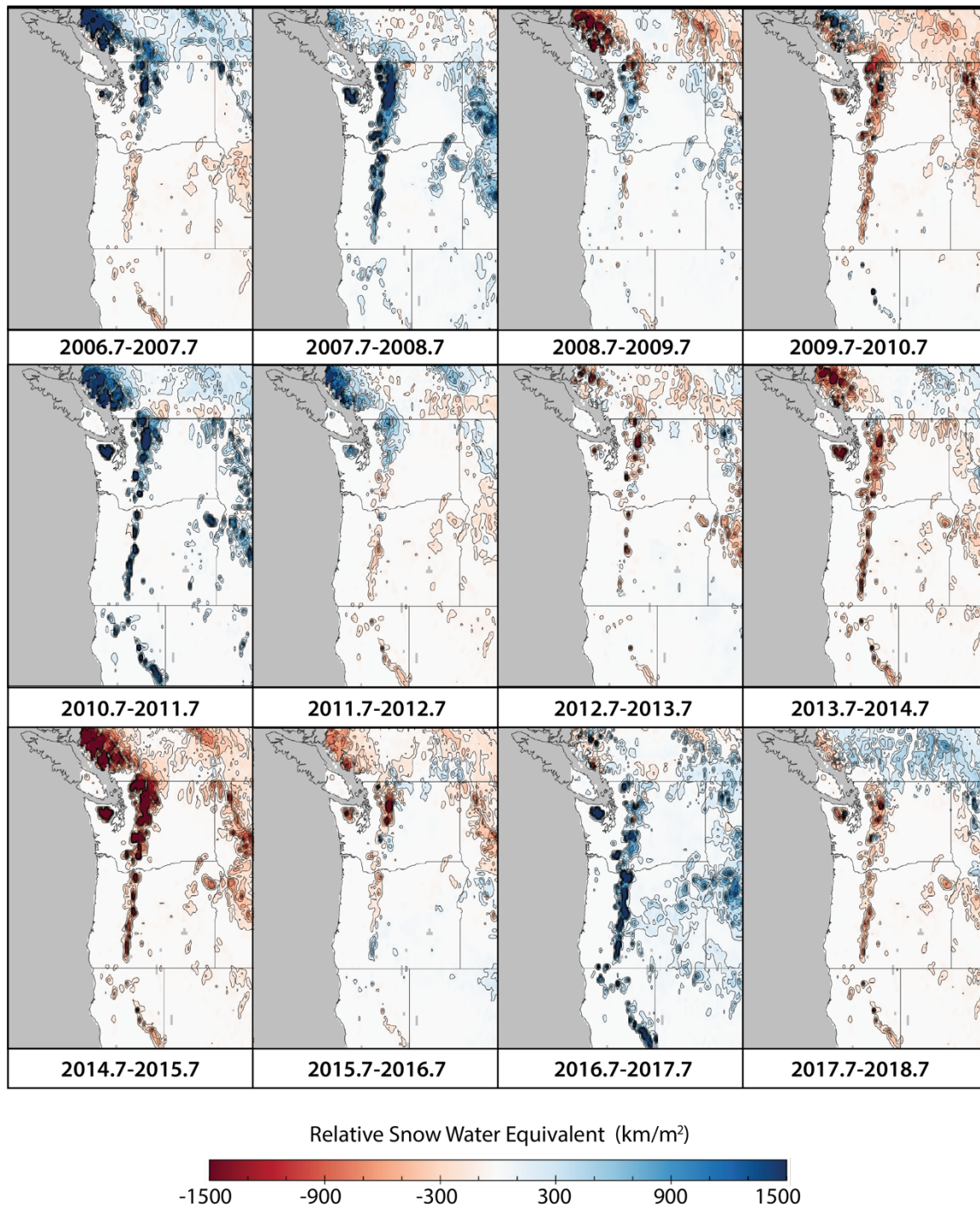


Figure A1.10 Yearly snapshots of variable snow water equivalent (SWE) in Cascadia relative to the 2006-2019 average. Blue colors indicate years with a relatively large snowpack. The thirteen-year mean has been removed from each epoch. SWE data was obtained from the land surface model for Phase 2 of the North American Land Data Assimilation System (NLDAS-2) (Mocko et al., 2012; Xia et al., 2012).

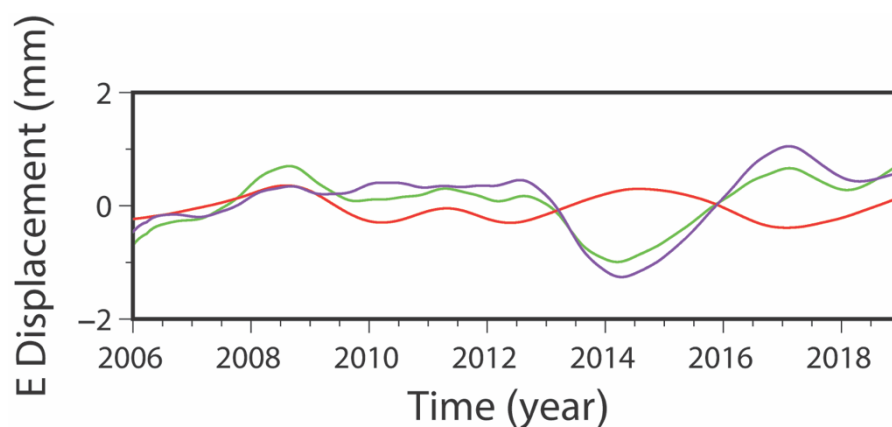


Figure A1.11 Comparison of the GNSS time series data at station P332 before (green) and after (purple) the removal of the common mode signal (red). Note how the removal of the common mode signal enhances the 2012.45 - 2014 transient at this station, but only by a fraction of the initial transient amplitude. Therefore, the common mode only acts to slightly increase the amplitude of the transient and is not responsible for creating the signal entirely.

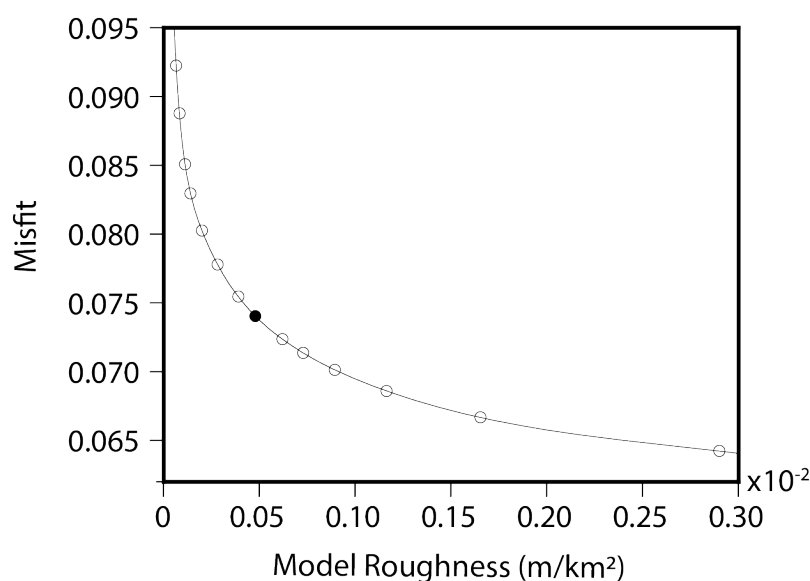


Figure A1.12 Trade-off curve between misfit and model roughness, where each circle denotes an inversion result using a given smoothing parameter. The filled circle represents the smoothing parameter chosen for the final inversion.

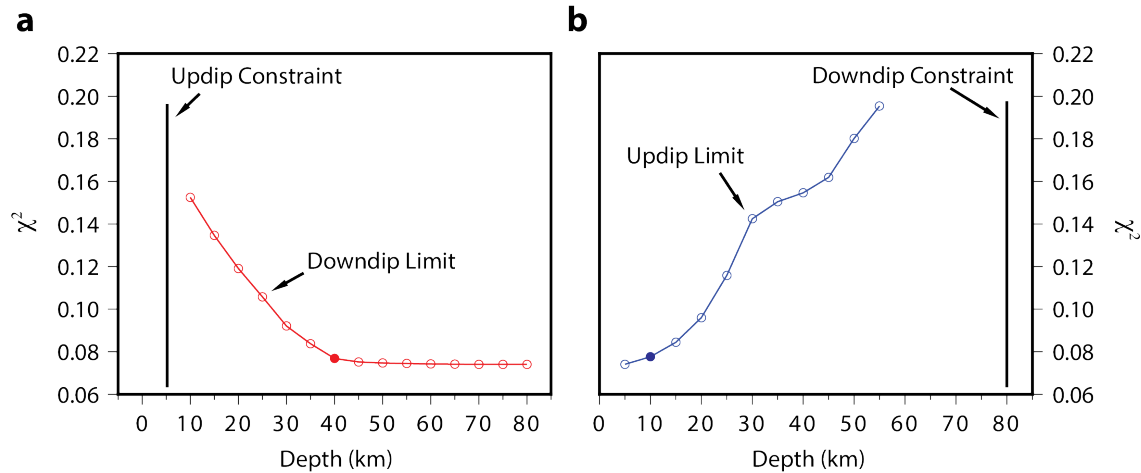


Figure A1.13 Sensitivity test results used for determining updip and downdip depth limits for the fault model used in the inversion. **a** Results for the series of slip inversions where the updip depth of the fault model was held constant and the downdip depth of the fault model was varied. Chi-square is plotted as a function of downdip depth and the filled circle represents the chosen downdip limit for the fault model used in our inversion. **b** Results for the series of slip inversions where the downdip depth was held constant and updip depth was varied. Chi-square is plotted as a function of updip depth and the filled circle represents the chosen updip limit for the final fault model used in our inversion.

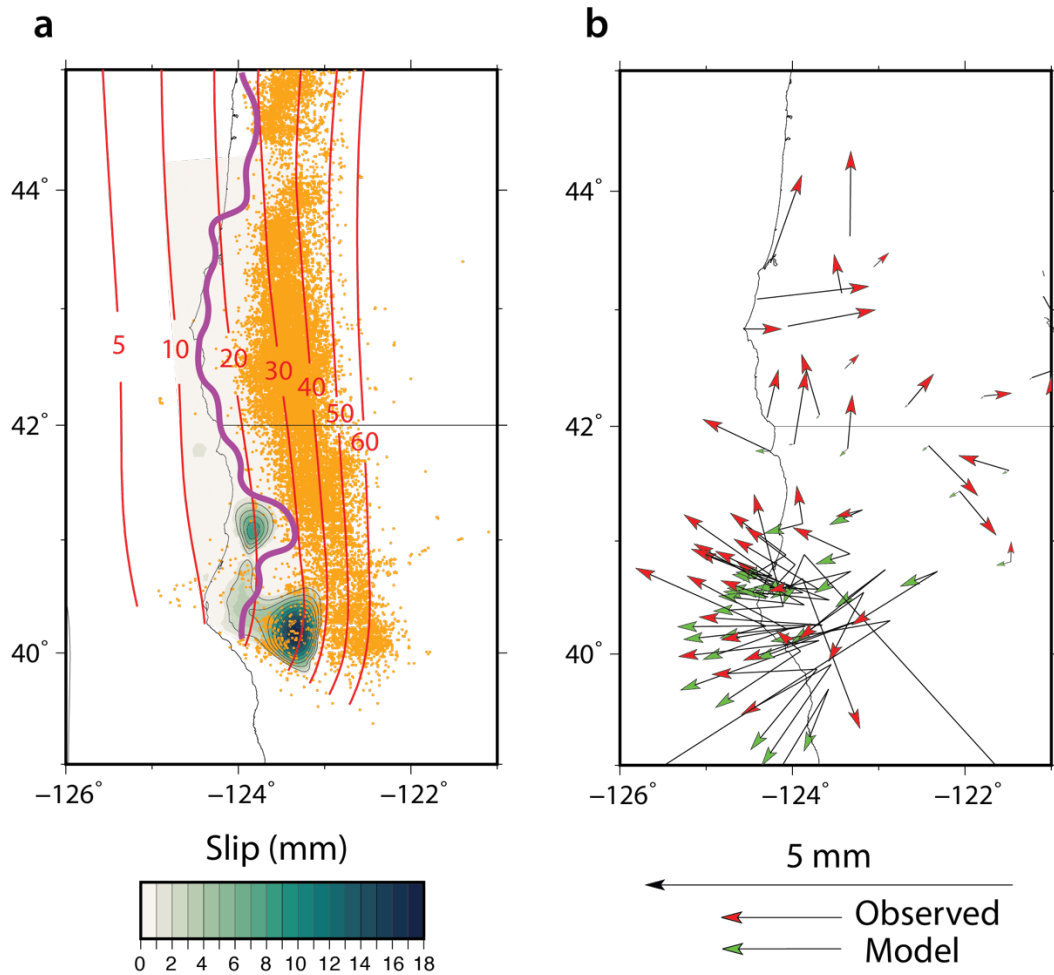


Figure A1.14 Inversion results from horizontal NGL GNSS displacements from 2012.2-2019. **a** Final slip model from our inversion with contours (black lines with green shading) outlining every 2 mm of fault slip, slab depth contours (red lines, km), tremor locations during the event (orange dots), and the contour of 20% locking (Schmalzle et al., 2014), representing the base of the seismogenic zone (thick purple line). **b** Vector map comparing observed (red) and modelled (green) GNSS displacements.

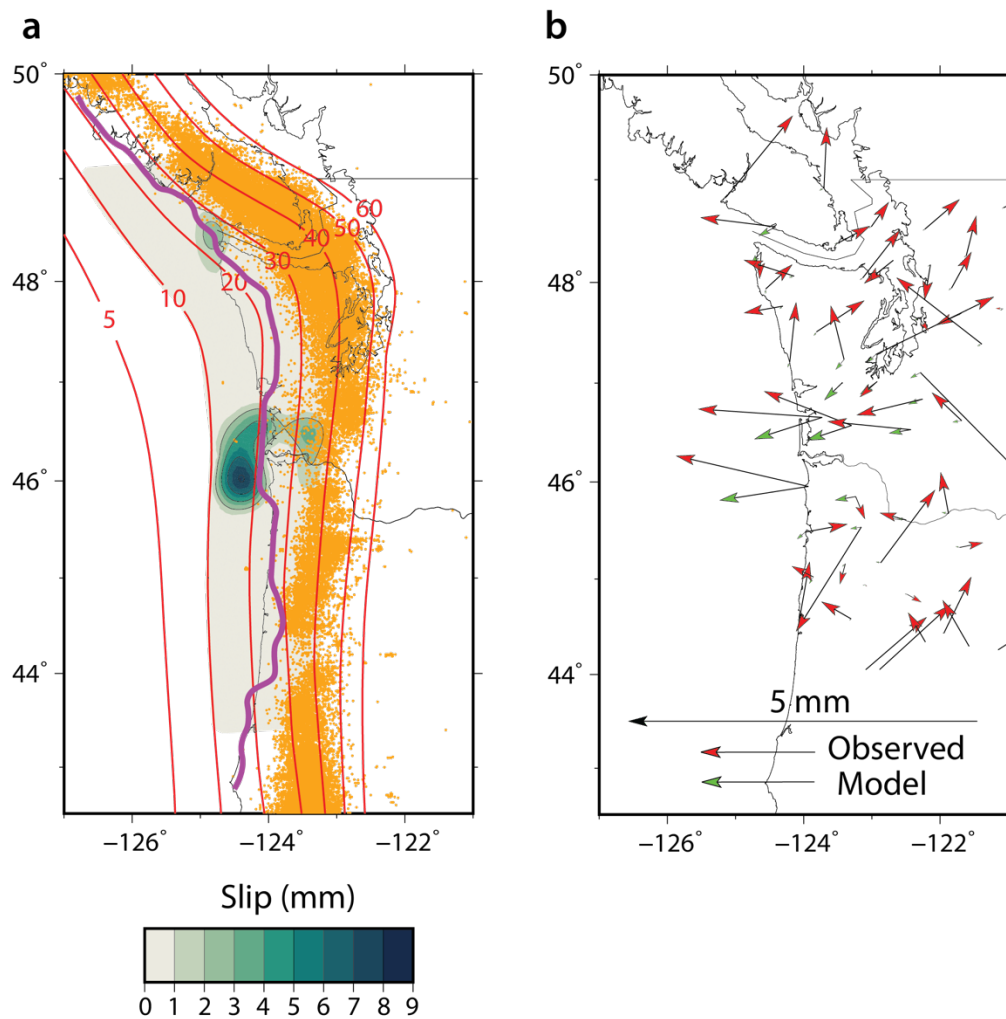


Figure A1.15 Inversion results from horizontal NGL GNSS displacements from 2017.6-2019. **a** Final slip model from our inversion with contours (black lines with green shading) outlining every 2 mm of fault slip, slab depth contours (red lines, km), tremor locations during the event (orange dots), and the contour of 20% locking (Schmalzle et al., 2014), representing the base of the seismogenic zone (thick purple line). **b** Vector map comparing observed (red) and modelled (green) GNSS displacements.

Date	Region	Displacement	Observed Maximum Surface Displacement* (mm)
2008.75 – 2009.75	Northern Oregon	Westward	1.4
2012.45 – 2014	Northern California & Southern Oregon	Westward	3.7
2014 – 2017.2	Northern California	Eastward	4.1
2017.2 – 2019 ^a	Northern California	Westward	2 ^b
2017.6 – 2019 ^a	Southern Washington	Westward	1.9 ^b

* We are most confident in transients with ≥ 2 mm of maximum surface displacement.

^a This represents a lower limit for duration, as our study period ended while this transient was possibly ongoing.

^b This represents a lower limit for maximum displacement, as our study period ended while this transient was possibly ongoing.

Table A1.2 A summary of the transient events found in this work.

APPENDIX 2: SUPPLEMENTARY MATERIALS FOR CHAPTER 3

Text A2.1 We calculate independent GPS station velocities to ensure that the published velocities are not biased by short-term tectonic signals such as slow slip events. In order to calculate station velocities, we start with the original horizontal GPS time series data provided by the Central Washington University in the NAM14 stable North American reference frame. We process these time series data using the following steps:

- 1) Outliers (positions more than 50 mm away from adjacent epochs) were removed.
- 2) Steps due to earthquakes and equipment changes were removed. The timing of steps was determined by the GAGE offsets from the Central Washington University (found at <ftp://data-out.unavco.org/pub/products/offset>). Step offsets were estimated by differencing the average positions 21 days before the event and 21 days after the event.
- 3) A 30-day moving average was applied to filter out unwanted high frequency noise.
- 4) Each time series was decomposed into seasonal (S), trend (T), and remainder (R) components using a seasonal trend decomposition method based on loess (STL) (Cleveland et al., 1990). Previous applications of the STL algorithm on GPS time series data shows that the S and R components are able to successfully remove signals from seasonal processes and short-term slow slip events, respectively (Nuyen & Schmidt, 2021). Therefore, the T component captures the background trend in GPS time series data and is not strongly influenced by seasonal and short-term slow slip signals.
- 5) We fit a line to the T component, and use the slope of this line as the horizontal velocity.

Our resulting independent station velocities are compared to the published velocities in Figure A2.1. Aside from a coherent NW signal at stations along the coast (Figure A2.1B), our velocity estimates are similar to the published velocities.

