Is there a discrepancy between geological and geodetic slip rates along the San Andreas Fault System?

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Abstract

Previous inversions for slip rate along the San Andreas Fault System (SAFS), based on elastic half-space models, show a discrepancy between the geologic and geodetic slip rates along a few major fault segments. In this study we use an earthquake cycle model representing an elastic plate over viscoelastic half-space to demonstrate that there is no significant discrepancy between long-term geologic and geodetic slip rates. The model includes 41 major fault segments having steady slip from the base of the locked zone to the base of the elastic plate and episodic shallow slip based on known historical ruptures and geologic recurrence intervals. The slip rates are constrained by 1,981 secular velocity measurements from GPS and L-band Intereferometric Synthetic Apeture Radar (InSAR). A model with a thick elastic plate (60 km) and half-space viscosity of 10¹⁹ Pa·s is preferred because it produces the smallest misfit to both the geologic and the geodetic data. We find that the geodetic slip rates from the thick plate model agrees to within the bounds of the geologic slip rates, while the rates from the half-space model disagree on specific important fault segments such as the Mojave and the North Coast segment of the San Andreas fault. The plate models have generally higher slip rates than the half-space model because most of the faults along the SAFS are late in the earthquake cycle, so today they are moving slower than the long-term cycle-averaged velocity as governed by the viscoelastic relaxation process.

1 Introduction

Geodesy has become an increasingly important tool for recovering crustal strain rates in tectonically active regions. In California, the high-accuracy GPS velocity field from continuous and campaign networks, such as the Plate Boundary Observatory (PBO), and Southern California Integrated GPS Network (SCIGN), and field surveys have been used to estimate fault slip rates along the San Andreas Fault System (SAFS) [*McCaffrey*, 2005; *Meade and Hager*, 2005; *Bird*, 2009; *Zeng and Shen*, 2010], especially for those faults where geological estimation is lacking or inaccurate. The long term fault slip rates on these major faults are an important component in earthquake hazard analysis because one can estimate moment accumulation rate when combined with estimates of the seismogenic depth.

The common approach for inverting for fault slip rates is through application of elastic half-space models [e.g. *McCaffrey*, 2005]. An important assumption in the half-space model is that the observed velocity field is steady over time and the transient effects from past earthquakes can be neglected. This assumption is valid if the relaxation times of the lower lithosphere are much larger than half of the recurrence interval of a given fault [e.g. *Meade and Hager*, 2005]. When this model applies to 3-dimensional problems, the interseismic velocity field can be explained by a combination of rigid block rotation with kinematically consistent fault slip rates and fault locking within the interseismic period.

Recent studies [e.g. *Bird*, 2009; *Zeng and Shen*, 2010] suggest an apparent discrepancy between the geologic and geodetic slip rates along the SAFS, although the uncertainties resulting from both of these estimates are quite large. The fault segments that are currently in debate include: the Imperial fault [*Dawson and Weldon*, 2013] the southernmost San Andreas fault and the San Jancinto fault [*Van der Woerd et al*., 2006; *Lundgren et al*., 2009; *Lindsey and Fialko*, 2013], the San Bernardino segment [*Loveless and Meade*, 2011; *Spinler et al*., 2011; *McGill et al*., 2013], the Mojave segment [*Savage and Lisowski*, 1998; *Chuang and Johnson*, 2011], the Eastern California Shear Zone [*Oskin et al*., 2008], the creeping section [*Titus et al*., 2006; *Toke et al*., 2011], and the Peninsular segment of the San Andreas fault [*Geist and Andrews*, 2000; *McCaffrey*, 2005].

The discrepancy can be possibly resolved by introducing viscoelastic relaxation [*Nur and Mavko*, 1974; *Savage and Prescott*, 1978] to the interseismic velocity modeling. It has been observed that following large earthquakes, the steady state motion is perturbed by the viscoelastic response: the surface strain rate will increase immediately following an event and diffuses away slowly over years or decades. The 1906 San Francisco earthquake and the 1857 Fort Tejon earthquake may have had a long lasting postseismic effect depending on the rheological properties of the lithosphere [*Savage and Lisowski*, 1998; *Pollitz et al*., 2004; *Chuang and Johnson*, 2011]. A range of viscosity structures have been estimated from postseismic deformation following recent M>7 earthquakes [*Pollitz et al.*, 2001; *Kenner and Segall*, 2003; *Freed and Bürgmann.*, 2004; *Smith and Sandwell*, 2004].

The goal of this study is to estimate fault slip rates using both the viscoelastic coupling model and the elastic half-space model to answer the following questions: How do the fault slip rates inferred using geodesy compare to the geological estimates? Along which fault segment of the SAFS are the geodetic and geologic slip rates are incompatible? Can these discrepancies be reconciled by the viscoelastic earthquake cycle model?

In order to answer the above questions, we incorporate GPS velocity data, InSAR line-of-sight (LOS) velocity data and geological data to construct a high resolution deformation model of the SAFS, starting from the Cerro Prieto fault to the south to the Maacama fault to the north (Figure 1). We simultaneously solve for the long-term fault slip rates of 41 major faults using a 3-dimensional earthquake cycle model. Then we focus on key faults where discrepancies between the geodetic and geologic slip rates are significant.

2 Data

2.1 GPS velocities

The GPS data used in this study include 1,981 horizontal velocity vectors covering major faults along the SAFS (Figure 2). 1,863 velocity vectors were used from the SCEC UCERF3 GPS velocity solutions [*Herring*, 2013, *personal communication*]. These measurements are a blend of eight different analyses including the Plate Boundary Observatory (PBO) GPS sites and continuous and campaign GPS sites from the SCEC Crust Motion Model (CMM4) [*Shen et al.*, 2011] and the Measures solutions [*Herring*, 2013, *personal communication*].

When modeling the interseismic velocity field, it is important to keep in mind how the GPS solutions are derived and what they represent. The GPS velocities span the years from 1996 to present. Secular velocity terms are estimated, along with any postseismic signals (logarithmic functions) for the events after 1996. The postseismic signals from earthquakes before 1996, like the 1992 Landers earthquake, 1906 San Francisco earthquake, and 1857 Fort Tejon earthquake are not accounted for in the GPS data analysis [*Herring*, 2013, *personal communication*]. In this study, we correct the secular velocity field solution for postseismic relaxation following the 1992 Landers earthquake using a slip model from *Fialko* [2004]. Our model predicts that the postseismic velocity after approximately 20 years of the Landers earthquake is ~1.5 mm/yr at maximum for the sites surrounding the rupture. This corrected secular velocity field is used in the slip rate inversion. The postseismic relaxation from the 1906 and 1857 events and other past earthquakes are treated systematically in our earthquake cycle model (Section 3).

In addition, we added velocities from 8 campaign sites in Central California [*Rolandone et al.*, 2008] that cover the central portion of the creeping section. A new velocity field from 110 campaign sites near Salton Trough in Southern California [*Crowell et al.*, 2013] were also included to provide a dense coverage of the near-fault deformation near the Imperial fault. Both of the campaign GPS results were rotated into the reference frame of the continuous GPS sites to yield a consistent velocity field. Because these GPS solutions are different in terms of observation duration, uncertainties, and processing technique, we quantify their importance by assigning a weighting factor to the different data sets. The weight to the PBO data set from T. Herring is 1. The weight to the campaign data set from *Rolandone et al.* [2008] and *Crowell et al.* [2013] is 0.5

and 0.33, respectively, because of relatively short observation periods of each. We focus on the horizontal GPS velocity data only; no vertical velocities are used in this study.

2.2 InSAR LOS velocities

The InSAR data used in this study were obtained through an L-band radar onboard Adavanced Land Observing Satellite (ALOS) launched by Japanese Space and Exploration Agency (JAXA), which can maintain good temporal coherence in vegetated areas compared to C-band radar. The InSAR data (spanning 4.5 years from 2006.5 to 2011) were acquired along the ascending orbits (351° flight direction in azimuth with 34° look angle). The InSAR LOS velocities (Figure 2) are derived from integration of the radar interferogram stacking and GPS velocities [*Tong et al.*, 2013]. In this previous study, the long wavelength of the velocity field (>40 km) was constrained by GPS and InSAR and was used to retrieve the short wavelength (<40 km) features of the deformation spectrum. A detailed description of the integration method can be found in *Tong et al.* [2013]. The main contribution of the InSAR data is to recover details of the aseismic fault creep on the creeping section and the faults of the northern SAFS.

For aspects of this study, we first made a mask for the InSAR LOS velocity data to isolate non-tectonic effects. We identified 47 anomalous areas that exhibit anthropogenic-related ground motion, most likely caused by groundwater extraction, along the major faults in California. These anomalous areas are evident because they produce large vertical motion either confined by known aquifers or bounded by faults. The data within these anomalous areas were not used. The remaining LOS velocity data were down sampled to 53,792 points based on the second invariant of the strain rates. This subsampled dataset provides full resolution in high velocity gradient area near the faults and lower resolution in areas of low strain rate far away from the faults. The 3-component look vectors and the standard deviations for each LOS velocity data point were subsampled in the same manner. The uncertainties of the LOS velocities are larger than GPS measurements, typically 3-4 mm/yr. Data accuracy would be greatly improved if the InSAR mission had a longer duration and if a second LOS direction along descending orbits were available. The ALOS-2 mission, scheduled for launch at the beginning of 2014, could provide two look directions (ascending and descending) so future data will likely resolve these issues.

2.3 Geological data

The fault slip rates of closely-spaced parallel faults such as the Elsinore, San Jacinto, and San Andreas in Southern California and the San Andreas, Maacama, and Green Valley faults in Northern California are difficult to resolve using geodesy alone. To make a kinematically consistent model, we introduced three types of geological conditions to loosely constrain the fault slip rates. First, we attemptted to constrain the recovered slip rate to be within the upper and lower bounds of the quaternary fault slip rates. The quaternary fault slip rates used in this study are from the Working Group on California Earthquake Probabilities (WGCEP) [*Dawson and Weldon*, 2013]. We assigned each estimate an uncertainty to account for the variability in quaternary fault slip rate derived by different investigators compiled in Appendix B in UCERF3 (Table 1). Second, we introduced a closure criterion at fault branching points such that when two fault strands join into a single strand, the sum of the two strand rates should match the single strand rate. This condition has an analog to the classic triple junction closure criteria at plate intersections, except that all the faults in this case are approximately parallel to each other. Third, we required that the sum of slip rates on parallel strands should approximately match the overall relative slip rate along the plate boundary (e.g., ~45 mm/yr).

In order to make a fair comparion between the recovered geodetic slip rates and the geological estimates, we treated the geological constraint with caution. The best approach we found was to apply a weighting paramter in the inversion to quantify the significance of the geological constraints. The best-fit weight for the geological constraints was found through a grid search (see Section 4 for details). In fact, this approach can be deemed as conservative because the recovered geodetic slip rates are required to match the geologic slip rates in the inversion. Thus the difference between the geologic and geodetic slip rate results is more likely to be caused by real discrepancy instead of non-uniqueness inherent in the inversion.

3 Earthquake cycle model

To calculate surface velocities from locked faults, we used a fully 3-dimensional, time-dependent earthquake cycle model [*Savage and Prescott*, 1978; *Smith and Sandwell*, 2006]. The model comprises an elastic plate overlying a viscoelastic half-space (here we refer to it as the "plate model" in contrast to the "half-space model"). The earthquake cycle effect produces time-dependent deformation by viscoelastic relaxation of the asthenosphere. This model assumes a linear rheology of the viscous layer corresponding to diffusion creep in the laboratory derived flow law.

Figure 3 shows an example of the surface interseismic velocity predicted by our earthquake cycle model [*Smith and Sandwell*, 2006]. The difference between this model and the elastic model are temporal variations of the present-day surface velocity. When the observation time is earlier than the relaxation (or Maxwell) time, which is defined as twice the effective viscosity divided by the shear modulus, the velocity is generally higher than the cycle average (gray line), while for later times (when the observation time is significantly later than the Maxwell time), the velocity is generally lower than the cycle average. This comparision serves as a validation of our 3-dimensional forward model against the 2-dimensional (2D) analytic solutions from *Savage and Prescott* [1978]. What deivates from the original 2D model is that we incorporated realistic curved faults in our 3-dimensional (3D) model and we use appropriate earthquake sequences based on geologic records. Compared to the 2D model, the 3D model predicts a reduced viscoelastic effect. Viscoelastic relaxation is proportional to the length of the fault segment thus only significant earthquakes produce long lasting transient deformation. This model is different from the traditional block models that use the "back-slip" approach [*McCaffrey*, 2005; *Meade and Hager*, 2005; *Chuang and Johnson*, 2011] in that this model describes faults as buried dislocations along block boundaries, i.e. "forward slip", to account for the interseismic locking effect [*Smith and Sandwell*, 2006].

In this model, the right lateral shear between the North American and Pacific plates is taken up by several major strike-slip faults (Figure 1). In the long term, the crust is displaced at the fault boundaries, behaving like rigid blocks. We restrict our analysis to the simple fault geometry adopted from *Smith-Konter and Sandwell* [2009]. The modeled faults include the entire trace of the San Andreas faults from Point Arena to Bombay Beach: the San Jacinto fault, Elsinore fault, Imperical fault, and Cerro Prieto fault in Southern California; the Hayward fault, Calaveras fault, Rodgers Creek fault, Maacama fault, Hunting Creek - Bartlett Springs fault, and Concord fault in Central and Northern California. In East California Shear Zone, we consider the Lenwood - Lockhart - Old Woman Springs fault, Helendale fault, and Calico-Hidalgo fault. We also include the Owens Valley fault and Death Valley fault with an aim to balance the slip budget across the plate boundary.

In summary, the fault model consists of 41 fault segments, each having uniform slip rate, locking depth, and earthquake history. Each segment is further sub-divided into smaller patches (~ 5 km length) following the curvature of the fault trace at surface. Each fault segment slips at a steady velocity from its locking depth to the base of the elastic plate. The coseismic rupture is assumed to extend from the surface to the locking depth prescribed for each fault segment. The locking depth of each fault is estimated by the seismogenic depth and GPS observations [*Smith-Konter et al.*, 2011].

Our experiment explores four different rheological models: an elastic half-space model and three elastic plate models (Table 2). We use two possbile thicknesses (thick versus thin plate) for the elastic plate in an attempt to understand the behavior of the viscoelastic relaxation in relation to the lithosphere's rheology. The half-space model has localized slip from the locking depth to infinite depth. In contrast, the plate models have localized slip from the locking depth extending to the base of the elastic layer. This localized slip surface approximates a deep-rooted shear zone beneath the fault zone in the upper crust. An important difference between the plate model and half-space model is that the elastic strain in the interior of the plate is much greater than the elastic strain in the interior of the half-space blocks.

We note that the earthquake reccurence along many fault segments of the SAF is irregualar based on the paleoseismological record. Oversimplification of the earthquake sequence using a characteristic earthquake model may not be appropriate. We used realistic earthquake sequences based on a recent compilation of all the historical and prehistorical earthquakes dated from the year 1000 to present [*Smith and Sandwell*, 2006 and references therein; *Solis*, 2013] to "spin up" the earthquake cycle. When the information on the past earthquake sequences are lacking, we prescribed periodic earthquake cycles according to the estimated recurrence interval [*Dawson and Weldon,* 2013]. Because the magnitude of the slip along each segment for each event is usually not known, we assume that the shallow slip events "catch up" with the deep slip over an earthquake cycle to satisfy block motion on the fault. This earthquake recurrence concept is directly derived from the slip-predictable model. In Section 6.1 we investigate the effect of this slip-predictable assumption in detail.

The deep-dislocation based earthquake cycle model cannot accurately resolve the surface velocity due to aseismic creep in the upper crust. It is generally thought that the fault creep is confined within the shallowest sedimentary layer of the crust (1 or 2 km depth). However, it has been found that fault creep can occur within the brittle upper crust along several major faults in Central and Northern California [*Rolandone et al.*, 2007]. We augmented this model using shallow dislocations in an elastic half-space [*Wang et al.*, 2003]. The creeping faults modeled in this study include the Hayward, Calaveras, Maacama, Concord, Bartlett Springs, Rodgers Creek fault, Parkfield, the creeping segment, and Santa Cruz Mountain segment of the San Andreas fault of the Northern SAFS and the Imperial, Superstition Hills, and the Brawley Seismic Zone of the Southern SAFS [*Tong et al*., 2012]. These fault segments are discretized into small rectangular dislocation patches extending from the surface to 12 km deep in the upper crust. We jointly solved for the aseismic creep rates of these fault segments along with 41 long term fault slip rates in the inversion as described in the next section. The details of the aseismic creeping faults in central and northern SAFS are out of the scope of this paper. These results are summarized in a companion paper [*Tong et al*., manuscript in preparation].

4 Inversion method

In this section, we describe the system of linear equations used to estimate slip rates on 41 fault segments $\overline{s}$ and 66 creep rates $\overline{p}$ from a combination of 1,981 GPS vector velocity measurements $\overline{v_g}$, 53,792 line-of-sight (LOS) InSAR measurements $\overline{l}$ and geologic constraints. This linear system consists of four subsystems of equations representing the GPS data, InSAR data, geological constraints and smoothing constraints, respectively:

$$\begin{bmatrix} \overline{\overline{G_g}} & \overline{\overline{E_g}} & \overline{\overline{I}} & \overline{\overline{r}} \\ \overline{\overline{G_i}} & \overline{\overline{E_i}} & \overline{\overline{I}} & \overline{\overline{r}} \\ \overline{\overline{C}} & 0 & 0 & 0 \\ 0 & \overline{\overline{S}} & 0 & 0 \end{bmatrix} \begin{bmatrix} \overline{s} \\ \overline{p} \\ \overline{v_0} \\ \overline{w} \end{bmatrix} = \begin{bmatrix} \overline{v_g} \\ \overline{l} \\ \overline{s_c} \\ 0 \end{bmatrix} \qquad (1)$$

where $\overline{\overline{G}}$ and $\overline{\overline{E}}$ are the Green's function for modeled surface velocity. The subscripts *g* and *i* refer to GPS and InSAR data, respectively. $\overline{\overline{G}}$ is derived from the earthquake cycle model and it depends on the elastic plate thickness, effective viscosity, locking depth of the fault, and the earthquake sequence of the segment. $\overline{\overline{E}}$ is derived from the dislocation model depending on the elastic property of the material. $\overline{\overline{C}}$ is the constraint matrix, which includes the geologic slip rate estimates, the triple junction closure constraint, and the far-field velocity constraint. $\overline{\overline{S}}$ is the smoothing matrix applied only to the shallow dislocations representing the aseismic creep. In order to separate the effect of the plate rotation from the interseismic signal, we introduce $\overline{v_0}$ and $\overline{w}$ representing the translation term and the rotation term of the velocity field in a cartesian coordinate. $\overline{v_0}$ has two unknowns denoting two translation terms in the east and north velocities. $\overline{w}$ describes the rotation rate (one unknown) around a prescribed rotation axis that is orthogonal to the east and north velocity direction. $\overline{\overline{r}}$ represents the location of the velocity measurements with respect to the rotation axis. We do not intend to solve for the location of the rotation axis and the rotation rate simultaneously because of the strong trade-offs between these two quantities. $\overline{\overline{I}}$ is the identify matrix. After running the inversion we found that the rotation term w can absorb the residuals of the observed velocity field although it is relatively small in this particular tectonic setting.

In the second subsystem that incorporates InSAR data, variable look vectors in the east and north component are used to project the horiztonal velocity into radar line-of-sight direction even though they are not shown explilcitly in equation (1). The third subsystem $\overline{\overline{C}}\overline{s} = \overline{s_c}$ represents three types of geological constraints represented by the following three matrix: $\overline{\overline{I}}$, $\overline{\overline{C_{tot}}}$ and $\overline{\overline{C_{tri}}}$, respectively. 1) Matrix $\overline{\overline{I}}$ denotes the estimates of slip rate from the geologic data on 41 segments $\overline{s_{geol}}$. 2) Matrix $\overline{\overline{C_{tot}}}$ represents the constraint that the sum of slip rate on sub-parallel fault strands must equal the total slip rate across the plate boundary ($\overline{s_{tot}}$ = 45 mm/yr). 3) Matrix $\overline{\overline{C_{tri}}}$ represents the constraint that at the fault junctions where two or more sub-parallel faults connect and converge into one main fault, the slip rate on the main fault must equal to the sum of the sub-parallel faults ($\overline{s_{tri}}$ = 0). We can represent this as:

$$\begin{bmatrix} \overline{\overline{I}} \\ \overline{\overline{C_{tot}}} \\ \overline{\overline{C_{tri}}} \end{bmatrix} \overline{s} = \begin{bmatrix} \overline{s_{geol}} \\ \overline{s_{tot}} \\ \overline{s_{tri}} \end{bmatrix} \qquad (2)$$

Equation (1) was normalized by the uncertainty in each component of the geodetic measurement. In addition, we introduced three weighting constants to the four sub-systems of equations to have a sense of control on the slip rates solutions. The relative weights were determined by a grid search method to minimize the RMS misfit of the GPS and InSAR data (Figure 4). The best-fit weighting paramter is set to ensure that the inversion gives the smallest misfit to the geodetic data.

We added Gaussian random noise to the input data and repeated the inversions 10 times. Then we computed the mean and the standard deviations as the final fault slip rate results. The amplitude of the random noise was chosen according to the uncertainties of the geodetic measurements. Like other studies [*McCaffrey*, 2005] we modified the formal uncertainties of the GPS data to have more realistic slip rate uncertainty estimates from the inversion. The minimum uncertainties of the GPS velocity mesurements are set to be 1 mm/yr.

5 Results

5.1 The quality of fit

Table 2 shows the statistics of the misfits for four different models: an elastic half-space (HS) model, a viscoelastic model with a relatively thick elastic plate (60 km) and moderate viscosity of 10^19 Pa • s (PL6019), a viscoelastic model with a thin elastic plate (30 km) and moderate viscosity of 10^19 Pa • s (PL3019), and a viscoelastic model with thin elastic plate (30 km) and relatively high viscosity 10^20 Pa • s (PL3020). The χ^2 misfit is defined as the squared sum of the residuals normalized by the standard deviation for each velocity measurement $\chi^2 = \frac{1}{N}\sum_{i=1}^{N}(\frac{o_i - m_i}{\sigma_i})^2$ where o_i is the data, m_i is the model, and σ_i is the uncertainties for *N* measurements. The χ^2 misfit to the GPS data is 2.67 for HS, 2.56 for PL6019, 2.74 for PL3019, and 2.68 for PL3020 model. The formal uncertainties of the InSAR data are relatively large ,thus the χ^2 misfit to InSAR are approximately 0.27.

Figure 5a shows the 1,981 GPS velocity vectors and the predicted velocity from model PL6019. Our model is able to reproduce the right-lateral shear motion across the Pacific-North American plate boundary from the Cerro Prieto fault to the south to the Maacama fault to the north. The model captures the pronounced westward rotation of the velocity field along the Big Bend and Mojave segment of the SAFS in large scale. Figure 5b shows the residual GPS velocity field for the same model PL6019. For illustrative purposes, only significant residuals that are greater than two times of the standard deviations are shown. There are residuals along the southern tip of the creeping section and the Mojave desert, which could be due to complicated postseismic signals from recent earthquakes. The residuals near the Channel Islands to the west of the California coast are probably caused by off-shore faults not included in our model. In general, we found that the secular velocity field observed by GPS is explained well by this 3-dimensional earthquake cycle model.

Figure 6 shows 53,792 InSAR LOS velocity point measurements, the prediction and its residuals from model PL6019 (Table 2). The InSAR observations added in the inversion provides improved resolution of the near-fault (<10 km from the fault trace) deformation. Our model can reproduce both the broad-scale deformation and the sharp velocity gradients at the creeping faults in California.