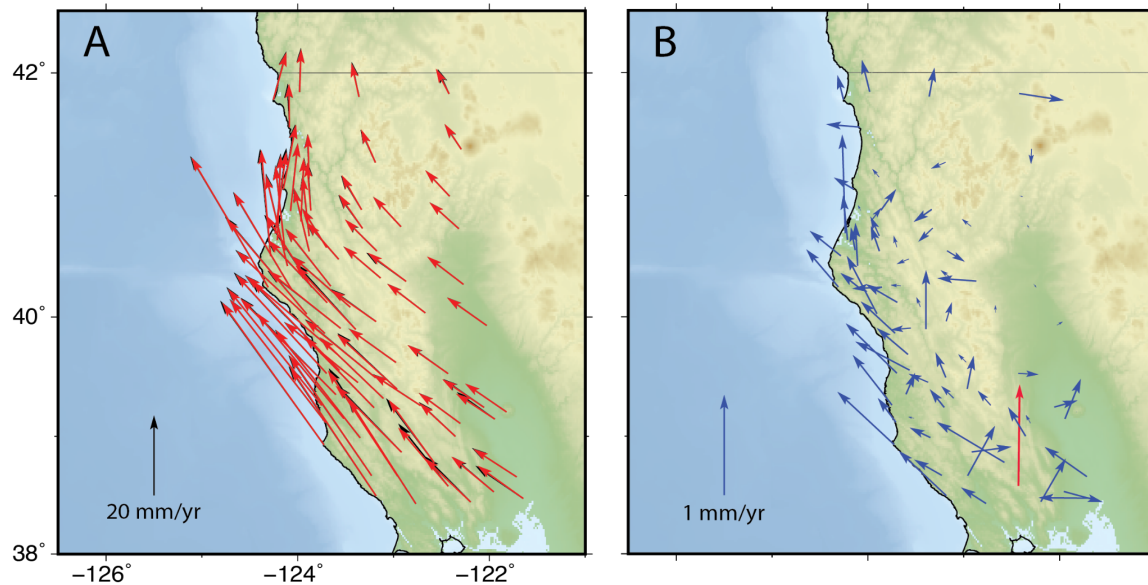


Figure A2.1 Comparisons between our independent and published GPS station velocity estimates. **A** Map comparing our independent (red) and published (black) velocity vectors. The vectors are similar enough that the red vectors overlie and hide the majority of black vectors. **B** Map showing residual vectors (blue) between our independent and published velocities. The single red vector delineates the station where the residual is equal or greater than 1 mm/yr, all other residuals are less than 1 mm/yr.

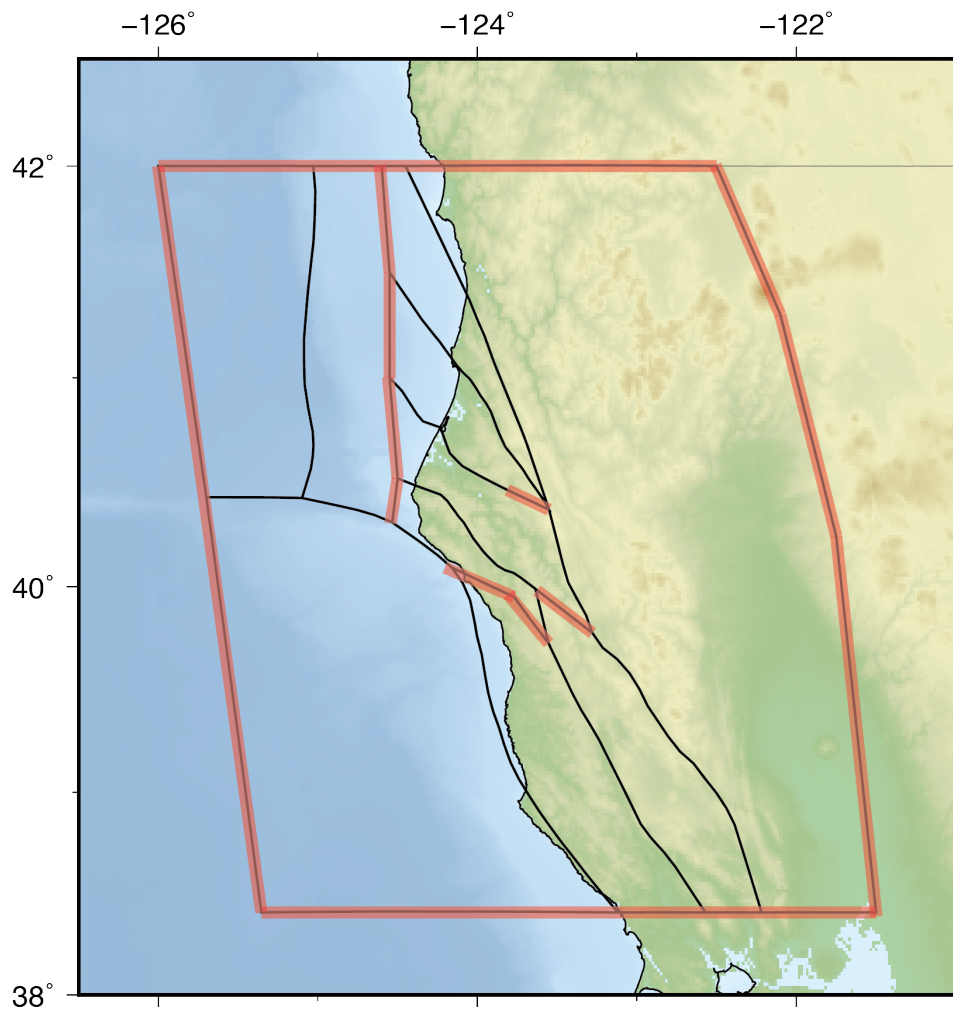


Figure A2.2 Map of Model 8 block geometry showing the locations of all pseudo-faults. The block boundaries are delineated by black lines and the block boundaries that are composed of pseudo-faults are overlain with red. Pseudo-faults are constructed using only surface nodes, so they do not have any depth component. Additionally, these ‘faults’ are forced to have zero locking, making them free-slip boundaries.

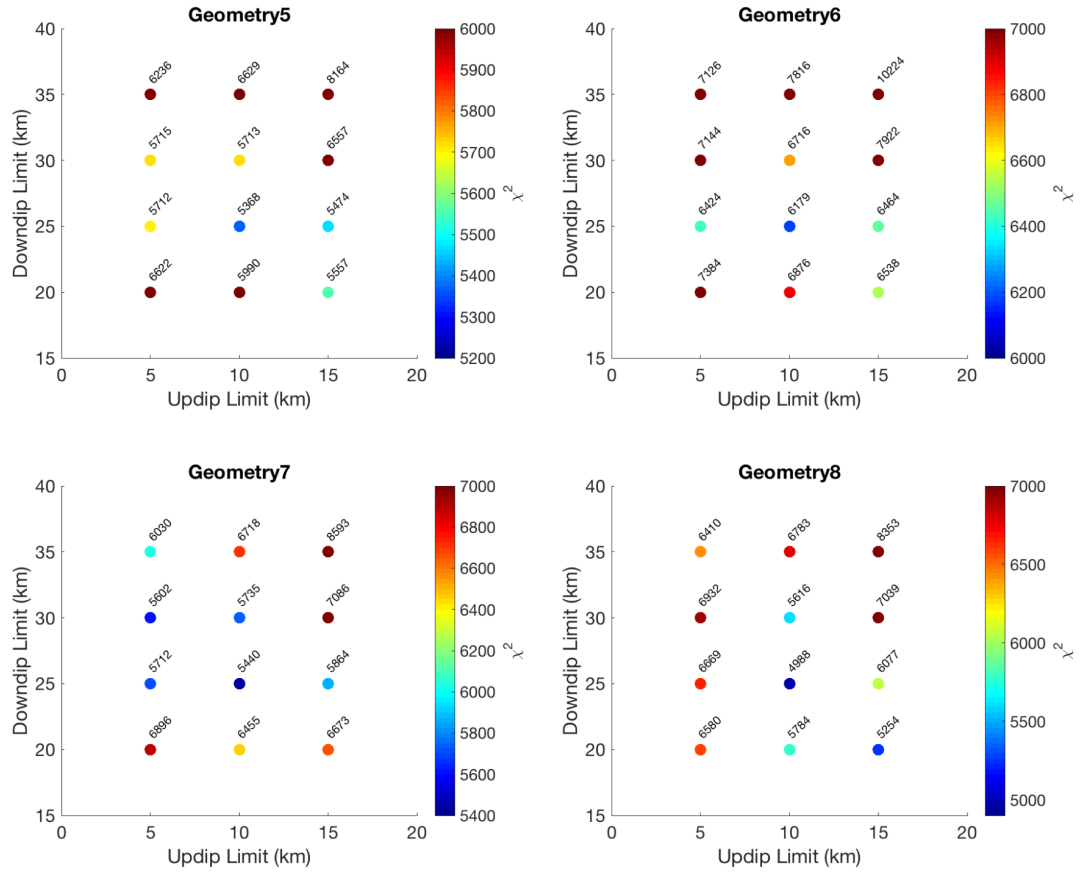


Figure A2.3 Results from our megathrust locking analysis. Each panel shows the results for a given block geometry, which is denoted by the panel title. Each point represents a full model run, and the location of a point describes the updip limit of full locking (x-axis) and the downdip limit of full slip (y-axis) used to constrain the model. Points are colored by the resulting χ^2 misfit for each model. χ^2 misfit values are provided for points with relatively high values that fall beyond the scale to the right. All four panels show an updip limit of 10 km and a downdip limit of 25 km yields the lowest χ^2 misfit for all four block geometries.

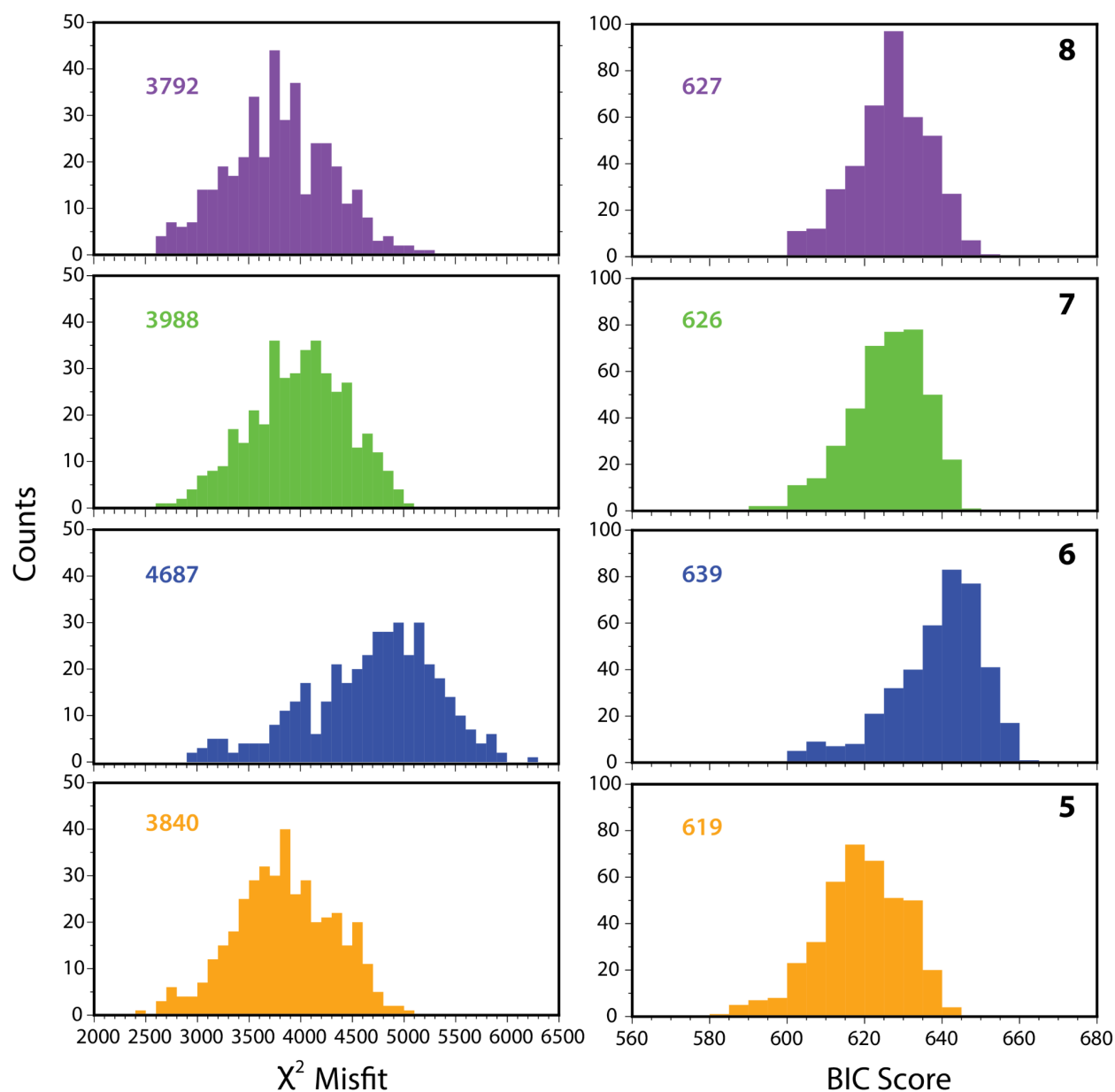


Figure A2.4 Histograms showing the distribution of χ^2 misfit (left) and BIC scores (right) from the bootstrap analysis. Histograms are colored by geometry and labeled in the top right-hand corner of the panels in the right-hand column. The values in the top left-hand corner of each panel represent the mean value for each distribution.

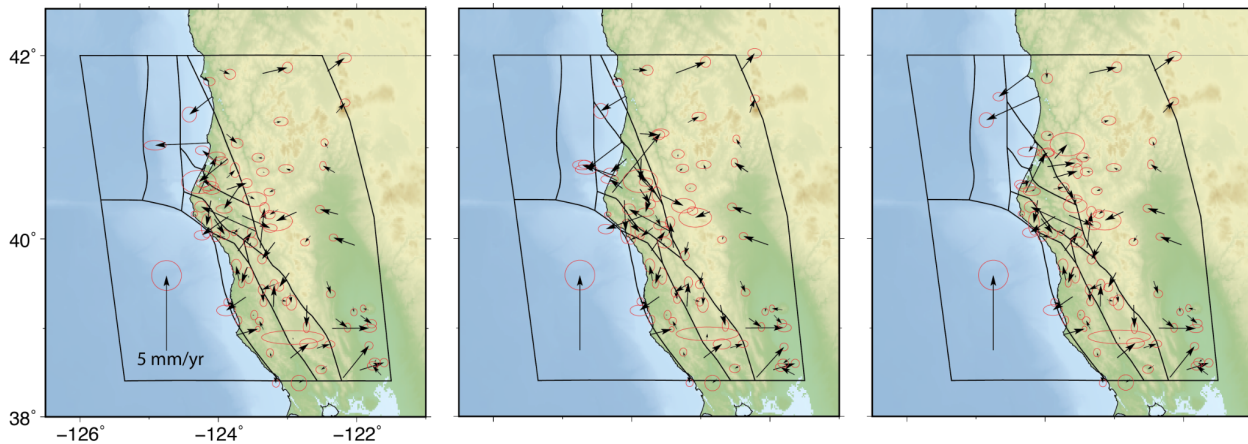


Figure A2.5 Maps showing the average misfit between observed and modeled GPS displacements from the bootstrapping analysis for geometries 5 (left), 6 (middle), and 7 (right). Red ellipses represent 2σ error and all panels share the same scale in the bottom left-hand corner, showing a 1 mm/yr error ellipse.

Text A2.2 Slip rate results for the main strands of the San Andreas fault system (SAFS) are shown below in Figures A2.6, A2.7, and A2.8, and Tables A2.1, A2.2, and A2.3. Overall, these results show consistent slip rates for the SAFS across all four of the model geometries. We observe multiple peaks in many of the slip rate distributions and attribute this multimodal behavior to changes in the fault strike from north to south. For example, while the majority of the San Andreas fault (SAF) is transtensional in these models, reverse slip rates are observed west of longitude -124° (Figure A2.6D). These reverse slip rates manifest as the negative peak in Figure A2.6C and may reflect a change in the obliquity of plate motion along the SAF (Beeson, et al. 2017). Variability in the sign (+/-) of slip rate values along the length of a fault can also lead to misleadingly low median values. In that case, it is helpful to inspect the along-strike line plots in panel D of each figure to see which portions of the fault experience which type of slip.

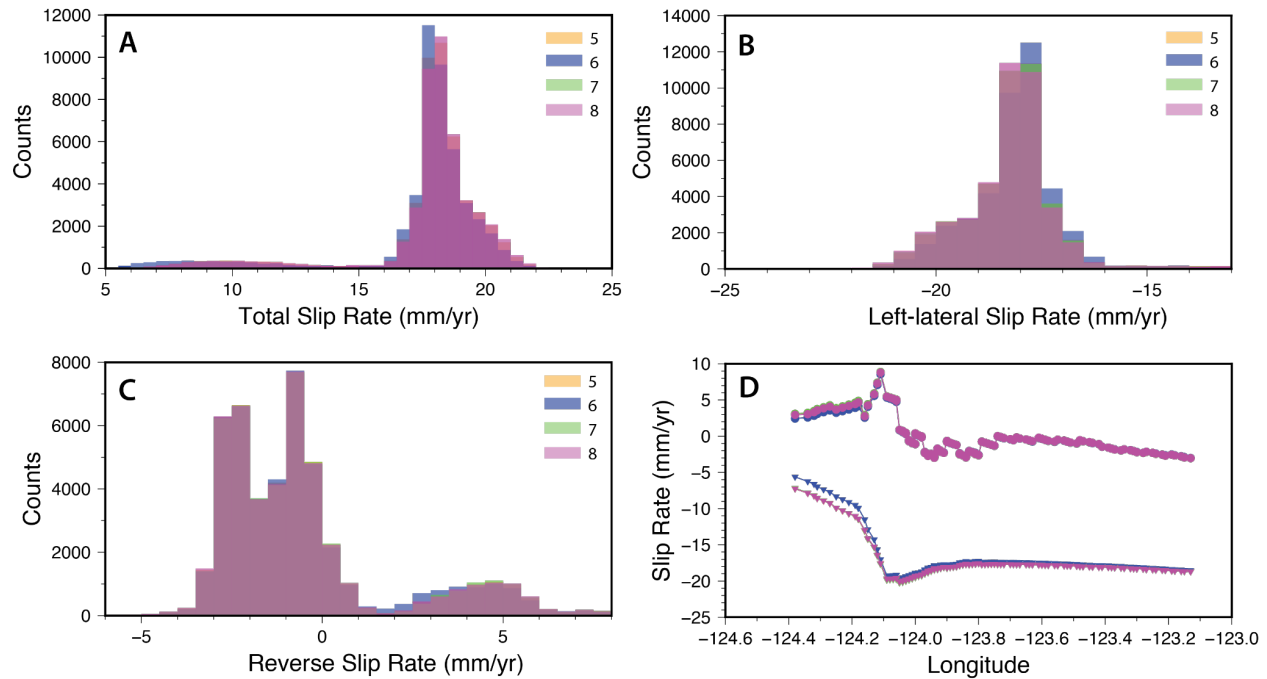


Figure A2.6 Bootstrapping fault slip rate results for the San Andreas fault. **A** Histograms of total slip rate. Histograms are colored by geometry, shown in the upper-right-hand of the panel. **B** Histograms of the left-lateral component of slip rate. Negative values indicate right-lateral slip rates. **C** Histograms of the reverse component of slip rate. Negative values indicate normal slip rates. **D** Line plots showing the median slip rate values at nodes along the fault. Inverted triangles and circles delineate left-lateral and reverse slip rates, respectively. Lines are colored by geometry and match the coloring in panels A, B and C.

	5	6	7	8
Total Slip (mm/yr)	18.2 (17.3 - 19.4)	18.0 (17.0 - 19.2)	18.2 (17.2 - 19.4)	18.2 (17.3 - 19.5)
Reverse Slip (mm/yr)	-1.0 (-2.6 - 2.1)	-1.0 (-2.6 - 1.5)	-1.0 (-2.6 - 1.8)	-1.0 (-2.6 - 1.1)
Left-lateral Slip (mm/yr)	-18.0 (-19.2 to -17.0)	-17.9 (-19.0 to -16.8)	-18.0 (-19.2 to -17.1)	-18.0 (-19.3 to -17.1)

Table A2.1 Bootstrap slip rate results for the San Andreas fault. Model geometry is indicated in the top row and all ranges in parentheses represent the 68% confidence interval. Red values indicate slip rate values from our preferred geometry.

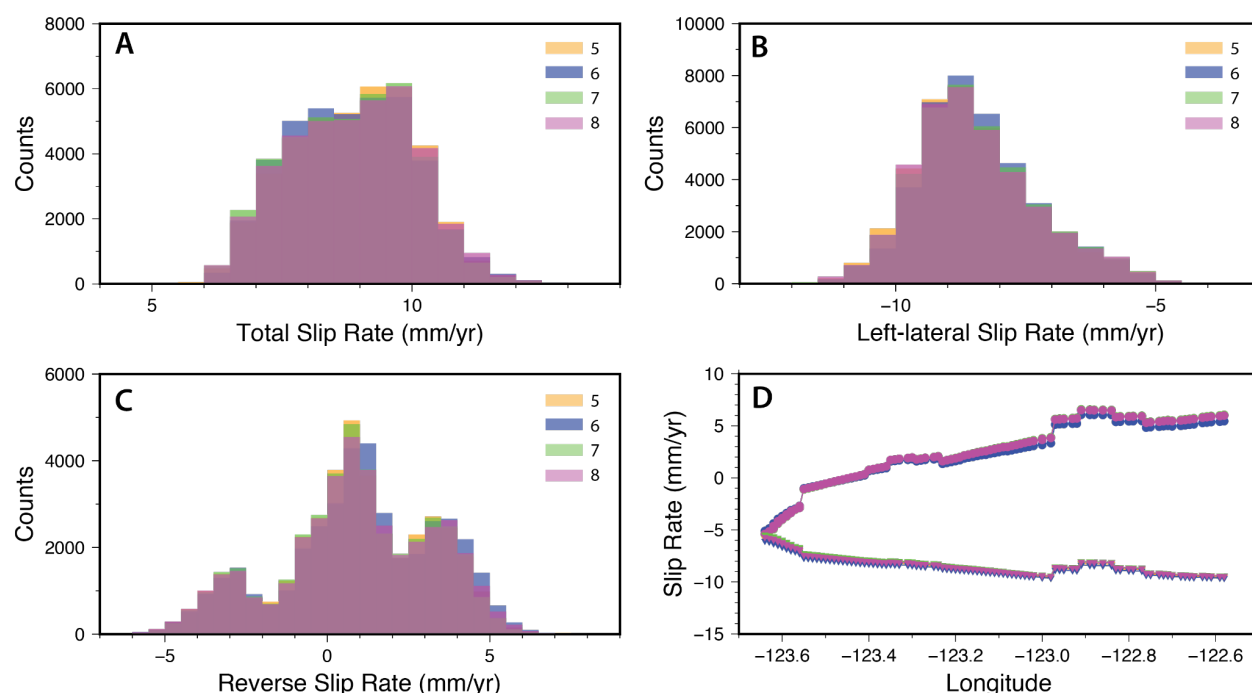


Figure A2.7 Bootstrapping fault slip rate results for the Maacama fault. **A** Histograms of total slip rate. Histograms are colored by geometry, shown in the upper-right-hand of the panel. **B** Histograms of the left-lateral component of slip rate. Negative values indicate right-lateral slip rates. **C** Histograms of the reverse component of slip rate. Negative values indicate normal slip rates. **D** Line plots showing the median slip rate values at nodes along the fault. Inverted triangles and circles delineate left-lateral and reverse slip rates, respectively. Lines are colored by geometry and match the coloring in panels A, B and C.

	5	6	7	8
Total Slip (mm/yr)	8.9 (7.6 - 10.0)	8.8 (7.5 - 10.0)	8.9 (7.5 - 10.0)	8.9 (7.5 - 10.1)
Reverse Slip (mm/yr)	0.9 (-1.5 - 3.4)	1.1 (-1.5 - 3.7)	0.9 (-1.6 - 3.4)	0.9 (-1.5 - 3.5)
Left-lateral Slip (mm/yr)	-8.7 (-9.6 to -7.3)	-8.6 (-9.5 to -7.3)	-8.6 (-9.6 to -7.3)	-8.6 (-9.6 to -7.3)

Table A2.2 Bootstrap slip rate results for the Maacama fault. Model geometry is indicated in the top row and all ranges in parentheses represent the 68% confidence interval. Red values indicate slip rate values from our preferred geometry.

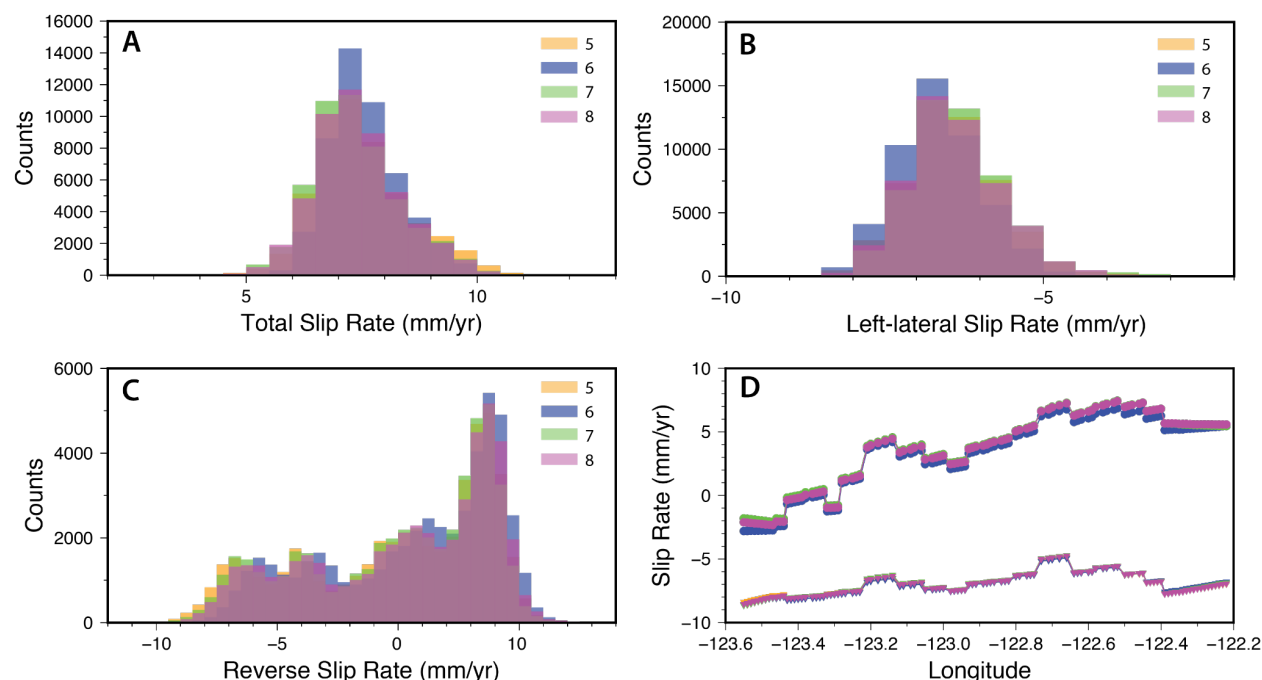


Figure A2.8 Bootstrapping fault slip rate results for the Bartlett Springs fault. **A** Histograms of total slip rate. Histograms are colored by geometry, shown in the upper-right-hand of the panel. **B** Histograms of the left-lateral component of slip rate. Negative values indicate right-lateral slip rates. **C** Histograms of the reverse component of slip rate. Negative values indicate normal slip rates. **D** Line plots showing the median slip rate values at nodes along the fault. Inverted triangles and circles delineate left-lateral and reverse slip rates, respectively. Lines are colored by geometry and match the coloring in panels A, B and C.

	5	6	7	8
Total Slip (mm/yr)	7.3 (6.6 - 8.5)	7.5 (6.8 - 8.4)	7.2 (6.5 - 8.3)	7.3 (6.5 - 8.3)
Reverse Slip (mm/yr)	0.9 (-4.9 - 3.8)	1.6 (-3.9 - 4.1)	0.9 (-4.7 - 3.7)	1.2 (-4.4 - 3.9)
Left-lateral Slip (mm/yr)	-6.5 (-7.1 to -5.7)	-6.7 (-7.3 to -6.0)	-6.4 (-7.0 to -5.6)	-6.5 (-7.1 to -5.7)

Table A2.3 Bootstrap slip rate results for the Bartlett Springs fault. Model geometry is indicated in the top row and all ranges in parentheses represent the 68% confidence interval. Red values indicate slip rate values from our preferred geometry.

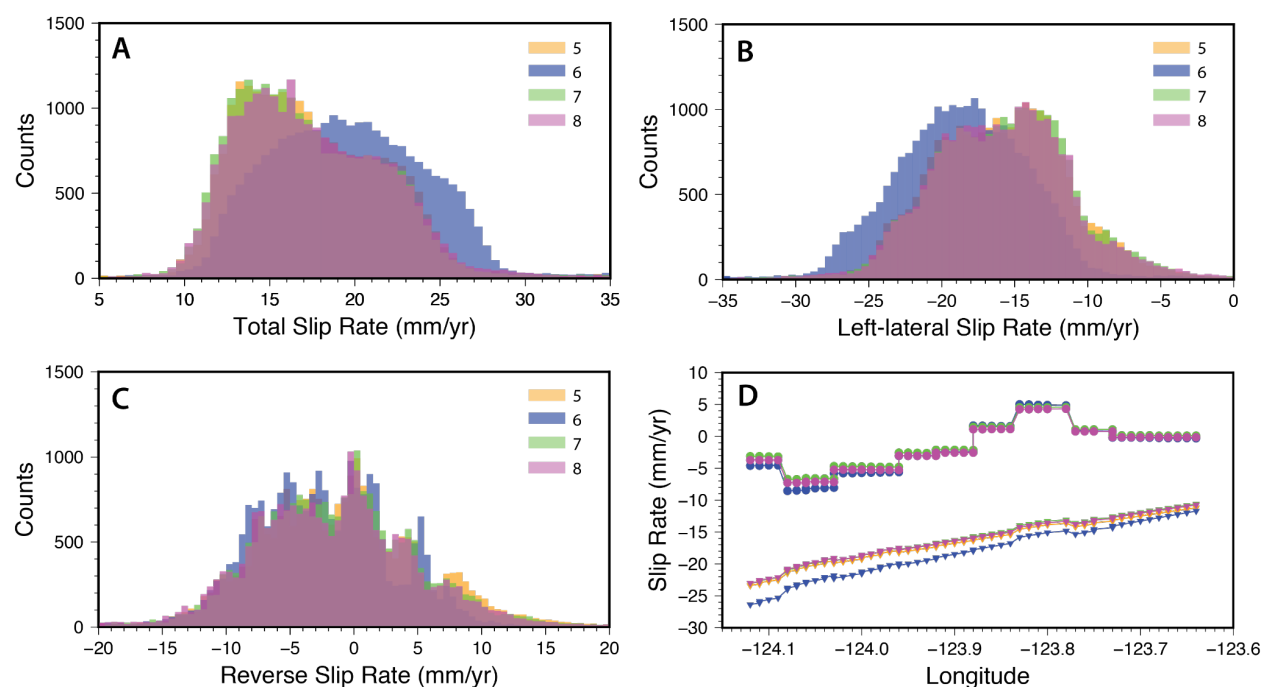


Figure A2.9 Bootstrapping fault slip rate results for the Garberville fault zone. **A** Histograms of total slip rate. Histograms are colored by geometry, shown in the upper-right-hand of the panel. **B** Histograms of the left-lateral component of slip rate. Negative values indicate right-lateral slip rates. **C** Histograms of the reverse component of slip rate. Negative values indicate normal slip rates. **D** Line plots showing the median slip rate values at nodes along the fault. Inverted triangles and circles delineate left-lateral and reverse slip rates, respectively. Lines are colored by geometry and match the coloring in panels A, B and C.

	5	6	7	8
Total Slip (mm/yr)	16.9 (13.2 - 22.2)	19.5 (14.8 - 24.7)	16.6 (12.9 - 21.9)	17.0 (13.1 - 22.3)
Reverse Slip (mm/yr)	-1.3 (-7.4 - 5.1)	-2.7 (-8.1 - 2.8)	-1.4 (-7.4 - 4.6)	-1.8 (-8.0 - 4.4)
Left-lateral Slip (mm/yr)	-16.0 (-20.7 to -11.6)	-18.6 (-23.2 to -14.1)	-15.7 (-20.5 to -11.4)	-15.9 (-20.6 to -11.4)

Table A2.4 Bootstrap slip rate results for the Garberville fault zone. Model geometry is indicated in the top row and all ranges in parentheses represent the 68% confidence interval. Red values indicate slip rate values from our preferred geometry.

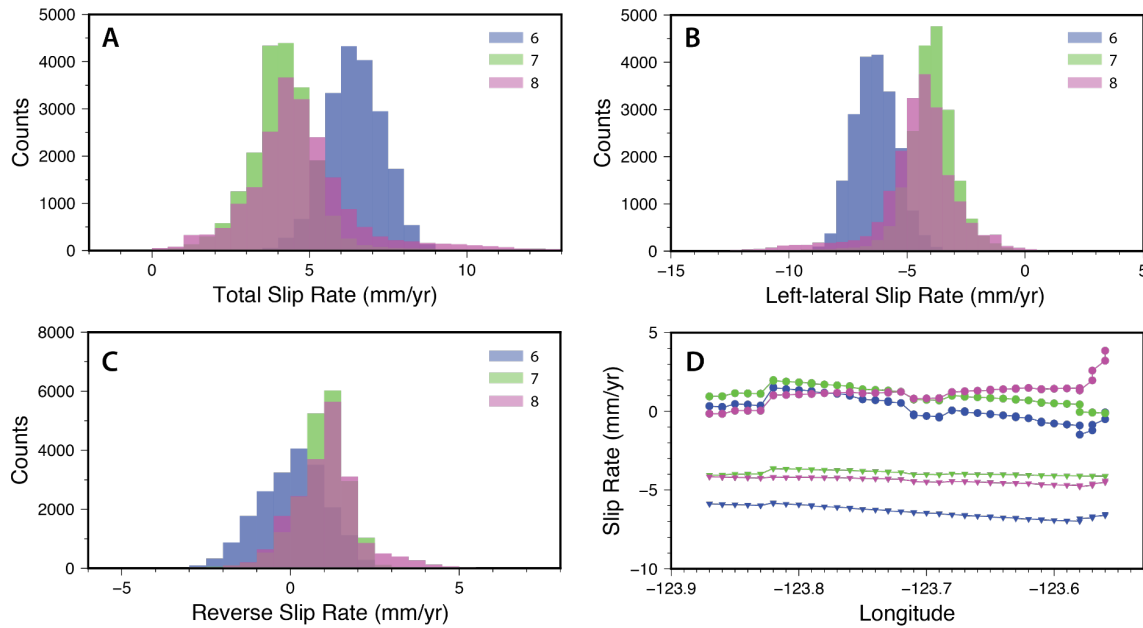


Figure A2.10 Bootstrapping fault slip rate results for the Eaton Roughs fault zone. **A** Histograms of total slip rate. Histograms are colored by geometry, shown in the upper-right-hand of the panel. **B** Histograms of the left-lateral component of slip rate. Negative values indicate right-lateral slip rates. **C** Histograms of the reverse component of slip rate. Negative values indicate normal slip rates. **D** Line plots showing the median slip rate values at nodes along the fault. Inverted triangles and circles delineate left-lateral and reverse slip rates, respectively. Lines are colored by geometry and match the coloring in panels A, B and C.

	5	6	7	8
Total Slip (mm/yr)	-	6.4 (5.5 - 7.3)	4.1 (3.2 - 5.0)	4.5 (3.4 - 5.9)
Reverse Slip (mm/yr)	-	0.1 (-1.0 to 1.1)	1.0 (0.3 - 1.6)	1.1 (0.1 - 1.8)
Left-lateral Slip (mm/yr)	-	-6.4 (-7.3 to -5.5)	-3.9 (-4.8 to -3.1)	-4.4 (-5.6 to -3.2)

Table A2.5 Bootstrap slip rate results for the Eaton Roughs fault zone (ERFZ). Model geometry is indicated in the top row and all ranges in parentheses represent the 68% confidence interval. Red values indicate slip rate values from our preferred geometry. Geometry 5 does not include the ERFZ.

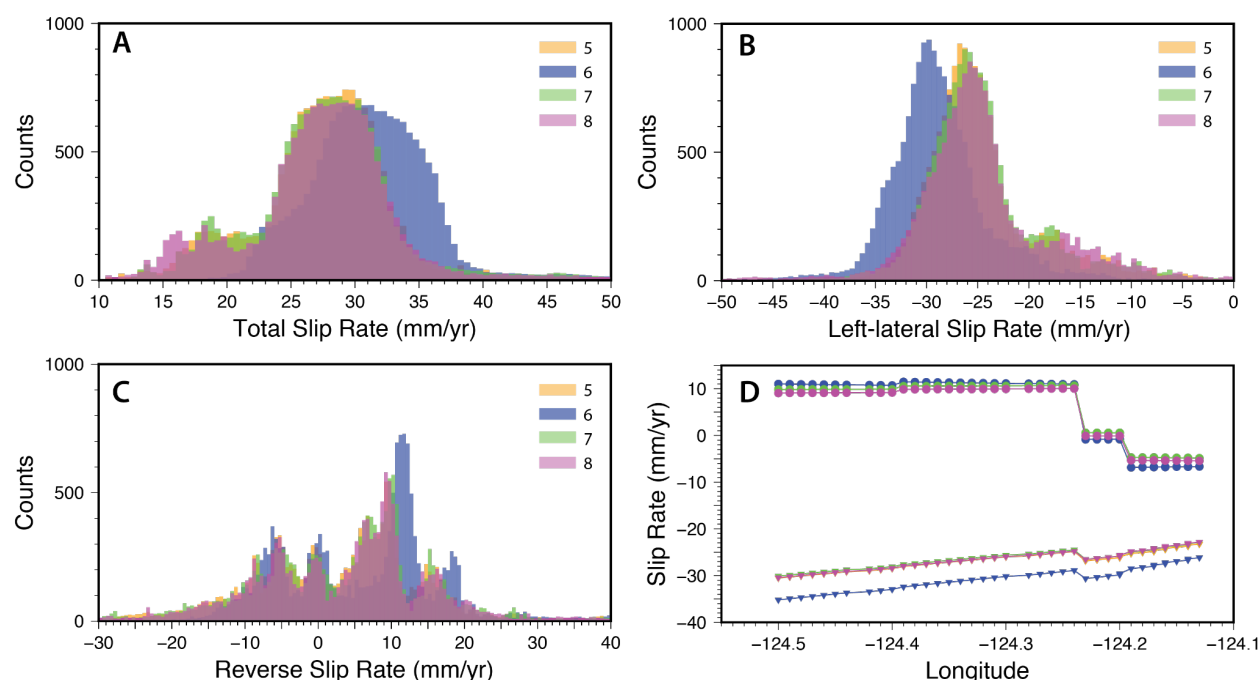


Figure A2.11 Bootstrapping fault slip rate results for the Bear River fault zone. **A** Histograms of total slip rate. Histograms are colored by geometry, shown in the upper-right-hand of the panel. **B** Histograms of the left-lateral component of slip rate. Negative values indicate right-lateral slip rates. **C** Histograms of the reverse component of slip rate. Negative values indicate normal slip rates. **D** Line plots showing the median slip rate values at nodes along the fault. Inverted triangles and circles delineate left-lateral and reverse slip rates, respectively. Lines are colored by geometry and match the coloring in panels A, B and C.

	5	6	7	8
Total Slip (mm/yr)	27.9 (22.6 - 32.0)	31.2 (26.8 - 35.7)	27.8 (22.6 - 31.9)	27.9 (22.3 - 32.3)
Reverse Slip (mm/yr)	4.6 (-8.4 to 14.1)	9.2 (-5.9 to 16.6)	5.2 (-8.4 to 14.4)	5.6 (-8.6 to 15.0)
Left-lateral Slip (mm/yr)	-25.7 (-29.3 to -19.4)	-29.3 (-32.9 to -24.9)	-25.5 (-29.1 to -19)	-25.4 (-29.5 to -17.6)

Table A2.6 Bootstrap slip rate results for the Bear River fault zone. Model geometry is indicated in the top row and all ranges in parentheses represent the 68% confidence interval. Red values indicate slip rate values from our preferred geometry.