In addition we calculated the weighted RMS defined as $WRMS = \sqrt{\frac{\sum_{i=1}^{N}(\frac{o_i - m_i}{\sigma_i})^2}{\sum_{i=1}^{N}\frac{1}{\sigma_i^2}}}$. The weighted RMS residual to the GPS data are found to be 1.71 mm/yr for HS, 1.68 mm/yr for PL6019, 1.73 mm/yr for PL3019, 1.72 mm/yr for PL3020 model. The weighted RMS residuals to the InSAR data are less sensitive to different models: 1.34 mm/yr for HS, 1.30 mm/yr for PL6019, 1.31 mm/yr for PL3019, and 1.34 mm/yr for PL3020. We found that the PL6019 model produces the smallest misfit to both of the GPS and InSAR data. The PL6019 model is marginally superior in matching observations to the PL3019 and PL3020 models, which indicates that the elastic thickness of the lithosphere underneath the SAFS is relatively thick. Different viscosities have minor influence on the model residuals.

In summary, the four models that we tested all yield satisfactory fit to the geodetic observations. We can see that the above statistics of the quality of fit is not adequate to differentiate the plate models from the half-space model. In Section 5.2, we investigate the half-space and plate models using profiles of the GPS velocity measurements. In Section 5.3 we compare the geodetic slip rates from the four models to the geologic slip rates to identify statistically significant mismatches.

5.2 GPS velocity profiles

Figures 7a and 7b show 16 fault-perpendicular profiles of the GPS velocities at different locations along the SAFS. These velocity profiles are plotted against the velocities predicted by the half-space model (HS), a thick plate model (PL6019) and a thin plate model (PL3019). We decomposed GPS velocities into two components, parallel and perpendicular to the plate motion, using an Euler pole (-74.4° W, 50.1° N). This pole is determined based on the pole of rotation analysis from *Wdowinski et al.* [2007]. The paraellel components, (shown as triangles) are compared to the modeled velocities (colored solid lines). We selected those GPS sites that lie within 10 km of the northern and southern sides of the profile. We have also tried to include more GPS sites by enlarging the width of the profiles, but this increases the scatterness of the GPS velocites. Since the 3D model has along-strike variation and the differences among the three models are subtle, we decided to limit the width of our profiles to be 20 km. We computed a weighted RMS mifit for each profile for each model with an aim of differentiating the plate models from the half-space model and to explore spatial variations of the plate thickness of the SAFS.

For profile (a) that crosses the Brawley seismic zone to the south of the Salton Sea in Southern California, the thin plate model yields the best fit to the GPS data at a RMS of 2.85 mm/yr. Our fault geometry is based on the seismicity location in the crust even though there is no evidence of surface breaks of an active fault. The model fit to the velocities is good to the west of the Brawley seismic zone but gets worse to the east. The Coachella segment of the San Andreas fault and the San Jancinto fault (profile b) is fit within an RMS of ~ 1.4 mm/yr for all the three models. Profiles (c) and (d) suggest that over the San Bernardino Mountain region, the geodetic data favor the thick plate model. The residuals 100 km to the east of the SAF are likely due to unmodeled postseismic signals from either the Landers or the Hector Mine earthquakes. The GPS velocites are matched well by the half-space model (RMS=1.56 mm/yr) and the thick plate model (RMS=1.61 mm/yr) at the Mojave segment (profile e), while the thin plate model gives a misfit of 1.77 mm/yr, slightly worse than the other models. At the northern tip of the Mojave segment (profile f), the difference in the plate thickness becomes more evident; the RMS misfit of the thin plate model is 0.2 mm/yr greater than the thick plate model. It is worth noting that approximately 50 km on either side from the SAF the thin plate model predicts slower velocities than what is observed by GPS. GPS observations provide evidence for the exsitance of a relatively thick (>60 km) plate underneath the Mojave segment. Profile (g) crossing the central section of the Cholame-Carrizo segment of the SAF reflects a strong asymmetry of the GPS velocities. We tested an alternative fault geometry and discuss the results of this in Section 6.3. For profile (h), there is no longer an asymmetry in GPS velocities.

For profile (i) (Figure 7b) that transects the locked portion of the Parkfield segment, we infer that the thin plate model fits the GPS observations best with an RMS of 1.82 mm/yr. It suggests the existence of anomlous lithospheric structure underneath Parkfield. From profile (j) it seems that the half-space model more appropriately represents the creeping section and no earthquake cycle model is needed at the central portion of the creeping section to explain the present-day GPS velocities. For the profile (k) that crosses the Santa Cruz mountain, there are two closely spaced paralleling creeping faults, the San Anreas fault and the Calaveras fault. They are well resolved by our model because of the constraints provided by InSAR. The thick plate model provides the best fit (RMS = 1.34 mm/yr). The GPS data are fit almost perfectly at profile (l) crossing the southern portion of the Pennisular segment of the SAF. The two steps in the velocity are due to surface creep of the Hayward and Calaveras fault. At the north portion of the Pennisular segment (profile m), the thick plate model produces the best fit (RMS = 1.51 mm/yr), compared to 1.68 mm/yr RMS from the half-space model and 1.61 mm/yr RMS from the thin plate. At profile (n) that crosses the North Coast segment and the Rodgers Creek fault, the half-space model predicts significant larger misfit than the plate models (RMS = 1.77mm/yr versus 1.6 mm/yr). Surface creep is recovered along the Rodgers Creek fault and the Hunting Creek fault.

Among all the profiles, the last two profiles (o and p) crossing the North Coast and the Maacama fault are the most intriguing ones. Due to lack of constraints, the three models we tested predict drastically different secular velocities. Our models deduce significant aseismic creep on both of the Maacama and the Bartlett Springs fault. There is little shear motion within the crustal block between the Maacama and Bartlett Springs fault as constrained by the four GPS sites (between distance 425 km to 475 km) in profile (o). To the north however, profile (p) reveals uniform shear within the same block. From profile (p), the RMS misfit is smallest, favoring the thick plate model (2.06 mm/yr). These models should be re-evaluated when more accurate geodetic data is available.

In summary, an analysis of the 16 GPS velocity profiles and the earthquake cycle models across the entire SAFS suggests that the thick plate (60 km) is a more appropriate representation in California, with exceptions at three important locations: the Brawley seismic zone, the Parkfield segment and the creeping section. The modeling favors a relatively thin (30 km) elastic plate near the Brawley seismic zone and the Parkfield segment. The half-space model is preferred over the earthquake cycle models over the center portion of the creeping section of the SAF. This section is known to be devoid of large historical earthquakes and the plate motion is mainly accomodated by aseismic creep.

5.3 Long-term slip rate

Here we compare the geodetic slip rates estimated from the inversion with geologic slip rates. We incoporated the best estimate rates, as well as the upper and lower bounds, derived from a recent compilation of the UCERF3 geologic slip rates [*Dawson and Weldon*, 2013]. The geodetic slip rates inferred from the half-space model and four plate models are shown separately (Figure 8). First, we evaluated a general misfit between the geodetic slip rates and the best estimate geologic rates. These are: HS (3.5 mm/yr), PL6019 (3.4 mm/yr), PL3019 (5.3 mm/yr) and PL3020 (4.0 mm/yr). From these results we infer that agreement between geology and geodesy is better for the thick plate model and the HS model.

Next, we focus our attention on fault segments where there are significant discrepancies between the the geodetic slip rates for the half-space model and the geologic rates. From Figure 8a and Table 1, we show that the HS model provides a reasonally good job of matching the geologic rates. However, there are two interesting anomalies: (1) the North Coast segment of the SAF is slipping at 14 mm/yr, much slower than the preferred geologic rates at 24 mm/yr, and (2) the Mojave segment of the SAF is slipping at 25 mm/yr while the geologic best estimate is 34 mm/yr with rather large uncertainties (25-40 mm/yr).

We next examined the slip rates inferred from the plate models (also incorporating a variable viscosity) to see if these differences could be explained. From Figure 8b and 8c, we conclude that the earthquake cycle model could, in general, resolve the discrepancy between the geodetic and geologic slip rates for both the Mojave segment and the North Coast segment of the SAF. The thick plate model (PL6019) yielded a slip rate of 23 mm/yr on the North Coast segment, and a slip rate of 27.8 mm/yr of the Mojave segment. The thin plate model (PL3019) resulted in a higher slip rate 36.5 mm/yr along the North Coast segment and 33.1 mm/yr along the Mojave segment. It can be seen that the plate thickness plays a key role in the recovered geodetic slip rate. For the North Coast segment, the thick plate model is the best in terms of matching the geologic slip rates, but for the Mojave segment, the thin plate model is the best. The effects of the viscosity on the recovered slip rates can be interpreted by comparing Figure 8c and 8d. For the Mojave segment, for example, increasing the mantle viscosity from 10^19 Pa • s to 10^20 Pa • s results in a dramatic decrease in the geodetic slip rates from 33.1 to 22.9 mm/yr. Similarly, for the North Coast segment, the viscosity change resulted in a reduction in the slip rate estimation from 36.5 to 33.2 mm/yr.

Because of time-dependent viscoelastic relaxation effects, the interseismic velocity in the early earthquake cycle is always faster than the cycle average [*Savage and Prescott*, 1978]. Likewise, the interseismic velocity in the late cycle is always slower than the cycle average (Figure 3). The recovered slip rates for the plate models are strongly influenced by the time at which the fault is in its earthquake cycle. The last event that occurred on the North Coast segment of the SAFS was the 1906 San Francisco earthquake and the last event on the Mojave segment was the 1857 Fort Tejon earthquake. Given a recurrence interval of 200 years, both the Mojave section and the North Coast section of the SAF are late in the earthquake cycle, so to fit the observed velocities, the model requires higher fault slip rates. The earthquake cycle effect gets stronger as the elastic plate gets thinner. The response time of the earthquake cycle effects are determined by the half-space viscosity: a high viscosity implies a longer response time than a low viscosity.

We compared our results with recent findings by *Chuang and Johnson* [2011] and *Hearn et al*. [2013]. Both of these detailed studies focused on the discrepancy along the Mojave segment in Southern California. *Chuang and Johnson* [2011] estimated a slip rates of 26 mm/yr along the Mojave segment assuming a three layer model. *Hearn et al*. [2013] deduced the slip rates of the Mojave segment to be 27-29 mm/yr assuming a four layer rheological model. Our results compared to these previous results shed new light on the importance of the rheology in estimating the slip rate parameters. Because the elastic plate thickness depends on the temporal characteristics of the loading, the elastic thickness inferred from the earthquake cycles should be much greater than the ones infered from the isostatic rebound or gravity studies [*Watts*, 2007]. Using a plate model and varying the elastic thickness of the plate, we demonstrated that the earthquake cycle model could agree with the geologic slip rates of 34 mm/yr along the Mojave segment if the elastic plate is relatively thin (30 km) and the half-space visocisty is 10^19 Pa • s . The difficulty of accepting this solution is that the elastic plate underneath the Mojave segment has to be unusually thin compared to the other regions of the SAFS. Another possibility is that the geological slip rates of the Mojave segment is overestimated by about 6 mm/yr.

We note that there are other anomalies between the half-space rates and the geologic rates, which our earthquake cycles models cannot explain. The geodetic slip rates of the Imperial fault and the Cerro Prieto fault are 10 mm/yr faster than the geologic estimates of 35 mm/yr. The Lenwood - Lockhart - Old Woman Springs fault and Calico-Hidalgo fault in the Mojave desert are also significantly faster than the geologic rates. The causes of these inadequacies are probably due to poor knowledge about the fault structures and the chronological sequence of the past events in those regions.

6 Discussions

6.1 Past earthquakes assumption

In this study, the timing of past events are derived from a compilation of historical and prehistorical earthquake records [*Smith and Sandwell*, 2006 and reference therein]. We assume that the coseismic slip of past events completely releases the slip deficit accumulated since the last events. This assumption orginated from the elastic rebound theory which is probably correct over geological time scales. However, it is rather difficult to collect a complete record of all the past earthquakes. There might be significant deviations from a periodic behavior over a time span of several earthquake cycles. For example, a study from *Sieh et al*. [2008] implied earthquake supercycles since past 700 years at the Sumatra subduction zone.