Text A2.3 The average strain rate results and average block motion results from the bootstrap analysis are displayed in Figure A2.12 for geometries 5-8. Overall, the results are fairly consistent across all of the geometries. For geometry 8, all of the blocks within the North American plate are moving to the Northwest with varying velocities. This motion is consistent with the northwest migration and impingement of the Sierra Nevada-Great Valley block (Williams et al., 2006; McKenzie & Furlong, 2021). In terms of block strain, the two southernmost blocks associated with the San Andreas fault system show shear-to-extensional strain rates. Moving to the north, the forearc blocks begin to show large components of ENE-oriented compression. However, the MAD block (Figure A2.16) exhibits extensional strains. Inspection of the bootstrap results shows that there is a bimodal distribution of strain for this block, wherein one of the peaks lies near zero and the other lies in the extensional regime. Therefore, although the average strain rate results favor extension, the alternate result provides strain rates that are more consistent with shear. This alternate result is consistent with our regional strain rate maps (Figure 3.2) and with non-zero right-lateral slip-rates on the MRFZ and Little Salmon systems.

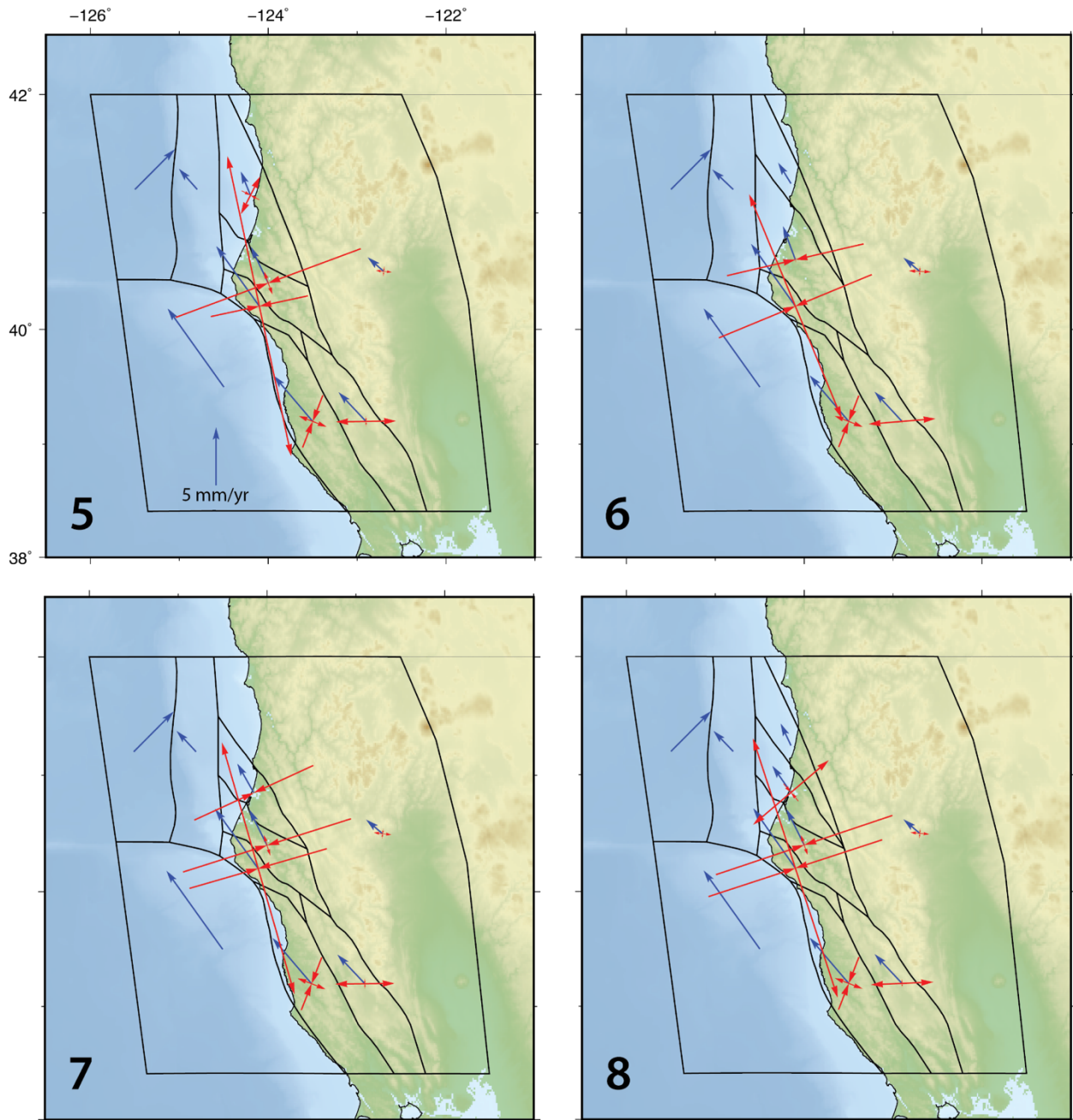


Figure A2.12 Bootstrap results for average strain rates (red arrows) and block motions (blue vectors) for block geometries 5-8. Geometry number is indicated in the bottom left-hand corner of each panel and all maps share the same scale shown for geometry 5.

Text A2.4 Figures A2.13, A2.14, A2.15, and A2.16 show the boundaries and names of the blocks used for geometries 5-8. All of the models share identical Juan de Fuca (JDF), Pacific (PAC), Offshore (OFFS), BEAR, MAC, and BART blocks. With the exception of the North America block (NA), all of the onshore blocks are named after the north-bounding fault: Maacama (MAC), Bartlett Springs (BART), Bear River fault zone (BEAR), Little Salmon (SALM), Mad River fault zone (MAD), and Grogan (GROG). Tables A2.7, A2.8, A2.9 and A2.10 provide the average Euler pole and strain rate bootstrap results for geometries 5-8, where ϵ_1 is the most compressive strain rate, ϵ_2 is the least compressive strain rate, and Azimuth is the azimuth of ϵ_1 in degrees clockwise from North. For these strain rate values, negative values correspond to compression and positive values correspond to extension.

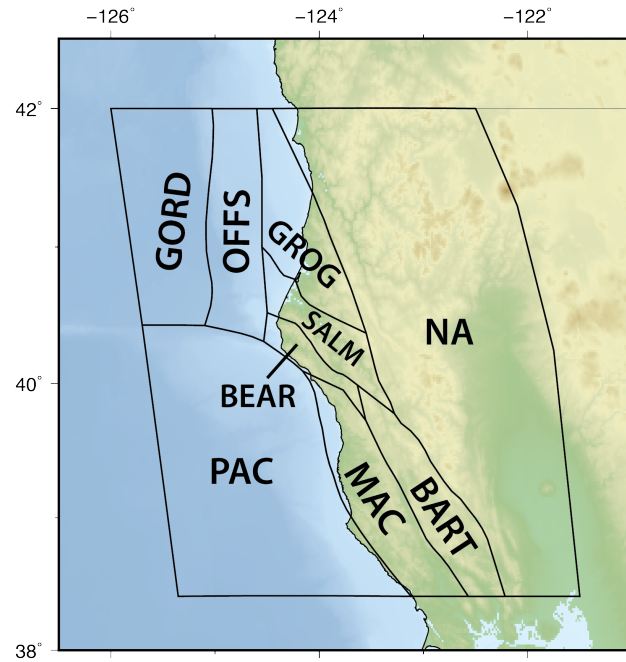


Figure A2.13. Block boundaries (black lines) and names used for geometry 5.

Block	Latitude (°N)	Longitude (°E)	Rotation Rate (°/Ma)	ϵ_1 (μ -strain)	ϵ_2 (μ -strain)	Azimuth (°)
BART	6.23 ± 4.47	-152.46 ± 3.31	0.28 ± 0.03	-12.9 ± 10.7	78.4 ± 22.6	179.2 ± 8.2
BEAR	47.24 ± 1.40	-107.10 ± 2.12	-1.42 ± 0.05	-129.2 ± 363.4	398.0 ± 398.2	78.0 ± 33.9
GROG	43.01 ± 0.56	-117.22 ± 0.83	-1.25 ± 0.08	-25.6 ± 26.2	55.3 ± 87.8	117.2 ± 32.0
MAC	4.64 ± 2.80	-156.62 ± 2.11	0.39 ± 0.02	-72.6 ± 10.7	38.9 ± 6.7	21.2 ± 3.6
NA	43.69 ± 0.07	-118.76 ± 0.06	-1.31 ± 0.01	-11.9 ± 1.4	21.1 ± 2.9	6.0 ± 2.3
OFFS	45.18 ± 0.61	-118.13 ± 1.06	-1.19 ± 0.15	-	-	-
SALM	43.45 ± 0.28	-114.03 ± 0.57	-1.35 ± 0.03	-256.0 ± 79.1	32.1 ± 21.0	69.6 ± 8.1

Table A2.7. Average Euler pole and strain rate results for geometry 5 from bootstrap analysis.

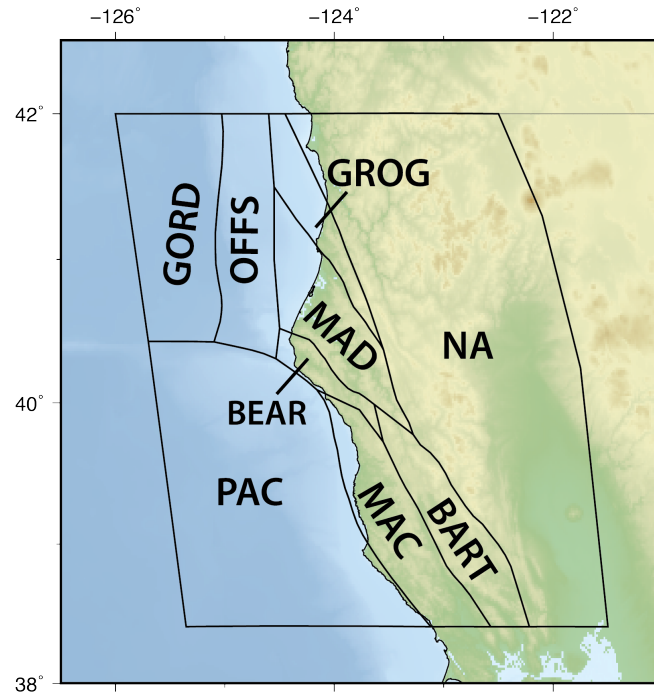


Figure A2.14 Block boundaries (black lines) and names used for geometry 6.

Block	Latitude (°N)	Longitude (°E)	Rotation Rate (°/Ma)	ϵ_1 (μ -strain)	ϵ_2 (μ -strain)	Azimuth (°)
BART	10.38 ± 4.34	-149.89 ± 3.17	0.31 ± 0.04	-6.0 ± 8.6	87.6 ± 25.22	175.3 ± 8.2
BEAR	47.24 ± 1.99	-106.90 ± 1.76	-1.45 ± 0.06	-214.9 ± 679.4	314.9 ± 362.8	67.5 ± 20.7
GROG	43.75 ± 0.44	-118.21 ± 0.41	-1.29 ± 0.09	-	-	-
MAC	1.70 ± 2.89	-158.88 ± 2.16	0.37 ± 0.02	-71.8 ± 11.4	39.4 ± 6.7	21.4 ± 3.6
MAD	42.77 ± 0.34	-114.81 ± 0.54	-1.31 ± 0.04	-182.1 ± 45.2	-6.5 ± 11.5	76.9 ± 7.6
NA	43.77 ± 0.06	-118.86 ± 0.08	-1.31 ± 0.02	-12.6 ± 1.3	29.1 ± 2.7	3.2 ± 1.9
OFFS	45.65 ± 1.16	-118.52 ± 0.75	-1.23 ± 0.18	-	-	-

Table A2.8. Average Euler pole and strain rate results for geometry 6 from bootstrap analysis.

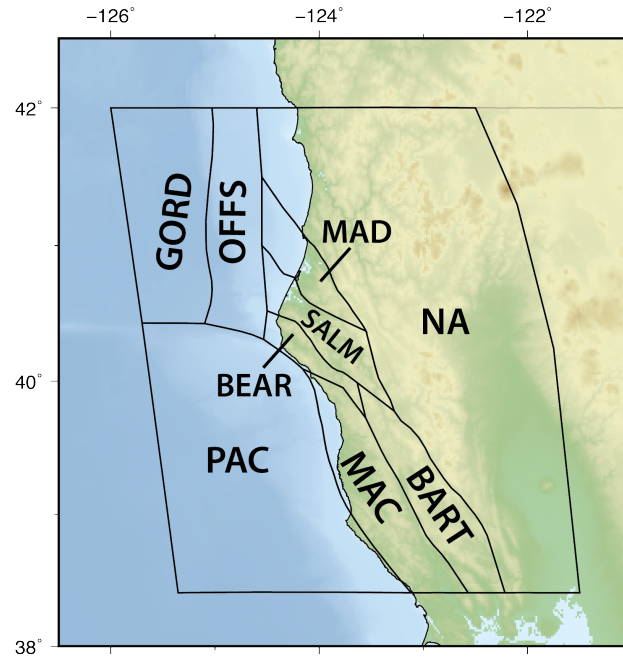


Figure A2.15. Block boundaries (black lines) and names used for geometry 7.

Block	Latitude (°N)	Longitude (°E)	Rotation Rate (°/Ma)	ϵ_1 (μ -strain)	ϵ_2 (μ -strain)	Azimuth (°)
BART	8.86 ± 4.35	-150.58 ± 3.16	0.30 ± 0.03	-12.5 ± 10.4	77.5 ± 21.6	179.4 ± 8.2
BEAR	47.30 ± 1.12	-107.32 ± 1.47	-1.42 ± 0.12	-185.7 ± 538.2	340.4 ± 326.3	73.9 ± 31.3
MAC	2.47 ± 2.88	-158.18 ± 2.17	0.38 ± 0.02	-73.4 ± 11.6	39.1 ± 6.9	21.2 ± 3.7
MAD	44.06 ± 0.49	-115.37 ± 0.90	-1.25 ± 0.10	-169.9 ± 45.3	2.5 ± 35.5	65.4 ± 18.7
NA	43.78 ± 0.07	-118.76 ± 0.07	-1.31 ± 0.00	-12.5 ± 1.4	23.4 ± 2.7	7.0 ± 2.1
OFFS	45.42 ± 0.64	-118.07 ± 1.07	-1.19 ± 0.15	-	-	-
SALM	43.52 ± 0.30	-114.24 ± 0.65	-1.35 ± 0.05	-228.9 ± 82.0	28.1 ± 25.5	72.4 ± 12.5

Table A2.9 Average Euler pole and strain rate results for geometry 7 from bootstrap analysis.

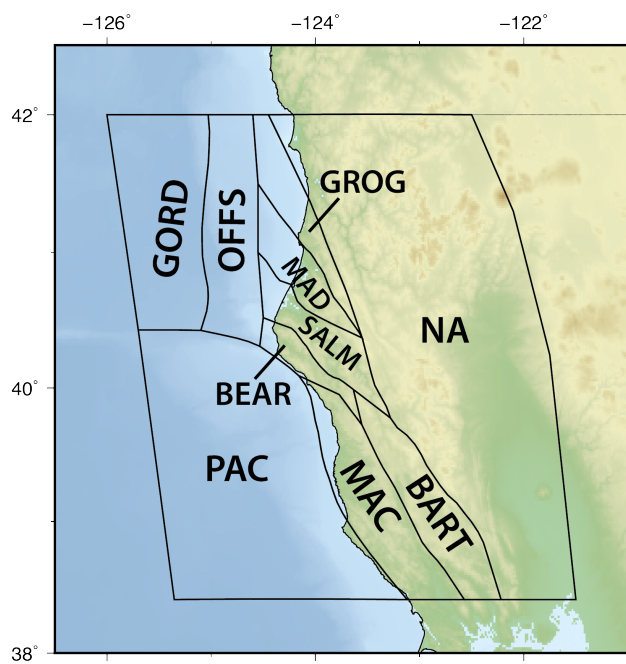


Figure A2.16. Block boundaries (black lines) and names used for geometry 8.

Block	Latitude (°N)	Longitude (°E)	Rotation Rate (°/Ma)	ϵ_1 (μ -strain)	ϵ_2 (μ -strain)	Azimuth (°)
BART	11.20 ± 3.54	-149.02 ± 2.60	0.32 ± 0.03	-10.5 ± 9.8	81.3 ± 23.3	177.6 ± 8.4
BEAR	47.18 ± 1.58	-107.02 ± 1.80	-1.4 ± 0.09	-237.3 ± 658.9	354.5 ± 360.9	71.9 ± 28.4
GROG	43.10 ± 0.32	-118.29 ± 0.37	-1.37 ± 0.08	-	-	-
MAC	4.69 ± 2.77	-156.55 ± 2.10	0.39 ± 0.02	-73.3 ± 10.8	38.7 ± 6.6	21.2 ± 3.6
MAD	44.54 ± 1.05	-116.23 ± 1.04	-1.26 ± 0.07	-29.3 ± 61.1	129.2 ± 113.4	139.6 ± 35.9
NA	43.74 ± 0.07	-118.81 ± 0.07	-1.32 ± 0.02	-12.1 ± 1.4	24.9 ± 2.7	5.1 ± 2.1
OFFS	45.58 ± 0.67	-117.99 ± 1.13	-1.17 ± 0.17	-	-	-
SALM	43.58 ± 0.37	-113.97 ± 0.62	-1.32 ± 0.04	-214.8 ± 90.8	27.2 ± 30.1	71.5 ± 12.1

Table A2.10 Average Euler pole and strain rate results for geometry 8 from bootstrap analysis.

	BRFZ	BS	ERFZ	Ga	Gr	LS	Ma	MRFZ	Off	Sa
Dip Angle (°)	60	89	89	89	89	30	60	35	45	89

Table A2.11 Dip values for the crustal faults in our block models. Note that TDEFNODE does not allow for dips of 90°, so vertical faults are assigned dips of 89°.

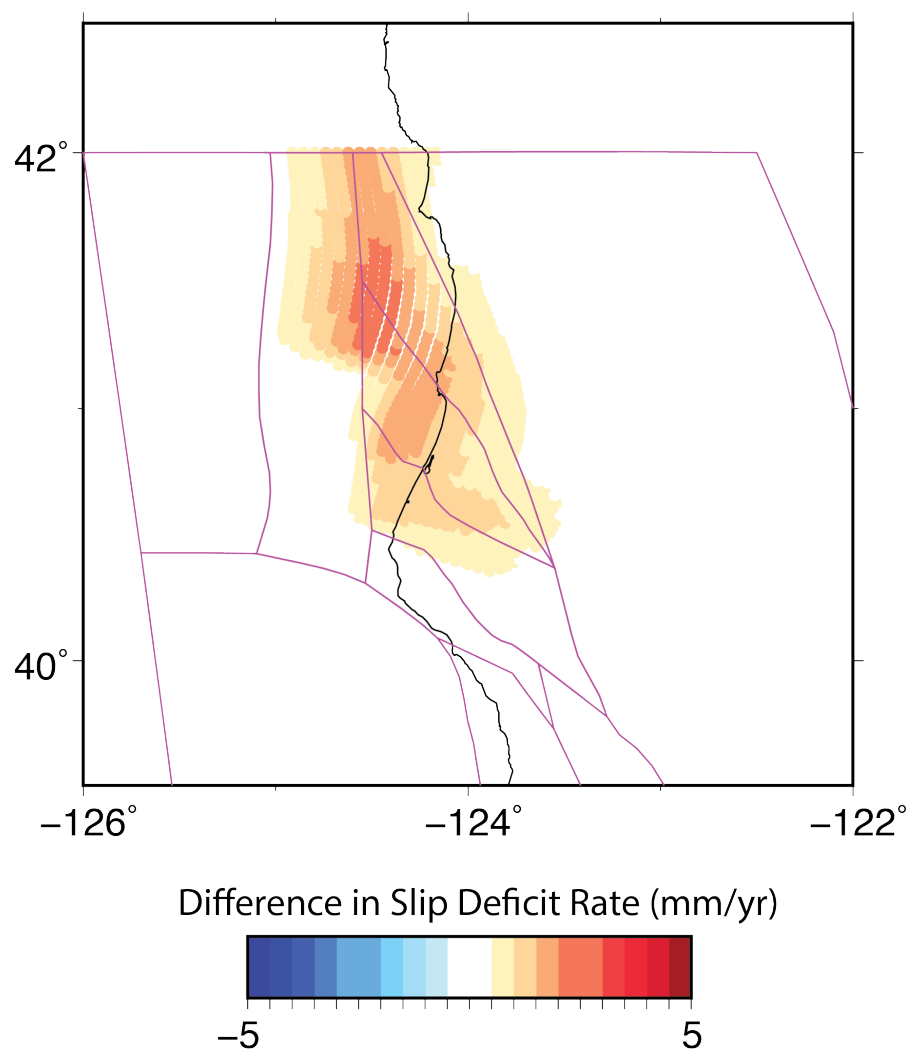


Figure A2.17 Map of the MTJ region showing the difference in slip deficit rates on the megathrust between geometries 5 and 8, where positive values indicate higher slip deficit rates in geometry 8 compared to geometry 5. Changes in slip deficit rate due to the OFFS block are not included. Magenta lines delineate the block boundaries for geometry 8.