Here we investigated the effect of past events on the geodetic slip rate using the Mojave segment of the SAF. The Mojave segment has experienced the 1812 Wrightwood-Santa Barbara earthquake (M=7.5) and the 1857 Fort Tejon earthquake (M7.9). Prehistorical earthquakes since the year 1000 A.D. are estimated to have occured around years 1016, 1116, 1264, 1360, 1487, 1536, and 1685. Under the slip predictable model assumption, the slip magnitude of the 1812 event is three times greater than the 1857 event. Since the transient velocity from the postseismic relaxation is proportional to the magnitude of the coseismic slip, the postseismic contribution following the 1857 rupture is three times smaller than the 1812 rupture.

We tested the influence of the 1857 rupture on present day velocities by considering three scenarios: a) increasing the coseismic slip of the 1857 event by a factor of four; b) removing the 1857 event from the earthquake sequence; c) adding a synthetic earthquake in 1957 on this segment. The new slip rates of the Mojave segment are shown in Table 3. Scenario a) and b) show that the 1857 event has little effects on the slip rate estimation for a viscosity of 10^19 Pa • s . For a viscosity of 10^20 Pa • s , magnifying the 1857 postseismic signal by a factor of four can decrease the slip rate estimate by ~2 mm/yr;

removing the 1857 event can increase the slip rate estimate by ~1 mm/yr. Scenario c) shows that the slip rate could be underestimated systematically, depending on viscosity, if the timing of the last event is set to be later than its real occurence.

We found that the influence of the slip-predictable hypothesis to the geodetic slip rates is not significant, given a moderate viscosity of the substrate. However, it is important to estimate the timing of the most recent events in order to determine whether the earthquake cycle is in an early or late stage. It should be noted that the slip magnitude of previous events is needed if one wants to evalute the absolute magnitude of stress in the lithosphere. *Hetland and Hager* [2006] considered such a model to investigate the influence of the initial stress on the interseismic strain accumulation.

As pointed out by *Hearn* et al. [2013], the postseismic effect resulting from a single earthquake is different from the long-lived transient effect resulted from multiple earthquake cycles. For a finite length rupture in 3-dimensions, as considered in our model, the postseismic effect is limited to a distance that is approximately the rupture length. However, the cumulative earthquake cycle effect from all past events reaches beyond the rupture length. This is because the viscoelastic relaxation effect is no longer resultant from one particular fault segment but rather the contributions from all the other fault segments in the region. This effect highlights the importance of incorportating realistic past earthquake sequences into 3-dimensional earthquake cycle models.

6.2 Spatial variations in elastic plate thickness inferred from secular GPS velocites

For the plate models, the deformation beneath the elastic layer within the lower crust and upper mantle is distributed by ductile flow. The viscosity of the lower crust/upper mantle is found to be 9.5 x 10^19 Pa • s [*Kenner and Segall*, 2003] using post-1906 earthquake deformation data. Studies following earthquakes occuring in the Mojave desert suggests time-dependent rheology consistent with power-law creep [*Pollitz et al*., 2001; *Freed and Burgman*, 2004]. Because the rheological properties underneath California are not known, we experimented with an effective viscosity of 10^19 Pa • s and 10^20 Pa • s for the plate models. In earlier studies, *Hearn et al*. [2013] and *Chuang and Johnson* [2011] has assumed different elastic layer thickness and a more sophisticated viscosity structure representing lower crust, uppermost mantle and the asthenosphere. *Chuang and Johnson* [2011]'s model consists of 20 km elastic plate, 10 km thick lower crust with a viscosity of 2 x 10^20 Pa • s and a half-space viscosity of 6 x 10^18 Pa • s . *Hearn et al*. [2013]'s model consists of 25 km elastic plate, 5 km thick lower crust with a viscosity of 3 x 10^19 Pa • s , 20 km uppermost mantle with a viscosity of 10^21 Pa • s and a half-space viscosity of 3 x 10^18 Pa • s . It is yet to be determined whether the lower crust or the uppermost mantle has stronger viscous strength in supporting the tectonic stress [*Bürgmann and Dresen*, 2008]. In this study, we do not intend to differentiate the lower crust from the upper-most mantle; there are trade-offs between the elastic thickness and the effective viscosity below [*Watts*, 2007], and in general a thicker elastic plate implies higher viscosity in the lower crust and the upper most mantle.

Along the SAFS where the fault geometry is simple, the effect of the plate thickness on the internal elastic strain within the blocks is readily discernable (Figure 7a, profiles g and h). The plate model produces prominent interseismic strain within internal elastic blocks away from the fault in contrast to the half-space model. The amount of internal elastic strain distributed within the block increases as the elastic plate thickness decreases. This can be seen clearly that the thin plate model predicts significantly slower secular velocities ~50 km away from the SAF than the thick-plate model. In the far-field, at distances greater than ~100 km away from the fault, velocities inferred from the aforementioned three models are essentially indistinguishable. We deduce the effective elastic thickness to be 60 km over the Cholame-Carrizo segment.

Currently there is no concensus on the elastic plate thickness of California. Previous studies using viscoelastic earthquake cycle models generally assume that the rheological structure underneath California is uniform everywhere. *Smith and Sandwell* [2006] deduced an overall elastic thickness of 70 km for the SAFS, in good agreement with 44-100 km found by *Johnson and Segall* [2004]. In this study, we attempted to probe the elastic thickness of the lithosphere using precise secular GPS velocity. By carefully examining 16 GPS velocity profiles we infer that the elastic thickness is generally large (~60km) over most of the SAFS but is relatively small (~30km) near the Brawley Seismic Zone and the Parkfield. The anomalously thin elastic thickness implies low effective viscosity in the lower crust/upper mantle, which can be caused by localized thinning of the lithosphere due to high heat flux or presence of partial melt or fluid.

6.3 Effect of dipping fault geometry

From Figure 7a, the GPS velocity data along the central Carrizo segment appear asymmetric across the SAF . It is possible that our slip rate inversion could be biased by inaccurate representation of fault geometry. It has been proposed that the geometry of the SAF is significantly different from vertical. In the southern SAF near the Coachella valley, the fault is dipping towards the northeast [*Lindsey and Fialko*, 2013] and near the "Big Bend" region the fault is dipping towards the southwest [*Fuis et al.*, 2012], where the overall shape of the fault surface is similar to a "propeller". The dipping geometry can be further tested using the deformation models because a dipping fault will shift the center of the strain concentration, which is observable in geodetic data. The gravity and electromagnetic data suggest that the Carrizo segment maybe dipping to the west at 60 deg. We tested this dipping fault hypothesis using local GPS velocity data and the elastic half-space model. The GPS data within 10 km from profile *g* shown in Figure 7a is used in evaluating the model misfit. As shown in Figure 7a, the deformation model with the SAF dipping to the west remarkably reduces the RMS misfit of the GPS data from 1.7 mm/yr to 0.96 mm/yr. This model comparison suggests that the dipping SAF hypothesis is supported by the geodetic data. An alternative explanation of the asymmetric strain at the Carrizo segment is through laterally varying crustal properties [*Schmalzle et al.*, 2006]. In their model, a weak zone with 10-25 km width to the northeast of the SAF is required to explain the observed GPS velocity.

7 Conclusions

Since long-term slip rates estimated from geology are subject to uncertainties, present-day geodetic measurements have been employed to estimate slip rates. We investigated the geodetic slip rates of the SAFS using both a viscoelastic earthquake cycle model and the elastic half-space model and compared them with geologic slip rates. Incorporating 1,981 GPS velocity vectors, 53,792 InSAR velocity points, and comprehensive geological information into a constrained least-square problem, we examined 41 fault segments along the SAFS to identify anomalous geodetic slip rates. We found that the geodetic slip rates from an elastic half-space model are significantly lower than the geologic estimates along the North Coast segment and the Mojave segment of the SAF by about 10 mm/yr. This apparent discrepancy can be reconciled by introducing time-dependent deformation governed by the viscoelastic earthquake cycle effect.

The influence of the earthquake cycle on geodetic slip rates depends strongly on the rheologic structure of the lower crust and upper mantle. A 60 km thick elastic plate with viscosity of 10^19 Pa • s provides the best fit to the geodetic and geological data. It is observed that the earthquake cycle effect gets stronger as the elastic plate gets thinner. For the Mojave segment the inferred slip rate is 27.8 mm/yr for the thick plate model and is 33.1 mm/yr for the thin plate model. For the North Coast segment the infered slip rate is 23 mm/yr for the thick plate model, and is 36.5 mm/yr for the thin plate model. We identified discrepancies on other faults along the SAFS such as the Imperial fault, the Cerro Prieto fault and the faults in the East California Shear Zone (ECSZ), which cannot be explained by the viscoelastic effect. Finally, we found that the influence of the slip-predictable hypothesis to the recovered geodetic slip rates is not significant. This finding implies that the present-day surface velocity measured by geodesy is not sensitive to the magnitude of historical earthquakes hundreds of years ago.

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Figure captions

Figure 1. A regional topography map of the San Andreas Fault System in California shown in an Oblique Mercator projection. The black line segments represents the fault segments studied in this paper. Each fault segment is labeled by a 3-characters name (Table 1). The whilte stars represent three major earthquakes that are believed to cause significant postseismic relaxation in the lower crust and upper mantle.

Figure 2. 1989 GPS (triangles) velocity vectors and 53,792 InSAR LOS velocity data points (colored grid) along the SAFS in California, shown in an Oblique Mercator projection. The projection pole (-74.4° W, 50.1° N) is from *Wdowinski et al*. [2007]. The InSAR data spans 4.5 years (2006.5-2010) and are derived from 1100 ALOS radar interferograms. The radar flight direction (ascending) and look direction are provided. Positive velocities (red) represent ground motion away from the satellite. The geological fault traces are shown as black lines. The thin black lines with alphabet letters corresponds to the profiles shown in Figure 7A and 7B.

Figure 3. Comparison between our earthquake cycle model (solid lines) and 2D analytic models (dash lines). The cross-sections of the fault parallel velocity at surface were shown. We considered a 60 km thick plate with an effective viscosity of approximately 10^19 Pa s in this comparison. The fault slips at plate rate from the locking depth (15 km) to the bottom of the elastic plate. We prescribed 20 earthquakes with recurrence interval of 100 years. The time is normalized by the Maxwell time (20 years). The colored lines represent different time periods during the earthquake cycle. It is clear that the numerical model accurately reproduces the analytical solution at different times within an earthquake cycle.

Figure 4. Grid-search to determine the relative weighting factors used in the inversion. a) weight for the C-matrix; b) weight for GPS; c) weight for InSAR. The triangles are the weighted RMS misfit to the GPS data and the circles are the weighted RMS misfit to the InSAR data. The actual weights used in the slip rate inversion is 0.3 for GPS, 0.2 for InSAR and 0.1 for the geological constrain.

Figure 5. The fit to the GPS data. In the top figure a) the black arrows show the observed GPS horizontal velocity vectors with 95% confidence interval. The blue arrows show the predicted horizontal velocity vectors from the preferred plate model PL6019. The bottom figure b) shows the GPS residual (observation - model) velocity vectors. Note the different scales used in the two figures. The thin gray curves denote the fault segments.

Figure 6. The fit to the InSAR LOS velocity data. Positive velocities (red) represent ground motion away from the satellite. The radar look direction and flight direction is marked in Figure 2. a) Observed InSAR LOS velocity. b) Predicted InSAR LOS velocity from the prefered plate model PL6019. c) Residuals (observation - model) of the InSAR LOS velocity. The thin black curves denote the fault segments.

Figure 7A. Cross-sections showing the GPS velocity data and its fit to the deformation models. The GPS velocity vectors are decomposed into two components (parallel and perpendicular to the plate motion) using an Eular pole (-74.4° W, 50.1° N). The paraellel components are shown as triangles, compared to the modeled velocity (solid line). The profiles are all running in N45°E. The profile is labeled by alphabet letter in the upperright of each subfigure (Figure 2). The geographic region of the profile is named at the top of each subfigure. Three models are considerred here: HS model (black line), PL6019 model (blue line), and PL3019 model (red line). The RMS misfit of each profile to the models are show in the followin order HS/PL6019/PL3019. The locations of the major faults are marked by black diamonds at the bottom of each subfigure with explanations below: BSZ-Brawley seismic zone, SAF-San Andreas fault, SJF-San Jacinto fault, HAY-Hayward fault, MAA-Maacama fault, ROD-Rodgers Creek fault. The left side of the subfigure is the west side of the profiles. The error bar of the GPS data show one standard deviation. Inside profile g, we tested a dipping fault model shown as dashed black lines. See text for details.