APPENDIX 3: SUPPLEMENTARY MATERIALS FOR CHAPTER 4

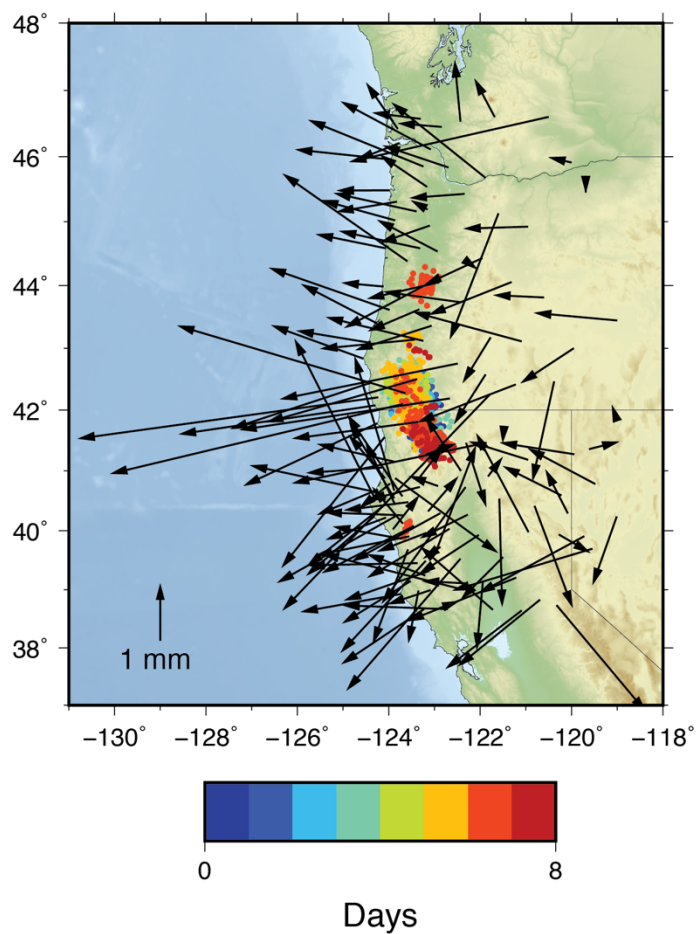


Figure A3.1 Map of displacements for ETS event 9 before the removal of the common mode signal.

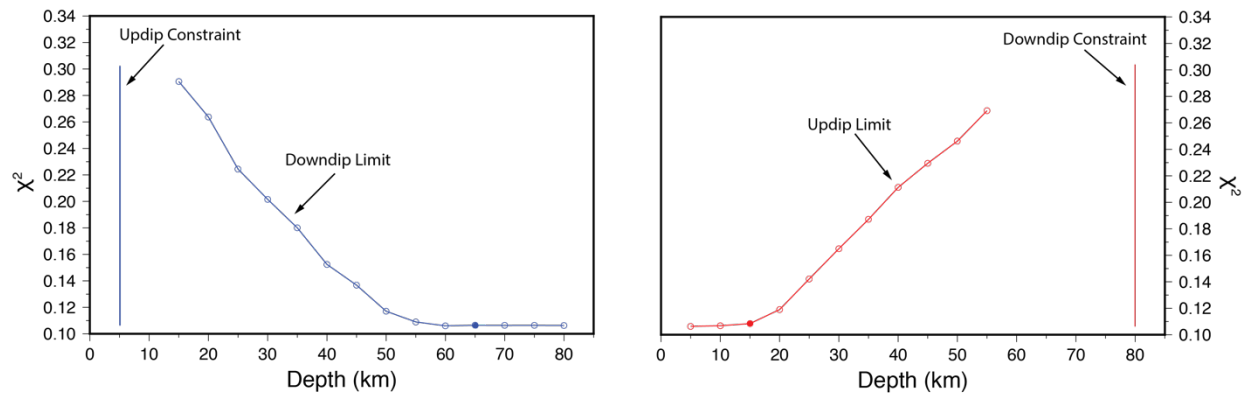


Figure A3.2. Results from our megathrust depth analysis. **Left** Results for the downdip limit of the megathrust fault. The vertical bar indicates the updip depth constraint and each empty circle represents an inversion run with a different downdip depth limit. The filled circle indicates the downdip depth that we use for the final inversion. **Right** Results for the updip limit of the megathrust fault. The vertical bar indicates the downdip depth constraint and each empty circle represents an inversion run with a different updip depth limit. The filled circle indicates the updip depth that we use for the final inversion.

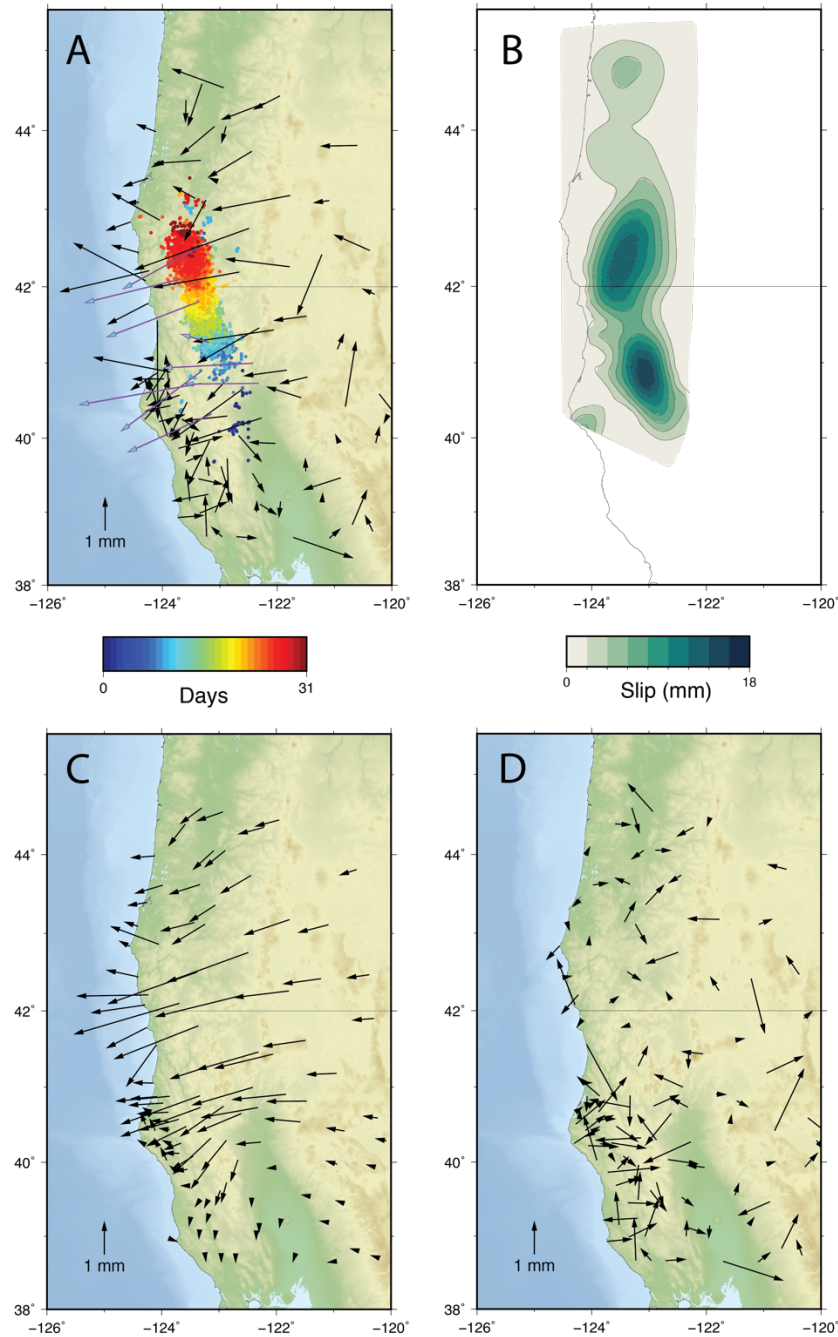


Figure A3.3 Model results for ETS Event 1 (August 2016). **A** Vector map showing horizontal surface displacements for ETS event 15. Purple vectors with cyan filling show displacements at index stations. Circles show the location of tremor during the event and are colored by the number of days that passed since the first instance of tremor. **B** Distributed slip model from our inversion with the contours (black lines with green shading) outlining every 2 mm of fault slip. **C** Vector map showing the predicted surface displacements resulting from the slip model in panel B. **D** Vector map showing the misfit between the observed and predicted surface displacements.

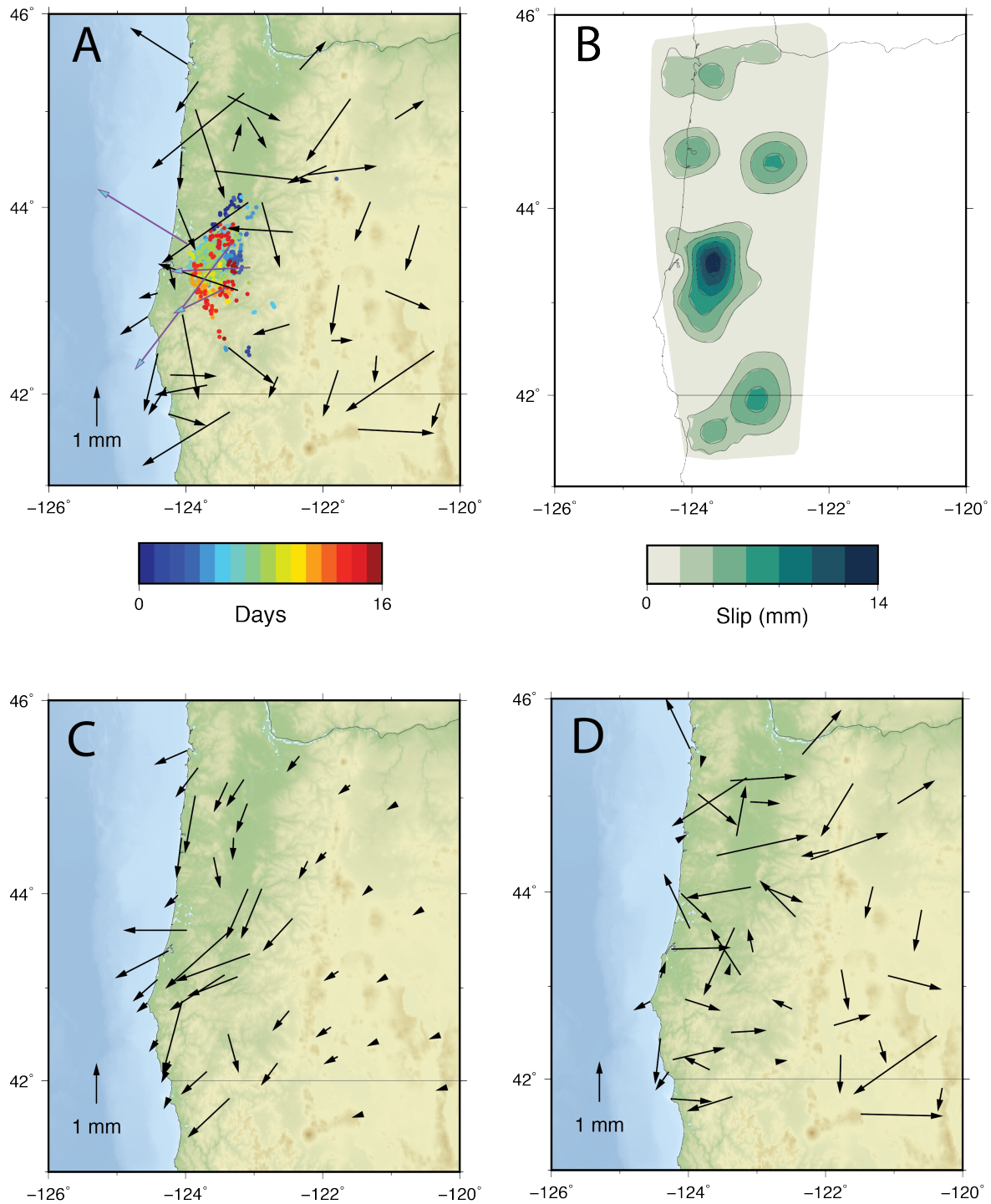


Figure A3.4 Model results for ETS Event 2 (September 2016). Panels are the same as Figure A3.3.

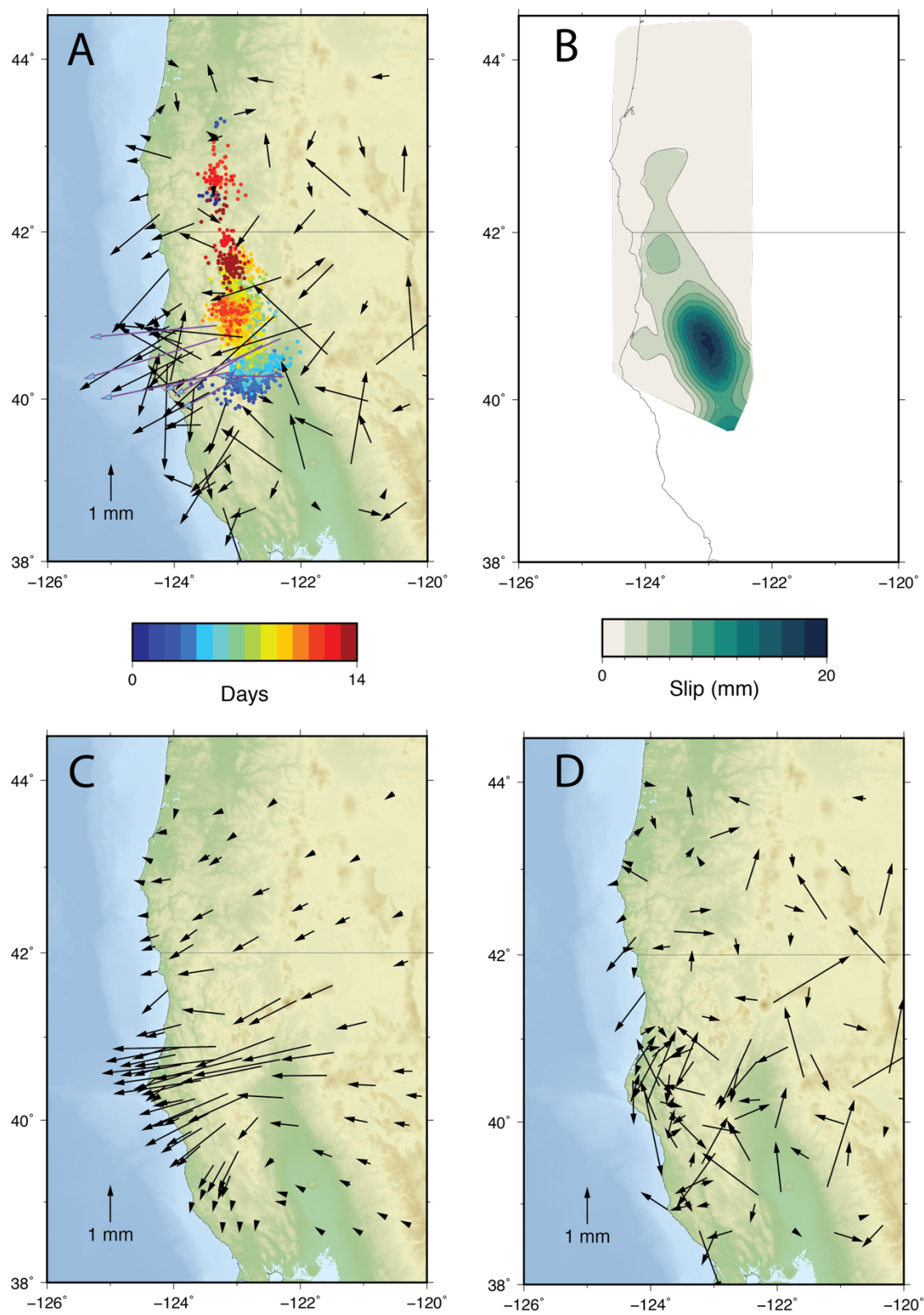


Figure A3.5 Model results for ETS Event 3 (April 2017). Panels are the same as Figure A3.3.

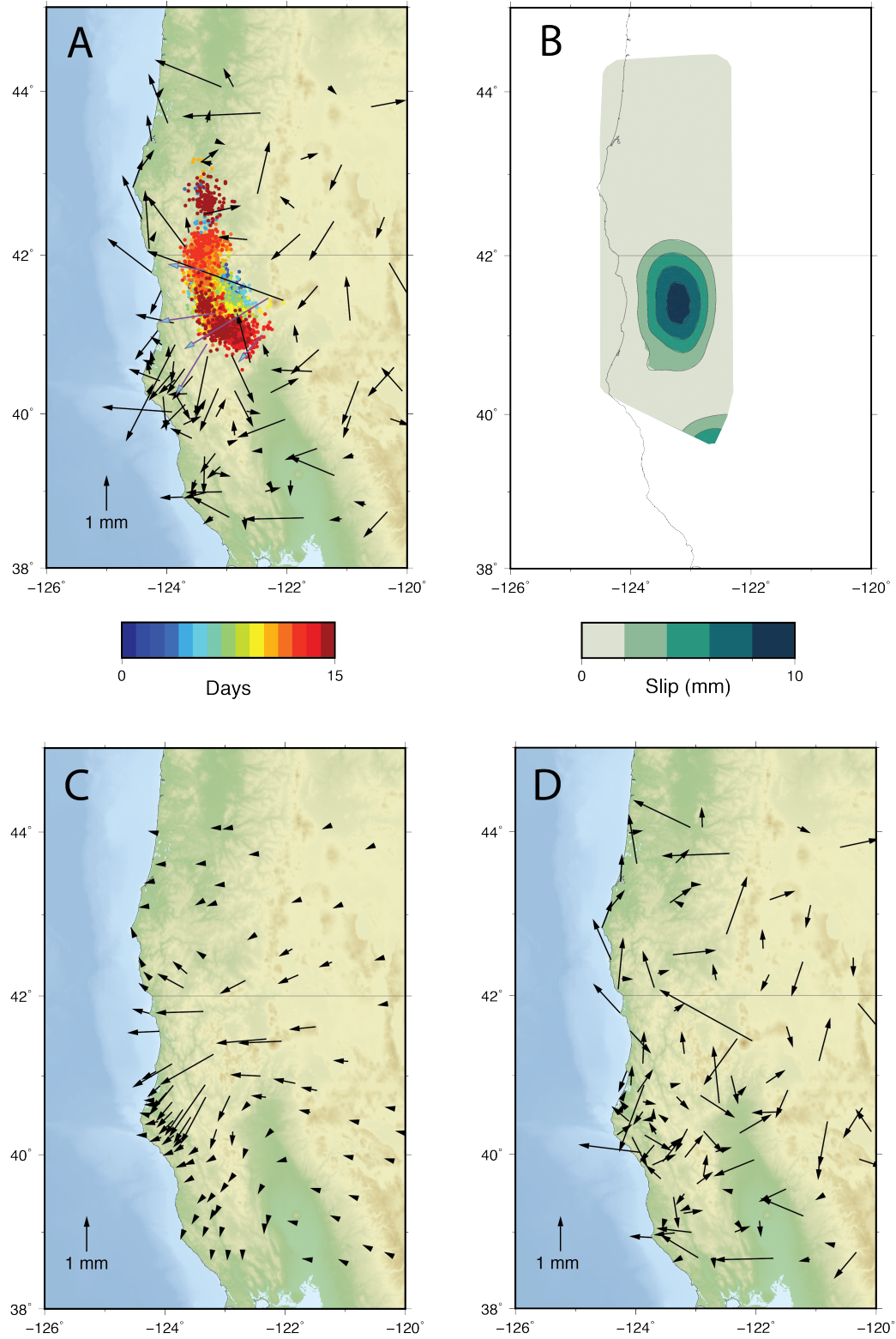


Figure A3.6 Model results for ETS Event 4 (July 2017). Panels are the same as Figure A3.3.