Figure 7B. Continuation of Figure 7A.

Figure 8. a) Geodetic slip rates in comparion to the geological slip rates. Geodetic slip rates are from HS model. The 3-character labels for fast slipping fault segments (slip rates > 10 mm/yr) are shown (Figure 1, Table 1). The horizontal error bars represent the upper and lower bounds of the geological estimates, the vertical error bars represent the uncertainties estimated in the slip rate inversion. The overall RMS misfits to the GPS, InSAR, and preferred geological slip rates are shown in the upperleft. b) Geodetic slip rates are from PL6019 model. c) Geodetic slip rates are from PL3019 model. d) Geodetic slip rates are from PL3020 model.

Table 1. Summary of the fault segments, locking depths, past earthquakes, geologic and geodetic fault slip rates of the SAFS.

Fault label	Fault name	Recurrence interval (years)	Date of the historical earthquakes	Locking depth (km)	Preferred geologic slip rate (mm/yr)	Bounds on the geological slip rates (mm/yr)	Geodetic slip rate from HS model (mm/yr)	Uncertainties of the geodetic slip rate from HS model (mm/yr)	Geodetic slip rate from PL6019 model (mm/yr)	Uncertainties of the geodetic slip rate from PL6019 model (mm/yr)	Geodetic slip rate from PL3019 model (mm/yr)	Uncertainties of the geodetic slip rate from PL3019 model	Geodetic slip rates from PL3020 model (mm/yr)	uncertainties of the geodetic slip rates from PL3020 model (mm/yr)

												(mm/yr)		
CER	Cerro_Prieto	200	1825 1915 1927 1934 1966	10.0	35	30 ~ 40	42.1	0.8	41.7	0.8	37.1	0.7	39.5	0.8
IMP	Imperial	200	1525 1575 1650 1700 1875 1915 1940 1979	5.9	35	15 ~ 40	44.1	1.0	44.0	1.0	43.0	1.0	43.9	1.0
BSZ	Brawley	200	1700 1875 1906 1979	12.0	25	15 ~ 30	23.5	0.5	24.6	0.5	32.7	0.5	30.7	0.5
COA	Coachella	200	774 824 982 1020 1107 1230 1300 1347 1475 1503 1683	11.5	20	15 ~ 30	19.7	0.5	19.5	0.5	19.6	0.6	22.8	0.5
SSB	South_San_Bernardino	200	774 824 982 1020 1107 1230 1300 1347 1475 1503 1683 1987	16.4	13	5 ~ 20	17.4	0.2	21.1	0.5	25.5	0.7	19.2	0.4

NSB	North_San_Bernardino	200	774 931 1173 1107 1313 1347 1450 1475 1500 1619 1684 1704	17.8	19	13 ~ 28	16.0	0.4	15.2	0.4	13.5	0.5	14.1	0.5
SUP	Superstition_Hills and Superstition Mount combined	600	1050 1460 1540 1987	10.8	11	4 ~ 15	11.4	0.5	16.7	0.6	13.2	0.6	13.9	0.6
BOR	Borrego_ Mountain	550	1050 1460 1540 1987	10.0	5	1 ~ 10	7.1	0.3	10.8	0.3	14.3	0.3	9.1	0.3
COY	Coyote_Creek	500	1892 1968	10.0	5	1 ~ 10	8.9	0.4	10.6	0.4	11.9	0.4	8.8	0.4
ANZ	Anza	300	1892 1942 1954 1968 1969 1987	10.0	14	11 ~ 18	15.1	0.3	16.4	0.3	16.9	0.4	14.7	0.3
CLA	Clark	300	1020 1230 1290 1360 1630 1760	12.0	8	6 ~ 11	10.1	0.4	9.7	0.4	8.7	0.4	8.9	0.5
SJV	SJ_Valley	450	1770 1899 1918	15.0	16	12 ~ 24	15.1	0.5	15.9	0.5	17.8	0.4	16.4	0.4
SJB	SJ_San Bernadino valley	500	1770 1923	15.0	6	2 ~ 8	4.6	0.5	5.9	0.5	12.4	0.6	7.8	0.6

MOJ	Mojave	220	533 634 697 722 781 850 1016 1116 1264 1360 1487 1536 1685 1812 1857	15.0	34	25 ~ 40	25.5	0.3	27.8	0.3	33.1	0.4	22.9	0.3
SCZ	S_Carrizo (big bend)	250	599 1078 1247 1277 1310 1384 1393 1417 1457 1462 1565 1571 1614 1713 1749 1857	15.0	34	31 ~ 37	36.2	0.3	36.3	0.3	36.9	0.3	31.7	0.3

CAZ	Carrizo	200	599 1078 1247 1277 1310 1384 1393 1417 1457 1462 1565 1571 1614 1713 1749 1857	15.0	34	31 ~ 37	36.2	0.3	36.3	0.3	37.2	0.3	33.1	0.3
CHO	Cholame	200	1857	12.0	34	29 ~ 39	32.2	0.2	33.0	0.2	34.5	0.2	32.4	0.2
PAR	Parkfield	20	1857 1881 1901 1922 1934 1966	12.0	34	26 ~ 39	34.0	0.5	34.9	0.5	37.7	0.5	35.0	0.5
CRE	Creeping	250	N/A	12.0	34	29 ~ 39	34.5	0.3	38.7	0.4	43.8	0.4	36.3	0.4
SCR	Santa_Cruz_Mt	150	1300 1600 1838 1890 1906 1989	12.0	17	13 ~ 21	15.7	0.5	17.1	0.5	19.7	0.5	17.2	0.5
PEN	SA_Peninsula	230	1300 1600 1838 1906	16.2	17	13 ~ 21	18.1	0.4	20.5	0.4	25.4	0.4	17.6	0.4
SNC	S_SA_N_Coast	230	1300 1600 1906	15.5	24	16 ~ 27	14.0	0.4	23.0	0.5	36.5	0.7	33.2	0.6
NNC	N_SA_N_Coast	270	1300 1600 1899 1906	13.2	24	16 ~ 27	19.0	0.7	22.2	0.9	21.3	1.0	20.0	0.9

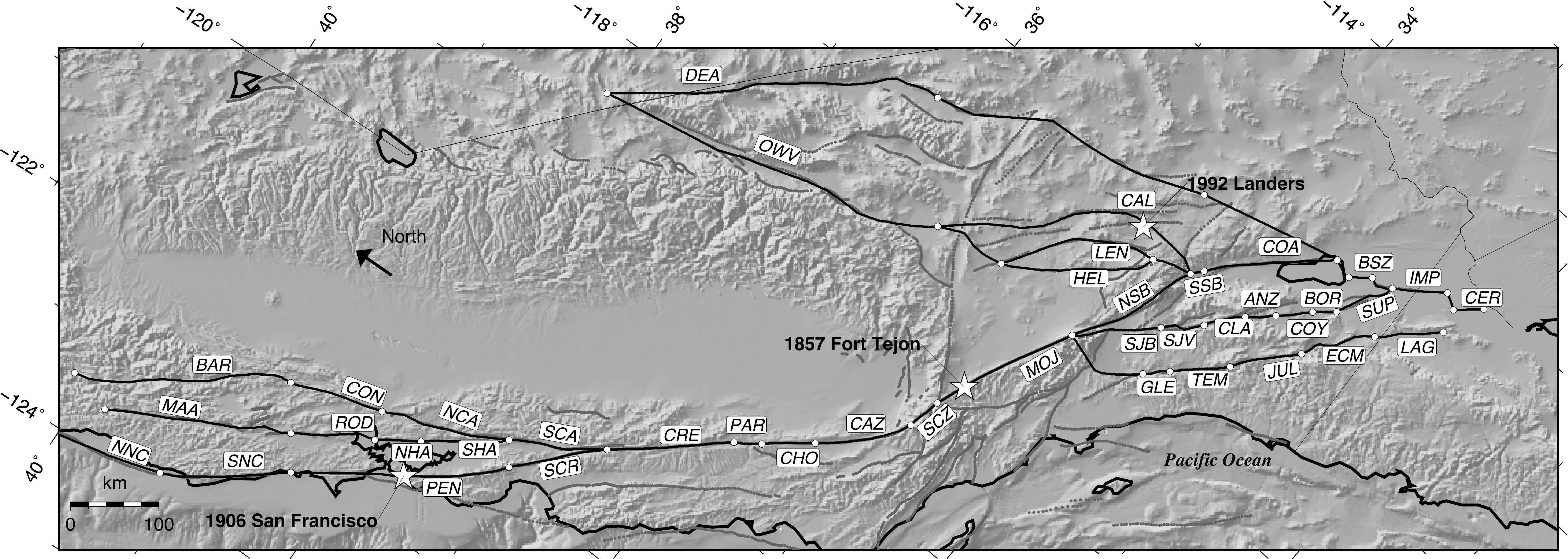
SCA	S._Calaveras	300	1864 1897 1911 1984	12.0	15	10 ~ 20	16.1	0.3	19.5	0.3	21.0	0.4	19.0	0.3
NCA	N._Calaveras	550	1864	12.0	6	3 ~ 8	9.0	0.3	10.4	0.4	10.0	0.5	9.0	0.4
CON	Concord	3000	510	12.0	4.3	3 ~ 9	7.7	0.5	9.0	0.5	10.7	0.5	9.1	0.5
BAR	Hunting Creek/Bartlett Spring	500	1760	12.0	4	2 ~ 9	7.3	0.4	8.5	0.4	8.5	0.5	10.1	0.4
SHA	S._Hayward	200	1470 1630 1730 1868	12.0	9	7 ~ 11	9.2	0.1	9.5	0.1	11.2	0.1	9.7	0.1
NHA	N._Hayward	300	1708	12.0	9	7 ~ 11	9.8	0.1	9.8	0.1	10.4	0.1	9.7	0.1
ROD	Rodgers_Creek	400	1898	12.0	9	6 ~ 11	12.0	0.2	12.0	0.3	12.6	0.3	10.2	0.3
MAA	Maacama	250	N/A	12.0	9	6 ~ 12	9.8	0.3	10.8	0.3	15.1	0.3	12.7	0.3
LAG	Laguna_Salada	1000	1892	9.0	3	1 ~ 5	2.9	0.2	5.4	0.2	7.2	0.3	5.4	0.2
GLE	Elsinore_GlenIvy	1000	963 1230 1283 1440 1627 1850 1910	10.0	5	3 ~ 7	4.7	0.1	4.7	0.1	4.6	0.1	4.6	0.1
TEM	Elsinore_Temecula	1000	1655 1810	10.0	5	3 ~ 7	4.0	0.2	4.0	0.2	3.2	0.2	3.6	0.1
JUL	Elsinore_Julian	1500	1655 1680 1753 1804	10.0	3	1 ~ 5	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ECM	Elsinore_Coyote_Mt	1500	1650 1892	10.0	3	1 ~ 5	2.1	0.3	1.2	0.3	0.4	0.3	5.5	0.4
LEN	Lenwood - Lockhart - Old Woman Springs	3000	1100	12.0	1	0.6 ~ 1.4	3.9	0.6	4.2	0.7	2.9	0.8	6.0	0.6
HEL	Helendale	3000	1100	12.0	0.6	0.2 ~ 1.0	0.4	0.4	0.8	0.5	1.1	0.6	0.0	0.0
CAL	Calico-Hidalgo	3000	510	10.0	1.6	1.0 ~ 2.8	6.9	0.3	7.3	0.3	7.7	0.3	6.6	0.4
OWV	Owens_Valley	3000	1100 1872	11.5	3.5	2 ~ 5	7.1	0.1	6.2	0.1	4.0	0.1	4.6	0.1
DEA	Death_Valley	1000	1100	12.0	3	2 ~ 4	5.0	0.3	5.0	0.3	6.5	0.3	5.7	0.3

Table 2. Fit to GPS, InSAR and geologic data for four different rheological models.