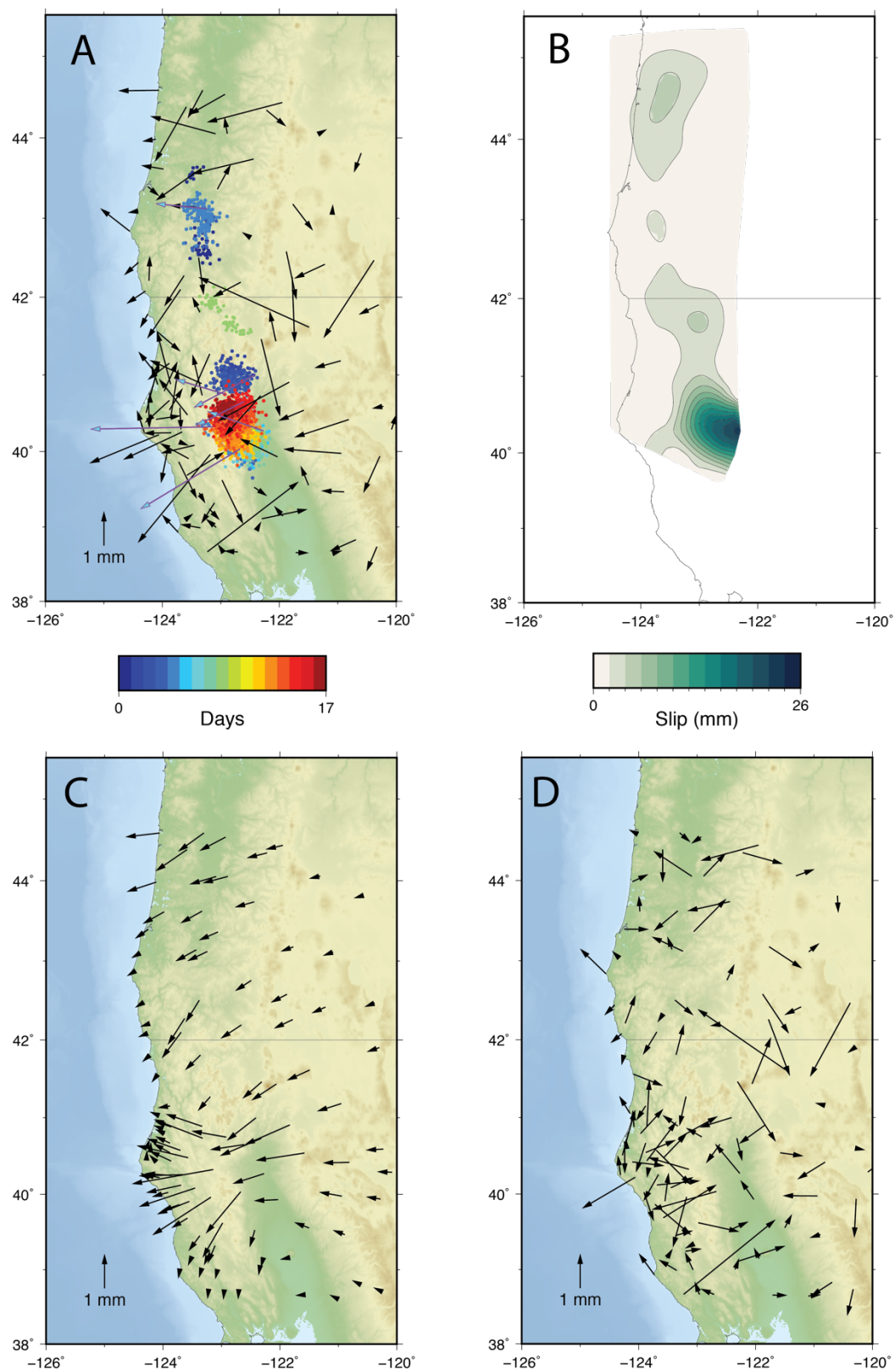


Figure A3.7 Model results for ETS Event 5 (October 2017). Panels are the same as Figure A3.3.

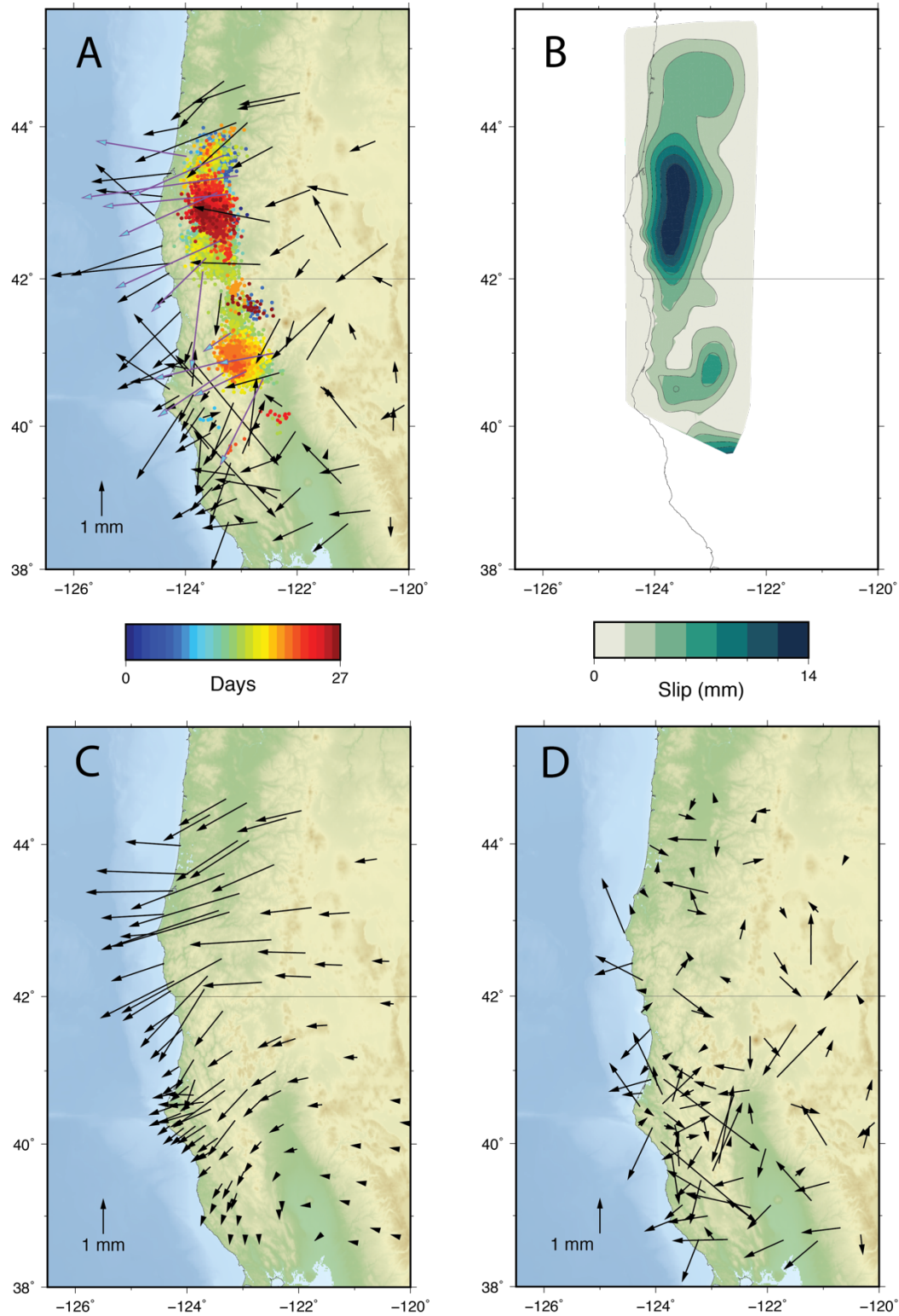


Figure A3.8 Model results for ETS Event 6 (December 2017). Panels are the same as Figure A3.3.

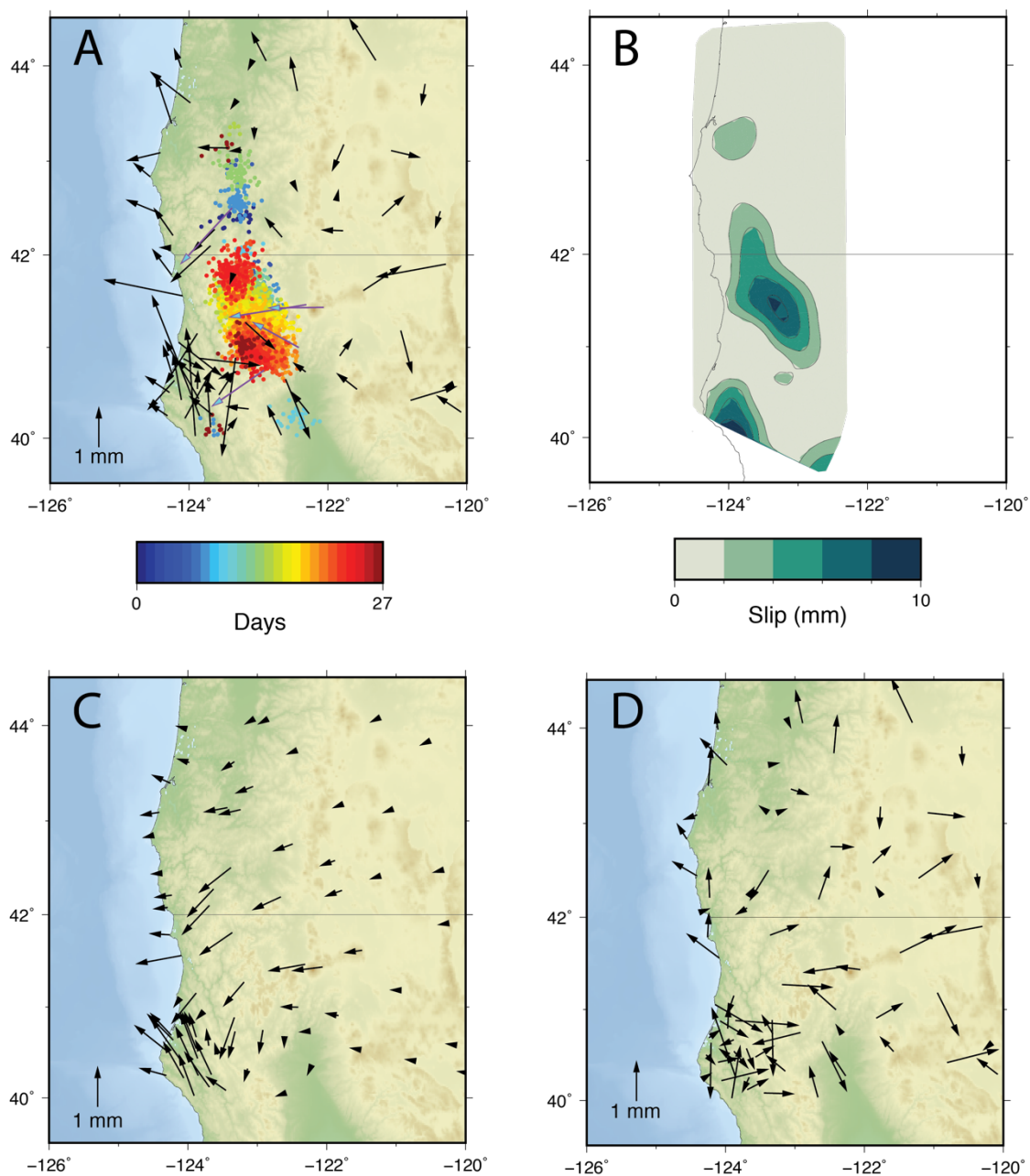


Figure A3.9 Model results for ETS Event 7 (May 2018). Panels are the same as Figure A3.3.

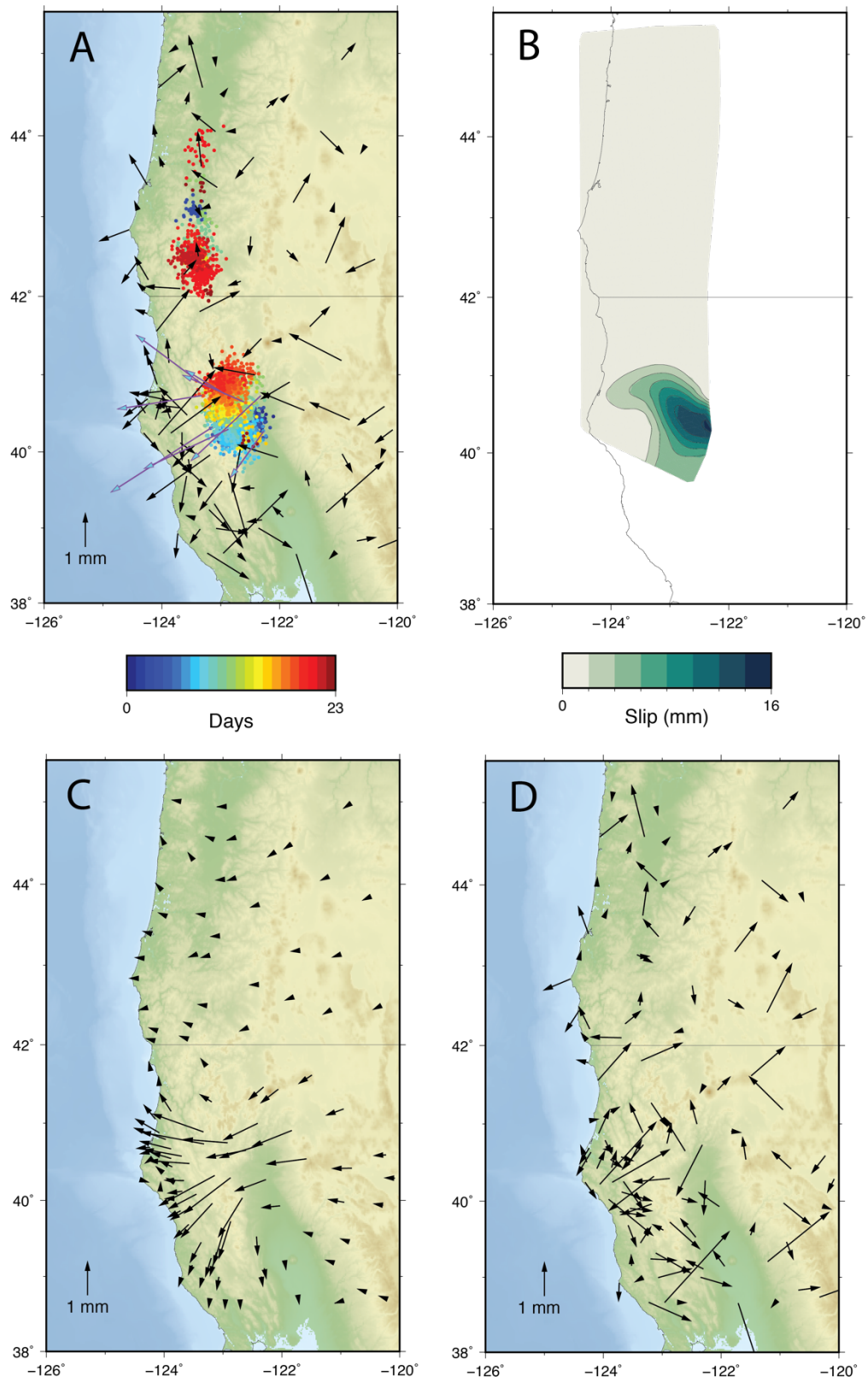


Figure A3.10 Model results for ETS Event 8 (June 2018). Panels are the same as Figure A3.3.

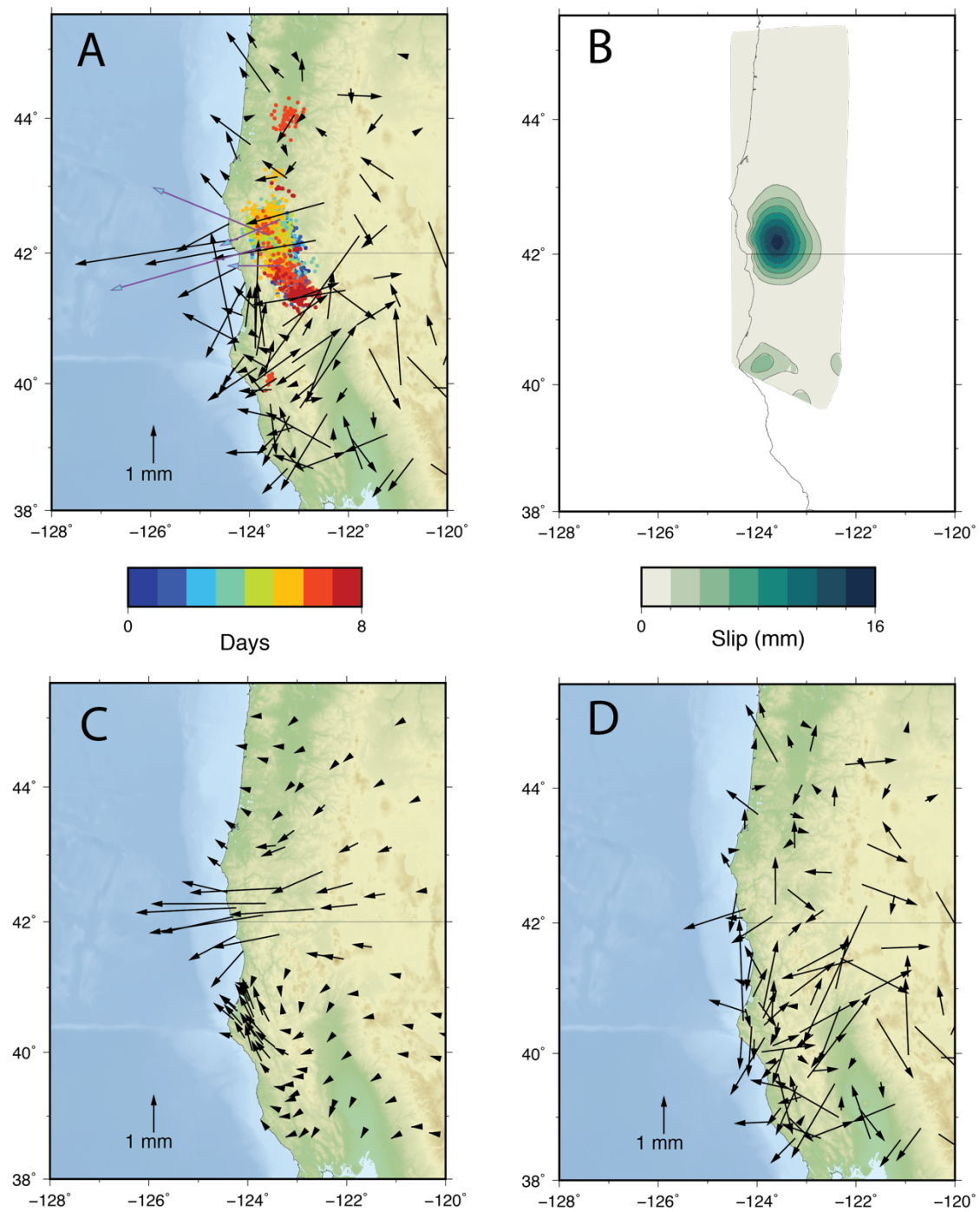


Figure A3.11 Model results for ETS Event 9 (November 2018). Panels are the same as Figure A3.3.

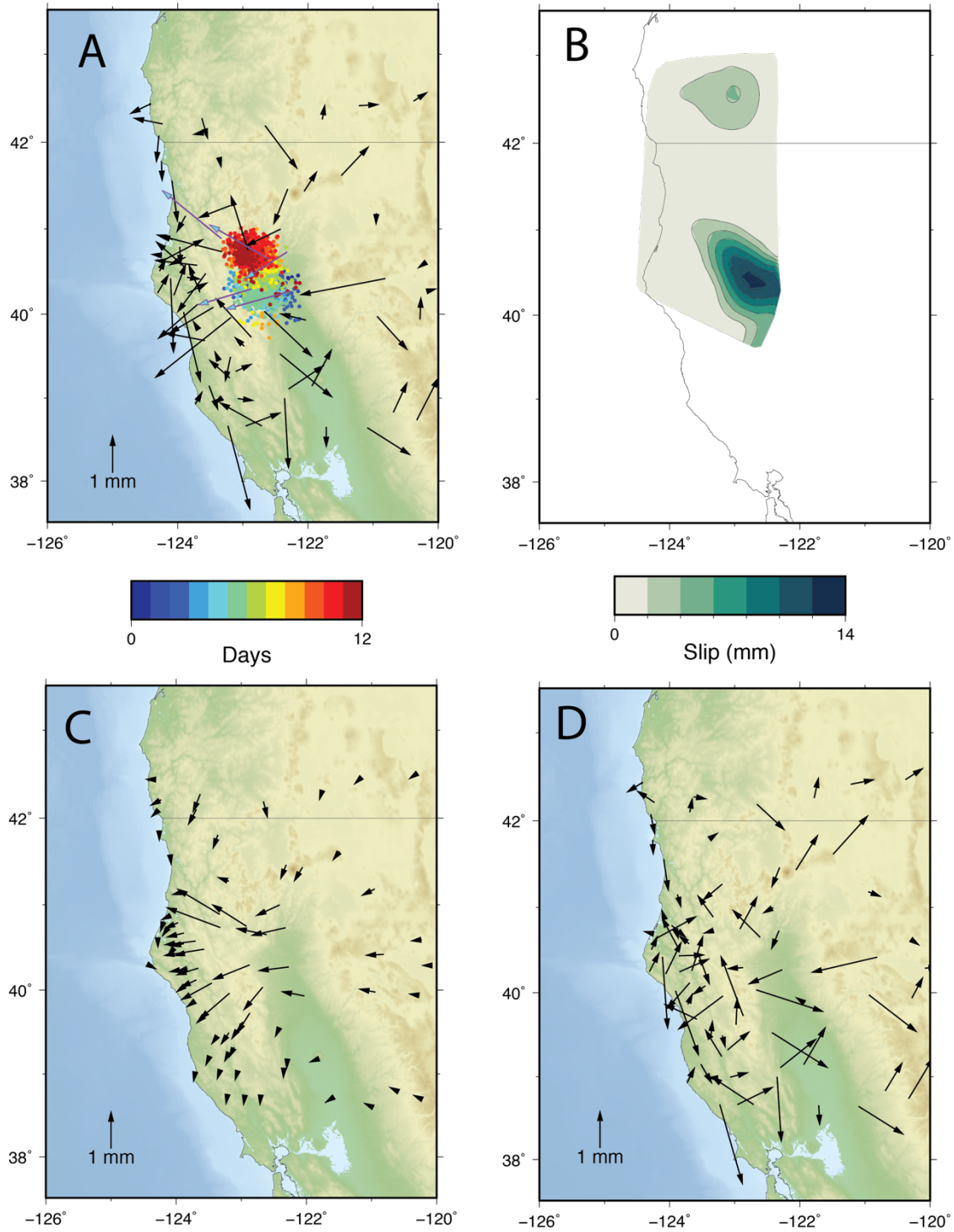


Figure A3.12 Model results for ETS Event 10 (February 2019). Panels are the same as Figure A3.3.

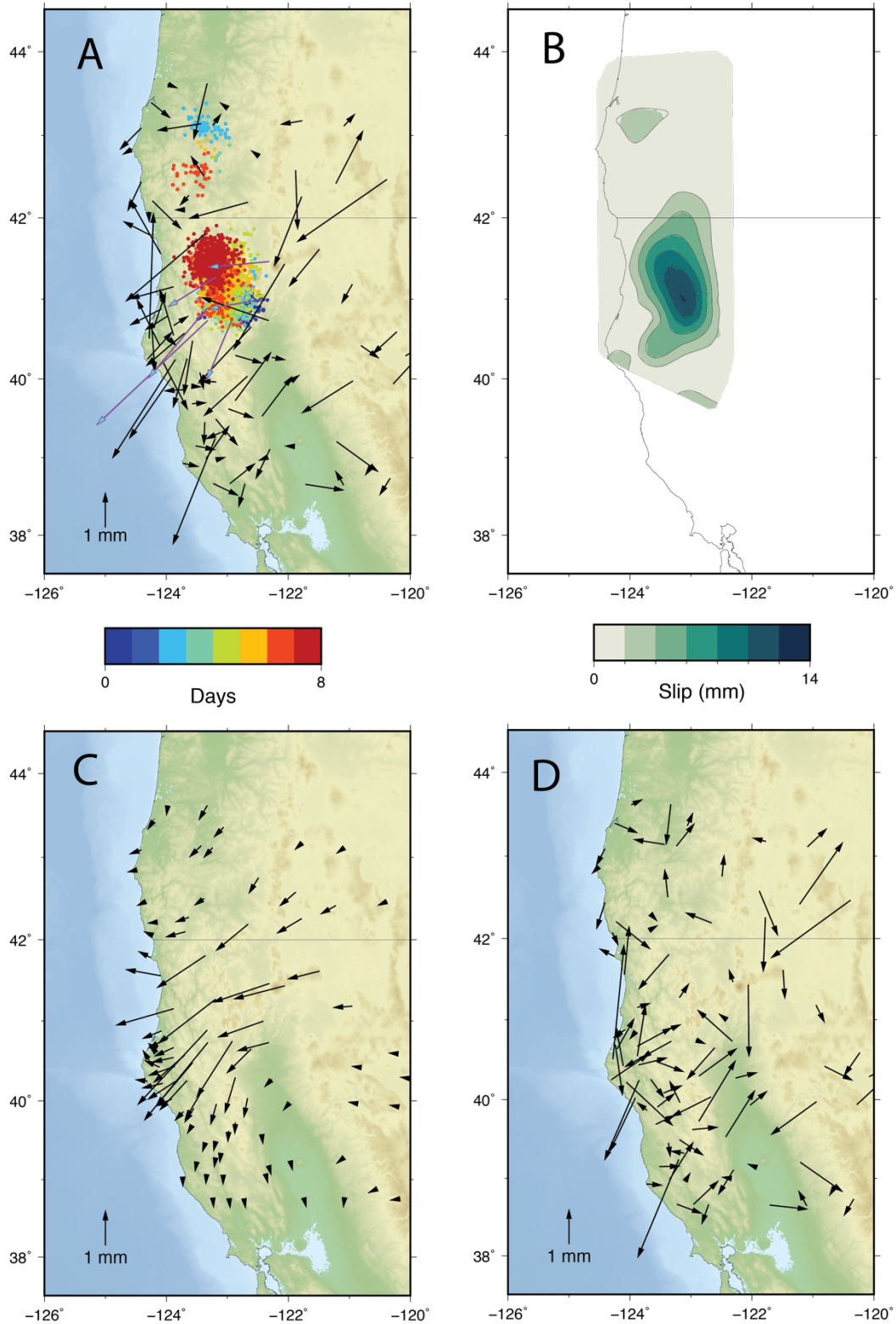


Figure A3.13 Model results for ETS Event 11 (March 2019). Panels are the same as Figure A3.3.

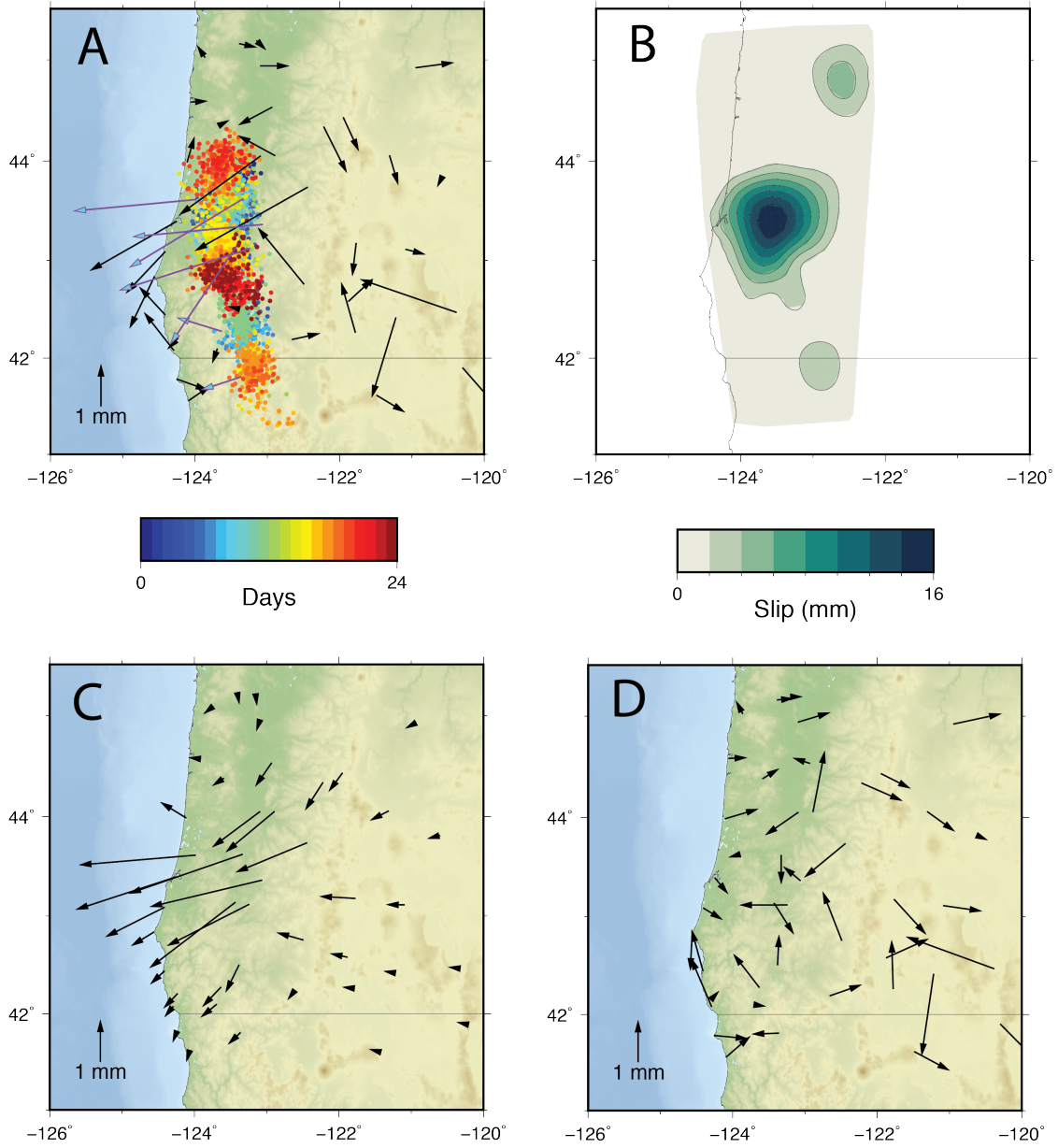


Figure A3.14 Model results for ETS Event 12 (July 2019). Panels are the same as Figure A3.3.

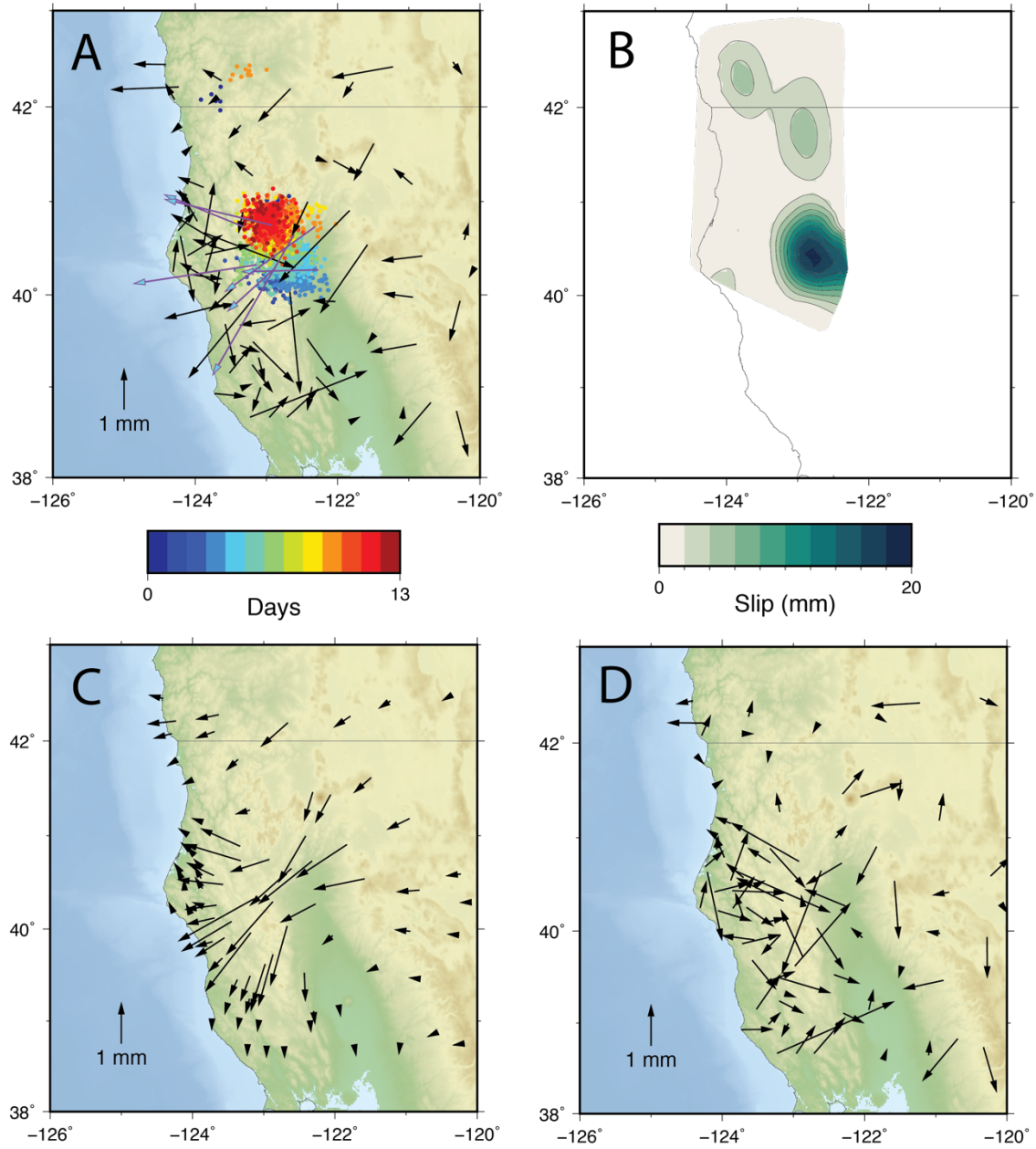


Figure A3.15 Model results for ETS Event 13 (November 2019). Panels are the same as Figure A3.3.

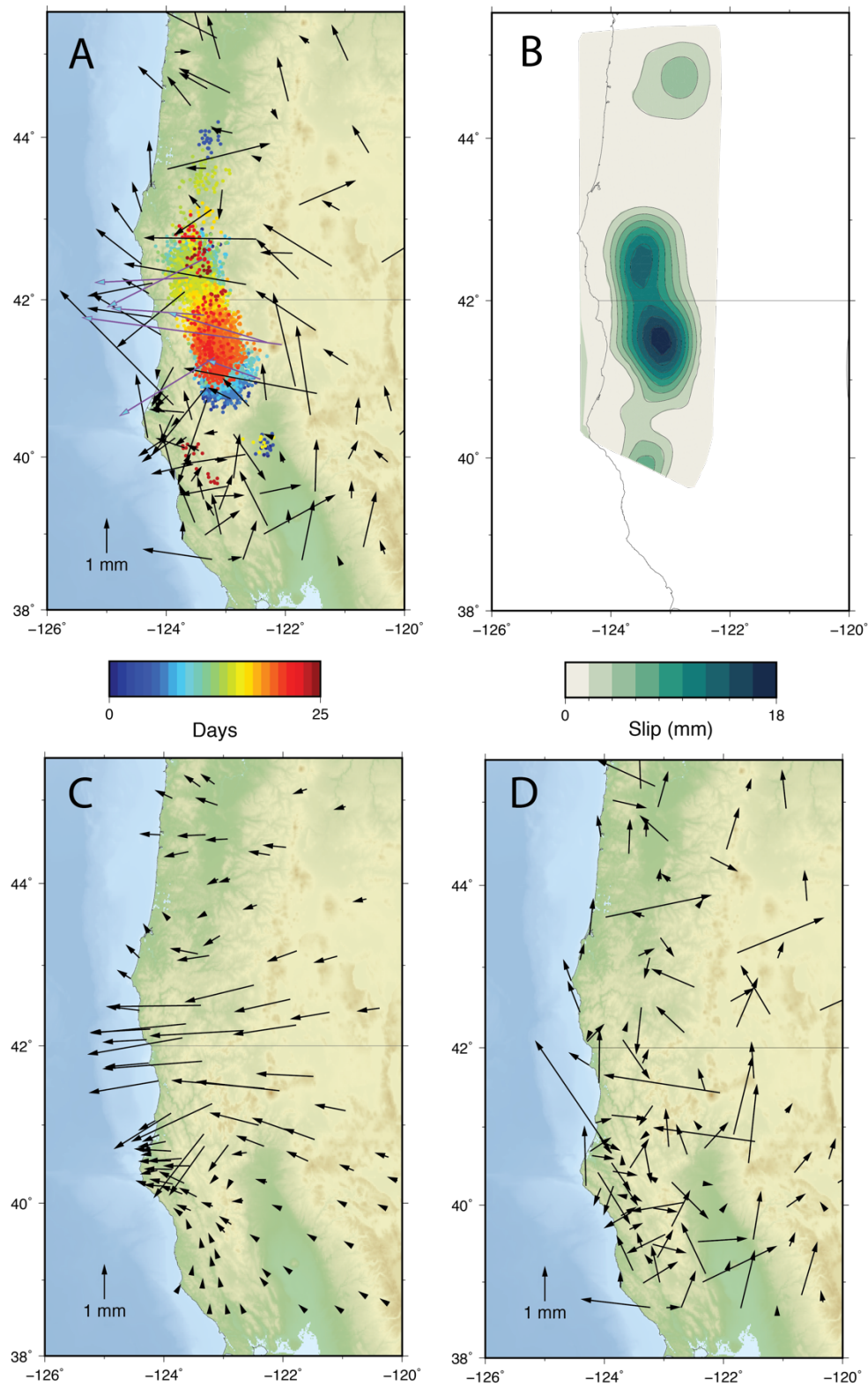


Figure A3.16 Model results for ETS Event 14 (December 2019). Panels are the same as Figure A3.3.