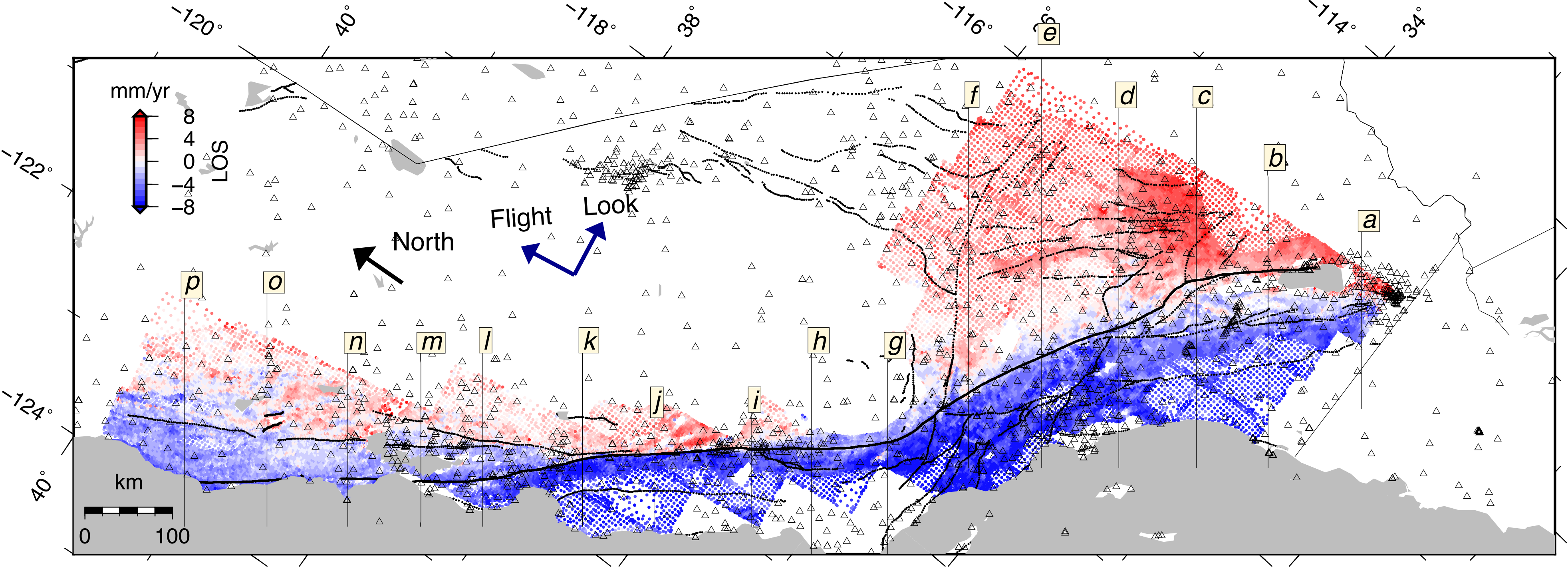
Model	HS	PL6019	PL3019	PL3020
GPS χ^2 misfit	2.67	2.56	2.74	2.68
GPS WRMS (mm/yr)	1.71	1.68	1.73	1.72
InSAR χ^2 misfit	0.28	0.27	0.27	0.28
InSAR WRMS (mm/yr)	1.34	1.30	1.31	1.34
RMS to the preferred geology slip rate (mm/yr)	3.50	3.70	5.30	4.00

Table 3. Summary of the synthetic test on the 1857 Fort Tejon earthquake.

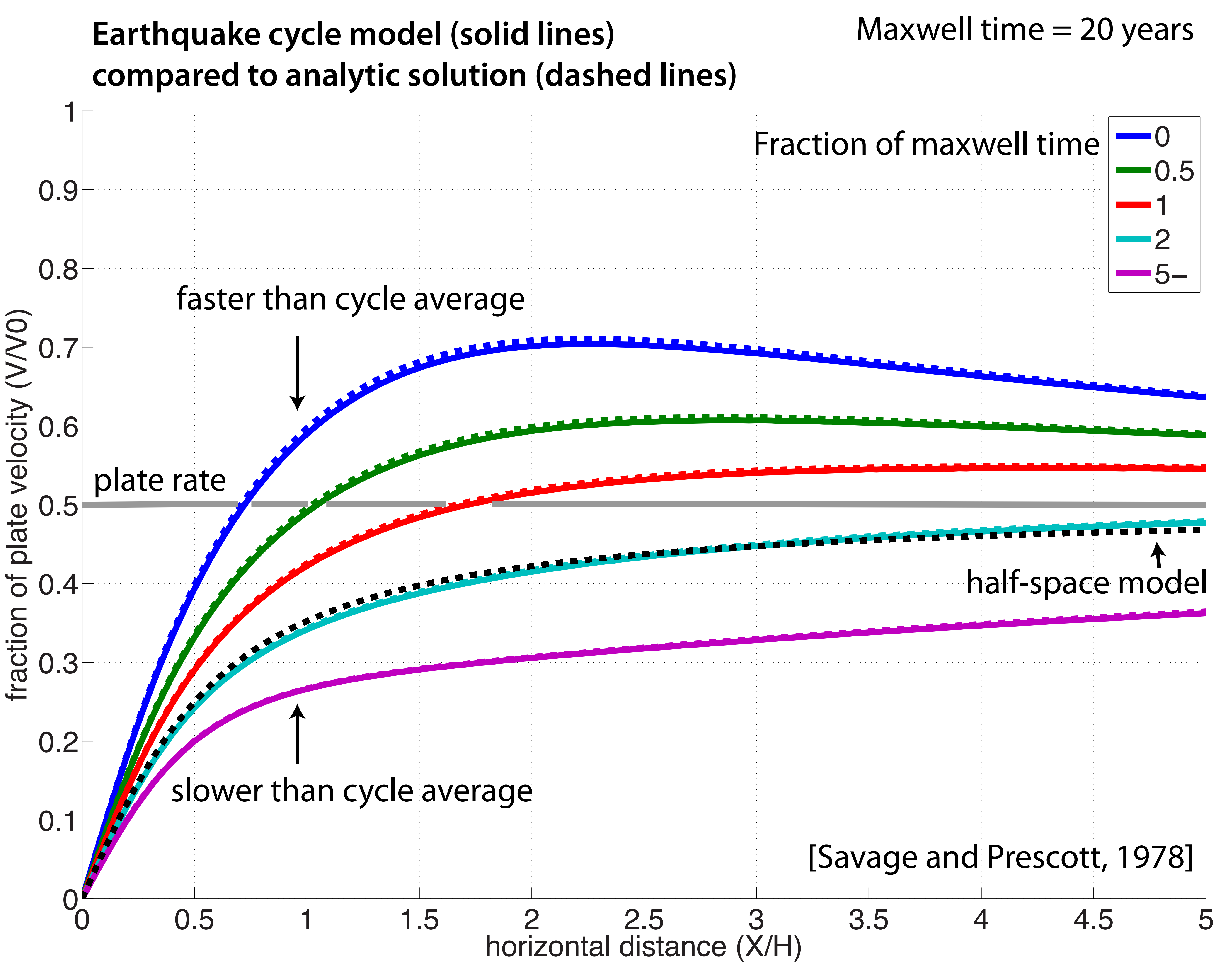
Mojave segment slip rate (mm/yr)	scenario A	scenario B	scenario C	no change
PL6019	27.5	27.8	26.4	27.8
PL3019	32.5	33	28.4	33
PL3020	20.8	23.6	20.8	22.9



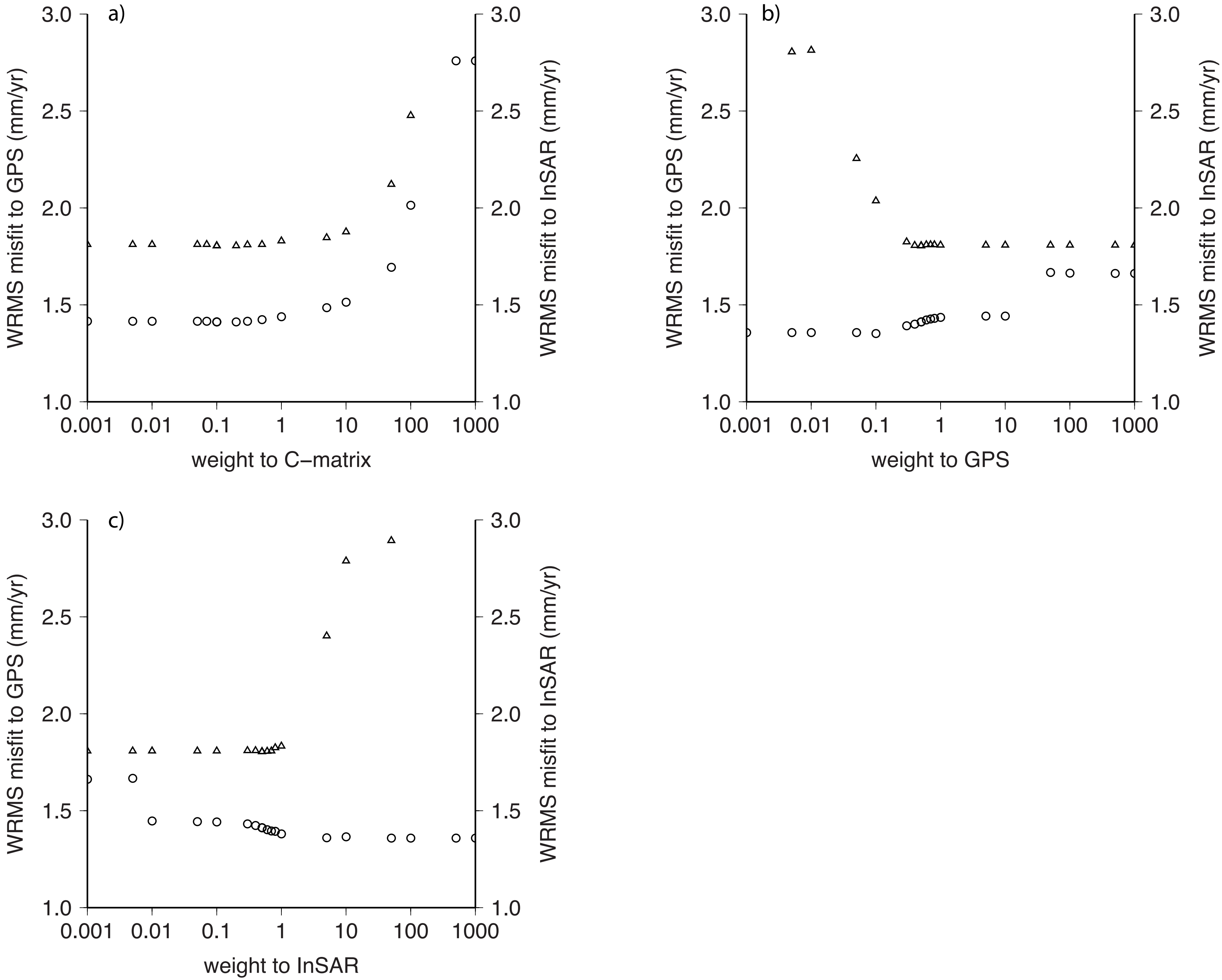
−120°
40°
−118°
38°
−116°
36°
−114°
34°
−122°
−124°
40°
DEA
OWV
1992 Landers
CAL
LEN
HEL
COA
BSZ
IMP
CER
SSB
NSB
ANZ
BOR
SUP
CLA
COY
SJB
SJV
LAG
ECM
JUL
TEM
GLE
MOJ
1857 Fort Tejon
North
BAR
CON
MAA
NCA
ROD
SCA
CRE
PAR
CAZ
SCZ
CHO
NNC
SNC
NHA
SHA
SCR
PEN
Pacific Ocean
km
0
100
1906 San Francisco