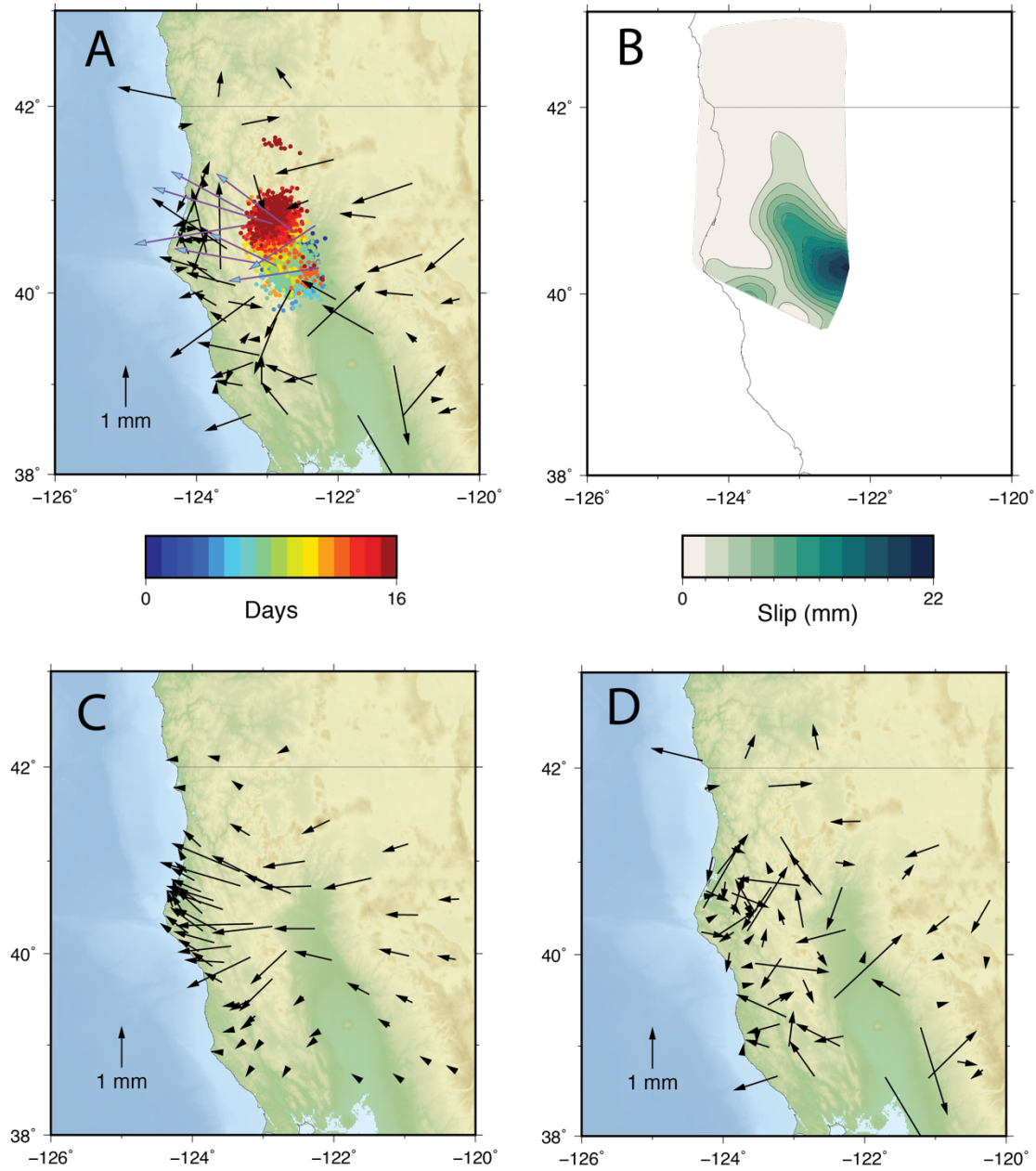


Figure A3.17 Model results for ETS Event 15 (June 2020). Panels are the same as Figure A3.3.

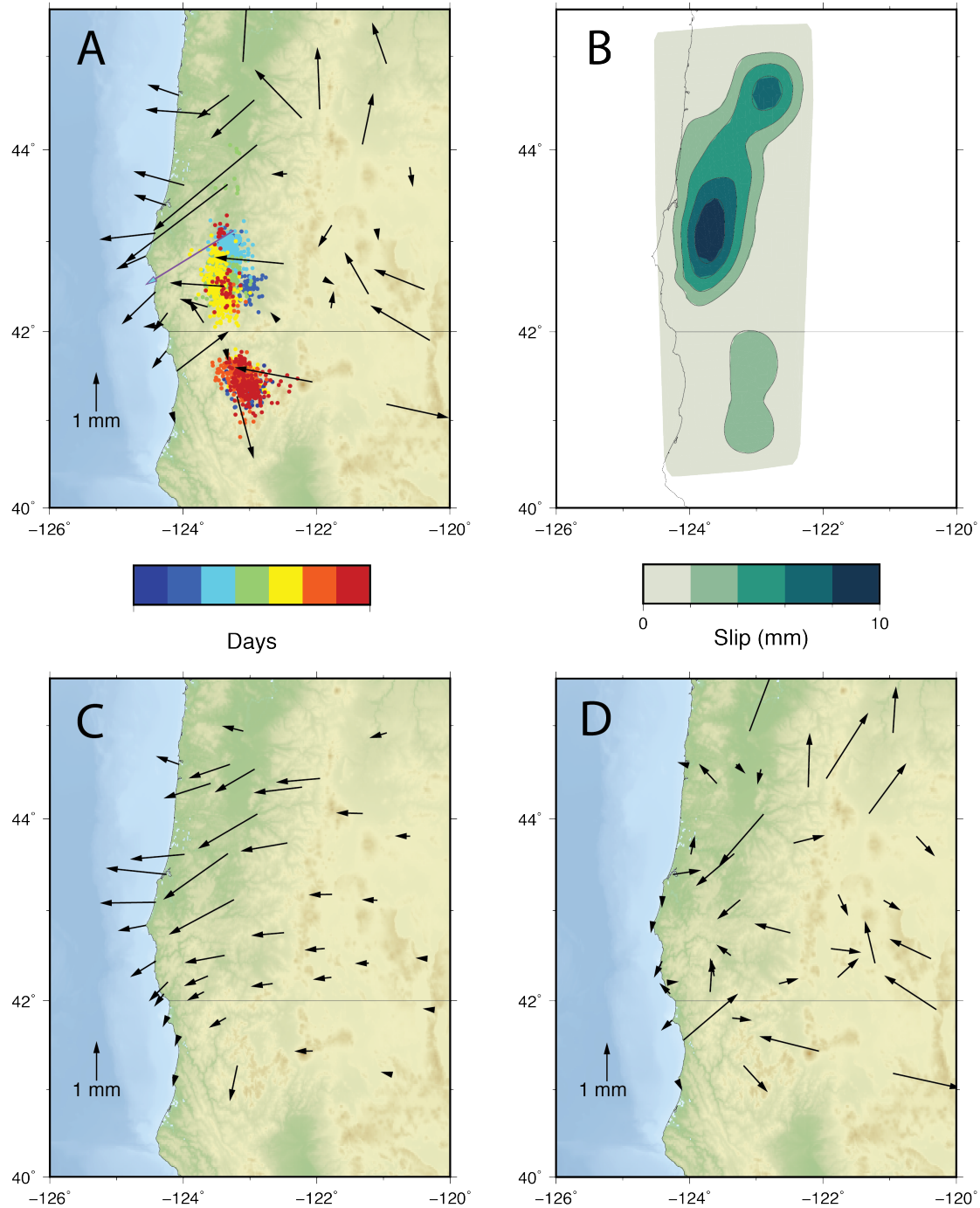


Figure A3.18 Model results for ETS Event 16 (September 2020). Panels are the same as Figure A3.3.

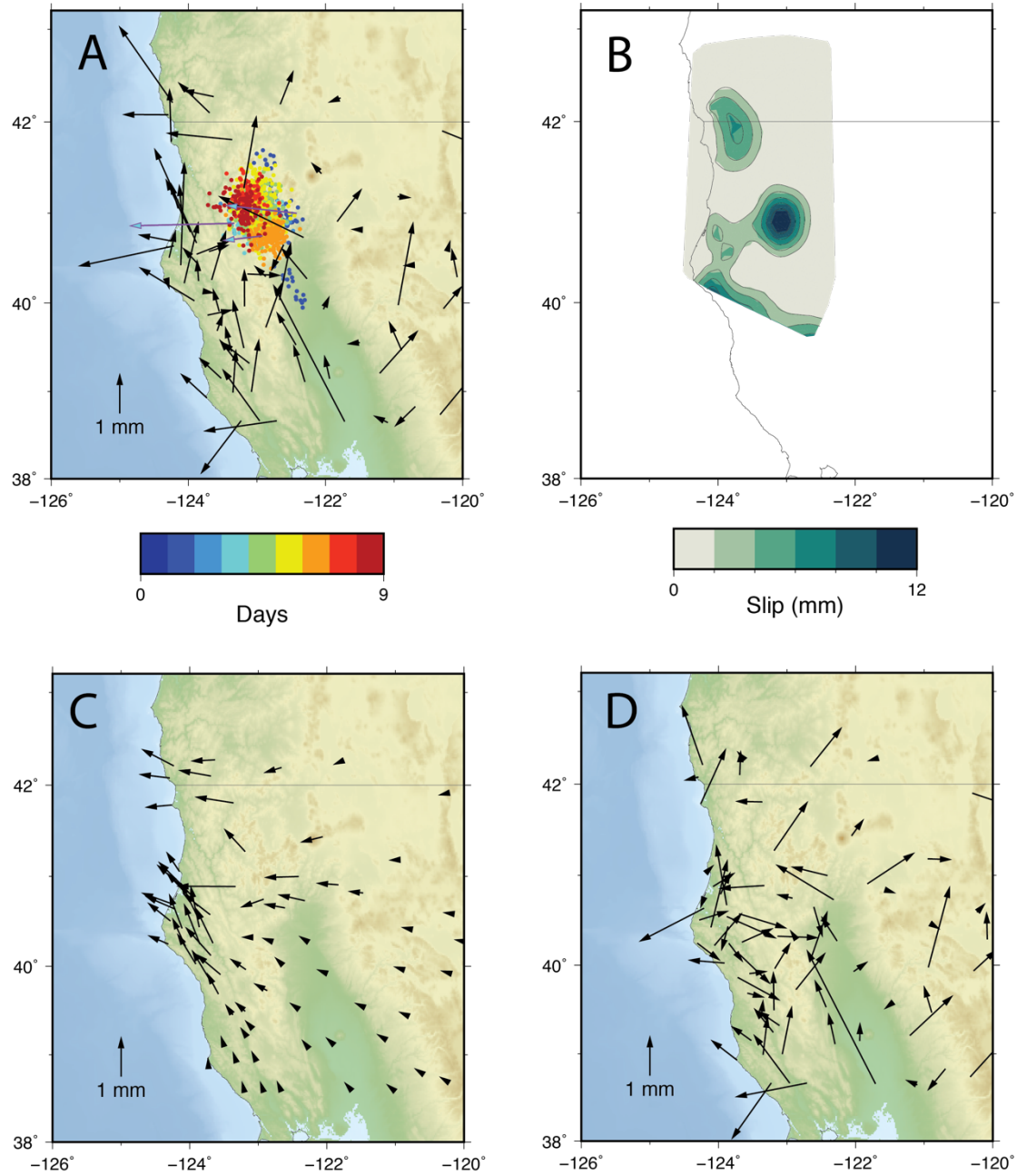


Figure A3.19 Model results for ETS Event 17 (November 2020). Panels are the same as Figure A3.3.

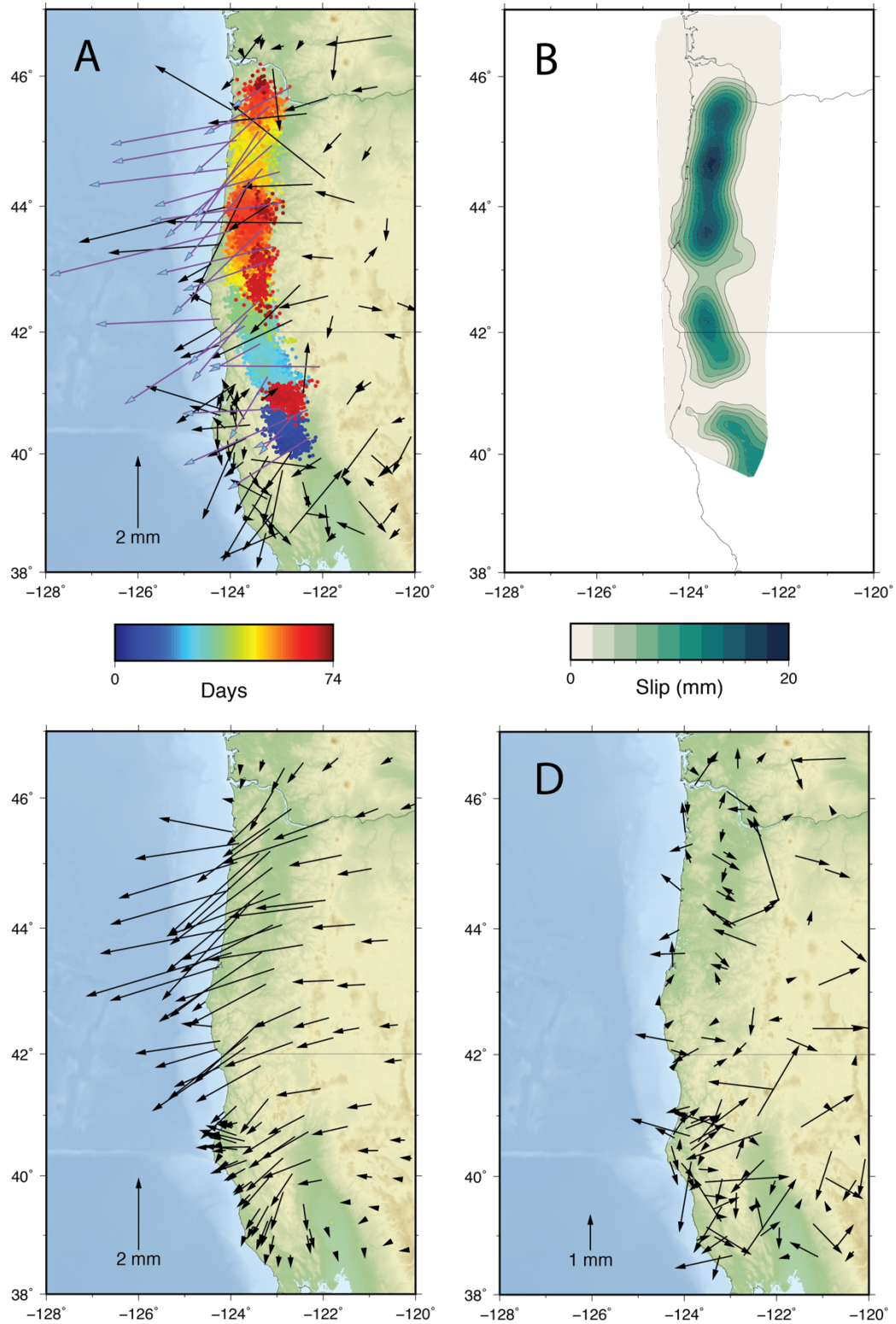


Figure A3.20 Model results for ETS Event 18 (March 2021). Panels are the same as Figure A3.3.

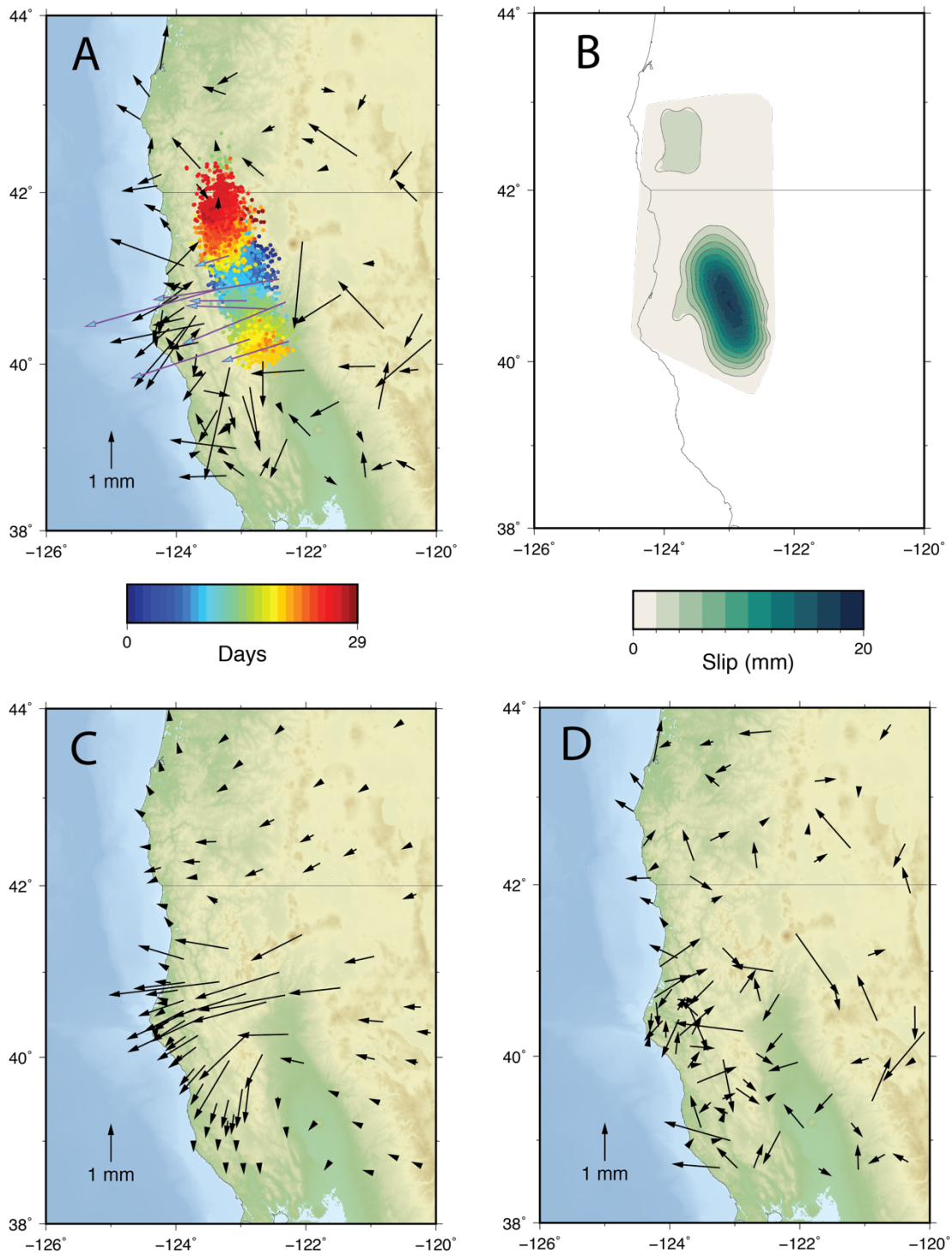


Figure A3.21 Model results for ETS Event 19 (September 2021). Panels are the same as Figure A3.3.

Event #	Moment Magnitude (Mw)	Tremor (hours)	Misfit (χ^2)
1	6.73	225	0.66
2	6.51	74	0.67
3	6.60	99	0.63
4	6.42	218	0.55
5	6.62	72	0.76
6	6.72	320	0.84
7	6.42	205	0.64
8	6.42	204	0.75
9	6.51	132	0.77
10	6.41	134	0.73
11	6.50	149	0.76
12	6.47	201	0.57
13	6.48	143	1.00
14	6.69	275	1.22
15	6.49	203	0.61
16	6.53	83	0.82
17	6.32	112	1.02
18	6.85	776	0.87
19	6.52	411	0.6

Table A3.1 Equivalent moment magnitude (Mw), tremor hours, and model misfit for each of the 19 ETS events in southern Cascadia.

VITA

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