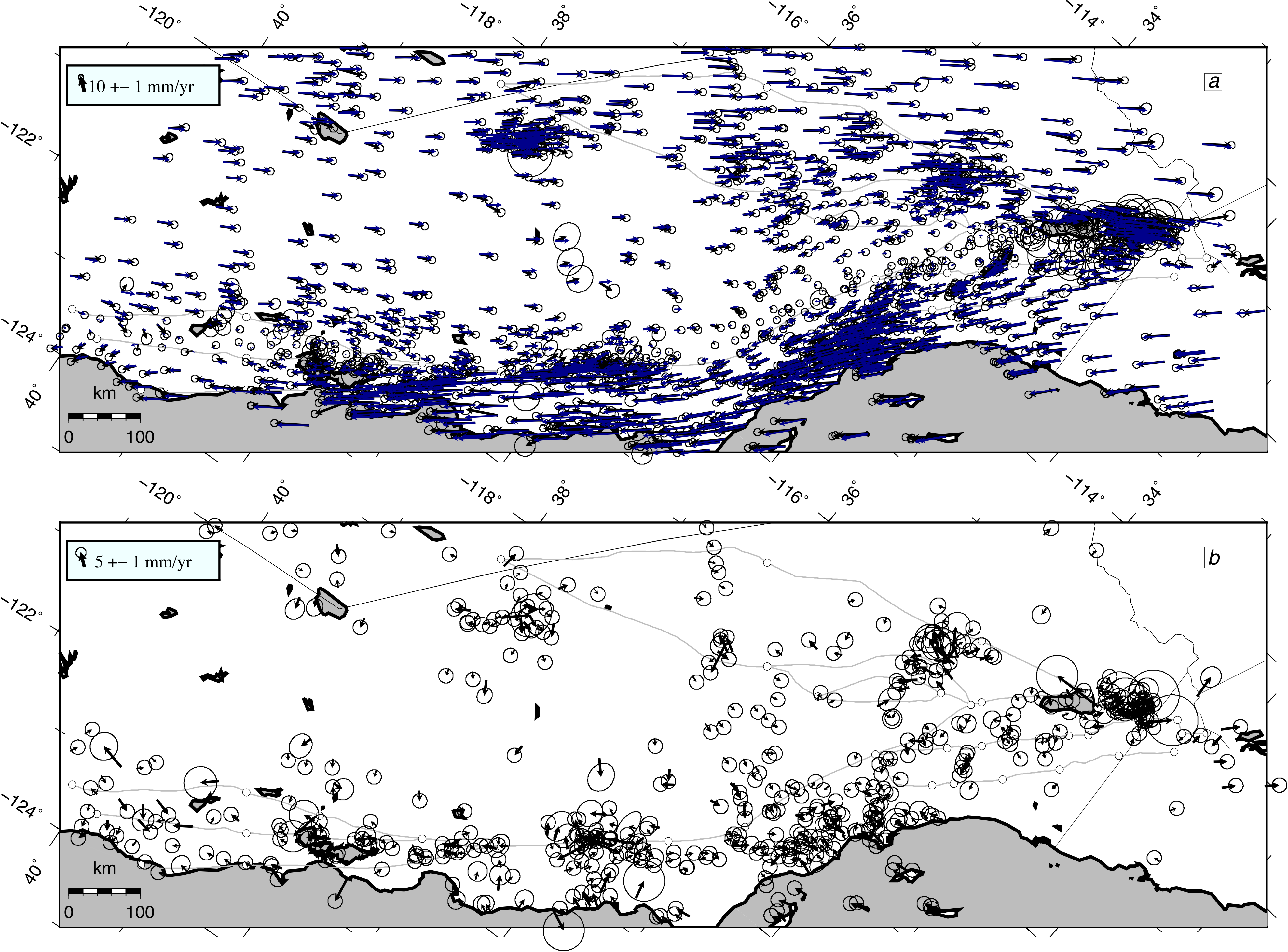
mm/yr
8
4
0
−4
−8
LOS
North
Flight
Look
a
b
c
d
e
f
g
h
i
j
k
l
m
n
o
p
km
0
100
−120°
40°
−118°
38°
−116°
36°
−114°
34°
−122°
−124°
40°



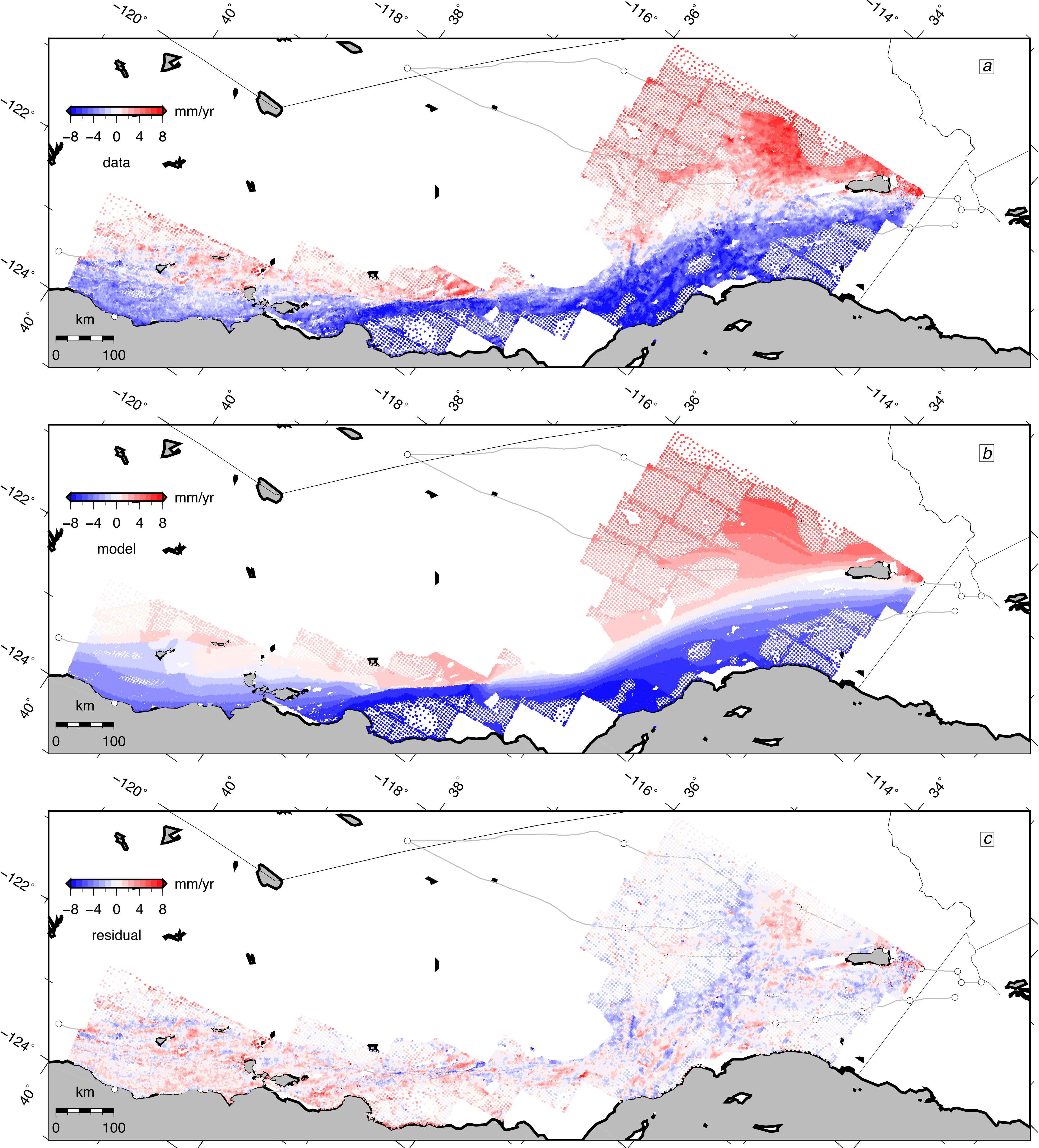
Earthquake cycle model (solid lines)
compared to analytic solution (dashed lines)
Maxwell time = 20 years
Fraction of maxwell time
0
0.5
1
2
5–
faster than cycle average
plate rate
half-space model
slower than cycle average
[Savage and Prescott, 1978]
fraction of plate velocity (V/V0)
horizontal distance (X/H)
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1
0
0.5
1
1.5
2
2.5
3
3.5
4
4.5
5



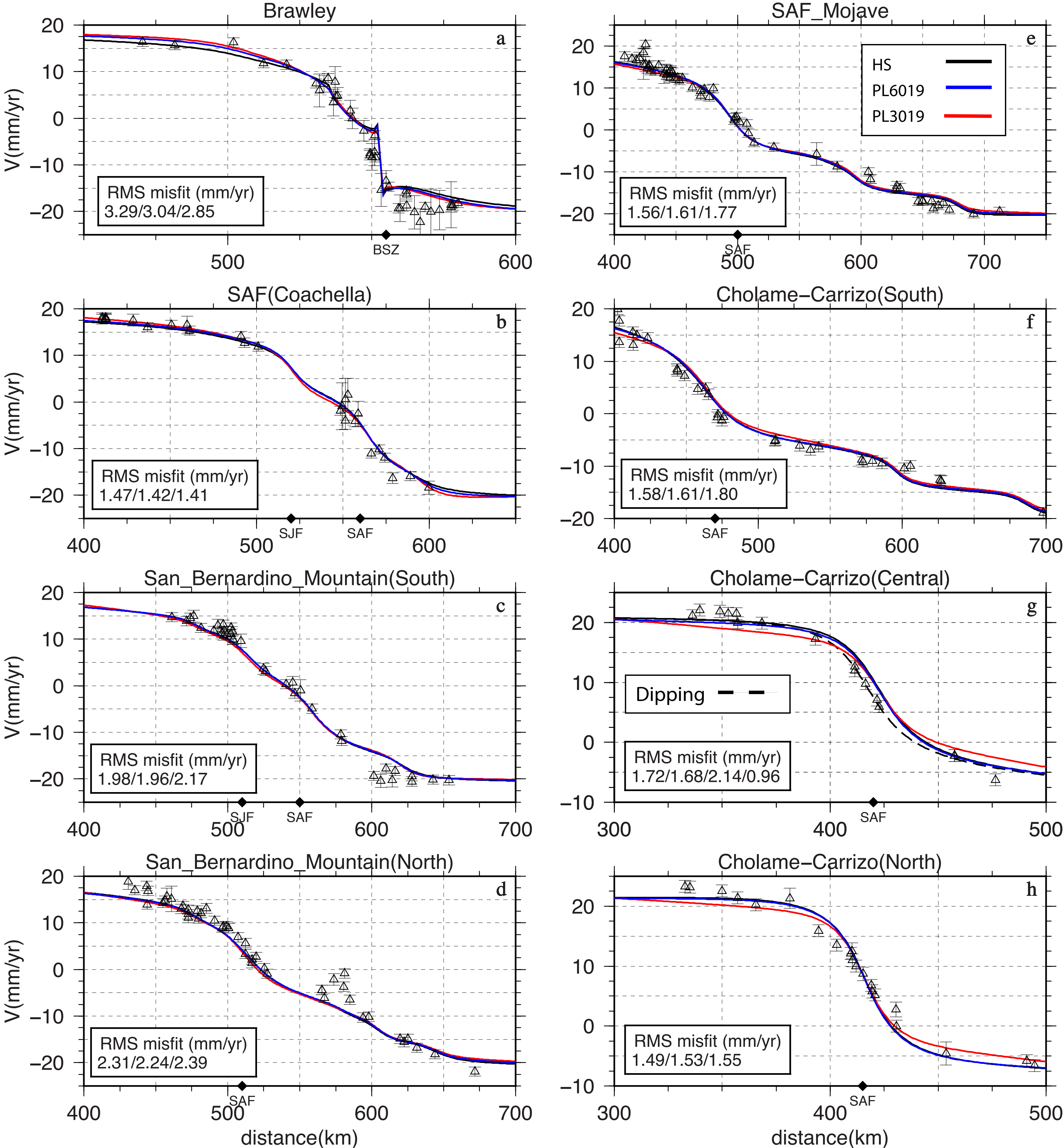
a)
b)
c)
WRMS misfit to GPS (mm/yr)
WRMS misfit to InSAR (mm/yr)
1.0
1.5
2.0
2.5
3.0
0.001
0.01
0.1
1
10
100
1000
weight to C−matrix
weight to GPS
weight to InSAR



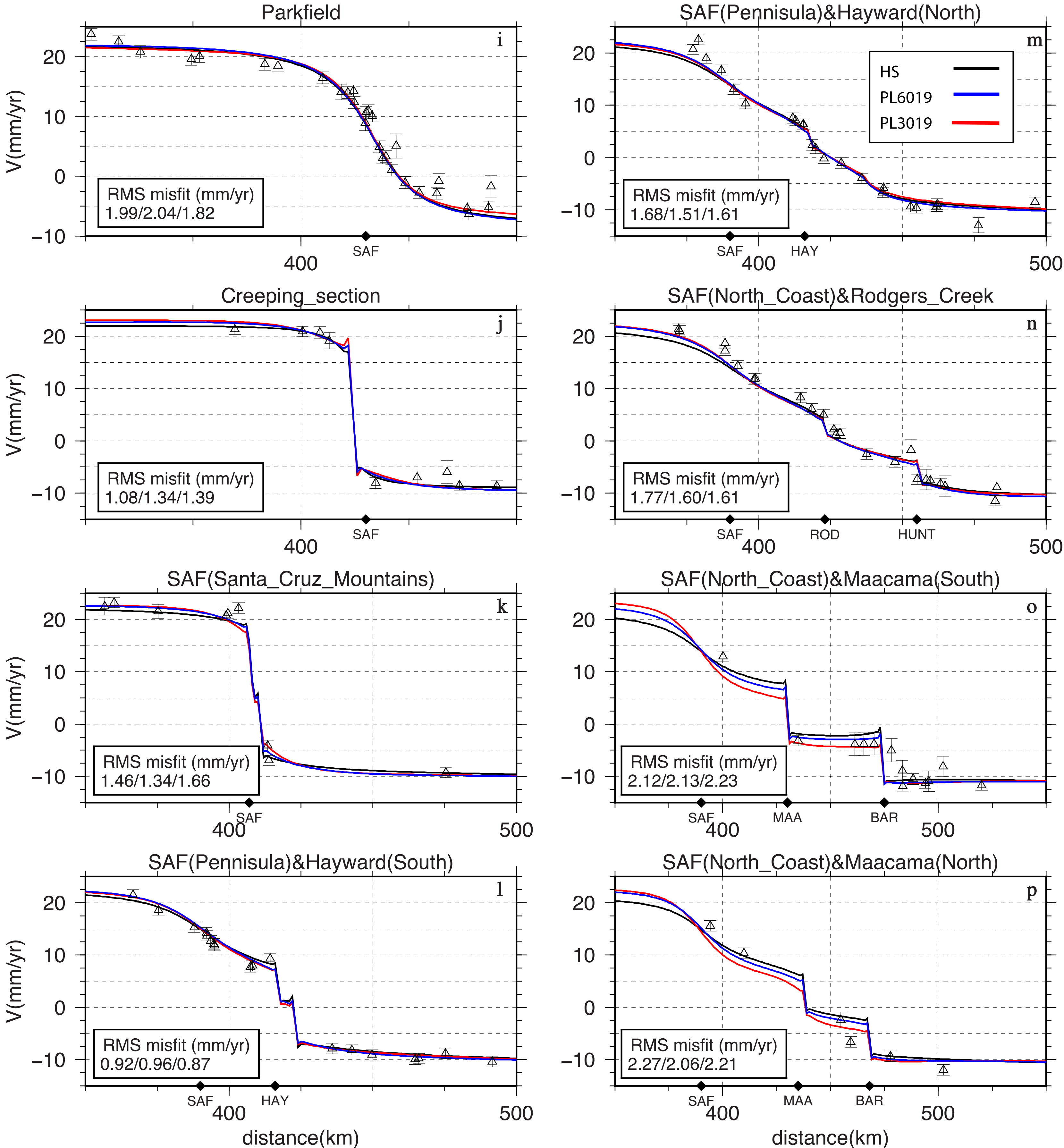
−120°
40°
−118°
38°
−116°
36°
−114°
34°
−122°
−124°
40°
10 +− 1 mm/yr
a
km
0
100
5 +− 1 mm/yr
b

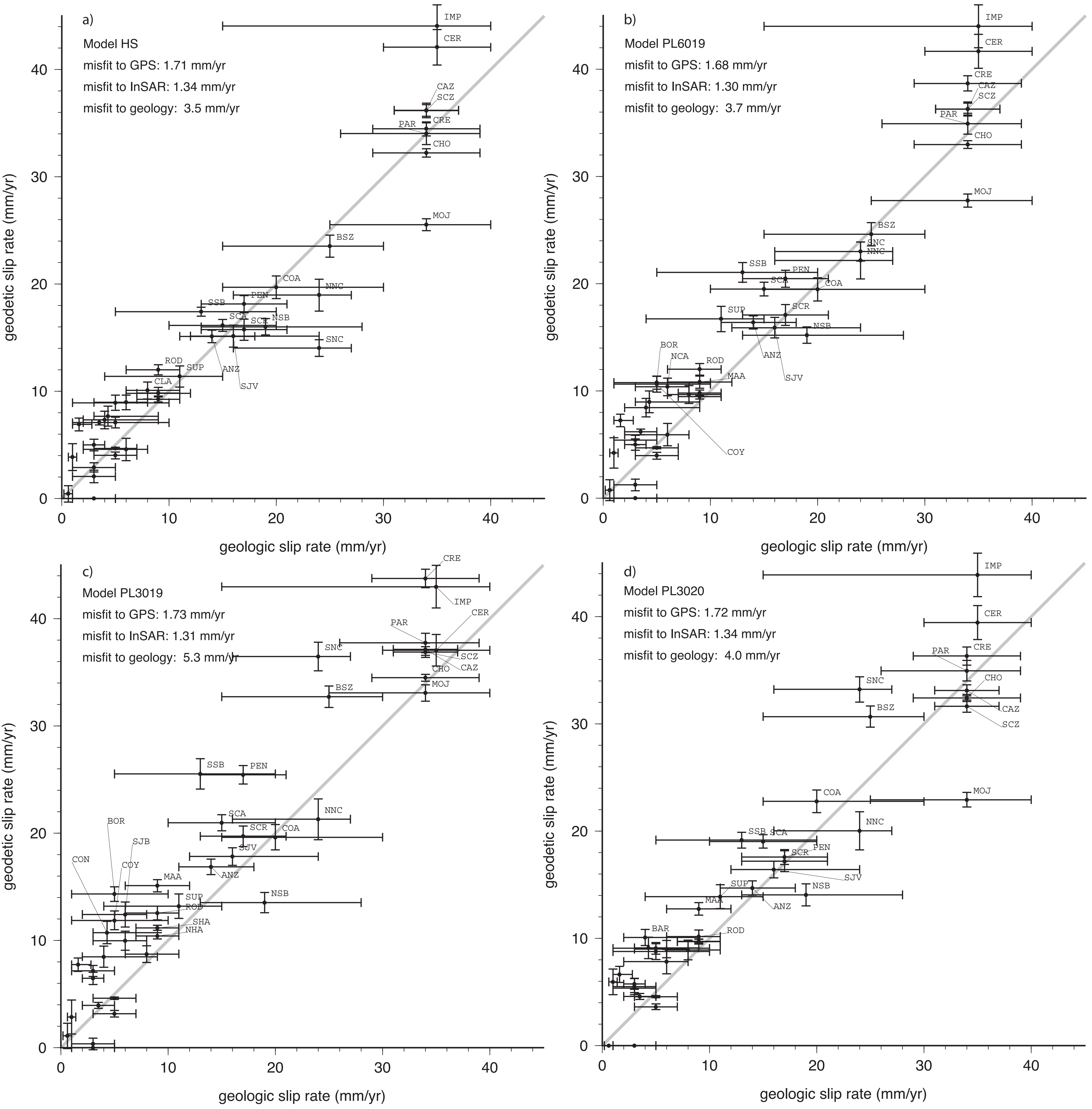


a
mm/yr
−8 −4 0 4 8
data
km
0 100
b
mm/yr
−8 −4 0 4 8
model
km
0 100
c
mm/yr
−8 −4 0 4 8
residual
km
0 100
−120°
−118°
−116°
−114°
−122°
−124°
40°
38°
36°
34°



Brawley
SAF_Mojave
SAF(Coachella)
Cholame−Carrizo(South)
San_Bernardino_Mountain(South)
Cholame−Carrizo(Central)
San_Bernardino_Mountain(North)
Cholame−Carrizo(North)
a
b
c
d
e
f
g
h
HS
PL6019
PL3019
Dipping
RMS misfit (mm/yr)
3.29/3.04/2.85
1.47/1.42/1.41
1.98/1.96/2.17
2.31/2.24/2.39
1.56/1.61/1.77
1.58/1.61/1.80
1.72/1.68/2.14/0.96
1.49/1.53/1.55
BSZ
SJF
SAF
V(mm/yr)
distance(km)


Parkfield
i
RMS misfit (mm/yr)
1.99/2.04/1.82
SAF
SAF(Pennisula)&Hayward(North)
m
HS
PL6019
PL3019
RMS misfit (mm/yr)
1.68/1.51/1.61
SAF
HAY
Creeping_section
j
RMS misfit (mm/yr)
1.08/1.34/1.39
SAF(North_Coast)&Rodgers_Creek
n
RMS misfit (mm/yr)
1.77/1.60/1.61
ROD
HUNT
SAF(Santa_Cruz_Mountains)
k
RMS misfit (mm/yr)
1.46/1.34/1.66
SAF(North_Coast)&Maacama(South)
o
RMS misfit (mm/yr)
2.12/2.13/2.23
MAA
BAR
SAF(Pennisula)&Hayward(South)
l
RMS misfit (mm/yr)
0.92/0.96/0.87
SAF(North_Coast)&Maacama(North)
p
RMS misfit (mm/yr)
2.27/2.06/2.21
V(mm/yr)
distance(km)



a)
Model HS
misfit to GPS: 1.71 mm/yr
misfit to InSAR: 1.34 mm/yr
misfit to geology: 3.5 mm/yr
IMP
CER
CAZ
SCZ
PAR
CRE
CHO
MOJ
BSZ
COA
NNC
SSB
PEN
SCA
SCR
NSB
SNC
ROD
SUP
ANZ
SJV
CLA
geodetic slip rate (mm/yr)
geologic slip rate (mm/yr)
b)
Model PL6019
misfit to GPS: 1.68 mm/yr
misfit to InSAR: 1.30 mm/yr
misfit to geology: 3.7 mm/yr
IMP
CER
CRE
CAZ
SCZ
PAR
CHO
MOJ
BSZ
SNC
NNC
SSB
PEN
SCA
COA
SUP
SCR
NSB
ANZ
SJV
BOR
NCA
ROD
MAA
COY
geodetic slip rate (mm/yr)
geologic slip rate (mm/yr)
c)
Model PL3019
misfit to GPS: 1.73 mm/yr
misfit to InSAR: 1.31 mm/yr
misfit to geology: 5.3 mm/yr
CRE
IMP
CER
PAR
SNC
SCZ
CHO
CAZ
BSZ
MOJ
SSB
PEN
NNC
SCA
SCR
COA
SJV
ANZ
BOR
SJB
CON
COY
MAA
NSB
SUP
ROD
SHA
NHA
geodetic slip rate (mm/yr)
geologic slip rate (mm/yr)
d)
Model PL3020
misfit to GPS: 1.72 mm/yr
misfit to InSAR: 1.34 mm/yr
misfit to geology: 4.0 mm/yr
IMP
CER
PAR
CRE
SNC
CHO
BSZ
CAZ
SCZ
COA
MOJ
NNC
SSB
SCA
SCR
PEN
SJV
SUP
NSB
MAA
ANZ
BAR
ROD
geodetic slip rate (mm/yr)
geologic slip rate (mm/yr)